V

Pedology

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PEDOLOGY IS THE EARTH SCIENCE that quantifies the factors and processes of soil formation including quality, extent, distribution, and spatial variability of soils from microscopic to megascopic scales (Sposito and Reginato, 1992; Wilding et al., 1994). Spatial distribution and variability of soils in landforms is governed by the processes of soil formation which are, in turn, interactively conditioned by lithology, climate, biology, and relief through geologic time. Soils form a continuous drape across the landscape and are intimately linked to one another; processes and impacts on higher topographic surfaces directly affect adjacent lower lying surfaces. This is because transfer of energy and mass, the driving forces of pedogenesis, occur within, over, and among 3D soil landform components. Renewal vectors of biomass input, rainfall, and dust counter constituent losses via drainage waters, lateral interflow, and downslope migration of erosion products.

The development of open versus closed drainage patterns during landform evolution strongly governs energy and mass flux in the system. In the closed drainage network, dispersal of chemical and erosion products is distributed to adjacent local sinks and depressions. In contrast, dispersal is mostly external to source areas in open drainage systems. In this case, distribution occurs via upland drainage ways to tributary streams, rivers, lakes, and oceans. Differences in drainage network are paramount when one considers the effect of landscape on soil moisture and nutrient regimes, environmental contamination, recharge of groundwater aquifers, crop production potential, etc. Hence, to adequately comprehend, interpret, and transfer knowledge of soil resources from one area to the next, a systematic soil/geomorphic landscape model must be applied.

Through knowledge of soil/landform relationships, pedologists have verified the occurrence, configuration, depth, and pedogenic formation of root and water restrictive layers; documented the origin and distribution of cracking and fissuring patterns in soils and geologic materials that govern bypass flow of nutrients, chemicals solutes, and fluids; identified the scale, mode, and occurrence of systematic soil spatial distribution to aid the design efficiency and sampling of soil units; and used soil color patterns on a macro- and microscale to infer major periods of soil aeration/reduction and relative periods of excess, sufficiency, an deficiency of soil moisture contents for specific land uses. In most soil systems, reaction kinetics and diffusion rates rather than thermodynamic equilibria control chemical reactions, solute movement, and precipitation of chemicals and minerals. These and other impacts of geomorphology and pedogenic processes reflected in soil landscape patterns are covered in Chapters 29 and 30, respectively.

Pedology is an integrative and extrapolative science. It provides an organizational framework to catalog modes, mechanisms, and magnitudes of soil spatial variability, and to generalize this knowledge for synthesis and data population for models. Pedology provides a vehicle for extrapolation and scaling of spatial variability from components of soils (individual particles, aggregates, hand specimens, and horizons) to the populations of soils comprising the landscape as a whole (pedons, toposequences, watersheds, physiographic regions, and the pedosphere) (Figure 34.1). Figure 34.2 illustrates hierarchical levels in this continuum of soil organization relative to tools used to generalize information content at multiple scales of resolution.

Various taxonomic systems have been developed to accommodate this cataloging of soil attributes and their landscape distribution. *Soil Taxonomy* (Soil Survey Staff, 1998, 2010) is such a system developed using morphogenetic indicators (diagnostic horizons and properties) as class criteria. Chapter 31 discusses the history leading to the development of *Soil Taxonomy*, an overview of class criteria, nomenclature, and interpretational inferences. Although *Soil Taxonomy* was developed to categorize soils from across the globe, the functions and objectives of it and other systems of soil classification systems, including the French, Canadian, Russian, and Chinese systems, often have a national bias. To enhance international understanding of and ability to communicate about the soil resource, FAO with support from the International Union of Soil Science (IUSS) is leading the development of the World Reference Base for Soil Resources (http://www.fao.org/docrep/w8594e/w8594e00.htm), and the IUSS is providing leadership for development of a universal soil classification system to enhance our ability to categorize, understand, and extrapolate knowledge of the global soil resource.

Spatial variability in soil systems belongs to two broad categories: systematic (structured) and random (unstructured and unknown causes). Systematic variability is a gradual or marked change in soil properties as a function of physiography, geomorphology, and the interactions of the soil forming factors (Wilding and Drees, 1983) and our ability to predict systematic soil change across the landscape is based on the soil–landscape paradigm (Hudson, 1991). Systematic variability permits pedologists to partition spatial variability in soils by subsets of properties that constitute soil survey map units corresponding to geomorphic landscape elements. The purpose of soil surveys is to partition the spatial variability of landforms into stratified subsets that are less variable than the landscape as a whole and to remove systematic error from the population of soils. Information gained from soil surveys on the quality and distribution of soils when correlated with their classification provides a power vehicle for transfer of knowledge of soil properties and appropriate technologies for sustainable management. Chapter 33 illustrates characteristics of soils in each of the 12 soil orders in *Soil Taxonomy*, their properties, processes, and distribution patterns on a global scale. Each of the 12 soil orders are addressed as a subset of this chapter.

Pedologists are being challenged to better integrate information on soil properties, processes, and distribution into assessments of food and fiber production, soil quality, environmental sustainability, ecosystem services, and risk avoidance. Land evaluation, the consideration of actual and potential land use as a function of land properties, can be realized with qualitative methods, but increasingly quantitative simulation models are used together with geographic information systems (GIS) as aids for land use decisions. These models and GIS systems need to be fed with adequate soil data, but use of databases with little attention to natural soil dynamics or landscape relationships may lead to unrepresentative results. Chapter 34 discusses both classical qualitative and modern simulation methods of land evaluation. This chapter illustrates land evaluations through case studies for both data-rich and data-poor scenarios, and outlines the role of pedology and soil survey in land use decisions for sustainable ecosystems.

Pedology is adapting to the demand for more data, better methods, and new approaches to address complex natural processes that strongly influence our ability to interpret soil behavior and design management systems that maintain or enhance soil function in an ecologically and environmentally sustainable manner. The remaining chapters in this section describe a selected set of these new areas of study and application that likely will be important pedological subdisciplines in the twenty-first century.

Hydropedology is a combination of pedology, hydrology, and soil physics disciplines that addresses soil and water interactions. It addresses multiscale basic and applied pedologic and hydrologic processes and their properties in the variably unsaturated zone (Lin, 2003). Hydropedology focuses on the synergistic integration of pedology and hydrology to enhance our understanding of soil–water interactions and landscape–soil–hydrology relationships across space and time as they relate to ecosystem functions. Chapter 35 provides an overview of the fundamentals and applications of hydropedology, including a review of principles and description of recent advances in soil architecture and preferential flow, soil hydromorphology, scaling, pedotransfer functions, and coupling biogeochemistry with hydropedology.

Subaqueous soils, those soils inundated with water almost continuously (positive water pressure at the soil surface for at least 21 h each day in all years), are found along oceanic coastal margins and in shallow water associated with lake margins. Historically, these subaqueous substrates were considered to be sediments instead of soils. Recent research in these important ecosystems, however, have shown that subaqueous soils are, in fact, soils with horizons and properties that are the result of pedogenic processes. Chapter 36 provides an overview of subaqueous soils including the history of their recognition as soils, descriptions of methods for characterization and mapping, subaqueous soil classification, genesis models, and examples of interpretation of properties and distribution for soil use and management.

The explosion in computation and information technology has motivated numerous initiatives around the world to build spatial data infrastructures including regional, continental, and worldwide soil databases. In soil science, the principal manifestation is soil resource assessment using GIS to produce digital soil property and class maps with the limited expensive fieldwork and laboratory analysis. Digital soil mapping refers to the production of spatial soil databases that are based on soil observations combined with environmental data through quantitative statistical relationships. It uses a range of technologies for more accurate and efficient prediction of soil properties through optimal sampling strategies, rapid analysis of soil properties, and rapid acquisition of environmental variables over large areas and can be thought of as a means for modernizing and systematizing traditional soil survey. Chapter 37 describes the development of and theory underlying digital soil mapping and reviews the various approaches with examples.

Pedologists have long recognized that soils change over time. With a few exceptions, however, this change has mostly been related to changes over the course of geologic time and landscape development. Recently, greater emphasis has been placed on the concept of soil change as it relates to the domestication of soil over historic time; the transformation of soil into a cultural–historic–natural system. To understand soil change and apply that knowledge to sustainable land management, we must understand how human activities impact the soil itself and soil's interactions with the wider environment. Chapter 38 discusses soil change including its impacts on soil function and ecosystem services, approaches to evaluate soil change, factors and mechanisms affecting soil change, and speculates on the future science and management of soil change.

The demand for more and more comprehensive data on soils as they vary across the landscape led to development and evaluation of numerous methods to rapidly evaluate soil properties on samples in the laboratory, at specific sites in the field, continuously across the landscape, and for fields or larger-sized areas. Among these techniques are diffuse reflectance infrared spectroscopy, gamma-ray spectroscopy, x-ray fluorescence, electromagnetic induction, and ground-penetrating radar. Advances in technologies to allow instrument portability and enhanced data analysis have made these technologies more commonly available for pedological research and soil inventories. Chapter 39 discusses two of these technologies, ground-penetrating radar and electromagnetic induction including the theory behind the technologies, initial uses, and their application to pedological studies.

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29

Geomorphology of Soil Landscapes

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29.1 Introduction

Soils form a continuum across the earth's land surface, that is, the interface of atmospheric, biological, and geological processes. Soils are more than a veneer of surficial alteration on landscapes or sediments. Soils, landscapes, and surficial sediments or rocks together comprise 3D systems that coevolve through the interaction and balance of physical and chemical weathering versus erosion and deposition. One must comprehend the relationships among soils, landscapes, and surficial sediments to fully understand soil landscapes, successfully predict soil occurrence, and anticipate soil behavior. Soil geomorphology is the scientific study of the origin, distribution, and evolution of soils, landscapes, and surficial deposits and the processes that create and alter them. As a science, soil geomorphology is directly linked to pedology, geology, hydrology, archaeology, geomorphology, physical geography, ecology, and geotechnical engineering (Figure 29.1). Soil geomorphology relies primarily, but not solely, upon geological principles and techniques (Daniels and Hammer, 1992). Geologic/geomorphic processes substantially, but not solely, determine the materials from which soils are derived via the nature and redistribution of sediments. Principles and techniques drawn from geology or from other sciences often have applications or expressions that are unique to soils and soil landscapes. This chapter summarizes the major geomorphic principles and techniques used to understand the relationships among soils, landscapes, surficial sediments (Hunt, 1986), and internal water movement.

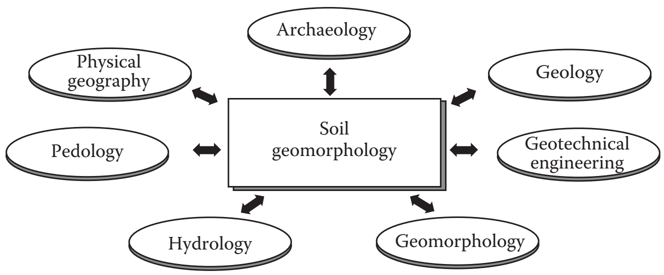


FIGURE 29.1 Conceptual diagram showing the relationships of soil geomorphology and other sciences.

29.1.1 Goals of Soil Geomorphology

Soil geomorphology serves two important capacities: (1) Geomorphic-based landscape models allow the segregation of the soil continuum into meaningful natural soil bodies that explain soil distribution from hillslope to continental scales. (2) It provides basic principles for understanding the geomorphic history of landscapes (e.g., spatial and time relationships among soils, landscapes, and surficial sediments).

29.2 Key Terminology

Before discussing the principles of soil geomorphology in detail, a few key definitions are needed. Although some terms may have alternate definitions, our discussion uses those following.

29.2.1 Depositional Surface

Depositional surface is a part of a geomorphic surface formed by the deposition of sediments (alluvial, colluvial, eolian, etc.) derived from an erosional surface or erosional processes (Ruhe, 1975). Water, wind, ice, and gravity are active agents that construct and shape depositional surfaces with water being the most universally prevalent agent.

29.2.2 Erosional Surface

An erosional surface is a part of a geomorphic surface formed by removal of rock, soil, or sediment under the wearing and transport action of water, wind, ice, and gravity (Ruhe, 1975). Running water is the dominant agent that forms and shapes most erosional surfaces (wind, ice, or gravity dominant agent in select settings).

29.2.3 Geomorphic Surface

A geomorphic surface is a definable land surface area that forms during a given time under a common set of erosional and depositional processes. A geomorphic surface may include multiple landforms, but it is a mappable spatial area based upon geomorphic techniques and field observations that has specific and identifiable borders (Ruhe, 1975).

29.2.4 Landform

A landform is any physical, recognizable feature on the earth's surface, having a characteristic shape and internal composition that is produced by natural processes (SSSA, 1997). Landforms are the individual features of the earth that together comprise the land surface (Ruhe, 1975).

29.2.5 Landscape

A landscape is a spatially adjacent collection or assemblage of landforms that can be observed in a single view or from a given vantage (Ruhe, 1975). The landforms that compose a single landscape may have the same age and origin or may vary in both age and origin. Another definition for landscape is the portion of the land surface the eye can comprehend in a single view (Ruhe, 1969).

29.2.6 Pedon

A pedon is a 3D soil body l with lateral dimensions large enough to permit the complete study of horizon shapes and relations. A pedon ranges from 1 to 10^2 m in land surface area and has a maximum defined depth of 2 m (SSSA, 1997).

29.2.7 Soil

Soil is a natural, 3D body with definable boundaries that occurs on the land surface composed of solids (mineral and organic), liquids (water), and gases. Soil is characterized by horizons distinguishable from the initial material as a result of additions, losses, transfers, and transformations in the surface environment and/or by the ability to support higher plant forms (Soil Survey Staff, 2010).

29.2.8 Soil Delineation

Soil delineation is an individual polygon shown on a soil map with a map unit symbol and name that defines a 3D soil body of specified area, shape, and location on the landscape (SSSA, 2010).

29.2.9 Soil Map Unit

A soil map unit is an aggregate of all soil delineations in a soil survey area that have a similar, established set of characteristics (Van Wambeke and Forbes, 1986). Each delineation of a soil map unit is identified by the same symbol and name in a soil survey.

29.3 Soil as a Landscape Unit or Body

Soil geomorphology is a field science that studies soils on landscapes. The soil continuum on a landscape, for ease of comprehension, is divided into discrete individuals. The pedon is the soil unit most commonly described, sampled, and classified by pedologists. The relative size or scale of a pedon is an intellectual

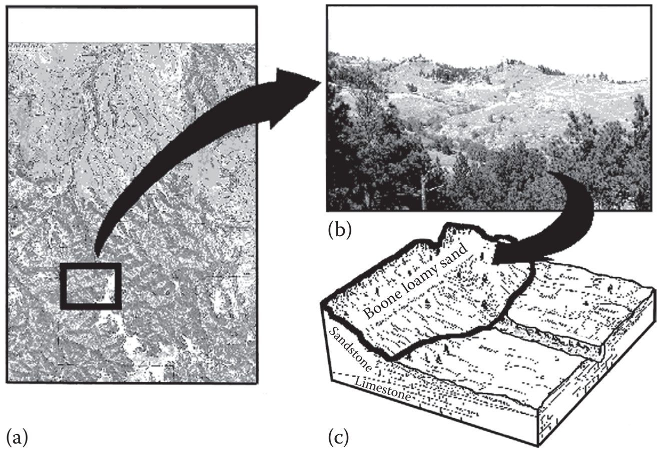


FIGURE 29.2 (a) Landscape image, (b) soil map image, and (c) block diagram showing soil as a landscape unit.

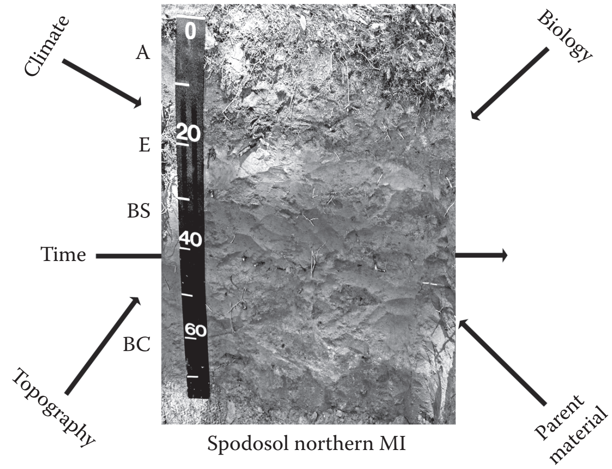


FIGURE 29.3 Conceptual diagram of Jenny's five soil-forming factors climate, biota, parent material, topography, and time.

construct useful and functionally necessary for description and classification. Pedons, however, lack distinct lateral boundaries and are, therefore, not natural landscape units. Geological processes, which are driven by the atmospheric agents of water, wind, ice, and gravity, impact landscapes as a continuum. Landscapes, however, possess natural internal boundaries that restrict or control mass and energy transfer. Examples of such boundaries are topographic divides, contacts between different rocks or sediment bodies, inflections in slope gradient or shape, and contacts between landforms of different age, origin, and internal structure.

29.3.1 Landscape Scale and Function

Soil delineations depicted on a soil map at or near a scale of 1:24,000 are more closely linked to geomorphic processes than are individual pedons. Accordingly, soil delineations (groupings of similar pedons) are better suited to geomorphic studies than individual pedons. Soil delineations depicted in a soil survey are landscape units (Figure 29.2). Boundaries between delineations are established by the soil surveyor, but not in an arbitrary fashion. Boundaries on a soil map delimit observable differences in soil morphology such as horizon type and thickness, and soil color, texture, and structure across the landscape. Changes in soil morphology across a landscape generally result from differences in the transfer of mass and energy driven by ecological, geomorphic, or atmospheric processes. Soil surveyors use observation, experience, and geomorphic landscape models to create soil maps.

29.4 Models of Soil Formation

Soils are complex systems that defy easy comprehension. Therefore, we use scientific models to help understand or explain soils. No single model provides complete understanding. Models of various form, function, and design are needed to understand systems as complex as soils (Dijkerman, 1974). Two important and well-known soil models are those of Jenny (1941) and Simonson (1959).

29.4.1 Jenny's Factors of Soil Formation

The eloquent, conceptual model of Jenny (1941) describes soil as a function of climate, biological influences, topography, parent material, and time (Figure 29.3). Implicit in this model are the distinct relationships among ecosystems (biological factor), landscapes (topography), surficial sediments (parent material), and landscape evolution (time). Stratigraphy of sediments or bedrock and surface contours strongly influence the movement of water within and over the landscape. Topography and parent materials have a strong control on both local (e.g., soil water tables, water holding, nutrient capacity, salt content, soil temperature) and regional soil environments (e.g., rain shadows, elevational induced climate zones, adiabatic winds) and therefore impact ecosystem form and function. All five soil-forming factors are linked either directly or indirectly to landscapes, surficial sediments, and landscape evolution.

29.4.2 Simonson's Process Model

Simonson (1959) explained soil formation through the interaction of four processes: additions, removals, translocations, and transformations (Figure 29.4). This model is more helpful than Jenny's (1941) for understanding the spatial relationships within soil landscapes. Geologic or geomorphic processes cause additions, removals, translocations, and transformations on a landscape scale that create and modify landforms, sediments, and soils. For example, sediment eroded from the flank of a hillslope is deposited as colluvium at the base of the slope or as alluvium in nearby drainageways or floodplains. The sediment is incorporated into the uppermost horizons of an existing soil or becomes fresh parent material in which a new soil begins to form.

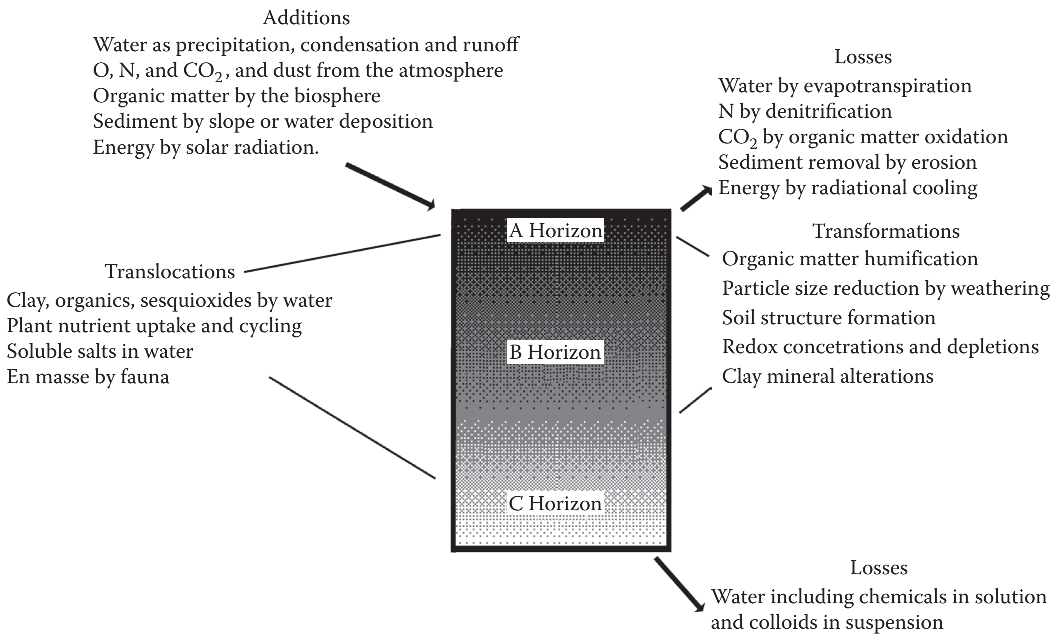


FIGURE 29.4 Simonson's (1959) process model of soil genesis showing the interactions of additions, removals, transfers, and transformations.

29.5 Soil Landscape Models

29.5.1 The Catena

The models of Simonson (1959) and Jenny (1941) provide an important conceptual framework for understanding soil formation. Neither model, however, establishes functional boundaries for segregating the soil continuum into natural landscape units. The catena is a fundamental concept that explains the soils pattern on hillslopes. Milne (1936a) coined the term catena to describe a repeating sequence of soils that occurs from the hillslope top to the adjacent valley bottom. Milne (1936a, 1936b) distinguished two catena types. The first type occurred on hillslopes developed over a single kind of parent rock. Despite the uniformity in parent rock, Milne observed a sequential change in soils along the slope gradient. Milne attributed the sequence of soils to variations in subsurface drainage, lateral transport of sediments, and the translocation of materials at or beneath the soil surface (Figure 29.5a).

In Milne's (1936a, 1936b) second example, the hillslope contained more than one type of parent rock (Figure 29.5b). An observable sequence of soils also occurred on this hillslope. Variations in drainage and lateral transport also produced this catena, but stratigraphic differences in the parent rock increased the complexity of the soil pattern. In this example, the surficial sediments form a drape on the landscape that is not coincident with the underlying rock strata. The catena concept includes both surficial stratigraphy and internal hillslope structure or lithology.

Furthermore, the catena concept is both a soil landscape model and a geomorphic model or system. Milne (1936a) recognized that lateral sorting contributed to the sequential soil variation down the hillslope. Erosion and sedimentation processes driven by relief and water movement redistribute sediments across hillslopes creating subtle lateral differences in soil parent materials (Kleiss, 1970). Recall that both parent material and topography are factors in Jenny's (1941) model of soil formation. The same erosional and depositional processes that drive landscape evolution influence the soil pattern on landscapes. The sequential change in soil morphology across a landscape is linked by process to landscape evolution on hillslopes both in time and space. In Jenny's model, landscape evolution means that parent material and topography are not independent variables, but rather are dependent variables that can covary in time.

A well-studied catena consists of the Clarion–Nicollet–Webster soils that occupy about 31,000 km^2 of the Des Moines Lobe in south-central Minnesota and north central Iowa (Figure 29.6). The Des Moines Lobe represents the last Late Wisconsinan glacial advance into Iowa about 14,000 YBP (Ruhe, 1969). Major topographic areas include hummocky, high-relief end moraines separated by undulating areas of low- to moderate-relief ground moraine (Ruhe and Scholtes, 1959; Ruhe, 1969; Kemmis et al., 1981). Closed, semiclosed, and linked depressions occur throughout the Des Moines Lobe, but closed depressions are most abundant in the end moraines (Kemmis, 1991). The Clarion–Nicollet–Webster soils are mollisols formed in a stratigraphic sequence composed of hillslope sediments, supraglacial sediments, and loam-textured till. The catenary relationships on the Des Moines Lobe landscape result from surficial sorting during deglaciation (Kemmis, 1991; Steinwand and Fenton, 1995), postglacial hillslope sorting (Walker, 1966; Burras and Scholtes, 1987; Steinwand and Fenton, 1995), and subsurface flow relationships (Steinwand and Fenton, 1995).

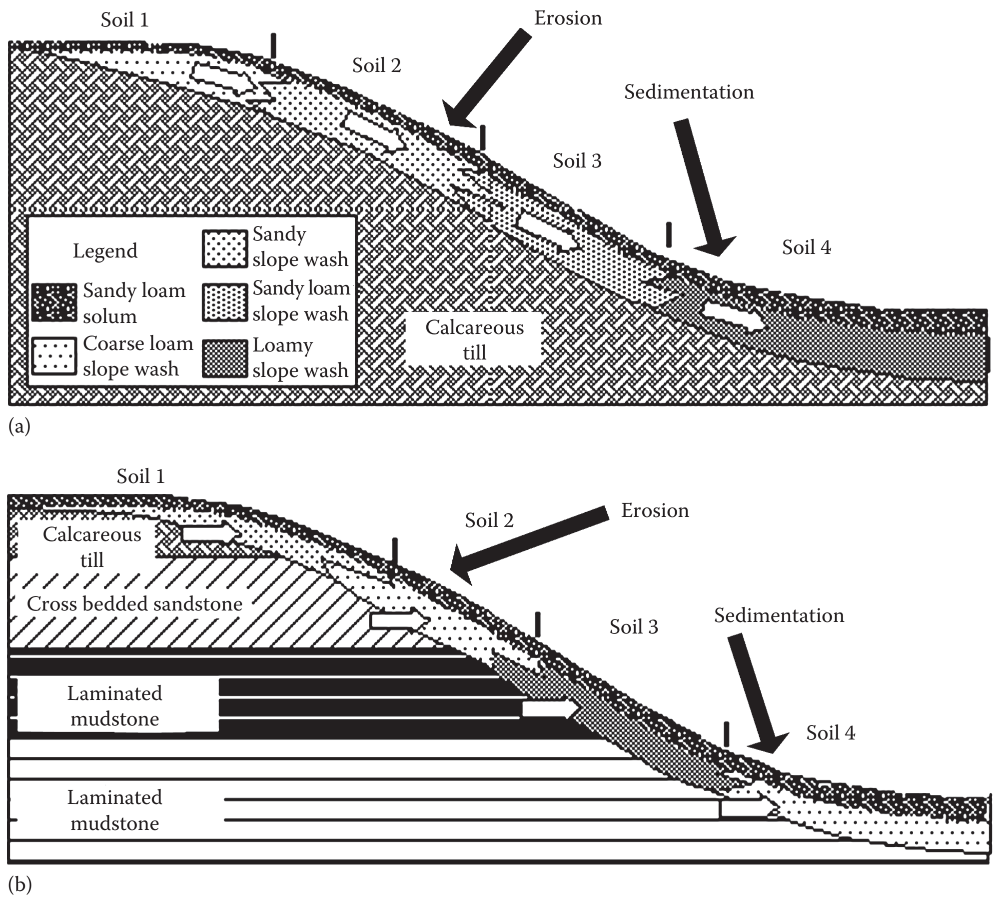


FIGURE 29.5 Two-dimensional diagram of Milne's (1936a, 1936b) catena showing idealized landscape relations.

Numerous studies (Dan and Yaalon, 1964; Blume and Schlichting, 1965; Blume, 1968; Walker and Ruhe, 1968; Huggett, 1975; Pennock and Vreeken, 1986; Pennock and Acton, 1989) have confirmed that catena relationships occur in various climates and landscapes. Conacher and Dalrymple (1977) and Dalrymple et al. (1968) provided a quantitative description of the catena. They defined the soil hillslope relationship as a 3D unit having arbitrary lateral dimensions extending from the hilltop to the valley bottom and from the soil surface to the base of the solum. They segmented the hillslope into nine land surface units (Figure 29.7) based on soil morphology, mobilization and transport of soil constituents, redeposition of soil constituents by overland and throughflow, or by gravity as mass movements.

29.5.2 The Toposequence

Bushnell (1942) studied morphological differences in soils across a hillslope gradient. The soils differed mainly in color. He attributed the changes in morphology to elevational position and local hydrology. This concept is commonly referred to as a toposequence. Unlike Milne, Bushnell did not recognize the influence of hillslope erosion and sedimentation. The terms catena and toposequence are often presently used as synonyms, but the original meanings are not identical.

29.5.3 The Valley Basin

Huggett (1975) expanded on the catena concept proposing that the basic 3D unit of the soil landscape is the first-order valley basin. The functional boundaries of this soil landscape are defined as the atmosphere–soil interface, the weathering front at the soil base, and the drainage basin divides. The topographic boundaries of a drainage basin define the physical limits and direction of overland flow and thus control geomorphic processes such as erosion, transport, and deposition. Also, groundwater divides are generally coincident with topographic divides that are boundaries for overland flow. Thus, the first-order basin forms the natural boundary conditions for chemical and colloidal transport and redistribution via water transport on most landscapes (Figure 29.8a). First-order basins connect to higher order basins forming a larger, linked system across landscapes.

The partitioning of water into surface and subsurface flow is an important component of both Huggett's (1976) watershed model and the catena. Surface runoff or overland flow is the mechanism that drives geomorphic processes. Water that enters the soil drives chemical and biological reactions and via throughflow transports soluble or colloidal materials. The movement of water and associated materials through the soil landscape occurs as both saturated and unsaturated flow. Transport

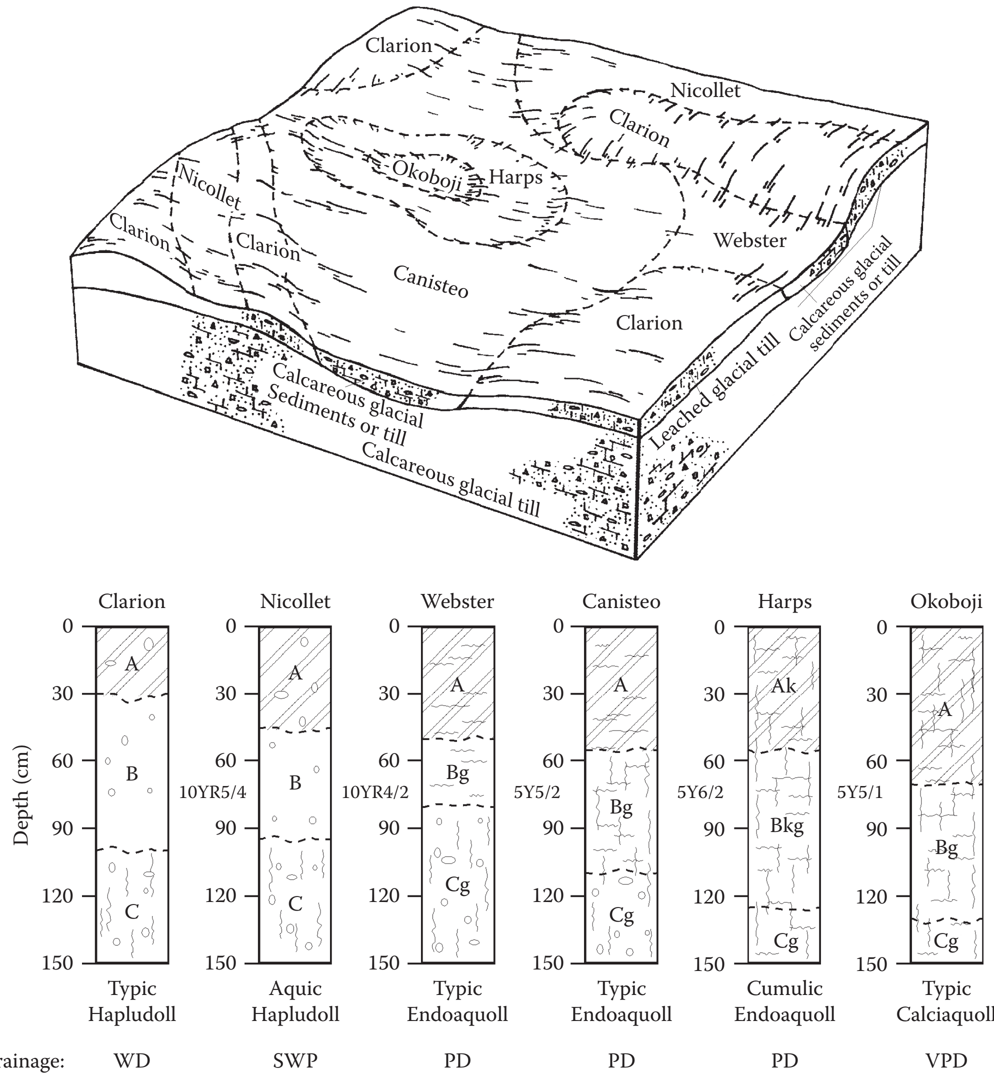


FIGURE 29.6 Two-dimensional diagram showing the landscape relations of the Clarion–Nicollet–Webster catena.

of material can occur between or within soil horizons, between soils at different landscape positions, or across an entire watershed. Deep percolation can transport material beyond the depth of a soil profile. A 3D approach must account for lateral, divergent, and convergent throughflow. Recent studies (Arndt and Richardson, 1989; Knuteson et al., 1989; Steinwand and Richardson, 1989), as well as earlier work (Glazovskaya, 1968; Cleaves et al., 1970; Crabtree and Burt, 1983), attest to the importance of throughflow and the geochemical link between soils, hydrology, and landscapes.

29.5.4 Open Basins

Drainage basins are either open or closed (Ruhe, 1969). An open drainage basin is confined by the head of the watershed and the perimeter divide. The mouth of the basin is open (Figure 29.8a). Surface water collected within the basin during precipitation events discharges through the outlet at the mouth. Eroded sediment can be transported and redeposited either within the basin or removed entirely from it. Both mass and energy can enter and leave an open basin by surface flow or runoff. The sediment retained within an open basin represents only a portion of the total sediment produced by erosion. The sediment package in an open basin is an incomplete record of the geologic history. A study of the sediment stratigraphy, therefore, yields only a partial understanding of the geomorphic evolution of the basin.

29.5.5 Closed Basins

Closed drainage basins lack a surface outlet. The drainage pattern within the basin flow to a central area or point (Figure 29.8b). No surface water leaves the basin as runoff. Water from precipitation is lost by evaporation or transpiration or infiltrates and becomes subsurface water. Mass and energy transfers driven by surface flow or runoff occur only within the basin. Surface flow does not remove mass from the basin in the form of clastic sediment. Mass can only

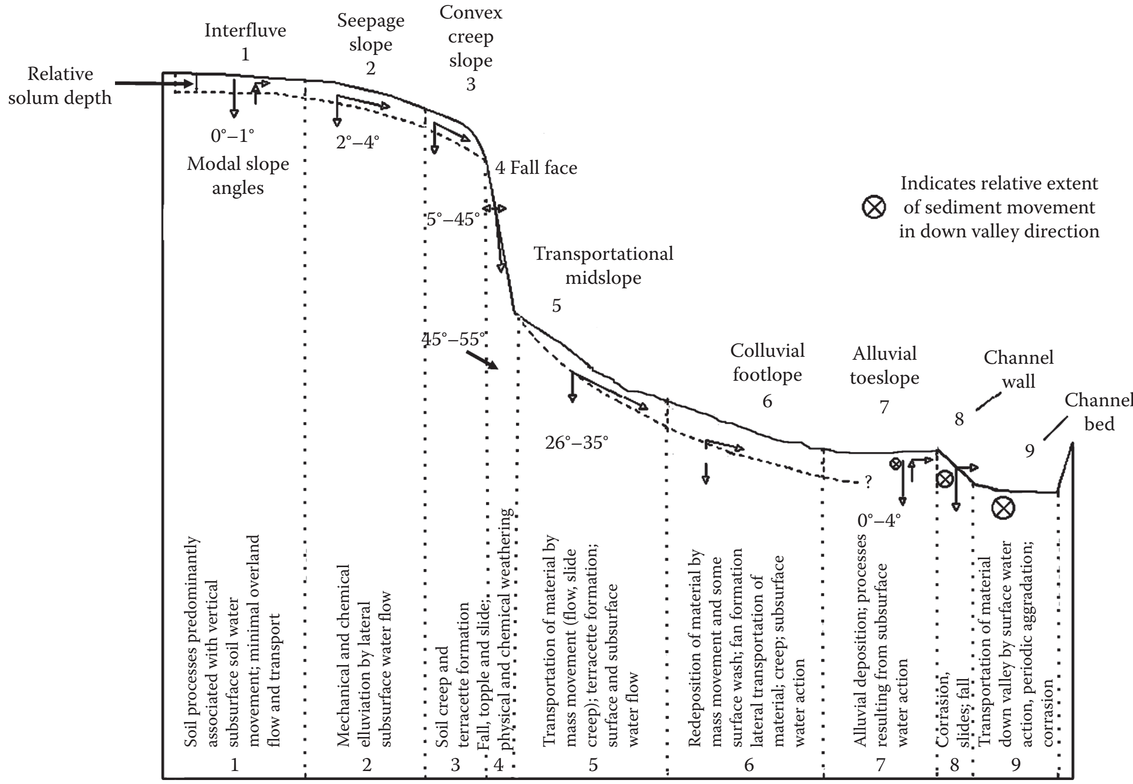


FIGURE 29.7 Two-dimensional nine unit soil and hillslope model (Dalrymple et al., 1968; Conacher and Dalrymple, 1977) showing soil hillslope relationships.

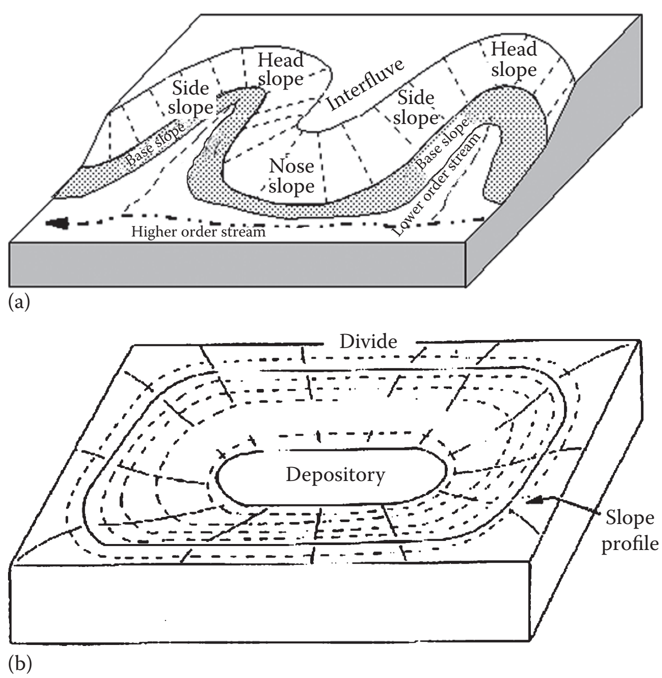


FIGURE 29.8 Three-dimensional diagram (a) open (Schoeneberger and Wysocki 1997) and (b) closed drainage basins as soil landscape models. (Modified from Walker, P.H., and R.V. Ruhe. 1968. Hillslope models and soil formation. I. Closed Systems. Trans. 9th Int. Congr. Soil Sci. 4:561–568.)

be lost as soluble or fine colloidal material in subsurface water. Mass can be removed or added to the basin by eolian processes, but the system is closed to surface runoff. The alluvial sediment contained in a closed basin forms a complete record of its erosional history.

The geomorphic concept of surface runoff that defines open versus closed basins is independent of scale. Open basins span the range from tiny, erosional rills on a hillslope to continental scale drainage basins like those of the Amazon or Mississippi rivers. Likewise, closed basins vary in size from small depressions like tree-tip pits to large tectonic basins such as Death Valley. The similarity is the lack of a surface outlet. The magnitude and complexity of the transfer processes within a basin increase with size. Surface flow never removes mass or energy from a closed basin regardless of size.

The lack of a surface outlet in closed basins precludes initial formation by water erosion. Closed basins originate through several mechanisms including deposition by ice or wind, dissolution of relatively soluble rock, subsidence due to structural failure of underlying rock often related to dissolution or groundwater fluctuations, and subsidence caused by faulting or tectonic downwarping. Closed basins are common in landscapes produced by these geomorphic processes. Examples include the "prairie potholes" and "chain lakes" of the glaciated midcontinent, depression lakes in dune topography such as the Sand Hills of Nebraska, karst topography created by dissolution of limestone, and the playa lakes of the southern Great Plains, which are partially related to wind deflation and deposition.

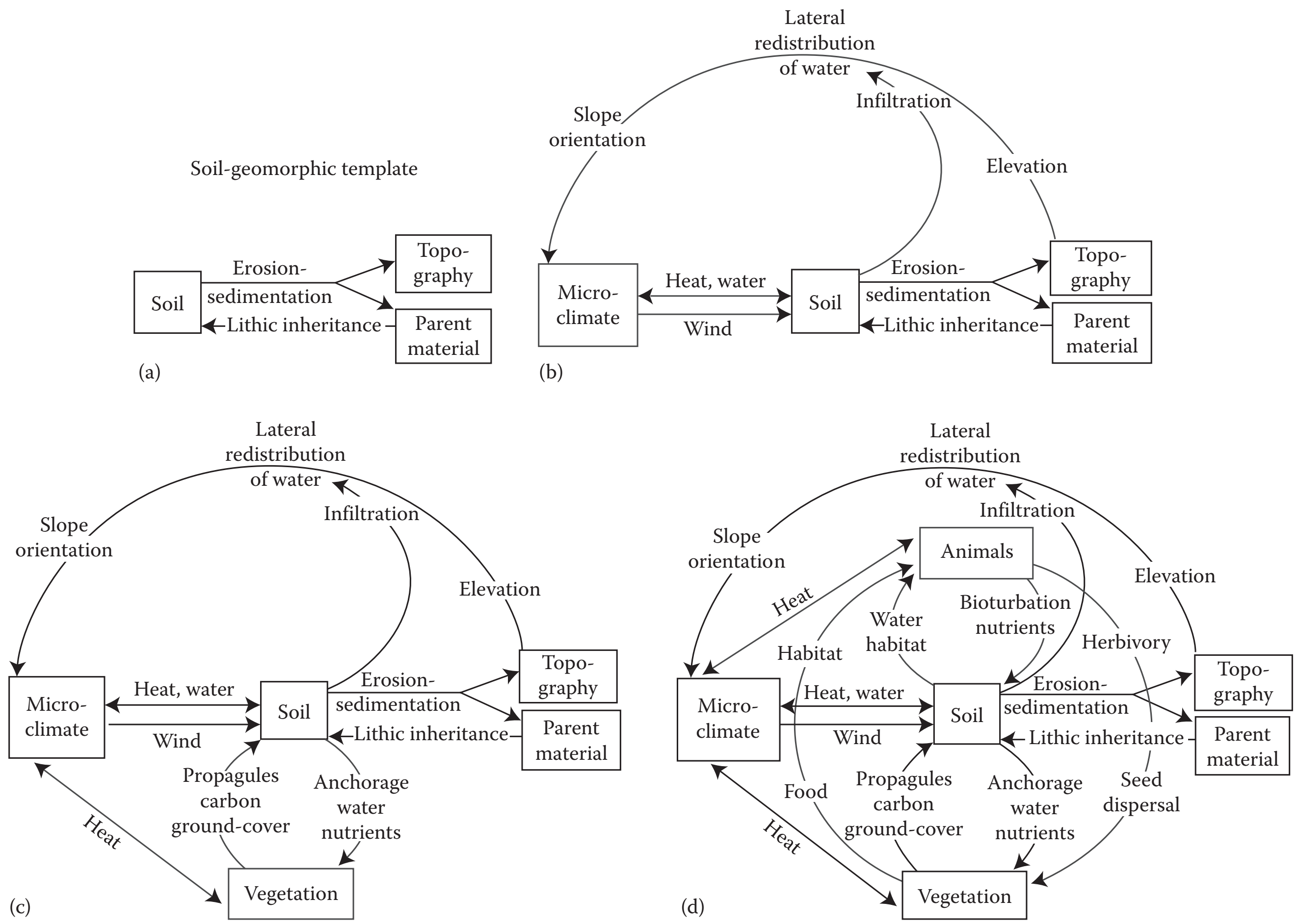


FIGURE 29.9 Linkages of the soil-forming factors in the soil-geomorphic template. (Modified from Monger, H.C., and B.T. Bestelmeyer. 2006. The soil-geomorphic template and biotic change in arid and semi-arid ecosystems. J. Arid Environ. 65:207–218.)

29.5.6 The Soil-Geomorphic Template

A more recent development in soil landscape models is the soil-geomorphic template introduced by Monger and Bestelmeyer (2006). It is a conceptual framework that depicts the interrelationships between the five soil-forming factors of Jenny (1941). Figure 29.9 illustrates these interrelationships. The soil-geomorphic template is defined as "…the soil, topography, and soil parent material" occurring on the landscape as a predictable pattern, which is reflected by the biotic community. Thus, it centers around the connection between topography and soil substrate on the one hand and ecological communities on the other. As such, it offers a framework to integrate various, often separated, disciplines to understand soil-geomorphic and ecological change across both short-term (centuries) and long-term (geologic) timescales.

The soil-geomorphic template has been incorporated in a proposed systematic procedure that can be used to unravel the complexity of landscapes (Peters et al., 2006; Figure 29.10). In this procedure, the researcher first identifies variation in the broad-scale environmental drivers and determines the spatial extent over which these drivers influence patterns on the landscape. Historical legacies are, then, identified and connections are made between these legacies and current landscape patterns.

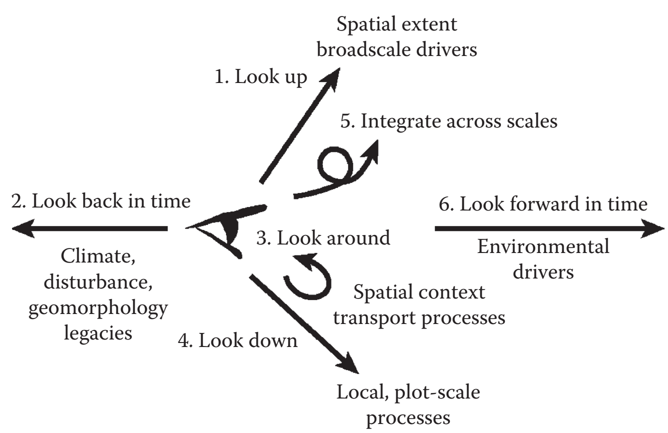


FIGURE 29.10 Six-step procedure to unravel landscape complexity. (Modified from Peters, D.P.C., B.T. Bestelmeyer, J.E. Herrick, E.L. Fredrickson, H.C. Monger, and K.M. Havstad. 2006. Disentangling complex landscapes: New insights into arid and semiarid system dynamics. Bioscience 56:491–501.)

Next, meso- and fine-scale spatial properties of the landscape units and the specific drivers connecting these units are identified. The information is integrated and key drivers and properties are identified at each spatial scale. Last, future effects from hypothesized changes in drivers and feedbacks are predicted.

29.6 Soil Hydrology

A major advancement in understanding soil systems has occurred in the last 20 years through insight into how water moves through landscapes: where the water goes, so goes soil development. Historically, the study of water flow in soils emphasized either overland flow, erosion, and sedimentation, or water movement within a pedon. Erosion and sedimentation control were foci of national soil programs and an integral part of agronomic applications of soil science. Efforts to understand and quantify water flow within a pedon or within fields (artificially delimited management areas versus naturally delimited bodies), such as for irrigation, have been an integral part of soil physics studies for over a century. However, water movement has not been pervasively integrated into the study of soil landscapes. To understand natural soil systems and thereby ecosystem behavior, conceptual and quantitative water-movement models must encompass water flow through soils, the vadose zone, and across and through landscapes.

Milne's concept of the catena (Milne 1936a, 1936b) recognized the existence of soils with different drainage classes across landscapes. His recognition of lateral transport as a contributing factor to landscape evolution has been de-emphasized over time. The main focus shifted to relatively static soil water conditions and vertical water flow (deep percolation) at a given position, hillslope or landscape position. Lateral transport was primarily recognized as overland flow with concomitant sediment sorting across the surface (Walker et al., 1966; Ruhe, 1975). The equally important processes of subsurface water flow and transport through soil landscapes were substantially ignored or forgotten. Soil hydrology studies in the last 20 years have expanded the catena concept to include water flow through landscapes (Arndt and Richardson, 1989; Steinwand and Fenton, 1995). There has been a simultaneous recovery from the historical emphasis on soil erosion caused by overland flow. The study of soil systems now includes the dynamic flow of water through as well as across soil and landscapes (Figure 29.11). This approach of dynamic water flow considers the entire vadose zone, not just the ground surface or the solum.

A variety of soil hydrologic terms have been developed or adopted, which partition water flow in soil systems. The terms can be placed in an idealized schematic to demonstrate relationships (Figure 29.12). These terms portray the potential fate of precipitation onto, into, and through soils. The emphasis is on water at or above the permanent groundwater table (i.e., the vadose zone). Water dynamics and aquifer conditions below the water table generally are the purview of groundwater hydrology. Water flow can be predominantly downward, lateral, or upward toward the ground surface, depending upon prevailing energy dynamics. This can be demonstrated by looking at prevailing water movement patterns in different climatic settings.

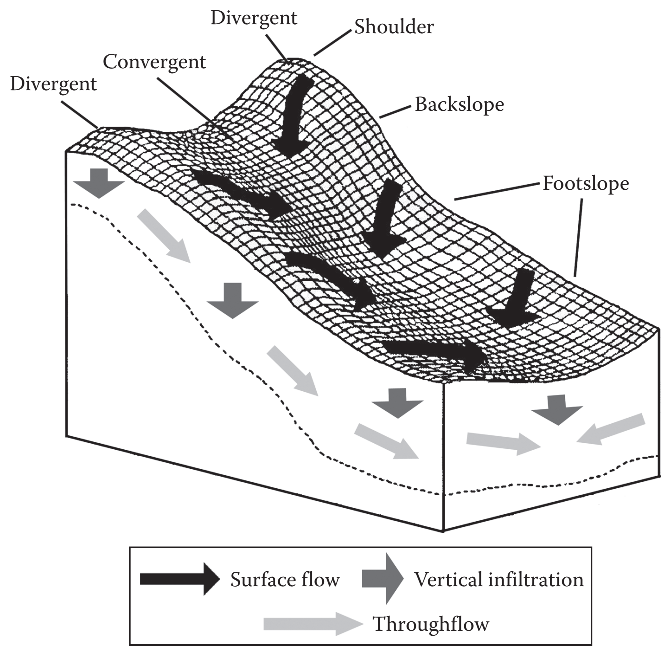
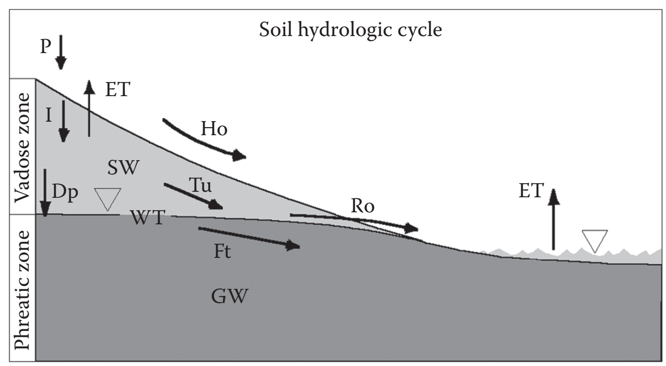


FIGURE 29.11 Three-dimensional hillslope with flow directions.



SW = Soil water	I = Infiltration
WT = Water table	Dp = Deep percolation
GW = Ground water	Tu = Throughflow
P = Precipitation	Ft = Flowthrough
ET = Evapotranspiration	Ro = Reflow
Ho = Overland flow (or runoff)	▽ = Zero tension water surface

FIGURE 29.12 Soil hydrologic cycle with terminology.

29.6.1 Climate Influence

Regional climate conditions control precipitation levels and patterns and evapotranspiration dynamics (either directly via solar radiation or indirectly through vegetation populations). All else being equal, regional climate determines the dominant water flow direction through the vadose zone. Within a given temperature range, precipitation levels can vary with predictable, generalized results. In humid environments, water tables and flow paths tend to loosely mimic the local topography with water generally infiltrating (i.e., *recharge*) in topographically high areas, flowing to and discharging from lower-lying areas (Figure 29.13a). In arid to semiarid environments, conditions are reversed, with recharge sites dominantly in low areas and subsurface water moving to and "discharging" from higher elevation areas (Figure 29.13b). Semiarid to subhumid environments behave as tension or transition zones between these two extremes. These transition zones temporarily follow either pattern depending upon short-term climatic conditions (e.g., seasonal climatic variations or annual cycles; Figure 29.13c).

Temperature also exerts a controlling climatic influence on soil hydrologic behavior both permanently and ephemerally. For example, permafrost effectively limits or precludes vertical water flow. If local conditions warm (human-induced or natural), permafrost melts and ceases to be a restriction to water flow. A seasonally ephemeral example occurs in areas that experience annual ground frost. The internal water flow behavior in soil is vastly different between winter, spring thaw, and summer (Emerson et al., 1990). Ground frost when present is an effective barrier to both subsurface water flow and the infiltration of surface water. During several weeks in the spring, thawed soil layers immediately above a remnant frost layer can become saturated and highly erosive (e.g., Willamette Valley, Oregon). After the soil completely thaws, subsurface flow is greatly enhanced and the surface layer is no longer saturated. The transport and fate of contaminants can be radically different depending upon seasonal soil conditions. This is the basis for discouraging land application of manure on frozen soil.

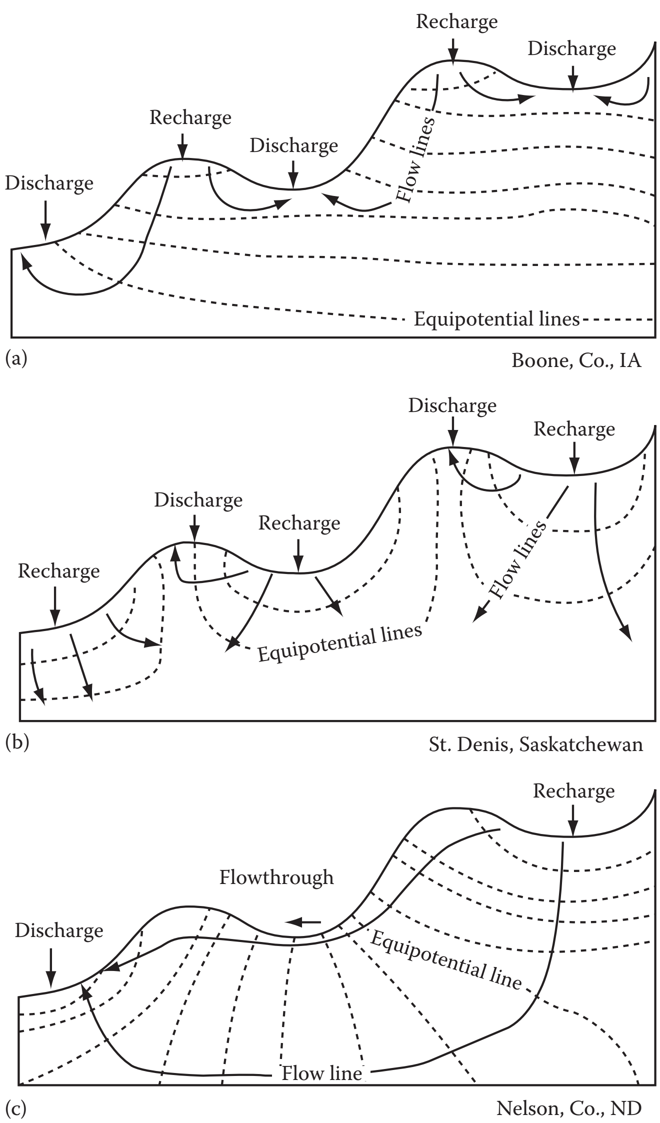


FIGURE 29.13 Two-dimensional landscape diagrams with recharge–discharge, flow lines, and equipotential lines in different climatic settings (a) humid, (b) arid to remiarid, (c) remiarid to subhumid.

29.6.2 Geostratigraphic Influence

Water restrictive (aquitard) or conductive (aquifer) sediments or rock strata can fundamentally alter subsurface water flow patterns. Tilted rock strata can preferentially redirect water such that infiltration in one location is shunted (laterally displaced) to an unlikely recharge or discharge site. Where bedrock strata approach or breach the ground surface, the vertical movement of subsurface water can be restricted, forcing water to flow back to the surface (i.e., *reflow*; Figure 29.12) and to form a spring, seep, or moist area. The composition of subsurface stratigraphic sequences and the extent to which they are connected to conductive materials also influence subsurface water flow. For example, a porous strata that is normally conductive will be nonconductive, if no outlets to other conductive material are available (e.g., a sand layer in a floodplain deposit confined by clay sediments; Figure 29.14). Similarly, a topographic position that precludes or minimizes water inputs may cause a strata that is normally conductive to be nonconductive.

29.6.3 Pedostratigraphic Influence

Some features unique to soil or derived from soil processes emulate geostratigraphic influences on water flow. Some pedologically derived layers such as an argillic layer (Schoeneberger et al., 1995), or duripan can restrict vertical water flow and enhance lateral flow (Figure 29.15). Water-repellent materials or layers (e.g., fire-induced hydrophobic layer in chaparral) can reduce infiltration of water into the soil or function as an aquitard and restrict vertical movement. Other pedogenic features or related phenomena such as pedogenic structure and biotic activity can greatly enhance water flow in upper soil horizons (Schoeneberger and Wysocki, 1996). Substantial differences in infiltration or internal water

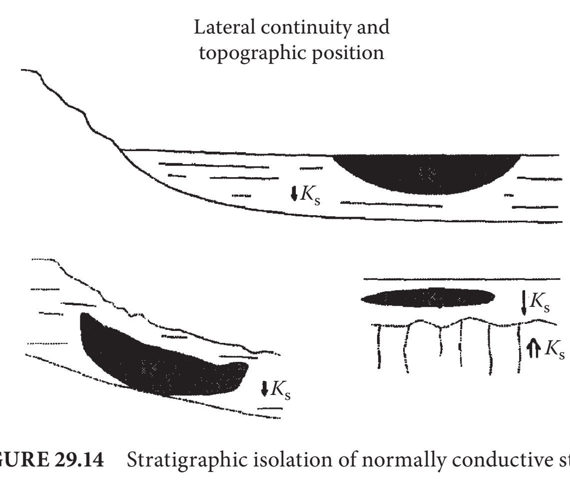


FIGURE 29.14 Stratigraphic isolation of normally conductive strata.

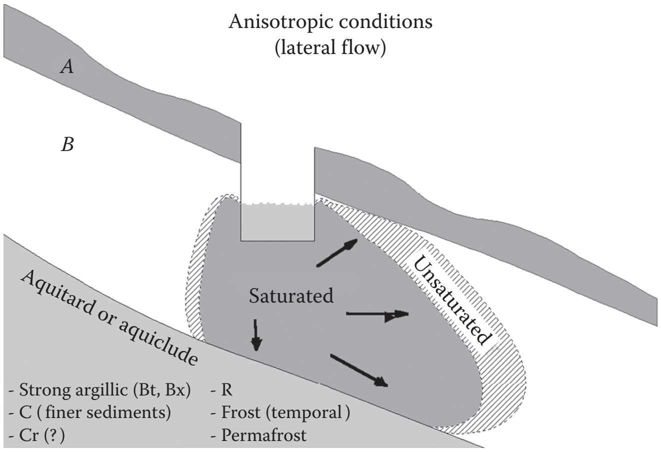


FIGURE 29.15 Pedogenic or geogenic acquitard restricting water flow through soil or strata. Note the A and B in the figure are soil horizon symbols. (From McAuliffe, J.R., E.P. Hamerlynck, and M.C. Eppes. 2007. Landscape dynamics fostering the development and persistence of long-lived creosote bush (*Larrea tridentata*) clones in the Mojave Desert. J. Arid Environ. 69:96–126. With permission from Academic Press.)

movement in a soil can result directly from different management practices (Franks et al., 1993, 1995).

29.6.4 Vegetative Pumping

Biotic activities, particularly plant respiration, can directly impact water flow patterns in soils. Plant communities can consume substantial volumes of water during transpiration. Water consumption may be sufficient to change local flow dynamics. Consider the case of phreatophytes fringing a marshes or riparian system (Figure 29.16). Minor differences in elevation (e.g., 10 cm) can promote plant establishment resulting in a disproportionately large reduction in water table, and subsequent differences in soils (e.g., the soil drainage class).

29.6.5 The "Soil Sponge"

An important but often overlooked function of soil is its role as a hydrologic "sponge." The porous character of soil allows much of the local precipitation to infiltrate. Obviously, the rate of water flow through the porous media of soil is less than that above ground. Infiltrated water is gradually transmitted vertically to recharge groundwater or laterally to be released as discharge to surface waters via throughflow. The hydrologic result is a delayed and dampened streamflow hydrograph. The importance of streamflow response can be seen from a practical, human perspective. The streamflow response for areas from which substantial amounts of soil are removed or sealed over (e.g., construction sites in urban areas) typically results in much higher and quicker peak flow compared to original conditions. This pattern can have a profound financial impact if structures are underdesigned (culverts, bridges, etc.) or if the new hydrologic dynamics initiates a new cycle of stream incisement or substantial change in sedimentation patterns (Hammer, 1995).

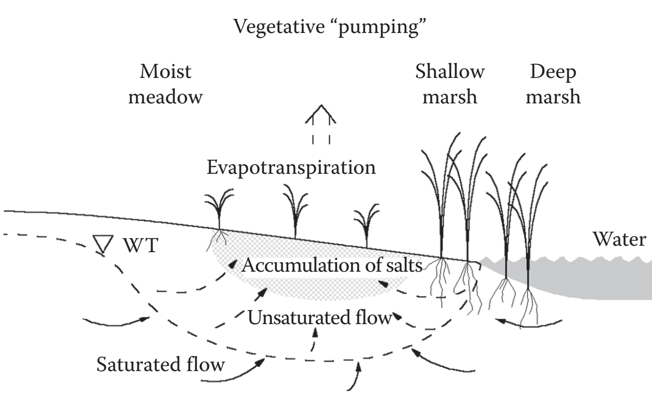


FIGURE 29.16 Diagram of vegetative pumping. (Drawn after Arndt and Richardson, 1989.)

29.6.6 Soil Morphology, Landscapes, and Water

Water is the dominant catalyst, mediator, and transport agent in most natural systems on earth. This is true for soils. Flora, fauna, and geologic substrate all affect soil and landscape development; water, however, is the controlling force providing both chemical and potential energy. The presence and flow of water in a landscape can be observed directly by excavation or piezometers. This is a tedious and time-consuming task given the seasonal or transient presence of water. The presence of saturated conditions and the movement of water in landscapes can be predicted from soil morphology (e.g., redoximorphic features). Soil morphology forms in response to long-term prevailing water state conditions. Soil patterns on a landscape result from differing water conditions and the prevailing water flow dynamics. Soil patterns on landscapes can explain where the water is and how it flows through landscapes. Soil morphology and soilscape models can explain and/or predict soil water dynamics and subsequent soil geography.

29.6.7 Eolian Influences

In addition to, or in lieu of water, wind can serve as a dominant mediator or transport agent on landscapes. Eolian processes directly affect soils and landscapes through the addition, redistribution, or loss of clastic materials, soil nutrients, carbonates, and soluble salts. The wind-blown sediments entrained through eolian processes can significantly alter soils and landscapes from individual landforms to entire regions. The effects of these sediments vary depending upon (among other factors) climate and localized landform characteristics and include the following: supplying essential nutrients to plants, influencing the character of near-surface horizons that directly affect surface hydrology, providing materials necessary for the development of certain morphologic horizons, and providing the parent material in which thick, well-developed soils have formed (Simonson, 1995). Examples of landscapes that are dominantly affected by eolian processes are those in which soils have developed in thick loess mantles, areas inundated with volcanic ash, dune systems, and regions with arid and semiarid climates. The impacts of loess, volcanic ash, and eolian sand (dunes) are generally geomorphically recognized. The influence of dust is less recognizable.

29.6.8 Arid and Semiarid Dust

Two profound ways in which dust can affect soil landscapes in arid and semiarid areas are as follows: the evolution of accretionary desert pavements and the formation of coppice dunes. Desert pavements are naturally occurring surficial features that consist of closely packed coarse gravel to cobble-sized rock fragments embedded into the underlying soil (Wood et al., 2005). These surfaces are often associated with silt-enriched horizons containing vesicular pores that directly underlie these pavements. Vesicular horizons are important because they significantly decrease infiltration rates into soils of these landforms thereby influencing the redistribution of water on these landscapes (Wells et al., 1985; Young et al., 2004). Desert pavements are the complex result of cumulic soil development whereby eolian fines are trapped by the surface, moved between peds into the soil, and into ped interiors through lateral conduits (Anderson et al., 2002). Both the surface armoring of the pavement and the low infiltration rate of the vesicular horizon limit the effective moisture in soils capped by desert pavement resulting in abundant quantities of near-surface nitrates and other soluble salts (Wood et al., 2005; Graham et al., 2008).

Coppice dunes form as a result of the trapping of entrained eolian silts and fine sands by shrubs. In the southwestern United States, these shrubs are commonly mesquite (*Prosopis pubescens*), greasewood (*Sarcobatus vermiculatus*), creosote bush (*Larrea tridentata*), or saltbush (*Atriplex* spp.) (Stuart et al., 1971; Rango et al., 2000; McAuliffe et al., 2007). Figure 29.17 illustrates the formation of coppice dunes. These dunes codevelop with the vegetation as the deposition of coarse silt and fine sand increases infiltration rates and enhances plant–water relationships. The dunes also provide ideal sites for burrowing fauna due to the lack of stones, abundant coarse material, and shading by the shrub. The burrowing activity further increases water infiltration into these areas. As plant-available water is increased in these areas, shrub survival and growth is enhanced so that more sediment is trapped causing the dune to further enlarge.

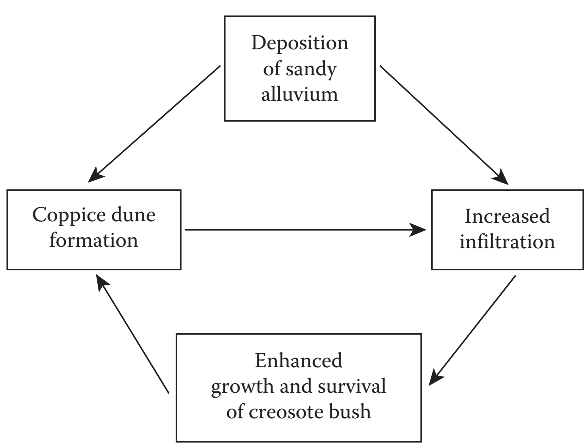


FIGURE 29.17 Diagram illustrating linkages in the codevelopment of creosote bush and coppice dunes in the Mojave Desert. (Redrawn from McAuliffe, J.R., E.P. Hamerlynck, and M.C. Eppes. 2007. Landscape dynamics fostering the development and persistence of long-lived creosote bush (*Larrea tridentata*) clones in the Mojave Desert. J. Arid Environ. 69:96–126.)

29.7 Geomorphic Description of Landscapes

29.7.1 Hillslopes

Soil patterns or sequences on hillslopes will in general follow a catenary relationship. The soil and landscape relationship inherent in the catena can be used to predict and describe soil occurrence. A fairly simple set of geometric or morphometric descriptors can be used to define hillslopes and therefore associated soil patterns. These descriptors include slope gradient, slope aspect, slope shape, slope complexity, slope position, and geomorphic component (position).

29.7.2 Slope Gradient

Slope gradient, which is measured along the vertical profile of a slope, is the angle of inclination of the ground surface from the horizontal plane. Commonly, slope gradient is expressed in degrees from the horizontal plane (e.g., 45°) or as a percent, which for a 45° slope is one unit of drop or rise per one unit of distance (slope gradient equals 100%). Slope gradient is a proxy measure of potential energy that drives mass movements and the erosive force of surface runoff on a slope.

Soil map units in a soil survey include a typical range for the slope gradient (e.g., Chemawa loam, 8%–15% slopes; Van Wambeke and Forbes, 1986). Inflection points or changes in slope gradient that repeat on a landscape are readily discernible and generally correspond to differences in the internal structure (underlying lithology), past erosion events (stream incision), or contacts between landforms or sediment bodies (Figure 29.18). Inflection points on slopes, therefore, often mark natural boundaries between soils. Soil surveyors use slope inflections as visible clues to changes in soils on landscapes.

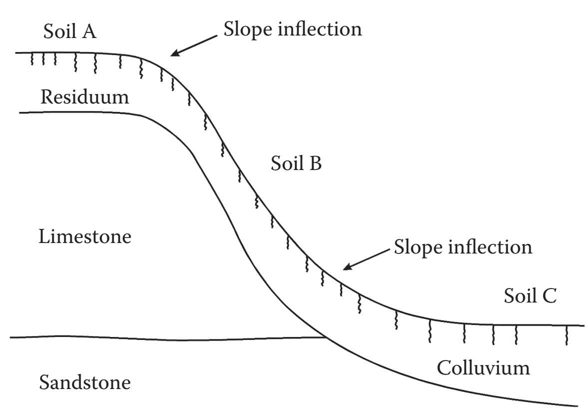


FIGURE 29.18 Diagram of natural landscape boundaries.

29.7.3 Slope Aspect

Aspect is the direction that a slope faces. Slope aspect is usually expressed as a compass azimuth (e.g., 215°) or as a cardinal direction (e.g., SW). Soil microclimates are strongly influenced by the amount of direct solar radiation, which is a function of slope aspect. In the northern hemisphere, north- and east-facing slopes are cooler and moister than south and west facing slopes. Distinct differences in soils occur as a result of aspect (Lotspeich and Smith, 1953; Finney et al., 1962; Franzmeier et al., 1969). The effects of slope aspect are more pronounced in mountainous or high-relief terrain than in low-relief areas. The influence of slope aspect is also more pronounced in temperate latitudes than in equatorial latitudes (Buol et al., 1989).

29.7.4 Slope Shape

Slope shape is the 3D geometry of a slope. The geometric form is obtained by combining the shapes of both the vertical profile (up and downslope) and the elevation contours (across slope). A 2D shape is either linear or curved. If curved, the shape can be convex or concave. This yields nine possible geometric forms to describe all slopes (Figure 29.19).

Slope shape is a property of hillslopes that strongly influences the movement of water both as overland flow and throughflow. For example, a slope that is linear in both vertical profile and contour shape creates parallel, lateral flow (Figure 29.19a). A slope that is convex in both profile and contour causes divergent flow (Figure 29.19d), and a slope concave in both profile and contour causes convergent flow (Figure 29.19h). Slope shape redistributes moisture received by precipitation, creating distinct microenvironments on the landscape. Areas of convergent flow are moister than the local climate whereas areas of divergent flow are dryer than the local climate. The influences of slope shape on water movement and soil moisture on landscapes in turn controls soil formation and vegetation.

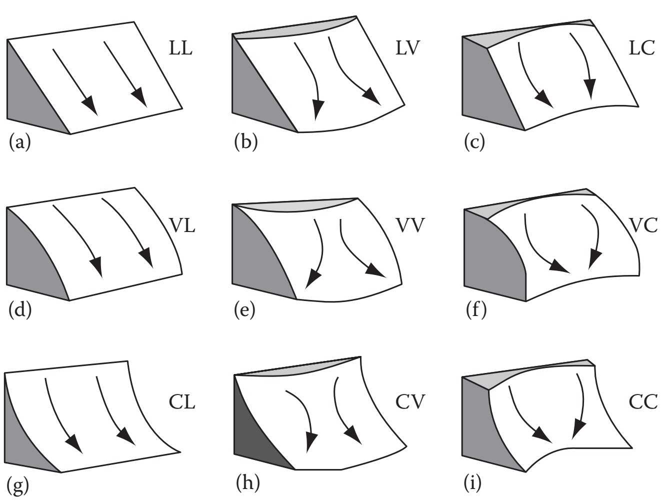


FIGURE 29.19 Three-dimensional diagram of the nine hillslopes shapes with flow lines shown: L = linear, V = convex, C = concave. (Schoeneberger and Wysocki, 1997).

In addition to its role in the redistribution of runoff, slope shape is a visual clue to the internal structure of a hillslope. Hillslopes that have a steeply convex, vertical profile are inherently resistant to erosion. Convex slopes occur predominately in landscapes where erosion is controlled by resistant bedrock. Soils on these slopes are shallow and usually display limited horizon development. In contrast, linear or concave slopes generally occur on unconsolidated material or weakly resistant rocks. Soils on these slopes are thicker and display greater horizon development.

Concave slopes or concave portions of slopes denote a decrease in slope gradient. Both potential energy and the velocity of overland flow decease at the inflection of a concave slope. An accumulation of colluvium or sediment derived from slope wash often occurs on concave slopes. Soils on these slopes often form by cumulic processes and usually have thick, poorly developed horizons compared to soils on adjacent uplands.

29.7.5 Slope Complexity

Slope complexity is a simplistic description of the ground surface in respect to the downslope path encountered by overland flow (Figure 29.20). Simple slopes are relatively smooth, with few obstructions to surface flow or decreases in slope gradient and thus nominal opportunities for sediment deposition. Complex slopes contain substantial irregularities in slope conditions.

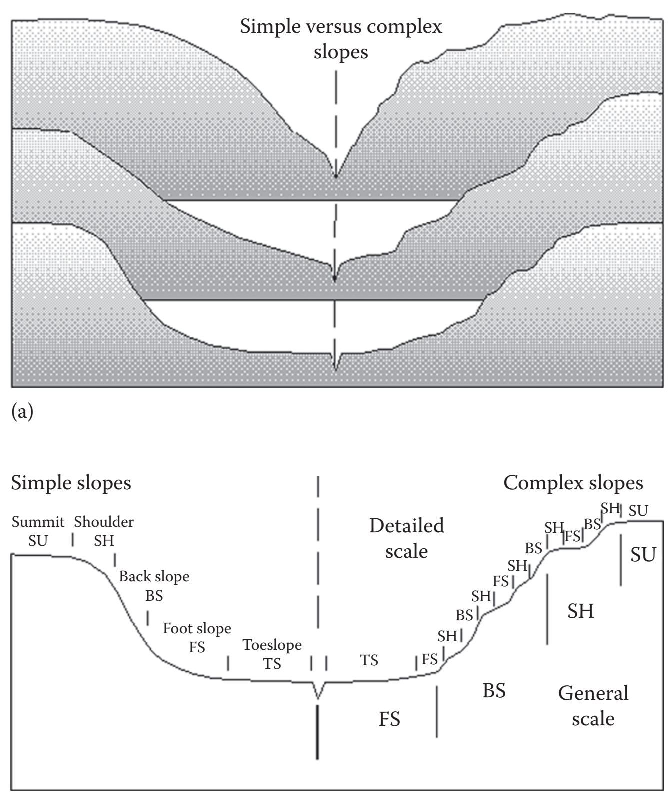


FIGURE 29.20 Two-dimensional diagram of simple versus complex slopes.

Surface water flow is apt to be interrupted and largely nonparallel with considerable changes in flow velocity and subsequent erosion and sediment transport capacity.

29.7.6 Slope Position

Slopes can be divided into segments or elements along a 2D, cross-sectional profile based on slope shape, the degree of erosion or deposition, the presence or absence of sediment, and the nature of the sediment. Wood (1942) identified four segments across a "fully developed slope" for a bedrock-controlled landscape. These elements are the waxing slope, free face, debris slope, and pediment. King (1957) elaborated on the geomorphic processes active on these four slope elements. The waxing slope is the convex crest of a hillslope dominated by chemical weathering rather than erosional removal. The free face is an outcrop of bare bedrock on the upper reach of a hill. Erosion is most active on this element. The debris slope occurs below the free face and is composed of material eroded and transported from the free face. Below the debris slope is the pediment, which is an inclined ramp extending from the hillslope base to an alluvial basin.

Ruhe (1960) modified Wood's 2D hillslope elements and applied them to the study of soil landscapes. Ruhe's hillslope profile elements include the summit, shoulder, backslope, footslope, and toeslope. These elements can be distinguished by inflections in slope gradient and line-segment shape (Figure 29.20). The *summit* is the relatively level, uppermost portion of a hillslope profile. It is the most geomorphically stable and least erosive part of a hillslope. The main vector of water flow is downward through the soil and erosional transport is minimal. Soils in this position display the greatest degree of profile development. The *shoulder* is the convex portion of the hillslope below the summit. The break between summit and shoulder is identified by an increase in slope gradient. The shoulder is subject to a greater degree of erosion, and greater lateral flow of water compared to the summit. Soils tend to be similar to, but thinner than those on the summit, and may appear to be vertically compressed or truncated. The shoulder descends to the steepest and more linear portion of the slope, the *backslope*, where surface runoff and erosional transport are greatest. The vector of water flow may be more lateral than vertical depending on the slope gradient. Soils may reflect inputs of less weathered parent material than soils on the summit and shoulder. On long slopes some degree of lateral sorting may be evident. The backslope descends to the concave portion of a hillslope, the *footslope*. The decrease in slope gradient reduces the carrying capacity of flowing water and increases sediment accumulation. Water flow vectors may be primarily lateral, but the position is characterized by the concentration of water from upslope. The footslope merges downslope with the *toeslope*. The toeslope is predominantly linear or slightly concave. The comparatively low slope gradient and low-lying position at the toeslope allows alluvial processes to dominate, given sufficient surface water from upslope or adjacent streams. Soils tend to be deep, comparatively moist, and composed of or strongly influenced by alluvial sediments. Toeslope sediments from lower order streams reflect short distance transport and a lower degree of fluvial modification (sorting, rounding, stratification, etc.) than toeslope sediments derived from higher order streams.

29.8 Geomorphic Components

A major focus of soil geomorphology is the movement of water through soil bodies and the way that a landscape sheds or concentrates water (surface flow). This focus is the result of the impact of surficial fluvial processes on major agricultural areas and population centers. Consequently, terms have evolved to describe portions of the earth's surface that share a common location, form, and geomorphic process and the practical emphasis on surficial water movement.

In addition to the 2D hillslope elements (Figures 29.20 and 29.21; Ruhe, 1960), Ruhe refined and popularized geomorphic descriptors for 3D pieces of landforms (Ruhe, 1969, 1975). These area descriptors, called *geomorphic components*, are based in part upon the convergent, linear (or parallel), or divergent nature of overland water flow (Figure 29.19) and associated sediment transport. The most widely used suite of geomorphic components was developed for and is most appropriately applied to hills. Historically, this set of terms has been applied to most landscapes, including mountains, due more to the lack of alternatives than to their utility. Other geomorphic settings (mountains, terraces, flat plains) have unique dynamics or complexity and warrant different area descriptors.

29.8.1 Geomorphic Components: Hills

The *interfluve* is the uppermost area of a hill, and represents the oldest, most stable part of the landscape, typically with the most developed soils. A noted exception is where opposing hillslopes have narrowed an interfluve or merged (e.g., a saddle or crest) to the extent that erosion begins to lower the crest. Hillsides are areas characterized by active backwearing (erosion) in general, and at the heads of streams in particular.

Hillsides can be divided into discrete parts (Walker and Ruhe, 1968) based on the dominant behavior of overland water

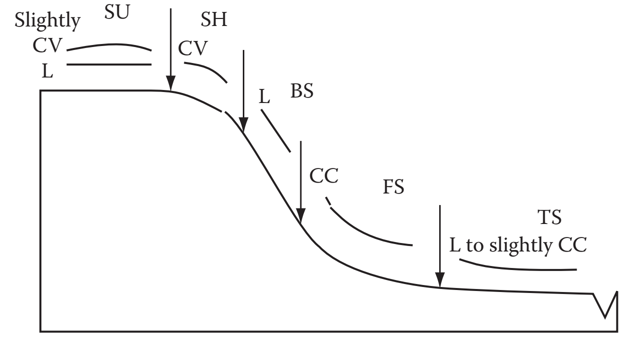


FIGURE 29.21 Hillslope profile terms and associated slope shapes.

flow (Figure 29.8a): converging (*head slope*), linear or parallel (*side slope*), or diverging flow (*nose slope*). Soils on hillsides are dominated by colluvial sediments in gradual transport down the slope and modified to varying degrees by slope wash (nonchannel, overland flow). Unless masked by changes in parent materials, some degree of lateral sorting downslope is usually present. Soil profiles on hillsides can range from thin to thick depending on the rate of erosion (high or low, respectively), the magnitude of the slope gradient, and the extent to which the bedrock is resistant to weathering and erosion. In many landscapes, the traditional assumption of a prevalence of shallow soils over hard rock on hillsides and mountainsides is erroneous (Knox, 1982; Graham, 1986). The *base slope* is commonly an apron or wedge of colluvium at the bottom of hillslopes (Schoeneberger and Wysocki, 1996). This drape of transported material can range from coarse debris to finer sediments that have been winnowed or sorted by slope-wash processes. The base slope does not typically include sorted and stratified alluvium associated with channel deposition. Distal base slope sediments commonly grade into or interfinger with alluvial fills.

29.8.2 Geomorphic Components: Mountains

Mountains represent a unique geomorphic setting due to their scale and slope complexity. Mountainsides commonly have long, complex backslopes up to thousands of meters long, steep slope gradients, highly diverse sediment mantles and complex, near-surface hydrology. Mass movement processes and features are more prevalent than in hills. Consequently, new area descriptors have been developed (Schoeneberger and Wysocki, 1996; Hirmas, 2008) that effectively identify and name geomorphic components for mountains.

The *mountaintop* (Figure 29.22) is the summit or crest of a mountain and is commonly characterized by comparatively short, simple slopes composed of bare rock, residuum, or short-transport (angular) colluvial sediments. In humid climates, soils on mountaintops can be quite thick (Oliver et al., 1997), whereas in arid climates, these soils may be capped by well-developed desert pavements and accumulate considerable quantities of carbonate, nitrate, and other soluble salts (Graham et al., 2008). The side of a mountain, the *mountainflank*, is characterized by comparatively long, complex slopes dominated by mantles of long transport (subangular) colluvium. Residuum, if present, is usually buried by 1–2 m of colluvium. Incised drainageways, structural benches, and mass movement features can be common. Rock outcrops, while visually prominent, are not a major portion of the land surface. The *mountainbase* is an apron of colluvium at the bottom of a mountain slope and analogous to the *base slope* in hills. It is marked by a substantial decrease in slope gradient compared to the mountainflank. The mountainbase is characterized by a thick mantle or wedge of colluvium and commonly contains a comparatively high percentage of coarse rock fragments. The colluvium can extend out onto more level land surfaces, and ultimately interfingers with or is buried by alluvium, or thins and joins re-emergent residuum. In desert environments, windward sides of these mountainflanks and mountainbases can trap significant amounts of dust within which, thick soils may develop (Blank et al., 1996; Hirmas, 2008). Hirmas (2008) described a feedback process in these landscapes where sediment is added to lower-lying landforms through alluvial and debris-flow processes while, concurrently, material entrained from these landforms is carried back to the mountains by wind to serve as parent material for mountain soils (Figure 29.23).

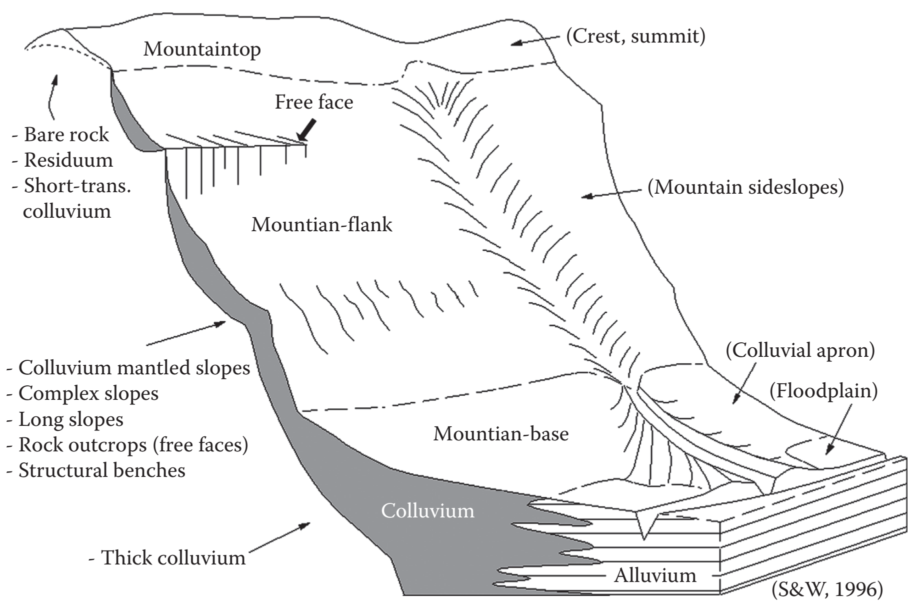


FIGURE 29.22 Geomorphic components for mountain landscapes. (From Schoeneberger, P.J., and D.A. Wysocki. 1996. Geomorphic descriptors for landforms and geomorphic components: Effective models, weaknesses and gaps. (Abstract). American Society of Agronomy, Annual Meetings. ASA, Indianapolis, IN.)

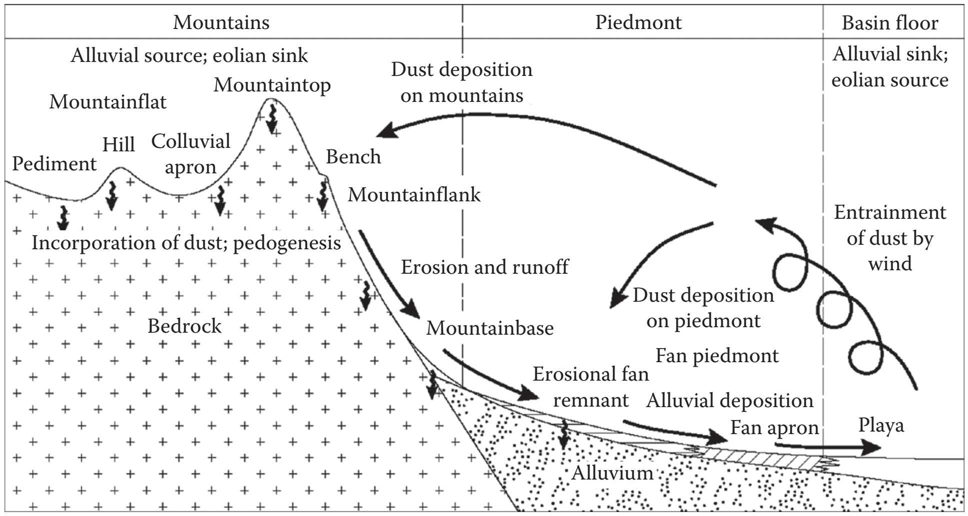


FIGURE 29.23 Idealized geomorphic model of desert mountain–piedmont–basin floor interactions. (After Hirmas, D.R. 2008. Surface processes, pedology, and soil-landscape modeling of the southern Fry Mountain Bolson, Mojave Desert, California. Ph.D. Dissertation. University of California. Riverside, CA.)

29.8.3 Geomorphic Components: Terraces

Terraces are landforms that form a relatively level or gently inclined surface, a constructional strip, or plain that borders a stream, lake, or sea. Terraces are a unique geomorphic setting and have dynamics quite different from those of hills or mountains. Stream terraces and floodplain steps (Figure 29.24) are originally developed by alluvial processes and sediments, rather than the dynamic slope processes and colluvial sediments that characterize hills and mountains or the localized, low energy processes typical of flat plains.

The *tread* is the comparatively broad, generally level part of a terrace or floodplain step. Treads can extend laterally for many kilometers (Gamble, 1993; Saucier, 1994). Treads are level or gently inclined (low gradients) and underlain by alluvial, lacustrine, or marine sediments. Strath terraces are similar in form to other stream terraces, but are erosional landforms characterized by thin alluvial sediments over an eroded bedrock bench or platform.

The *riser* is an escarpment that separates terrace or floodplain levels. The riser commonly consists of a short, steep, planar slope cut into the sediments that underlie the adjacent tread. The areal extent of a riser rarely exceeds tens of meters and therefore is typically depicted on a soil map with a spot symbol rather than a delineation. Geomorphically, risers represent an abrupt change to a lower hydrologic base level, which suppresses the water table along the edge of the adjacent, higher surface. This directly affects soil processes and subsequent soil geography. Soils above and adjacent to a riser tend to be better drained than those farther away from the escarpment (Figure 29.25). Daniels and Gamble, 1967 called this relationship the "red edge effect." Soils at the base of a riser tend to be as wet as or wetter than those farther from the scarp. This is especially true in floodplains.

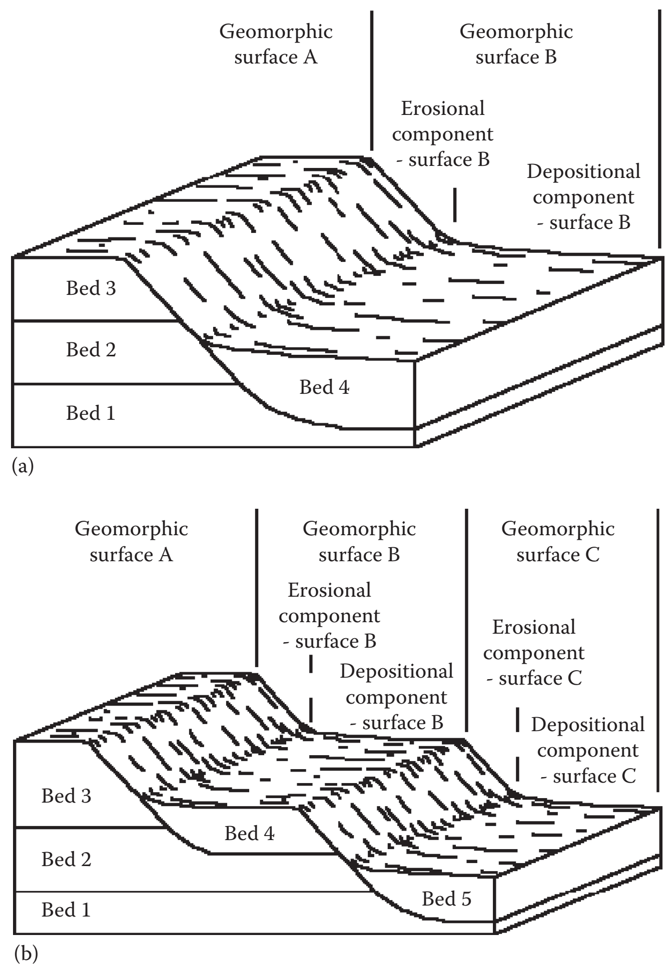


FIGURE 29.24 Geomorphic components for terraces and floodplains.

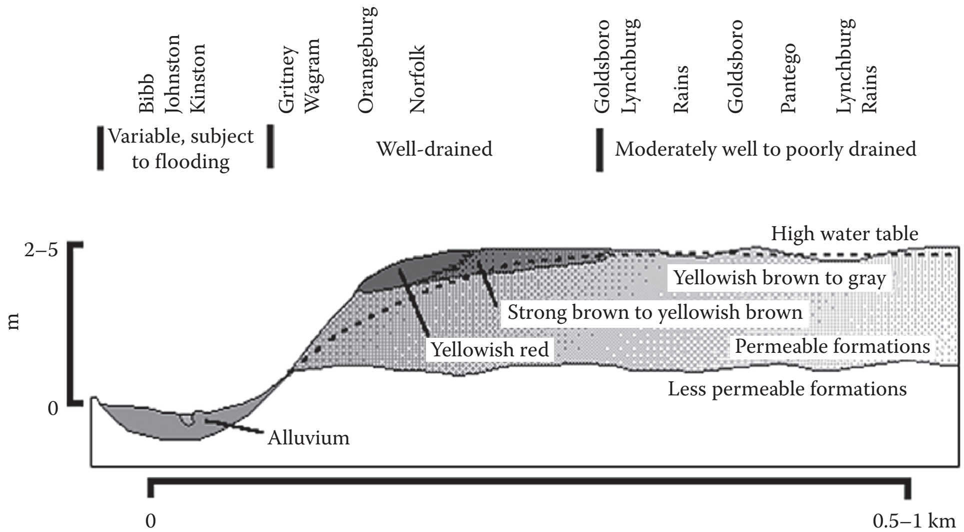


FIGURE 29.25 Two-dimensional landscape diagram showing drainage relationships on low-relief divides.

29.8.4 Stream Terraces versus Floodplain Steps

Fluvial landforms include the active channel, surfaces that are intermittently inundated, and higher areas that no longer flood or receive alluvial sediments. The distinction between stream terraces and floodplain steps is an arbitrary break in a natural continuum. There are practical pedological reasons for separating areas that are within the active floodplain from areas that flood or receive sediments in rare instances. For most land uses (e.g., home site), it is valuable to know the relative occurrence of flooding. Terraces are defined as a part of the fluvial system that no longer actively experiences fluvial modification (Ruhe, 1975), that is, no longer aggrading (receiving additional alluvial sediments), no longer experience significant flooding. Floodplain steps are morphometrically similar to terraces, except that they occur within the fluvially active portion of a floodplain and are subject to relatively frequent modification (experience relatively regular or significant flooding and alluvial sediment inputs). Pedologically, terraces commonly have more extensively developed soils (e.g., alfisols, ultisols) compared to lesser-developed soils (e.g., inceptisols, entisols) of floodplain steps. The practical separation of terraces from floodplain steps varies. Stream terraces that have not flooded during recorded history could be inundated during rare, catastrophic events (e.g., 500 year flood). Soil development on terraces is usually controlled more by moisture derived from precipitation than from the adjacent river. Terrace soils in cold or dry climates, despite a greater age and lack of new sediment, may closely resemble soils on adjacent floodplain steps that periodically receive fresh sediment. Conversely, soils on floodplain steps in hot, humid settings may exhibit extensive development and resemble soils on the older, adjacent terraces. In temperate North America, a practical separation between terraces and floodplain steps (Figure 29.24) is the 100 year flood stage (Soil Survey Staff, 1998).

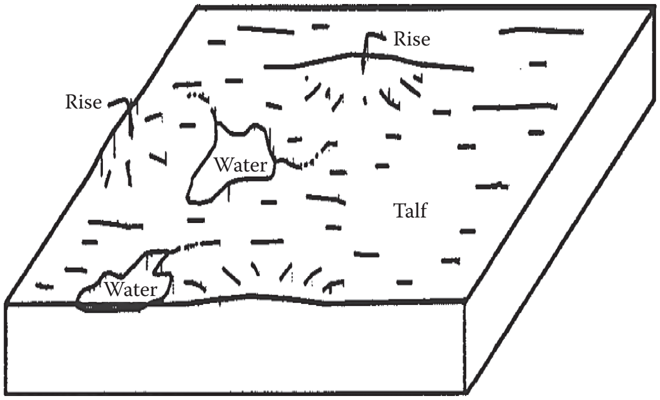
29.8.5 Geomorphic Components: Flat Plains

Broad, flat plains such as proglacial lake plains, low-lying coastal plains, and low-gradient glacial till plains are a geomorphic setting distinct from hills, mountains, or terraces. The primary differences are geomorphic origin (nonfluvial) and the low potential energy to drive water flow. Historically, generic terms (e.g., flat, broad interstream divide, rise) or terms developed for other settings have been applied to flat plains with unsatisfactory results. Recent interest and recognized value of wetlands and hydric soils (e.g., Mausbach and Richardson, 1994) highlights the need for unique descriptors for flat areas. Small differences in elevation (e.g., 15 cm or less) on a landscape can mark the change from moderately well drained to very poorly drained soils (Richardson, 1997). The search is presently on for better, unique descriptors to express the geomorphic components of flat plains. The following terms are proposed as provisional contenders.

Flat plains are subdued landscapes dominated by broad areas with low slope gradients (e.g., 0%–1%) called *talfs* (Figure 29.26), and closed depressions, with only nominal changes in local relief (i.e., microhighs, microlows). A *rise* is a slightly elevated area (i.e., microhigh), which tends to be broad with low slope gradients (e.g., 1%–3%). The low gradients of flat plains result in minimal erosion by running water, especially if vegetated. Consequently, fluvial drainage networks tend to be poorly developed: nonintegrated, incipient, or deranged. Precipitation tends to pond locally and lateral transport is slow both above and below ground. These conditions favor accumulation of organic matter (e.g., pocosins, upland bogs) and retention of water and fine earth (<2 mm diameter) sediments.

29.8.6 Microrelief

As used here, microrelief refers to slight variations in the height of a land surface that are too small to delineate on a



- Very low gradients (e.g., slope 0%–1%)
- Deranged, non integrated, or incipient drainage network
- "High areas" are broad and low (e.g., slope 1%–3%)
- Sediments commonly lacustrine, alluvial, or till

FIGURE 29.26 Geomorphic components for flat low-relief plains.

topographic or soils map at commonly used scales (e.g., 1:24,000 and 1:10,000). Examples of microrelief include microhighs and microlows. These minor elevational differences can have a surprisingly large impact on vegetation distribution, soil hydrologic dynamics (Hopkins, 1996), and sedimentation patterns. This approach to microrelief is strictly morphometric. We make a distinction between small elevational differences (*microrelief*) and small features (*microfeature*) with which these elevational differences are associated. An example is gilgai (a kind of microfeature), with the associated microrelief (mound areas = microhighs, bowl areas = microlows). It is an unfortunate yet common practice to confuse the two items and describe both as "microrelief."

29.9 Landscapes, Landforms, Microfeatures, and Anthropogenic Features

Armed with the ability to describe the most detailed morphometric nuances of the land surface, one is inevitably faced with some variation of the question: "Taken all together, what is this feature that we see and what should it be called?" A potpourri of names has been applied over time that connotes internal composition, form, arrangement, collective relationships, and origin. These terms range in scale from small, human scale features to continental-scale assemblages. Confronted with such complexity, some organization of terminology is reasonable and in fact, necessary. Various schema have been developed to achieve this, typically from a particular geographical perspective: Developed schema include physiography (Fenneman, 1931, 1938, 1946; Thornbury, 1965, 1969; Hunt, 1967), landuse (USDA-SCS, 1981), and ecology (Omernik, 1987, 1995; Bailey et al., 1994; USDA-EPA, 1996). A different approach is to de-emphasize the geographical context and focus on geomorphic process (Peterson, 1981). One pseudohierarchy of terms has been assembled specifically from and for soil survey and geographic applications (Schoeneberger and Wysocki, 1997). In addition to morphometric terminology and physiographic location, this system loosely arrays land surface features in a progression of scale in the following fashion.

A *microfeature* is a small, local, natural form (feature) on the land surface that is too small to delineate on a topographic or soils map at commonly used map scales (e.g., 1:24,000 to 1:10,000). *Note*: The conventional use of microrelief commonly encompasses some of the terms or features that in this system are contained within microfeature (see below).

A *landform* is any physical, recognizable form or feature on the earth's surface, having a characteristic shape and range in composition, and produced by natural causes; it can span a wide range in size (e.g., *dune* encompass both *parabolic dune*, which can be several tens-of-meters across, as well as *seif dune*, which can be up to 100 km long). Landforms provide an empirical description of similar portions of the earth's surface.

A *landscape* is a collection of spatially related, natural landforms, usually the collective land surface that the eye can comprehend in a single view (Soil Survey Staff, 1998).

An *anthropogenic feature* is an artificial feature on the land surface, having a characteristic shape and range in composition, composed of unconsolidated earthy or organic materials, artificial materials, or bedrock, that is the direct result of human manipulation or activities. Historically, landforms and the like have been defined as, and therefore restricted to, natural features. There are relatively consistent formational processes with common compositional, structural, stratigraphic results within a population of natural features (e.g., sand dunes). Human-made features may have a common theme in intent (e.g., water impoundment), but can have an almost limitless variety of formational processes, material composition, and internal structure. This simplifies the study and interpretation of the land's surface but ignores human-made features, which progressively apply to ever more of the earth.

29.10 Age Assessment of Soil Landscapes

Age assessment of landscapes is the single most important contribution that geomorphology makes to the understanding of soils. The law of superposition and the concept of geomorphic surfaces are the key principles for defining relative age relationships in a landscape (Daniels et al., 1971; Hall, 1983). Radiometric, isotopic, or paleontological dating combined with field studies using geomorphic and stratigraphic principles can establish the absolute age of a deposit, soil, or a landscape.

The longer a landscape is exposed to subaerial weathering, the greater the potential for soil development. The terms "weathered," "well developed," and "old" are relative indicators of time in soil formation, but none are accurate descriptors of age. *Weathered* refers to the relative stability stage (Goldich, 1938;

Jackson, 1968) of the minerals contained in a soil. The *weathering stage* is a function of age, weathering intensity, and mineral stage of the parent material. A chronologically young soil may be composed of highly weathered minerals if the parent material is preweathered.

Development is a relative measure of soil formation based on the type and degree of horizonation. Soil development is a function of both age and the rate of horizon formation. Rate of horizon formation depends on the intensity of soil-forming processes (see Chapter 30) and composition and resistance of the parent material to change. Soils of similar morphology can differ substantially in age. For example, Burges and Drover (1953) reported that spodic horizons in Australia formed in sandy beach deposits within 1000–3000 years. Franzmeier and Whiteside (1963) found that spodic horizons took 8000 years to form on dunes in northern Michigan. Holzhey et al. (1975) suggested 21,000–28,000 years for development of thick spodic horizons in North Carolina. Well-developed, weathered soils are described as *old*. In this context, *old* is a proxy for age based on weathering and degree of soil formation. Geomorphic principles are a means to assess soil and landscape age independent of horizon development or weathering stage.

The law of superposition specifies the sequence of deposition of sedimentary rocks and unconsolidated sedimentary deposits. Sediment can only be deposited atop preexisting materials. Younger beds or strata invariably overlie older beds, if not overturned. Bed 2 (Figure 29.27a) is younger than Bed 1. Once deposition ceases, the top of Bed 2 represents a depositional surface (surface A) equal in age to the youngest strata in the bed. Stabilization of the surface allows a soil to form in the sediment (Figure 29.27b). This is time zero for subaerial weathering and soil formation. The surface and soil age are the same. The passage of time is recorded by an increasing degree of soil development. Soil formation requires a period of nondeposition and therefore indicates an unconformity in the sedimentary sequence. A buried soil represents an unconformity in the sedimentary sequence. For example, deposition of Bed 3 buries the older units and the soil formed on them (Figure 29.27c). Bed 3 is the youngest deposit. The top of Bed 3 is a new, younger depositional surface B. The stratigraphic record contains three depositional beds and a buried soil.

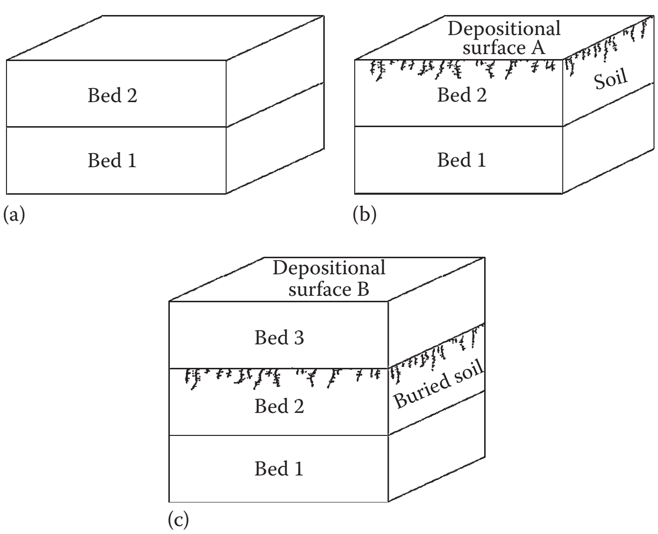


FIGURE 29.27 Three-dimensional block diagram depicting the law of superposition.

29.10.1 Geomorphic Surfaces

In a landscape, an erosion cycle or event (e.g., stream incision, hillslope retreat, pedimentation) that cuts existing beds creates a geomorphic surface of erosional origin (Figure 29.28). This erosion surface is younger than any strata, soil, or surface that it cuts. An erosional surface can only cut preexisting materials, strata, or surfaces. As you ascend a landscape geomorphic surfaces increase in age. This is the principle of ascendancy. Surface E1 is younger than surface A and younger than beds 1, 2, and 3. The cutting of an erosional surface produces sediment that is deposited downslope or down valley. The erosional surface descends or grades to a corresponding depositional surface and the underlying alluvium (Figure 29.28). The depositional

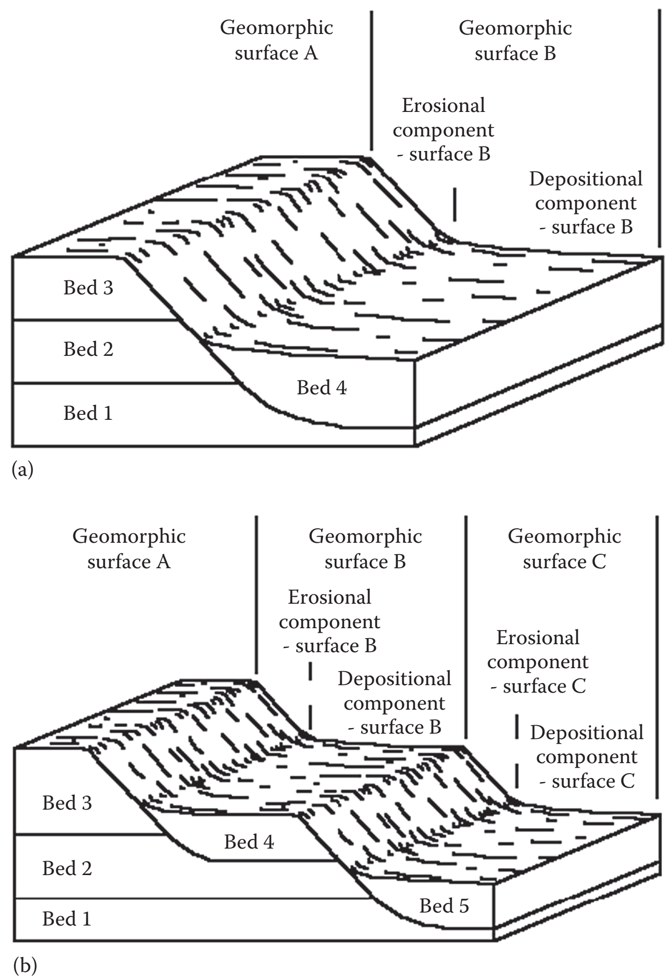


FIGURE 29.28 Three-dimensional block diagram showing geomorphic surfaces and associated time relationships.

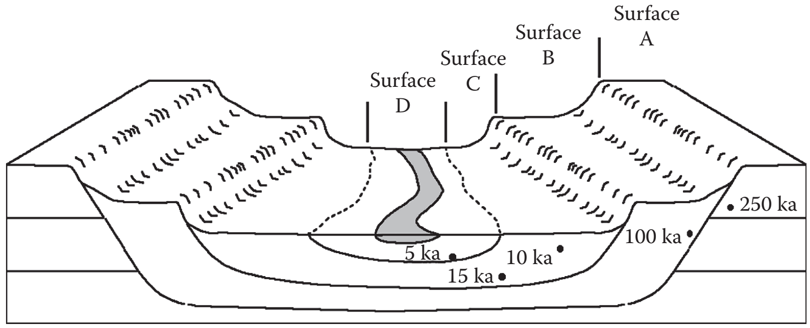


FIGURE 29.29 Three-dimensional block diagram showing geomorphic surfaces and dating of sediments.

and erosional surfaces and the top of the alluvium are all the same age. An erosional surface is the same age as the alluvium to which it descends. This is the principle of descendancy. Relative age of an erosional surface in the field can be determined by observing what it slopes downward to in a smooth, concave profile (Daniels et al., 1971).

29.10.2 Stepped Surfaces

Landscape evolution commonly proceeds episodically through time, which creates a series of stepped or faceted geomorphic surfaces along interfluves (Ruhe et al., 1968; Gamble, 1993; Figure 29.29). The age relationships of stepped surfaces follow the principles of ascendancy and descendancy. Surface A on the interfluve summit is the oldest, surfaces become successively younger as you descend the landscape. The age differences between surfaces depend on the timing of the erosional processes that produced them.

29.10.3 Absolute Age Assessment

Dating techniques along with the geomorphic relationships provide a framework for establishing a landscape chronology. For example, ages shown in Figure 29.29, which could be obtained by C14, luminescence techniques, or fossil analysis, establish the periods for deposition. Surface D is underlain by sediment with an age of 5 ka at the base. This unit is inset into an alluvium that has an age of 15 ka at the base and 10 ka near the top. Stream incision and back filling therefore occurred between 5 and 10 ka that deposited the alluvium under surface D. Surface C includes both an erosional depositional element. Cutting of the erosional surface occurred between 10 and 15 ka based on the alluvium to which it descends. The alluvium is a time transgressive deposit as is the cutting of the erosional surface. Alluvium under surface B has an age of 100 ka and the erosional element ascends to a surface of 250 ka. The erosional processes that created surface B began later than 250 ka and ended by 100 ka. Given this scenario, the absolute age differences between the surfaces and the soils on them can be established.

29.10.4 Geomorphic History

The concepts as displayed in Figures 29.27 through 29.29 are straightforward. Application to soil landscapes requires detailed stratigraphic control obtainable only by field observations from natural (e.g., stream banks) or man-made (e.g., road embankments) exposures or drill cores and the ability to think in three dimensions. The geomorphic history of a region must sometimes be established through detailed stratigraphic investigation before soil patterns are completely understood (Ruhe et al., 1967; Daniels et al., 1970).

Consider the geomorphic history of a hypothetical landscape in Figure 29.30. A landscape is composed of soil s1 on geomorphic surface A, which is formed in a single deposit (Figure 29.30a). The deposit is buried by a younger material (e.g., alluvium, glacial till, loess) with subsequent development of soil s2 on surface B (Figure 29.30b). The landscape now consists of two stacked stratigraphic units separated by a buried soil s1. Stream incision cuts an erosion surface across both units (Figure 29.30c). The landscape becomes a two stepped sequence comprising surfaces B and C. Soil s3 forms in surface C (Figure 29.30d). Two soils s2 and s3 of different age make up the landscape. Valley alluviation buries surface C and reduces relief in the landscape (Figure 29.30e). Soil s4 forms in surface D on the alluvium (Figure 29.30f). The soil landscape at this point consists of two soils s2 and s4, which again differ considerably in age. Stream incision and back filling inset a deposit into the preexisting alluvium (Figure 29.30g). The inset is denoted by surface E and soil s5. The landscape now includes three surface soils s2, s4, and s5, all of different age. Soil distribution on the landscape, soil age, and the geomorphic history of this landscape can only be established from multiple field observations and drill cores at critical locations that identify all the stratigraphic units and buried soils.

29.10.5 Soils and Geomorphic Surfaces

Despite the strong association between soils and geomorphic surfaces, soil patterns and morphology cannot be used to initially define or recognize geomorphic surfaces. Geomorphic surfaces must first be defined using geomorphic principles

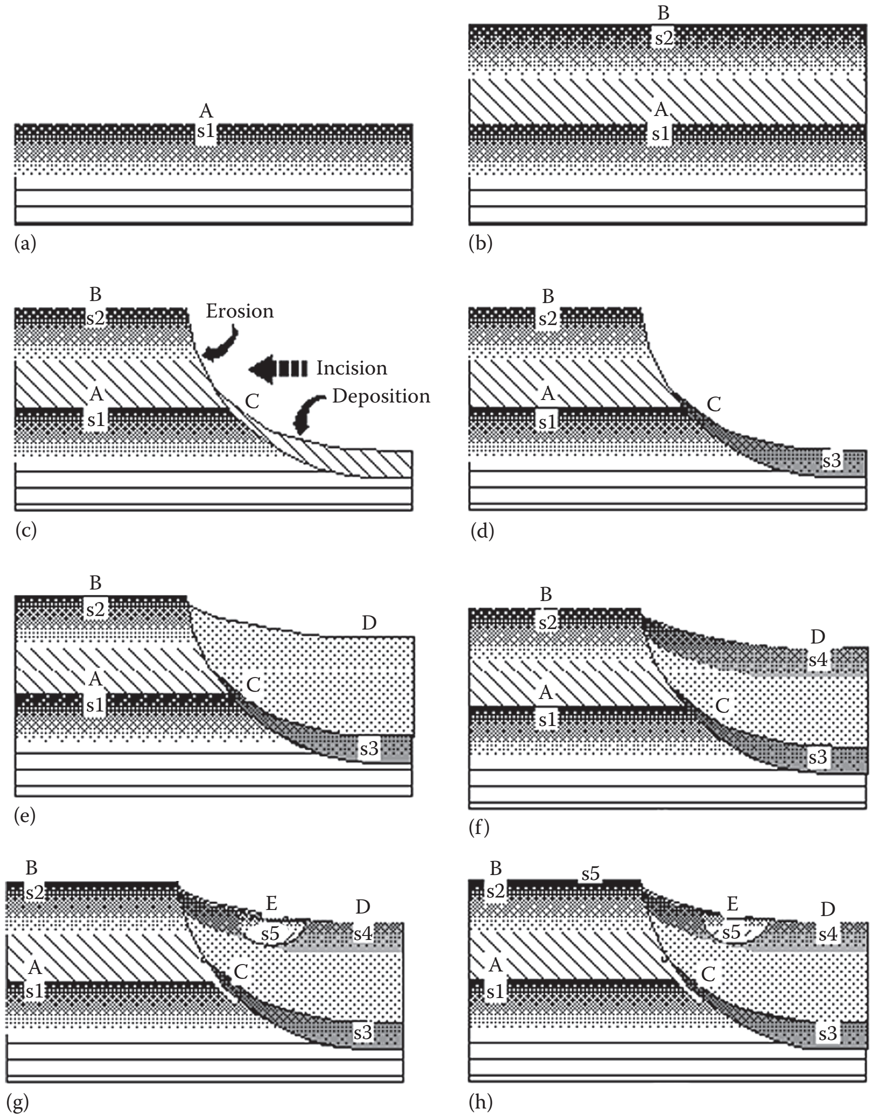


FIGURE 29.30 Two-dimensional diagram explaining hypothetical landscape evolution.

and then the linkages made to soil patterns. A soil or suites of soils may occur on more than one geomorphic surface and soil boundaries do not always correspond to geomorphic boundaries on a landscape. Consider Figure 29.31, which shows a stratigraphic sequence and soil horizons cut by an erosional surface. Soils 2, 3, and 4 on erosional surface B form over beds of varying composition and age and from preexisting soil horizons. Soils 2, 3, and 4, however, are all the same age (the age of surface B). In this case, three different soils occur on the erosional surface. The sequence of soils is determined by the stratigraphic units present and erosional sorting along the surface. The same erosional surface across its geographic occurrence can cut different strata or sediments. The soils in that case would differ.

The boundary between soil 1 and soil 2 (Figure 29.31c) lies below the slope break that denotes the erosional surface. Erosion has not removed enough of the original profile to cause a change in the soil. The boundary between the surfaces and the boundary between soils are not coincident. A soil map and a geomorphic surface of the same landscape would not have coincident boundaries.

29.11 Paleosols, Geosols, and Climate Interpretation

29.11.1 Paleosols

Geologic processes that create and destroy soils and landscapes vary in time and space. Continental-scale processes like glaciation can create or destroy entire landscapes. More commonly landscapes evolve by erosion and deposition on only parts of them. Erosion removes soils and sediments from the active parts of landscapes, while soils continue to form on the stable parts. Sediment from

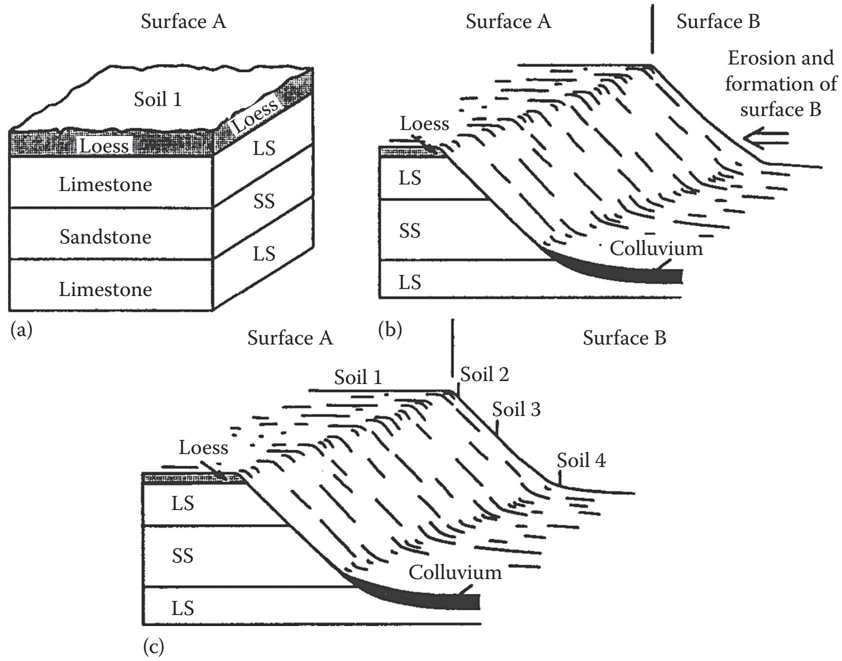


FIGURE 29.31 Two-dimensional diagram relating soils and geomorphic surfaces.

erosion buries preexisting surfaces, soils, and sediments. Buried soils and soils that endure on stable landscapes for long periods become part of the geologic record and are valuable for interpreting earth history. *Paleosols* are soils that formed on landscapes of the past (Yaalon, 1971; Valentine and Dalrymple, 1976). Age alone, however, does not denote a paleosol. The process or processes responsible for the soil morphology must no longer operate due to a change in climate, local environment, or because of burial.

Ruhe (1965) recognized three types of paleosols buried, exhumed, and relict. *Buried paleosols* form at the land surface, but later are covered by sediment, which removes the paleosol from the dominant soil-forming zone. Burial must be deep enough to suspend the soil-forming processes, and rapid enough to prevent formation of a cumulic soil profile. Postburial and diagenetic changes in soil properties and horizons are common (Yaalon, 1971; Olson and Nettleton, 1999) and must be considered when studying paleosol composition.

Exhumed paleosols form at the land surface, are buried, and later re-exposed on younger landscapes by erosion of the covering sediment (Ruhe, 1969). Erosion exposes both the paleosol and the former land surface. Exhumed paleosols rarely contain completely intact paleosolums. Former A horizons are usually truncated or mixed with the burial sediment and difficult to recognize. In contrast, B horizons are the best preserved and most easily discerned feature of exhumed paleosols. An exhumed paleosol differs from a buried one in that the outcrop can be traced across a significant area of the present land surface. An exhumed paleosol can be out of balance with both the existing climate and the adjacent soils on the landscape. Landscapes that include exhumed paleosols may not conform to the expected catena relationship in a geographic area (Figure 29.32; Ruhe, 1969).

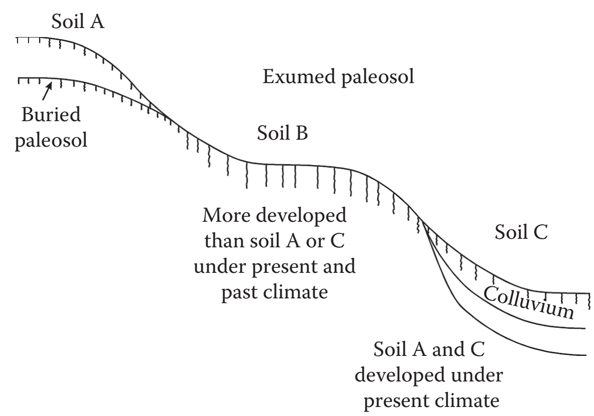


FIGURE 29.32 Landscape diagram showing buried and exhumed, paleosols.

Relict paleosols are soils formed on preexisting landscapes and were never buried (Ruhe, 1965). Therefore, identification of relict paleosols is somewhat problematic. A relict paleosol must have endured one or more shifts in regional climate or local environment (e.g., lowering of a water table by stream incision or baselevel change) such that it is no longer is in balance with the existing conditions. This concept requires that the age of the landscape be known or inferred and that we have a precise understanding of the processes responsible for soil morphology (see Chapter 30 for discussion). Soils on old, easily distinguishable geomorphic surfaces such as terraces (Figure 29.33) are the most commonly recognized relict paleosols.

Our present knowledge of soil processes is inadequate to determine relict versus nonrelict soil conditions in all situations.

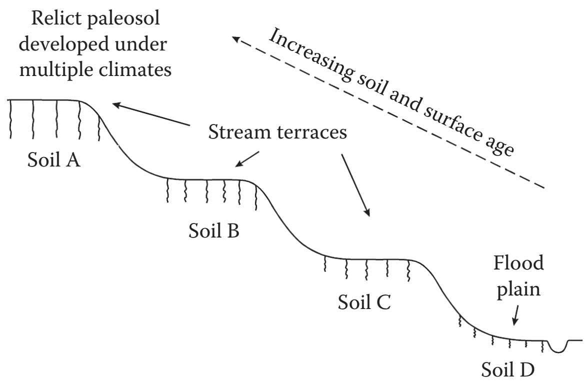


FIGURE 29.33 Landscape diagram depicting relict paleosol.

Soil-forming processes can be reversible self-terminating, steady state or metastable, or irreversible self-terminating (Yaalon, 1971). This means that all morphological features do not have equal utility as indicators of climate change. For example, the formation of silica-cemented soil horizons, or duripans, occurs in arid or semiarid climates (Soil Survey Staff, 1996). Once formed, a duripan is resistant to destruction even in humid conditions. Thus, the presence of a duripan in a humid climate is strong evidence of a relict condition. In contrast, some morphological features such as A horizon thickness and color, which have a strong correlation to present climatic regimes, can adjust rapidly to a shift in climate. Despite the difficulties in defining and recognizing relict paleosols, this concept is valuable because it focuses on the relationships of landscape age and soil processes.

29.11.2 Geosols

Buried paleosols are integral components of the stratigraphic record. Formation of a soil requires a period of landscape stability with exposure to subaerial weathering. A buried paleosol, therefore, (1) represents an unconformity in a sedimentary sequence and (2) marks a consistent stratigraphic position at the top of a sedimentary body or bodies. Buried paleosols that are laterally traceable with 3D form and have mappable distribution at a scale of 1:24,000 can be formally defined as pedostratigraphic units (Morrison, 1967) called *geosols* (NACSN, 1983). Geosols include only pedogenic horizons formed in preexisting rock or sediment bodies. Buried O horizons and some C horizons of paleosols are excluded from geosols. A geosol may form in parent materials of diverse ages and composition. Formal naming and establishing of geosols follows the same criteria of other stratigraphic units (NACSN, 1983). A geosol is named after a geographic feature in the type area.

29.11.3 Climate Change and Paleosols

Jenny's (1941) model can be solved for an individual soil-forming factor. For example, the equation

$$\text{Soil} = f'(c)_{p,r,b,t} \tag{29.1}$$

ascertains the climate during soil formation, where the subscripts denote a constancy in the factors of parent material, relief, biota, and time. There are problems with this approach, however, Equation 29.1 defines the soil-forming factors as independent variables, but in reality the factors covary. It is also difficult to assign quantitative values to some factors, particularly parent material and the biota. Nonetheless, this approach allows useful, semiquantitative comparisons between soil morphology and individual soil-forming factors.

Processes of soil formation occur over time scales of tens to thousands of years. Soil morphology forms a physical record of the climate, vegetation, and/or environment during the time of formation. Recent emphasis on global climate change has renewed interest in paleosols as indicators of past climates. Numerous studies have used soil morphology to interpret past climates or environments (Busacca, 1989; Driese et al., 2005; Retallack, 2005; Brock and Buck, 2009). The interpretation of climate based on paleosols requires one or more of the following: (1) that the paleosol existed on a stable landscape and had reached a steady state within a particular climate, (2) identification of the ancient soil environment (e.g., floodplain), and/or (3) identification of the catenary relationship or position of the paleosol on the ancient landscape.

For example, a climatic interpretation based solely on the properties of the Clarion profile (Figure 29.6), such as A horizon thickness, organic carbon content, depth to inorganic carbonate, and B horizon color are indicative of a well drained, prairie soil formed in the present humid climate of central Iowa. If only the Harps soil (Figure 29.6) is considered, which has a thicker, darker A horizon, a gray B horizon, and inorganic carbonates are at or near the surface, a climate more humid than the Clarion morphology is indicated. Carbonates near the surface suggest a climate more arid than Clarion and the A horizon thickness suggests a cooler climate. Despite occurring only tens of meters from Clarion soils on the landscape, the morphology of Harps soils leads to an ambiguous and a potentially erroneous interpretation of the regional climate. Climatic interpretations based on paleosol morphology must include paleocatenary relationships.

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30

Pedogenic Processes

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30.1 Introduction

Soil is a mixture of geological parent material, living organisms, and the colloidal residue of their interaction. The colloidal fraction is particularly important in imparting soil with its unique characteristics. The nature of soil exposes a large surface area that gives rise to adsorption of water and chemicals, ion exchange, adhesion and capillarity, swelling and shrinking, and dispersion and flocculation. Soil colloids include inorganic (clay minerals) and organic (humus) components. The nature and amount of these colloids vary in response to environmental stimuli and organic input. Their variation is responsible for lateral differentiation in soil properties. Soils are separated vertically into horizons that indicate differences in the internal soil environment, which, in turn, determines the amount and character of soil colloid accumulation. This chapter provides an overview of the way in which soil properties vary as soils develop under varying environmental stimuli.

Soil formation is powered by gradients in chemical and physical potential as Earth's atmosphere and biosphere interact with rocks and minerals. Rock minerals are formed at high temperature and pressure, which makes them susceptible to attack by water and the biospherically derived acids existing in soils. The fundamental mechanisms of physical, chemical, and biological processes in soils are discussed in Parts I, II, and IV. Here the focus will be on the matrix of processes that produce 3D soil bodies called pedons, which have differentiated horizons and lateral variability.

30.2 Environmental Factors That Drive Pedogenesis

The concept of soil-forming factors is one of the earliest and most important of soil science. It defines soil as a component of ecosystems that must be characterized in terms of both geological substrate and biological input (Jenny, 1941; Amundson and Jenny, 1997). It provides a far-reaching description of the controls on soil processes that allows prediction of general soil properties based on the interaction of a relatively small number of driving variables. In the words of Hans Jenny (1980), "Pedogenic order in a landscape is unraveled by stratified random sampling along vectors of the state factors." Knowledge of the soil properties produced by the interaction of environmental variables allows prediction of soil characteristics in detail, locally, or in general, regionally and globally. The fact that characteristics of soil properties can be predicted based on environmental variables is critical to soil resource mapping because it is impossible to observe soil profiles everywhere.

The interaction of environmental variables establishes the soil-forming processes whose actions on parent material manifest themselves in characteristic soil morphologies, which may, in turn, alter the nature of the ongoing processes. Environmental variables are broadly grouped as geological, climatological, biological, and topographical. The geological characteristics constitute a site factor that sets the initial condition for soil formation, whereas the climatological and biological factors represent energy input that drives soil development. The energy factors share a strong spatial covariance because of the dependence of organisms on climate. Within any of these groupings, local differences in topography modify the activity of the more broadly defined variables. Human manipulation of edaphic environments is significant enough to warrant close consideration of their effect on soil properties (Amundson and Jenny, 1991). Dynamic soil properties (e.g., soil organic carbon, bulk density, pH, salinity, aggregate stability) that change on timescales from decades to centuries are most affected by human land use (Tugel et al., 2005).

30.2.1 Geologic Foundation: Lithology, Mineralogy, and Topography

Geological processes ranging from plate tectonics and volcanism to deep ocean carbonate sedimentation to glaciations create conditions that eventually provide substrate for soil development. Plate tectonics drive global processes that differentiate lighter, silica-rich rocks from the mantle into the crust and create relief that produces erosion, accumulation of alluvium, and formation of sedimentary rocks. Continental areas that are actively uplifted, such as the Tibetan plateau and the cordillera of South and North America, support thin, young soils. In contrast, those areas on stable cratons, such as central West Africa, Brazil, and Australia, have deep soils that are superimposed on thick, highly weathered saprolite zones (Paton et al., 1995). In these ancient cratons, depth to unweathered rock can exceed 100 m. Volcanism brings locally melted silica-rich rock in subduction zones to the surface in areas such as Japan, Indonesia, and Washington state. Hot spots, such as Hawaii, the Canary Islands, or Yellowstone, provide a direct conduit from mantle to surface for Fe- and Mg-rich rocks. A dramatic example of mantle-fed volcanism is exhibited by flood basalts on the Columbia and Deccan plateaus. On ocean floors, far from continental inflow, deposition of carbonate produces limestone that, when subaerially exposed, has distinct parent material properties when compared with igneous rock.

In polar and temperate zones, glacial advances deposit moraines and outwash composed of the mix of lithologies crossed by the glacier. The more finely ground glacial drift is lofted into the atmosphere and deposited as loess. Numerous Pleistocene glaciations have left distinctly younger soils than the deeply weathered soils of stable craton areas in the tropics. Even within temperate zones, there are strong differences in soil properties between recently glaciated areas and those that have not been overrun. Globally, much of the variation in soil properties can be attributed directly to differences in parent material mineralogy and the dynamics of Earth's history. Locally they are important as well. For instance, stream channel and associated hillslope erosion may cut through an extensive surficial geological unit into an underlying lithology that has different mineralogical and chemical properties, which affect soil properties of soils that form on that unit (Chapter 29).

Parent material lithology determines the physical and mineralogical nature of soil. For example, soil formed in marine clays or shale inherits a large quantity of colloids that lead to poor water infiltration, relatively low plant-available water, and substantial shrink/swell behavior. In contrast, soils formed on quartz sandstone or weathered granite are likely to be sandy with little inherited clay. Commonly, they are excessively drained and often have low fertility and water-holding capacity. Fracture patterns in the underlying bedrock provide preferential lines of water flow, which enhance weathering and soil formation. The stratigraphic juxtaposition of soft and hard lithologies provides gentler and steeper relief and less or more rock fragments in the overlying soil.

Parent material mineralogy determines ecological soil properties, such as nutrient supply and retention, and water movement. For example, mafic mantle-derived rocks typically weather to a smectite and Fe (hydr)oxide-rich colloidal fraction with the simultaneous release of Ca and Mg for use by plants and soil organisms. The smectites provide high cation exchange capacity (CEC), which can easily retain nutrient cations. Felsic crust-derived rocks tend to weather to a kaolin- and gibbsite-rich colloidal fraction with release of Ca and K for use by the biota. Kaolin and gibbsite provide only a small amount of cation exchange capability, leading to relatively rapid leaching of nutrient cations. The presence of quartz in the felsic rocks provides a skeletal framework of resistant sand size grains that tends to promote water percolation. In contrast, the smectites derived from mafic rocks tend to form clay-rich horizons. The clays tend to swell when wet, which can reduce water flow by closing off macropores. Often igneous rocks contain a broad suite of minerals that weather to provide a mixture of nutrients and a diverse array of secondary minerals. In contrast, limestone is quite pure calcite, which should only release Ca and H_2CO_3 when it reacts with water. Soils forming on limestone often have inherited their colloidal fraction from weathering of impurities in the rock or from eolian addition. Similarly, soils forming on quartz sandstone or quartzite must inherit their colloidal fraction from eolian input.

30.2.2 Energy: Water, Temperature, and Biology

Soil development is strongly dependent on energy that is ultimately provided by the sun and Earth's gravitational field. Soil is heated directly by energy absorption at the surface and indirectly by transfer from the atmosphere. These processes set up superimposed diurnal, seasonal, and annual waves of energy moving through soil profiles. Soil is cooled by latent heat transfers or through long-wavelength emissions. Soil provides short-term storage of water in the global hydrological cycle, which moves water from oceans to the atmosphere and back. Rain and snowmelt provide water to soils through downward infiltration under the influence of gravity and adhesive forces. Within soil profiles, water is held by adhesive and cohesive forces; however, once the water-holding capacity is exceeded, water can move downward to the groundwater table or laterally to river systems. Evaporation and transpiration remove water directly from the soil surface and from plant leaves. Plant roots collect water from much greater depths, thus extending greatly the role of latent heat flux in driving land–atmosphere hydrologic transfers.

During Earth's history, evolution of photosynthesis and establishment of rooted plants on land initiated a powerful source of energy to drive pedogenesis. Photosynthesis produces complex organic molecules, which are shed to the soil either directly or after they have cycled through animals. In the soil, organic matter is metabolized by microorganisms; this process (often called soil respiration) releases the CO_2 and water that was utilized in photosynthesis. The paired photosynthesis and soil respiration reactions set up a giant planetary reduction–oxidation cycle involving carbon and oxygen. Carbon in CO_2 is reduced to organic compounds; O in water is oxidized to molecular form. In turn, microbial breakdown of organic compounds oxidizes C back to CO_2, transferring electrons to O_2, which is reduced to water. Photosynthesis captures massive amounts of solar energy and much of the ensuing organic matter is transferred to the soil, where its breakdown by soil's vast microbial populations provides most of the energy that drives pedogenesis. Without water, nutrients, and an appropriate temperature range, biological reactions cannot function, but once these environmental conditions are met, it is the C–O redox cycle that releases organic acid into soil, which, in turn, drives specific pedological transformations.

Organic acids lower the pH of soil solution thereby enhancing mineral weathering, but the release of cations during weathering acts to neutralize the acidity. The amount of water percolation and cation removal determines the extent to which weathering buffers soil pH against acidification and determines the nature of secondary mineral synthesis. In well-drained humid environments, dilute soil solution leads to strong chemical gradients, intense weathering, efficient leaching of soluble components (basic cations and silicic acid), and secondary mineral synthesis involving only the least soluble components (Fe and Al (hydr) oxides and kaolin minerals). In poorly drained sites, soil solution accumulates highly soluble ions, as well as sparingly soluble ions, leading to synthesis of smectite (in environments with high Si and Mg activity) and other minerals. In these soils, Fe and Mn may be reduced and slowly leached away. In contrast, arid environments often have soil solutions that contain large concentrations of ions, leading to minor chemical gradients, short leaching distances for only the most soluble ions, and secondary mineral synthesis that is dominated by ionic salts (calcite, gypsum, halite, etc.). Thus, long-term average soil solution concentrations determine most pedochemical properties. Furthermore, they help to condition the terrestrial portion of many biogeochemical cycles because elements released by weathering and not utilized biologically or assimilated into pedogenic minerals are rapidly leached from soil into groundwater, rivers, lakes, and oceans (Berner and Berner, 1996).

The annual amount and seasonal distribution of precipitation and solar energy define broad life zones or biomes based on ecosystem composition and productivity (Aber and Melillo, 1991). Up to some maximum, increasing rainfall leads to greater net primary production of C, which, in turn, leads to greater production of organic matter and more rapid rates of nutrient cycling (Schlesinger, 1997). Highly productive ecosystems often have high levels of CO_2 and enhanced organic acid production, which lead to low soil solution pH values and high weathering rates that release more nutrients until their store of primary minerals is exhausted. Temperature controls the rate of biological processes and mineral weathering/synthesis. At 0°C, and below, little water is available to mediate chemical reactions. In extremely cold environments, physical disintegration of rock material often predominates over chemical breakdown of minerals. For each 10°C increase, biologically mediated reaction rates double until the upper limit of enzymatic functioning is reached. This implies that, all other things being equal, soils in tropical environments will receive far more yearly heat energy and will be more chemically weathered than those in arctic environments (Ugolini and Spaltenstein, 1992; van Wambeke, 1992). In hilly and mountainous terrain, precipitation and solar energy inputs are modified by local topographic position. Soil moisture varies along hillslopes because of redistribution from convex to concave positions, with consequent impact on vegetation type and productivity. Locally, sites are drier or moister than are predicted by average climatic parameters. Similarly, local relief modifies the angle and duration of solar illumination and hence temperature.

In addition to capturing solar energy by photosynthesis, plants have more direct influences on soil formation. The locus of deposition of dead plant material is a factor determining the nature of near-surface horizons. In forests, periodic litter fall deposits organic matter on the soil surface resulting in O horizons. Litter partly decomposes in the O horizon, releasing residues that are carried to the mineral soil in solution as colloidal flow or mixed into the mineral soil by soil fauna. Downward movement of organic acids provides chelating power, which can produce horizons of depletion in the mineral surface soil (E horizon), where one might expect to find organic accumulation (A horizon). In contrast, grasslands tend to be dominated by fire, which releases nutrients as ash at the soil surface, and they shed more organic matter as dead roots and root exudates in the mineral surface soil horizons (producing A horizons). When trees fall, soil and underlying geologic material are often pulled up with the root mass, acting to mix and destroy existing horizons. Roots break up geological substrate through physical expansion and enhance chemical weathering by exuding acidic molecules. In this way, they produce preferential flow paths where pedogenic processes are intensified. Roots fuel nutrient cycling by providing a direct conduit for nutrient movement from soil to aboveground plant parts, which are eventually returned to the upper soil horizons. Mycorrhizal fungi associated with the roots of many plant species are particularly good at decomposing mineral substrate to release P (Chapter 27).

Roots are an important source of food for animals that burrow for food or shelter. Fossorial mammals are effective at mixing parent material fabric and soil horizons. They can sort particle sizes and create macropores that direct preferential water flow. Smaller animals are equally effective at driving processes that enhance or retard specific vectors of soil formation. Ants, termites, springtails, and earthworms move organic matter from the surface downward, mix soil horizons, and create macropores. On hillslopes, animals that deposit soil material on the surface can significantly enhance downslope redistribution of mass through their diggings.

30.3 Pedogenic Processes

In combination, environmental variables drive processes that act on geological or preexisting soil substrate to effect pedogenic alteration at scales ranging from the microscopic parts of soil fabric to watershed soil mantles. Commonly, there will be many processes acting partly in conjunction and partly in opposition. In their totality, they produce soil bodies with recognizable horizons.

In some cases, continued enhancement of horizon properties can, itself, modify ongoing processes, changing the vector of soil formation, and initiating horizon deterioration.

Understanding soil formation requires measurement of present processes, prediction of future trends based on these processes, and interpretation of past processes based on present, relict morphology. A large body of pedological research informs this chapter. The authors have not referenced most of it, choosing instead to focus on essential concepts and specific examples.

30.3.1 Conceptual Process Model

A given body of in situ soil material, regardless of size, can be altered by additions, losses, transformations, or translocations (Simonson, 1959). These processes interact differently depending on the depth from the soil surface and the mix of environmental variables at a specific location. Some of these processes lend themselves to mass balance calculations, others to use of specific tracers or extractions, but, in general, there is no single, universal approach that will provide knowledge of these four processes.

30.3.1.1 Additions

Soil is formed when C (and N) is added from the atmosphere by biological fixation processes. Enzymes in plants and associated microorganisms reduce atmospheric components into biologically useful forms. Dead plants and animals feed a host of microorganisms, which utilize their C and N for growth and energy. Residues from these processes accumulate in soils in complex forms collectively known as soil organic matter (SOM). Since addition of organic residues is greatest near the surface, soil functioning can sometimes be conceptualized into two compartments: an upper part where organic acids and microbial dynamics hold sway and a lower part where inorganic processes dominate, though still influenced by organic processes. It is the mixture of SOM with the porous matrix of inherited and newly synthesized minerals that fundamentally defines Earth's soils.

Solutes and particulate matter are added to soils either directly from the atmosphere (Simonson, 1995; Derry and Chadwick, 2007) or by movement from topographically higher points in the soil landscape (Paton et al., 1995). Although these added components are often incorporated with little outward evidence of their influence, their role can be significant, such as when ecosystems growing on highly weathered soils are sustained by nutrients added from atmospheric sources. Atmospheric additions provide nutrients to rainforests that have inherently low fertility, sometimes providing the majority of nutrient inputs to the ecosystem (Chadwick et al., 1999; Okin et al., 2004; Pett-Ridge, 2009). When additions are rapid and large, existing soils may be buried and new soils developed on the added material. An intermediate result is formation of cumulic soils where soil formation continues as new matter is added at the surface, thickening the soil, as may occur along the lower portions of hillslopes where matter eroded form higher positions accumulates or in areas receiving moderate amounts of airborne dust (Chapter 29).

30.3.1.2 Losses

Mass is lost from soils through erosion by wind and water at the surface and by leaching of solutes and colloids. Eroded soils often lie on the upper slope portions of hillslopes. Microbes break down SOM and minerals weather, releasing ions to the soil solution. In the organically dominated soil compartment, organic chelates facilitate movement of ions deeper into the profile. When excess water moves through soil pores, some elements are carried into streams or groundwater. Typical cations lost through leaching are Na^+, Ca^{2+}, K^+, Mg^{2+}, and as soil acidity increases, smaller amounts of Al are leached in the form of variously charged hydrolysis products. Except at pH values >9, $Si(OH)_4$ is leached as an uncharged compound. Anions leached from soil are dominated by HCO_3^-, which leads to significant amounts of atmospheric carbon sequestration on geological timescales (Chadwick et al., 1994). Lesser amounts of NO_3^- and SO_4^{2-} are leached from soils, but those receiving acid rain may show large losses of these anions (Sposito, 1989). Under aerobic conditions, little Fe is leached from soils and both Fe and Al tend to accumulate by residual enrichment because of leaching of other elements. Iron and Mn can be leached from soils under anaerobic conditions because their reduced forms are soluble.

30.3.1.3 Transformations

Organic matter added to soil is modified by microbial respiration leading to accumulation of more complex SOM or organic compounds that bond with inorganic colloids. Weathering of parent material minerals provides not only nutrients for biological activity but also building blocks for synthesis of secondary minerals. Compared with mantle- and crust-derived minerals that form soil parent material, secondary minerals are enriched in stable elements such as Al and Fe and depleted in easily solubilized elements such as base cations and Si. Under humid, well-drained conditions, where leaching is maximized, secondary mineral formation is characterized by synthesis of aluminosilicate clays and Al and Fe (hydr)oxides. In arid environments, leaching is negligible and even basic cations accumulate in the lower parts of soil profiles as soluble and semisoluble salts. These chemical transformations also lead to physical transformations. For example, wetting and drying results in swelling and shrinking that transform soil fabric, producing a structure reflective of its mineralogical, textural, and organic matter status. Smectite, vermiculite, and short-range order clays hold much more hygroscopic water than do the primary minerals from which they form. Chemical and mineralogical transformations can not only release nutrients for biological use but also sequester nutrients. Nitrogen can be immobilized by NH_4^+ fixation in clay minerals and P can be fixed by sorption onto Fe and Al (hydr)oxides at low to neutral pH values and precipitation with or adsorption to $CaCO_3$ at high pH values.

30.3.1.4 Translocations

Colloids, composed of low-molecular-weight organic matter and small clay particles, and dissolved constituents can move through soil pores. Commonly, the location of mobilization and deposition are controlled by complex chemical interactions between soil solution and the colloidal fraction. Translocation in humid environments usually is effected by downward flow of water. In arid environments, it is often downward as well, but salts can also be carried upward by capillary flow of water in response to evaporation. These processes lead to formation of zones of colloid and solute depletion, such as E horizons, and zones of colloid enrichment and solute precipitation, such as Bhs (≈spodic), Bt (≈argillic), Bk (≈calcic), By (≈gypsic), and Bqm (≈duripan) horizons. Another class of translocations includes mixing processes, which move bulk soil material upward, downward, or laterally in the soil. Mixing can be caused by activity of soil organisms, freeze–thaw cycles, or shrinking and swelling of expansive clays. These processes counterbalance colloid redistribution because they stir soil profiles, preventing or slowing concentration of translocated materials in subsurface horizons.

30.3.2 Dominant Processes

Locally, soil processes reflect a balance among gains, losses, internal redistribution, and chemical and physical changes. The resulting soil properties represent the long-term effect of these processes acting on a 3D reaction column that is open to exchange of matter and energy with the environment. Important soil-forming processes are described later, beginning with two universal aspects of soil formation: organic matter accumulation and formation of soil structure. Subsequent sections are organized to cover processes that are driven by water flux and to some degree, reflect the influence of progressively greater leaching: accumulation of soluble salts and gypsum, accumulation of $CaCO_3$, accumulation of opaline silica, redistribution of clay, complexation and redistribution of Fe and Al, leaching of Si, and concentration of resistant (hydr)oxides. Next, the processes in soil environments where the presence of water is so prevalent that it greatly limits the supply of O are discussed. In these soils, microbially induced reduction controls important aspects of the chemistry, mineral transformation, and morphology. Finally, biological and physical processes of soil mixing are considered. Soil mixing occurs to some degree in most soil-forming environments but is especially influential under some conditions, such as in the seasonally thawed layer above a permafrost table.

30.3.2.1 Organic Matter Accumulation and Alteration

The primary source of organic matter in soils is vegetation, with leaves, stems, and floral parts added to the surface as they drop from the plant, and roots added directly into the soil itself as they grow. Animals and microorganisms (Part IV) feed on the vegetable matter, and each other, decomposing the organic matter parent material to yield gases (e.g., CO_2, CH_4) and humus. The influence of SOM is out of proportion to its weight fraction, a reality that is recognized in the designation of soil materials as organic when they contain a minimum of only 12%–20% organic C. The organic fraction of soils can be extremely important in determining many aspects of soil processes. For example, even in soils with as little as 0.5%–1% organic C, SOM controls pesticide adsorption.

30.3.2.1.1 Organic Matter Decomposition

Soil organisms use organic tissues as a source of energy and C. Each type of organism has its own role in the decomposition process, from primary decomposers, such as beetles, ants, termites, earthworms, and fungi, to decomposers that feed on the feces or tissue of the primary decomposers, such as bacteria, fungi, and various meso- and macrofauna. A succession of populations operates, each using the altered material from the previously active population and further altering the organic substrate itself. The result is a successive depletion of chemical energy sources, increased resistance to microbial attack, and a lowering of the C:N ratio of the SOM.

The rate of decomposition is controlled by a number of factors related to organic matter quality, the interaction of organic matter with soil minerals, and environmental conditions that affect biological activity. The amount and kind of the various compounds in the organic matter are critical (Melillo et al., 1989). Simple organic acids, sugars, and starches are very quickly utilized by microorganisms (on the order of days to months). Protein, chitin, cellulose, and hemicellulose are used less rapidly and are listed in order of increasing resistance to decomposition. Lignin, fats, and waxes are most resistant to decomposition.

Association with soil minerals can be an important factor in stabilizing organic matter against decomposition. In a study of organic matter in subsoil horizons of acidic forest soils, 73% of stable organic matter was found to be mineral protected (Mikutta et al., 2006). Furthermore, most variability in the size of the mineral-protected organic matter pool was explained by the concentration of poorly crystalline minerals. Soils dominated by short-range order minerals have slowly decomposing organic carbon pools that are larger than those of other mineral soils (Rasmussen et al., 2006). The high surface area of these minerals adsorbs organic carbon, decreasing its availability for decomposition, resulting in relatively slow decomposition rates.

Biological activity in soils is greatest when the soil pH is between 6 and 8, soil temperature is 20°C–40°C, soil water potential is between field capacity and about −1.5 MPa, and there are no deficiencies in essential nutrients. Rates of decomposition are especially slow when the soil is saturated for long enough to become anaerobic and when the temperature is below biological zero (≈5°C). Under these conditions, thick layers of organic matter (>40 cm) may accumulate above the mineral soil. On the other hand, there are microbes adapted for virtually every condition that may come about naturally in soils (Atlas and Bartha, 1981).

30.3.2.1.2 Characteristics of Soil Organic Matter

Soils may contain organic matter ranging across the entire spectrum of decomposition products, particularly if a mixing mechanism, such as burrowing animals or surface cracks, operates to incorporate fresh organic material into the soil (Quideau et al., 1998). In the absence of aggressive mixing, much decomposition occurs within a litter layer. In general, SOM is increasingly decomposed with depth. The final product is humus: the dark brown, complex, microbially resistant, colloidal compounds that are highly modified original materials or materials synthesized by microbes. Humus has high surface area and chemical reactivity (CEC ranges from 100 to 300 $cmol_c\ kg^{-1}$), is very hydroscopic, and readily forms complexes with the inorganic fraction (Chapter 11). Humus is so complex that its composition is usually defined using the following operational criteria (Oades, 1989): fulvic acid fraction, which is soluble in both alkali and acid, humic acid fraction, which is soluble in alkali but insoluble in acid, and the humin fraction, which is insoluble in both alkali and acid. Fulvic acid is composed of the lowest molecular weight compounds and has the greatest number of acid groups and the highest CEC. In contrast, humin is composed of the highest molecular weight compounds and has relatively few acid groups and the lowest CEC. In soils of pyrogenic ecosystems, charcoal may be a significant component of the carbon pool (MacKenzie et al., 2008). Carbon in the form of charcoal is highly resistant to decomposition and influences nutrient availability.

30.3.2.1.3 Relationship of Organic Matter to Morphology and Function

The accumulation of humified organic matter is key to formation of an A horizon, which, if it becomes thick enough, may evolve to qualify as a mollic or umbric epipedon. A horizons are an important link between soils and plants, since most plant roots exist within the A horizon and most of the nutrients are extracted from there. Decomposition of organic matter in the A horizon, including the roots themselves, biocycles essential nutrients from vegetation back into plant-available pools. High levels of organic acids lead to maximum rates of primary mineral decomposition in A horizons as well. Addition of organic matter also influences the physical properties of soils. As the A horizon develops in young soils, there is a decrease in bulk density, and the soil color becomes darker and redder (Turk et al., 2008). Organic matter tends to promote development of soil structure and porosity, thereby enhancing water infiltration and water-holding capacity.

30.3.2.2 Development of Soil Structure

An essential component of soil development is the substitution of soil structure for the geologic structure of parent material. While geologic structure is inherited from jointing cracks and bedding planes, soil structure forms as a function of shrinking and swelling induced by change in moisture content and by the downward movement of water and colloids along preferential flow lines. Soil structural units, known as peds, become more strongly expressed as their faces are differentiated from their interiors. The surface/interior differentiation comes about through several mechanisms. Wetting-induced expansion forms pressure faces or stress argillans by orienting platy silt and clay particles along contact planes between peds. Subsequent drying causes contraction, which opens voids between peds. Upon rewetting, water flows preferentially through the contraction voids depositing clays and organic matter on the ped faces, further differentiating the outer from the inner parts of the peds. Other compounds, including calcite, silica, and Fe and Mn (hydr)oxides can also be deposited along ped faces. Conversely, under some conditions, outer margins of peds become depleted by preferential leaching of clay and soluble compounds. Roots follow paths of least resistance and concentrate in voids between structural units, enriching ped surfaces with organic matter and enhancing preferential drying of ped exteriors relative to interiors.

30.3.2.2.1 Size, Degree of Expression, and Reinforcement of Peds

A number of factors influence the size and degree of expression of soil structural units. The amount of volume change associated with wetting and drying cycles (swelling and shrinking) is very important. Fine-textured soils and those with smectitic mineralogy typically have relatively large volume changes and, therefore, develop the most strongly expressed structural units. Fine-textured soils also have less resistance to shear and so tend to have more cracks and finer structural units. Frequent shrink/swell cycles, which can be related to cyclic drying/wetting, tend to reinforce the expression of structural units (Southard and Buol, 1988). This is reflected in the development of soil structure in shallow agricultural drainage ditches that undergo cyclic drying/wetting but not in deeper ditches that are continually saturated (Vaughan et al., 2008). Slow drying allows a relatively uniform shrinkage and contraction of the soil mass, producing larger peds. Crystallization of salts in saline soils yields a finer structure than is found in similar nonsaline soils (Reid et al., 1993).

30.3.2.2.2 Granular Structure in Surface Horizons

The structure of surface horizons is often most strongly influenced by biological factors, including organic matter, microbes, plant roots, and soil fauna (Figure 30.1). Fecal casts of soil macrofauna, such as earthworms (Lee and Foster, 1991) and snails (Anderson et al., 1975), often form compact, stable aggregates, which dominate in some A horizons. Dense networks of fine grass roots and fungal hyphae cause uniform fine-scale shrinkage cracks and enmesh soil materials into granular or crumb-like units (Oades, 1993). At the most fundamental level, soil aggregates in surface horizons are usually held together by SOM–mineral complexes (Chapters 3 and 11).

30.3.2.2.3 Platy Structure

Platy structure typically forms in response to unidirectional compressional forces (Figure 30.1). It is most often produced in surface soils through compaction by animals or machines, or by raindrop impact. Vertical compaction yields horizontally oriented, platy structural units. Gravity can contribute to the necessary vertical force for platy structure development in the silty vesicular horizons that are common in arid and semiarid regions. In these surface horizons, vesicles are formed by escaping gases as the soil material is wetted. Repeated wetting and

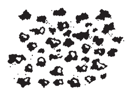
Type of structure	Soil environment	Formation processes
Granular 	A horizon	Aggregation by biological agents: organic compounds, fungal hyphae, fine roots, fecal casts
Platy 	A and other surface horizons Petrocalcic horizons Duripans	Vertical compression from compaction
Prismatic 	B horizons with uniform texture, fine roots, few coarse fragments, relatively slow drying	Horizontal compression from shrink-swell
Blocky 	B horizons with coarse fragments, heterogeneous root sizes	Shrink-swell interrupted by nonhomogeneous matrix

FIGURE 30.1 Types of soil structure, their occurrence within soils, and the processes responsible for their formation.

drying cycles enlarge the vesicles until they can no longer support the weight of the overlying material and collapse, forming planes that bound platy structural units (Miller, 1971). Platy structure is also common in highly developed petrocalcic horizons and duripans. The origin of platy structure in these subsoil horizons is less clear but may be caused by vertical compression generated as precipitation of secondary calcite and opal increases subsoil volume. In pergelic soils (Gelisols), platy structure is associated with the formation of lenticular ice (vein ice) (Section 33.6)

30.3.2.2.4 Prismatic and Blocky Structure

Shrinkage of moist soil material as it dries results in multidirectional compression centered around numerous loci within the mass. When shrinkage forces are largely resolved in lateral directions, the soil material may contract uniformly toward more or less equally spaced centers forming prismatic or columnar peds bounded by vertical cracks (Figure 30.1). In the formation of prismatic structure, uniform shrinkage is important because it develops the consistently spaced and arranged cracks that form ped boundaries; uniform shrinkage is favored by slow drying and homogeneous soil materials. It is often associated with uniformly fine-textured sediments in poorly drained, low-lying landscape positions. Prismatic or columnar structure is common in natric horizons because of enhanced swelling and poor drainage associated with Na-saturated clays. In contrast to the elongated character of prismatic structure, blocky structure forms as a result of combined lateral and vertical shrinkage forces (Figure 30.1), with neither predominating. Conditions favoring blocky structure development are rapid drying and factors that break continuity of soil fabric, such as woody roots and coarse fragments. Also, bedded moderately fine and fine-textured fluvial or marine deposits tend to form blocky structure in B horizons because horizontal planes of weakness, inherited from parent materials, are bisected by vertical shrinkage cracks. In permafrost soils, blocky structure is associated with ice nets, a cryostructure consisting of 2–5 cm diameter polygons separated by vertical or diagonal desiccation cracks (Ping et al., 2008).

30.3.2.2.5 Vertic Features

Vertisols and other clayey, expansive soils develop characteristic structural features as a result of their dynamic shrink/swell behavior (Section 33.7). Upon wetting, the fabric of these soils swells, commonly increasing its volume by about 10%–15%. In surface horizons, this increase in volume is accommodated by upward movement, but in the subsoil, overburden pressures confine vertical movement so that lateral swelling pressures are at least four times as great as the vertical pressures and exceed the soil shear strength (Wilding and Tessier, 1996). Under these conditions, shear failure takes place at angles between 10° and 60° from horizontal. The shear planes present slick, grooved surfaces, known as slickensides that bound characteristic wedge-shaped aggregates.

30.3.2.2.6 Aggregation by Fe and Al (Hydr)oxides

Highly weathered soils, such as Oxisols, have a fine (<2 mm diameter) granular structure that is the result of aggregation by Fe and Al (hydr)oxides (van Wambeke et al., 1983) (Section 33.13). Aggregation is affected by very small positively charged Fe and Al (hydr)oxide particles and surface coatings that form bridges between negatively charged clay particles to build up small aggregates (Hsu, 1989; Schwertmann and Taylor, 1989).

30.3.2.3 Accumulation and Redistribution of Salts

Gypsum and more soluble salts accumulate in poorly leached soils and soils where evapotranspiration (ET) greatly exceeds leaching. The composition and concentration of soluble salts influence the physical behavior of soil particles because of the effect that dissolved salts have on dispersion and flocculation. Ecosystems with salt-affected soils are subject to many plant physiological stresses, related to surface crusting, low osmotic potential of the soil solution, and specific ion toxicities.

30.3.2.3.1 Characteristics of Soluble Salts

Sulfate, carbonate, chloride, and nitrate salts occur in soils of arid environments (Table 30.1). The concentration and type of salts influence the physical behavior and chemical properties of the soils in which they occur. Salt-affected soils have major impacts on the flocculation or dispersion of clay minerals and organic matter (Chapters 17 and 18 of *Handbook of Soil Sciences: Resource Management and Environmental Impacts*). In general, high electrolyte concentrations keep soils flocculated, whereas colloidal dispersion occurs when the electrolyte concentration is low and Na predominates on exchange sites. Thus, sodic soils [exchangeable sodium percentage (ESP) > 15, electrical conductivity (EC) < 4 dS m^{-1}] are dispersed, resulting in clogged pores and very low infiltration rates, whereas saline (ESP < 15, EC > 4 dS m^{-1}) and saline-sodic (ESP > 15, EC > 4 dS m^{-1}) soils are flocculated, promoting porosity and higher infiltration rates.

Salts that occur in arid-land soils vary in their solubilities (Table 30.1). Salts of Na, Mg, and K are more than 100 times more soluble in water than calcite and gypsum and move readily with saturated and unsaturated water flow in soil. As a result of the high solubility of Na salts and the production of hydroxide ions when carbonate ions dissolve, soils containing Na_2CO_3 have high pH values (9–12):

$$Na_2CO_3 + H_2O \leftrightharpoons 2Na^+ + HCO_3^- + OH^- \tag{30.1}$$

Sodium carbonate can form in sodic soils through several mechanisms. Weathering of Na aluminosilicates releases Na, which reacts with HCO_3^- in water to form $NaHCO_3$, which concentrates to Na_2CO_3 under the influence of evaporation. Neutral Na salts can react with $CaCO_3$ to yield Na_2CO_3:

$$Na_2SO_4 + CaCO_3 \leftrightharpoons Na_2CO_3 + CaSO_4 \tag{30.2}$$

and Na can be replaced from the exchange by H or Ca from carbonates:

$$[CLAY] - 2Na + CaCO_3 \leftrightharpoons [CLAY] - Ca + Na_2CO_3 \tag{30.3}$$

TABLE 30.1 Some Common Salts in Soils and Their Solubilities in Water at 25°C

Name	Formula	Solubility (mol L^{-1})
Soda niter	$NaNO_3$	10.69
Niter	KNO_3	3.53
Halite	NaCl	6.15
Thenardite	Na_2SO_4	1.97
Mirabilite	$Na_2SO_4 \cdot 10H_2O$	2.74
Nahocolite	$NaHCO_3$	1.22
Soda	$Na_2CO_3 \cdot 10H_2O$	2.77
Trona	$Na_3H(CO_3)_2 \cdot 2H_2O$	2.56
Bloedite	$Na_2Mg(SO_4)_2 \cdot 4H_2O$	2.31
Hexihydrite	$MgSO_4 \cdot 6H_2O$	3.17
Epsomite	$MgSO_4 \cdot 7H_2O$	3.03
Gypsum	$CaSO_4 \cdot 2H_2O$	0.005
Calcite	$CaCO_3$	0.0006 (pH 8 and $CO_2 = 10^{-4}$ MPa)

30.3.2.3.2 Natural Sources of Soil Salinity

The ultimate source of salts is weathering of primary minerals (Figure 30.2), which produces dilute soil solution. Atmospheric deposition contributes appreciable quantities of salts to soils, particularly as dry fall (dust) in arid regions and as sea spray along coastal margins. Most saline soils receive salts from sources where salts have been concentrated by geological processes. For instance, dissolution of marine or closed basin lake sediments provides preconcentrated salts to near-surface waters. Isotopic studies in the Atacama Desert documented the transition from coastal soils receiving salts from marine sources to inland soils, in which salts have isotopic signatures similar to local surface and ground waters (Rech et al., 2003). The salts in groundwater are concentrated by evaporation within the capillary fringe and precipitate as evaporite minerals on playa surfaces. These evaporite minerals are highly susceptible to wind erosion (Reynolds et al., 2007), which is responsible for the dispersal of salts across the desert landscape. In a study of a Mojave Desert bolson landscape, salts were found to be concentrated in soils on slopes downwind from the playa (Hirmas and Graham, 2011).

30.3.2.3.3 Accumulation of Soluble Salts in Soils

Salts precipitate when their solubility is exceeded, usually as the soil solution is concentrated by ET. In well-drained soils where leaching is greater than ET, salts do not accumulate because the constituent ions are leached to groundwater. On the other hand, salts accumulate when leaching is minimal. Mineral leaching results from high ET rates and/or low rainfall, convex topography that disperses water flow, and soil conditions such as crusting that yield low infiltration rates. One special case of poor leaching conditions leading to salt accumulation involves the soils formed beneath desert pavement (a monolayer of closely packed clasts). The accumulation of salts beneath desert pavement corresponds strongly with the size, sorting, and percent cover of the surface clasts (Wood et al., 2005) and includes large nitrate pools in some soils in the Mojave Desert (Graham et al., 2008). In each of these cases, the areal and vertical distribution of salts reflects a balance between a downward or lateral flux of leaching water that removes salts, and the upward flux of ET that helps retain salts within the soil.

Although the dominant influence of plants on salt distribution is through their role in ET, some plants have a localized effect on soil salinity and sodicity. For example, surface soils under the canopy of grease-wood have EC values six times those of soils in interspaces between the shrubs, and ESP values are

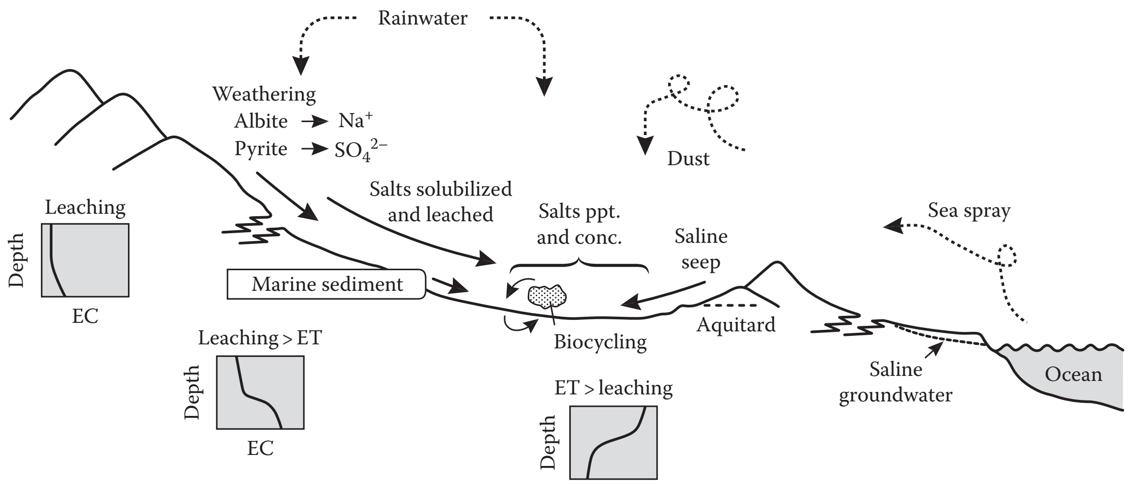


FIGURE 30.2 Sources and redistribution of soluble salts in soils in relation to hydrology, lithology, and landscape position. When the downward leaching flux of water greatly exceeds upward flux due to ET, soluble salts, as indicated by EC of saturation extracts, are minimal throughout the soil profile. When leaching is slightly greater than ET, EC increases with depth, since salts are leached from the surface soil. When ET is greater than leaching, EC is greatest near the surface due to the predominating upward flux of water, carrying salts, to the evaporative surface.

more than 10 times greater due to biocycling of sodium ions (Fireman and Hayward, 1952).

Bare patches, known as slickspots, caused by salt accumulations, have been shown to grow due to interactions between the vegetation and soil properties (Reid et al., 1993). The grass-covered soils surrounding the slickspots have adequate infiltration rates and grass traps saline dust from a nearby soda lake playa (Figure 30.3). Under the slickspots, the much lower infiltration rates at the surface causes the subsoils to remain dry even during the wet season. This, in turn, establishes a matric potential gradient from the moist subsoils under grass to the slickspots, so that salts are carried laterally in solution as water is drawn by capillarity into the slickspot soils. The buildup of salts in the slickspot soils eventually has a toxic effect on the grasses at the slickspot margin, which exposes bare soil to raindrop impact and wind erosion, thereby promoting expansion of the slickspot.

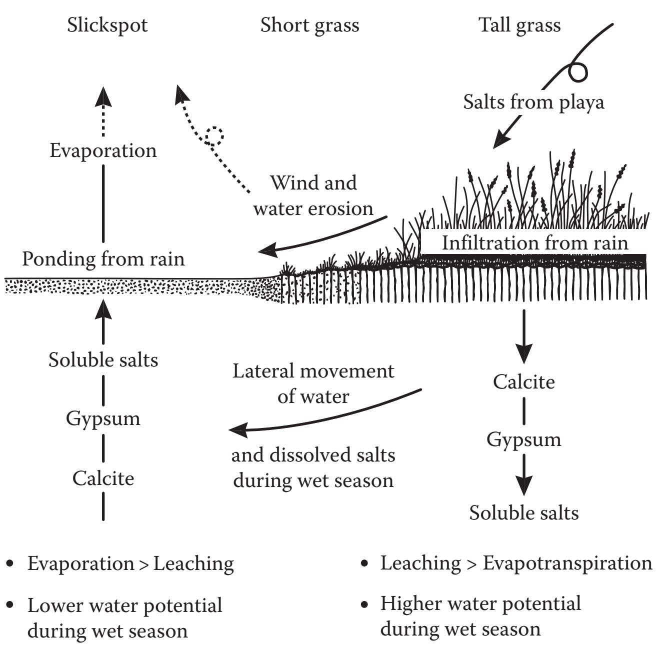


FIGURE 30.3 Diagrammatic representation of processes contributing to the salinization and expansion of slickspots on the Carrizo Plain, California. (Modified from Reid, D.A., R.C. Graham, R.J. Southard, and C. Amrhein. 1993. Slickspot soil genesis in the Carrizo Plain, California. Soil Sci. Soc. Am. J. 57:162–168. With permission of the Soil Science Society of America.)

30.3.2.3.4 Relation to Soil Morphology

Accumulation of soluble salts may be expressed in a morphological sequence similar to the stages of carbonate accumulation (discussed in the Section 30.3.2.4). Soils in stable landscape positions and hyperarid deserts, such as the Atacama Desert in South America and the Dry Valleys of Antarctica, can eventually develop horizons that are indurated with highly soluble minerals, including halite and soda niter (Bockheim, 1997; Rech et al., 2003). Soluble salt accumulation in moraines of the Transantarctic mountains progress with age from coatings on the bottoms of stones (stage I), to flecks (stages II–III), to a cemented pan (stages IV–V), and finally an indurated pan (stage VI) (Bockheim, 1990). In some cases, the morphology of salt accumulations may be a clue to their composition. Concentrations of gypsum may be identified in the field by a unique morphology, called gypsum snowballs, which are 0.5–1 mm, white, powdery spheres (Buck and Van Hoesen, 2002).

Exchangeable sodium can influence soil morphology by causing clay dispersion, which decreases aggregate stability and increases mobility of clays. Illuvial clay accumulates more rapidly in sodic soils, compared to nonsodic soils, thus complicating the relationships between soil morphology and surface age (Peterson, 1980). Strong development of vesicular horizons has been associated with sodic soils, due to dispersion and crust formation (Bouza et al., 1993).

The contrast between leached soils and salt-affected soils is well illustrated by the comparison of soils in slickspots with the surrounding vegetated landscape (Reid et al., 1993). The grass-covered soils have clayey Bt horizons with strong prismatic structure, which allows rapid infiltration of rainwater and leaching of salts. In these soils, less-soluble minerals occur near the surface and more-soluble minerals are concentrated deeper in the profile. With increasing depths, accumulations of calcite, gypsum, and soluble salts are observed, indicating a regime in which leaching predominates over ET. In contrast, <1 m away, the bare surface of the slickspot is crusted from raindrop impact, and the upper few centimeters are sufficiently leached of salts, so that the soil disperses. Crusting and dispersion result in very low infiltration rates, so that leaching of the profile is minimal. The predominance of evaporation over leaching is indicated by the accumulation of soluble salts above gypsum, and calcite accumulation deeper in the soil a trend opposite that in grassed areas (Figure 30.3).

30.3.2.4 Accumulation and Redistribution of Calcite

Where soil pH values are above about 7.2, $CaCO_3$ can precipitate in soils to form the mineral species calcite. Calcite contains <5 mol% Mg. At 25°C, pH 8, and P_{CO_2} of 10^{-4} MPa, calcite has a solubility of 0.0006 mol L^{-1} (0.06 g L^{-1}), which is much more soluble than silicate minerals but is considerably less soluble than the salts commonly found in soils. Under well-drained conditions, calcite is found in soils of arid, semiarid, and subhumid environments. With progressively greater leaching, it is found at greater depths below the soil surface. It can also accumulate in poorly drained soils where calcium and bicarbonate are concentrated in ponded water.

30.3.2.4.1 Natural Sources of Calcite in Soils

While calcite may be inherited in soils from parent materials such as limestone, pedogenic calcite precipitates from soil solution and requires sources of Ca^{2+} and CO_3^{2-} ions. Calcium ions may be derived from weathering of parent materials. Calcareous rocks provide the most abundant Ca, with limestone containing 30%–40% Ca and dolomite containing about 20% Ca. Calcareous tills, loess, and other sediments also provide a ready supply of Ca for pedogenic calcite formation. Calcium is present many in primary minerals, particularly Ca-rich plagioclase, but also in some amphiboles, pyroxenes, garnets, and epidote. Among igneous rocks, Ca is most abundant in the more mafic (but not ultramafic) rocks, such as gabbro and basalt, which contain about 7% Ca. Mafic rocks are also rich in Mg and soils formed on basalt may contain pedogenic high-Mg calcite and/or dolomite (Capo et al., 2000; Whipkey et al., 2002). Atmospheric deposition can be a substantial, even dominant, source of Ca, both in ionic form and as particulate $CaCO_3$. For example, in southern Nevada and California, dust contains 10%–30% $CaCO_3$, equating to a deposition rate of 1–6.6 g $CaCO_3$ m^{-2} $year^{-1}$ (Reheis and Kihl, 1995). In southern New Mexico, dust contributes about 0.4 g $CaCO_3$ m^{-2} $year^{-1}$, whereas rain delivers 1.2 g m^{-2} $year^{-1}$ (Birkeland, 1999). On the Edwards Plateau in Texas, no $CaCO_3$ is delivered as dust, but rain supplies Ca equivalent to 2.3 g $CaCO_3$ m^{-2} $year^{-1}$ (Rabenhorst et al., 1984). In southern Arizona, isotopic studies have suggested that most $CaCO_3$ in dust actually comes from the erosion of soil carbonates; thus, dust can be viewed as a recycled source of Ca (Capo and Chadwick, 1999; Naiman et al., 2000). Pedogenic carbonates have a Sr isotope signature between that of dust and parent rock, indicating a mixed source of Ca. Calcium can also be derived from biocycled plant material and groundwater, which may be highly calcareous, depending on the lithologies exposed in the aquifer. When Ca-laden groundwater lies within or just below the soil zone, calcite can form in the soil as it is precipitated within the capillary fringe. At sites near the coast, Sr isotope ratios in pedogenic carbonates are similar to those of ocean water, suggesting that some Ca may come from sea spray (Naiman et al., 2000; Whipkey et al., 2000).

The source of the carbonate anion in calcite is the dissolution of CO_2 in water, which yields the following species:

$$CO_2 + H_2O \leftrightharpoons H_2CO_3 \tag{30.4}$$

$$H_2CO_3 \leftrightharpoons HCO_3 + H^+ \quad pK = 6.4 \tag{30.5}$$

$$HCO_3 \leftrightharpoons CO_3^{2-} + H^+ \quad pK = 10.3 \tag{30.6}$$

The amount of CO_2 in the soil solution depends on the relationships expressed in Henry's law:

$$K_H = \frac{(H_2CO_3)}{P_{CO_2}} = 10^{-1.5} \tag{30.7}$$

An increase in the proportion of CO_2 in the gas phase increases the CO_2 concentration in the solution in contact with the gas. The P_{CO_2} in soils is on the order of 0.003–0.03, compared to 0.00033 in the atmosphere. The distribution of CO_2 in soils depends on a balance between production of CO_2 within the soil, by microbes and plant roots, and diffusion losses of CO_2 out of the soil to the atmosphere. As an example, the CO_2 concentration in a coarse-textured soil in southern Nevada was greatest in spring when biologic activity was at its peak (Figure 30.4; Terhune and Harden, 1991). Even at this time, CO_2 concentration in the surface soil was less than that in the subsoil due to diffusional losses to the atmosphere. Biologic activity and CO_2 production were low during the winter due to cold temperatures and during summer due to dryness.

30.3.2.4.2 Calcite Precipitation in Soils

The reaction for calcite precipitation can be expressed as follows:

$$Ca^{2+} + 2HCO_3 \leftrightharpoons CaCO_3 + H_2O + CO_2 \tag{30.8}$$

Precipitation of calcite is promoted by a number of factors that influence this relationship. An increase in pH drives the reaction to the right by supplying more HCO_3^-, as a result of CO_2 + OH^-. Decreasing the P_{CO_2}, through the loss of CO_2 to the atmosphere,

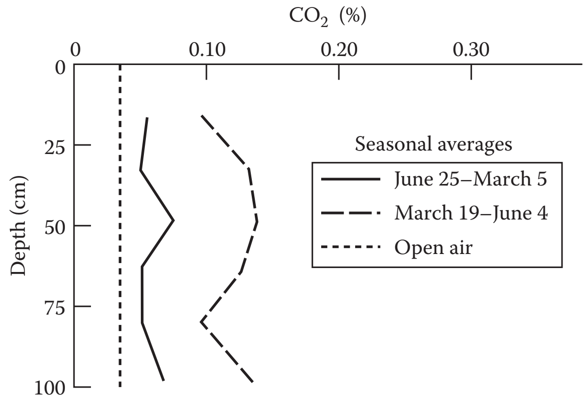


FIGURE 30.4 Seasonal changes of CO_2 with depth in an Entisol in southwestern Nevada. (Reprinted from Terhune, C.L., and J.W. Harden. 1991. Seasonal variations of carbon dioxide concentrations in stony, coarse-textured desert soils of southern Nevada, USA. Soil Sci. 151:417–429. Copyright Williams and Wilkins, Baltimore, MD. With permission.)

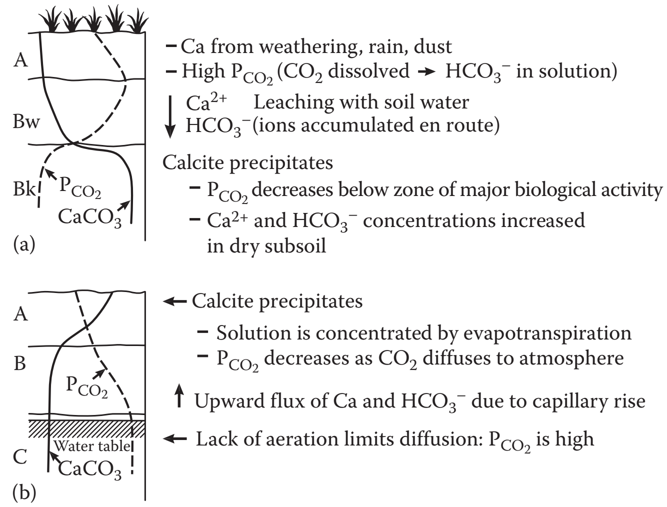


FIGURE 30.5 Idealized diagram of processes involved in calcite precipitation in (a) a well-drained soil (water flux is downward) and (b) a poorly drained soil (water flux is upward).

also drives the reaction to the right. Loss of water through ET increases the ionic concentration of the soil solution to the point where it exceeds the solubility product of calcite and results in precipitation. Precipitation of calcite in soils is not simply an inorganic chemical phenomenon. In fact, the role of microorganisms seems to be ubiquitous and perhaps essential in pedogenic calcite formation. Microbial influence is revealed by calcified biological structures, such as fungal hyphae, and experimental evidence that shows that both bacteria and, especially, fungi produce calcite as a byproduct of their metabolism. In Ca-rich soils, microbes excrete excess Ca, which concentrates on their external surfaces and reacts with HCO_3^- in the soil solution. In the laboratory, calcite precipitates rapidly in inoculated soil columns but not at all in sterile columns that were otherwise identical (Monger et al., 1991). Hyperarid regions of the Atacama Desert and Antarctica, in which no plant life is observed, have been found to contain only trace levels of $CaCO_3$ (Bockheim, 1990; Ewing et al., 2006; Quade et al., 2007). The isotopic signature of $CaCO_3$ in soils without plant life in the Atacama Desert is distinct from that in soils with plant life, suggesting possible low levels of abiotic precipitation.

Precipitation of calcite in soils occurs preferentially in certain types of morphologic sites. These favored sites include the vicinity of roots, where microbes and nucleation sites are abundant and plant water uptake concentrates the soil solution. Another favorable site is in large pores where drying is relatively rapid and the P_{CO_2} is lower than in the matrix due to effective gas exchange with the atmosphere. In gravelly soils, calcite tends to precipitate preferentially on the undersides of gravels, perhaps because water tends to collect and is last to dry in those sites. Once calcite precipitation is initiated at any of the sites described above, it is preferentially precipitated on preexisting calcite crystals.

30.3.2.4.3 Examples of Calcite Distribution and Precipitation in Soils

The processes governing precipitation of calcite in a well-drained soil, where water flux is downward, are illustrated in Figure 30.5a. Calcium in the soil originates from weathering, rain, and/or dissolution of dust. Organic matter decomposition in the A horizon produces a high P_{CO_2}, although it decreases substantially near the soil surface due to gas exchange with the atmosphere. The high P_{CO_2} yields a relatively large amount of CO_2 dissolved in the soil solution, producing HCO_3^-. The Ca^{2+} and HCO_3^- ions are leached with soil water to a depth at which calcite precipitates. This depth is determined by decreased P_{CO_2} below the zone of major biological activity and, perhaps more importantly, by the soil solution being concentrated as it enters the dry subsoil and is depleted by ET.

In poorly drained soils of arid and semiarid regions, dominant water flux is upward in response to evaporation (Figure 30.5b). The relatively high water content in subsoil limits gas diffusion, so P_{CO_2} is high and decreases near the soil surface as CO_2 diffuses to the atmosphere. Calcium and HCO_3^- originating from groundwater sources or in situ from the soil move upward with the flux of water. Calcite precipitates in an upper soil zone where the P_{CO_2} is relatively low and the soil solution is concentrated by evaporation.

30.3.2.4.4 Relation to Soil Morphology

Soils with pedogenic calcite commonly progress through distinct evolutionary stages (Table 30.2; Gile et al., 1966). The stages and morphologic expressions are different for gravelly soils compared to fine-textured soils. Because gravelly soils have less total pore space than fine-textured soils, segregated calcite accumulates more rapidly, precipitating readily on pebbles bounded by relatively large pores. One of the distinct morphologic expressions of pedogenic $CaCO_3$ in gravelly soils is pendants precipitated on the bottom of clasts. The pendants grow by precipitation of $CaCO_3$ both at the clast–pendant contact and at the base of the pendant (Brock and Buck, 2005).

TABLE 30.2 Stages of Carbonate Morphology in Soils

Stage	Gravelly Parent Material	Nongravelly Parent Material
I	Thin discontinuous clast coatings; some filaments; matrix can be calcareous next to stones; about 4% $CaCO_3$	Few filaments or coatings on sand grains; <10% $CaCO_3$
I+	Many or all clast coatings are thin and continuous	Filaments are common
II	Continuous clast coatings; local cementation of few to several clasts; matrix is loose and calcareous enough to give somewhat whitened appearance	Few to common nodules; matrix between nodule is slightly whitened by carbonate (15%–50% by area), and the latter occurs in veinlets and as filaments; some matrix can be noncalcareous about 10%–15% $CaCO_3$ in whole sample, 15%–75% in nodules
II+	Same as stage II, except carbonate in matrix is more pervasive	Common nodules, 50%–90% of matrix is whitened; about 15% $CaCO_3$ in whole sample
Continuity of fabric high in carbonate		
III	Carbonate forms an essentially continuous medium in 50%–90% of horizon; color mostly white; carbonate-rich layers more common in upper part; about 20%–25% $CaCO_3$	Many nodules, and carbonate coats many grains such that over 90% of horizon is white; carbonate-rich layers are more common in upper part; about 20% $CaCO_3$ in whole sample
III+	Most clasts have thick carbonate coats; matrix particles continuously coated with carbonate or pores plugged by carbonate; cementation more or less continuous; >40% $CaCO_3$	Most grains coated with carbonate; most pores plugged; >40% $CaCO_3$ in whole sample
Partly or entirely cemented		
IV	Upper part of horizon is nearly pure cemented carbonate (75%–90% $CaCO_3$) and has a weak platy structure due to the weakly expressed laminar depositional layers of carbonate; the rest of the horizon is plugged with carbonate (50%–75% $CaCO_3$)	
V	Laminar layer and platy structure are strongly expressed incipient brecciation and pisolith (thin, multiple layers of carbonate surrounding particles) formation	
VI	Brecciation and recementation, as well as pisoliths, are common	

Source: After Birkeland, P.W. 1999. Soils and geomorphology. 3rd edn. Oxford University Press, New York. By permission of Oxford University Press.

In fine-textured soils, calcite precipitates first as filaments in root pores, then as soft masses. Eventually a calcite-cemented (petrocalcic) horizon forms, in which carbonate plugs nearly all the pores and laminar plates of carbonate build up at the interface between the overlying horizons and the plugged ones. Formation of a petrocalcic horizon (stage IV or greater) has major implications for the direction of further development. Plugged horizons form a barrier to most root penetration. They also resist erosion and the overlying horizons may be stripped off, leaving the petrocalcic horizon exposed at the surface. Burrowing animals and erosion commonly bring fragments of petrocalcic horizons to the surface, where the calcite is dissolved and recycled back into the soil (Eghbal and Southard, 1993).

30.3.2.5 Accumulation and Redistribution of Silica

After oxygen, Si is the most abundant element in Earth's crust. It exists in many primary mineral forms and is incorporated into soil-formed minerals as well. When Si is released by weathering, it is leached and can be lost or can be incorporated into aluminosilicate clay minerals, such as kaolinite or smectite. In humid environments, most Si is either leached or consumed during clay mineral synthesis. In subhumid and semiarid environments where there is enough water to support mineral weathering but not enough to completely leach Si, it can also precipitate as nanocrystalline silica that cements soil horizons.

30.3.2.5.1 Characteristics of Silica in Soils

Silica refers to compounds consisting of SiO_2 in crystalline, poorly crystalline, or amorphous forms, and is sometimes hydrated to some degree. Quartz is a common form of silica in many soils. It is highly crystalline SiO_2 that is inherited from parent materials and usually does not form pedogenically. Opal-A, the common pedogenic and biogenic silica, is a hydrated, x-ray amorphous, form of SiO_2. A somewhat more crystalline form of opal, opal-CT, may form in some very old soils, but it is not nearly as common as opal-A.

30.3.2.5.2 Sources of Soluble Silicon

The weathering of silicate minerals releases Si to solution that can be precipitated, under appropriate conditions, as opal-A. Soils rich in easily weathered silicates, such as olivine ($[Mg, Fe]_2SiO_4$) and anorthite ($CaAl_2Si_2O_8$), release abundant Si into solution. The accumulation of soluble silica in soils of southern California was attributed to feldspars weathering in the upper horizons (Kendrick and Graham, 2004). More resistant silicates, such as quartz, release Si very slowly. Amorphous silica is much more soluble than crystalline forms (Figure 30.6). Volcanic glass is a primary form of amorphous silica, which is abundant in many soils or parent materials, sometimes even in those far from current volcanic activity, since volcanic ash can travel great distances downwind from eruptions. Another source of amorphous silica is

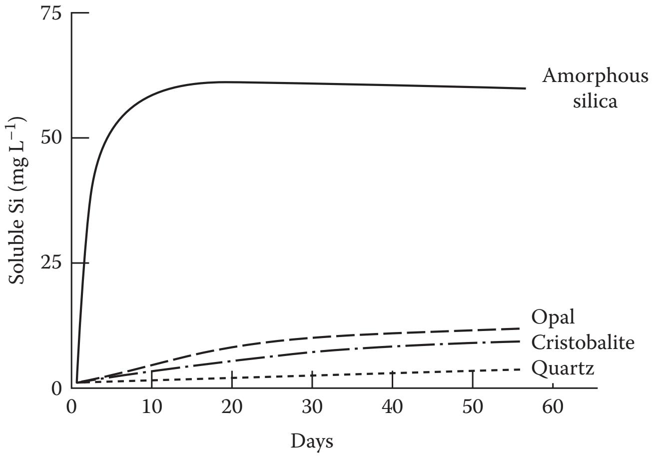


FIGURE 30.6 Dissolution of silicon in water from various silica phases as a function of time. (Reprinted from Drees, L.R., L.P. Wilding, N.E. Smeck, and A.L. Senkayi. 1989. Silica in soils: Quartz and disordered silica polymorphs, p. 913–974. *In* J.B. Dixon and S.B. Weed (eds.) Minerals in soil environments. SSSA, Madison, WI. With permission of the Soil Science Society of America.)

biogenic opal, which is actually opal-A produced as part of plant structures, known as phytoliths, or as part of aquatic organisms, such as diatom tests or sponge spicules. Phytoliths are a biocycled form of silica in soils and can be very abundant in A horizons, particularly in grasslands. In soils that are depleted of primary mineral silica, weathering of phytoliths provides a significant portion of the $Si(OH)_4$ in solution (Alexandre et al., 1997; Derry et al., 2005). Diatoms and sponge spicules are most often found in soils derived from lacustrine, deltaic, or floodplain sediments.

30.3.2.5.3 Dissolution of Silica

Release of Si into solution from a solid is controlled by a number of factors. The inherent solubility of the mineral or amorphous compounds varies widely, with opal-A being the most soluble source of silica (Figure 30.6). Solubility also increases dramatically for very small particles (<0.01 μm). External conditions play a large role, with Si solubilization increasing when the soil solution has a low ionic strength, high pH (>9), and relatively high temperature (Drees et al., 1989). Organic acids promote dissolution by complexing $Si(OH)_4$ released into solution, resulting in highest silica dissolution potentials in the root zone. Sorption of Si onto Fe/Al (hydr)oxides can make them a sink for soluble Si, thereby keeping solution concentrations of Si low and increasing dissolution. On the other hand, coatings of organic materials or Fe/Al (hydr)oxides can retard dissolution by isolating the reactive surface from the soil solution. Mobilization of silica in soils of the Pacific Northwest was decreased in response to management practices that promote accretion of organic carbon, presumably due to organic matter coatings on reactive surfaces (Gollany et al., 2005). Solubility of silica increases with increased pressure, a situation that may arise in petrocalcic horizons as calcite crystallization increases the subsoil volume and generates grain-to-grain pressure contacts (Monger and Daugherty, 1991).

30.3.2.5.4 Translocation and Precipitation of Silicon

In solution, Si exists mostly as silicic acid $[Si(OH)_4]$. It moves with percolating water and precipitates when conditions are favorable. The soil pH influences the way in which silica precipitates. At pH < 7, $Si(OH)_4$ precipitates as SiO_2 on adsorption sites as individual molecules or low-molecular-weight polymers, whereas at pH 7–10, Si in solution takes the form of high-molecular-weight polymers, and these are flocculated by cations in solution. Other factors may alter, or override, the effect of pH on Si precipitation in soils. Silica precipitation is promoted in soil solutions with high ionic strength and adsorption is promoted by a solid phase with a high surface area. Release, leaching, and precipitation of Si favors preservation of ^{28}Si in the solid phase of soils, creating strong negative shifts in $\partial^{30}Si$ signatures in soils during weathering (Ziegler et al., 2005a, 2005b).

In soils, $Si(OH)_4$ is adsorbed on exposed hydroxyl groups of clays, (hydr)oxides, and primary silicates (Figure 30.7). A high surface area (e.g., argillic horizon) or ionic strength (e.g., calcic horizon) promotes rapid adsorption of silica as individual or low-molecular-weight polymers. Adsorption is greatest at pH 7–9. Drying dehydrates the adsorbed $Si(OH)_4$ causing amorphous SiO_2 to precipitate on the surface (Figure 30.7a). The strong Si–O bond inhibits rehydration and desorption. The precipitated silica acts as a template for further precipitation (Figure 30.7b) and, eventually, opaline silica forms bridges between the grains on which it precipitates (Figure 30.7c; Chadwick et al., 1987).

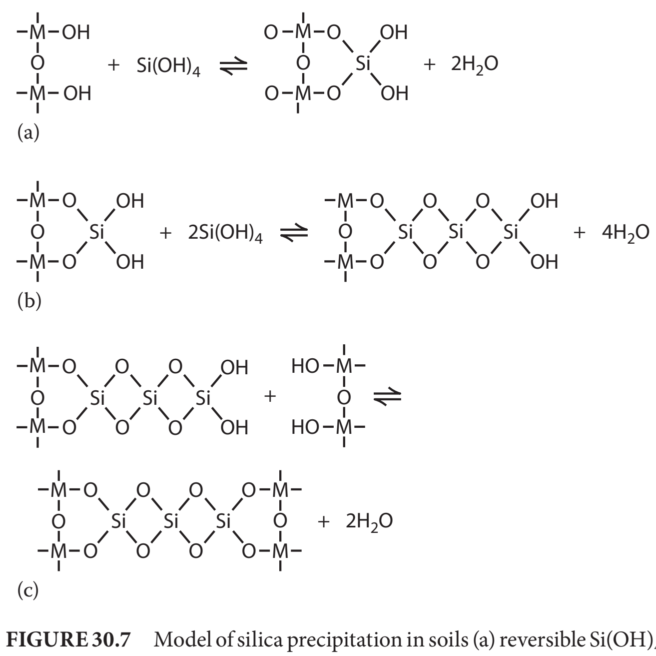


FIGURE 30.7 Model of silica precipitation in soils (a) reversible $Si(OH)_4$ adsorption on Fe and Al (hydr)oxides, silicates, and aluminosilicates (M may be Al, Fe, Mg, or Si); (b) further $Si(OH)_4$ adsorption followed by dehydration to SiO_2 during soil drying decreases the reversibility of the original adsorption process; (c) two soil components with $Si(OH)_4$ adsorption surfaces may be bonded together by opaline SiO_2 bridges. (Reprinted from Chadwick, O.A., D.M. Hendricks, and W.D. Nettleton. 1987. Silica in duric soils. 1. A depositional model. Soil Sci. Soc. Am. J. 51:975–982. With permission of the Soil Science Society of America.)

In fine-textured soil horizons, silica diffuses into smaller voids in the soil matrix, where there is more surface area for adsorption, and precipitates as polymerized layers. The zone of accumulation tends to correspond to the zone of clay accumulation because of the abundant sites for precipitation there (Kendrick and Graham, 2004). The silica polymers are found as flocs (≈1 μm diameter) on grains in the soil matrix and interlaminated with clay skins. The same mechanisms apply for coarse-textured soils, but silica not in direct contact with adsorption surfaces may precipitate by flocculation if the ionic strength is sufficiently high (Chadwick et al., 1987). Opaline silica may also precipitate preferentially on the undersides of gravels, even forming pendants (Munk and Southard, 1993).

In many soils of arid and semiarid regions, the depth distributions of silica and calcite overlap, but these pedogenic compounds are concentrated in different microsites by nature of their mode of precipitation. Calcite precipitates by self-nucleation and microbial processes, particularly in larger voids because these dry early, concentrating the soil solution. Larger voids also often have the most effective contact with the atmosphere, so the P_{CO_2} is relatively low. Silica specifically bonds with Al and Fe hydroxy compounds, whereas, under most soil conditions, calcium does not. Thus, silica precipitates primarily by adsorption on surfaces and so is found in the matrix, coating grains, and clay skins, rather than in voids. Furthermore, there is no direct chemical bonding between Ca and $Si(OH)_4$ at pH < 9, since $Si(OH)_4$ is neutral and nonpolar and cannot compete with polar molecules or anions for adsorption on calcite (Chadwick et al., 1987). Coatings of opal also occur as silans in macro voids when the soil matrix is plugged with calcite and silica cements.

30.3.2.5.5 Relation to Soil Morphology

A soil horizon that is thoroughly cemented by silica, such that it will not slake in water or HCl, is known as a duripan. Since the silica needs only to cement the matrix by bridging between grains, cementation can occur with very low concentrations of silica (e.g., ≈4%), particularly in coarse-textured soils. In a chronosequence study of silica accumulation in soils, an increase in rupture resistance was one of the first signs of silica precipitation, observed in 55 ka soils with no macroscopically visible silica precipitation (Kendrick and Graham, 2004). Duripans often have accessory cementing agents, such as Fe (hydr)oxides or calcite. Duripans exhibit two forms corresponding largely to the regional climate. In arid regions, duripans are commonly platy, with plates 1–15 cm thick. Pores and plates are coated with opal and are usually engulfed with pedogenic calcite. In Mediterranean climates, duripans take a different form in which opal coats the faces and pores of very coarse to extremely coarse (0.3–3 m diameter) prisms that make up the subsoil. The upper boundary is abrupt and the pan may have an opal coating on top in the strongly developed cases. Water often perches on top of the pan during the rainy season, so that Fe/Mn (hydr)oxide nodules and illuvial clay accumulate there. The matrix of this kind of duripan is extremely hard when dry but is brittle when moist and can be penetrated, with some difficulty, with a hand auger. Pedogenic calcite may, or may not, be present, but it is not a dominant component.

Reversible, weakly expressed silica cementation seems to play a role in the formation of fragipans. These horizons are dense and appear cemented when dry but will slake in water (Franzmeier et al., 1989; Marsan and Torrent, 1989). The strength of fragipans can be related to the ratio of $Si_d/(Si_d + Al_d)$, suggesting that an accumulation Si under low Al conditions is responsible for the hardening of fragipans, due to the precipitation of amorphous silica rather than aluminosilicates (Duncan and Franzmeier, 1999).

Silica cementation is often enhanced in soils near scarps or incisions, such as at terrace edges. Evaporation at the scarp surface concentrates the soil solution and promotes silica precipitation. The resulting induration impacts water movement and slope retreat, and renders the soil profile exposed along the scarp atypical of the soil under the broader geomorphic surface (Moody and Graham, 1997). Scarps in terrain with silica-cemented soils are often nearly vertical because cementation strengthens these edges (Kendrick and Graham, 2004).

30.3.2.6 Accumulation and Redistribution of Clay

Clay-sized particles have a dominant influence on soil properties due to their large, reactive surface area. Surfaces of clay minerals play a role in water retention, nutrient storage and exchange, and precipitation of secondary minerals. The redistribution of clay is a widely observed pedogenic process, which results in a horizon of clay accumulation (Bt horizon). The development of a Bt horizon, in turn, enhances other pedogenic processes, such as the reinforcement of soil structure and the precipitation of amorphous silica.

30.3.2.6.1 Sources of Clay

Clay (inorganic particles <2 μm diameter) is produced by weathering of primary mineral grains such as feldspars, micas, pyroxenes, and amphiboles. The ions released by weathering are carried in solution to other parts of the soil profile or the landscape and precipitated as colloids such as smectite, gibbsite, kaolin, or Fe (hydr)oxides. Commonly, as primary grains are weathered, secondary minerals replace them pseudomorphically. In this case, the secondary minerals are not initially dispersed as colloids in the soil (Nahon, 1991); they enter the clay size fraction when pseudomorph grains are crushed by turbation processes (Graham and Buol, 1990). Clay contained in some parent materials, especially sediments (e.g., alluvium, colluvium, lacustrine deposits, loess) and sedimentary rocks (e.g., shale, limestone, mudstone), is simply inherited in the soils derived from them. Another significant source of clay that is particularly important in arid regions is aerosolic input (Reheis and Kihl, 1995; Simonson, 1995). In the eastern Sierra, Nevada, California, spatial variability in eolian inputs helps to explain why pedogenic clay appears to increase with moraine age while other pedogenic processes are strongly confounded by erosion of the moraines (Rossi, 2009). In this chronosequence, older moraines are located closer to the valley floor, from which most eolian material is derived, and, therefore, the oldest moraines receive the greatest inputs of eolian-derived clays.

30.3.2.6.2 Mobilization and Transport of Clay

As the smallest particle-size fraction, clay is most susceptible to suspension in and transport by water percolating through the soil. The mobility of clay particles depends on the characteristics of the clay particles, the pore water chemistry, and the physical nature of water movement through soil. In general, the finest clay particles are more readily transported, moving even at low pore water flow rates (Kaplan et al., 1993), but even silt particles move in suspension if conditions are favorable. Clay-sized minerals with relatively strong negative charge, such as smectite, are most mobile.

Dispersion, the chemical repulsion of colloids in suspension, greatly enhances clay movement. Chemical conditions that favor dispersion in soils are those that result in an expanded diffuse double layer (Chapter 15), most notably Na-saturated clay and absence of soluble salts or partially soluble minerals (e.g., calcite, gypsum) that would increase the electrolyte concentration in solution or put Ca on exchange sites at the expense of Na. Dispersion is also enhanced by high pH because it imparts a negative charge to variable charge materials (e.g., (hydr)oxides, kaolin), which are then repulsed by each other and by the permanently negatively charged minerals. An example of this in natural systems is the effect of wood ash from forest fires, which may raise the surface soil pH above 10 (Ulery et al., 1993), causing kaolinite to disperse (Durgin and Vogelsang, 1984). Low-molecular-weight organic acids can be effective dispersive agents (Jenny and Smith, 1935; Kaplan et al., 1993, 1997; Kretzschmar et al., 1993, 1995). They complex cations in solution, thereby keeping them from flocculating clays, and they can specifically adsorb to positively charged mineral edges, preventing edge-to-face bonding with negatively charged minerals (Durgin and Chaney, 1984; Heil and Sposito, 1993a, 1993b). On the other hand, dispersion is inhibited by processes that produce stable aggregates. Thus, interparticle bonding that is enhanced by organic matter and noncrystalline inorganic compounds minimizes clay redistribution.

Mobilization and transport of clay particles by water is affected by the velocity of pore water flow and by the wetting and drying of pores. Slaking, which is the physical detachment of clay, is enhanced by increased pore water flow rates because it increases shear stress needed to detach particles from the matrix (Kaplan et al., 1993). Water flow through macropores occurs at greater velocities than micropore flow. As a result, macropore flow can move more clay and larger particles to greater depths. In fact, most clay movement is driven by macroscopic flow, even though it occurs less frequently than micropore flow. Wetting–drying cycles also enhance slaking. As a soil rewets, moving water can dislodge loose particles and carry them in suspension (Hudson, 1977). Another mechanism by which wetting–drying cycles promote clay transport is enrichment of clays with pH-dependent charge at the air–water interface in partially drained pores (Wan and Tokunaga, 2002). This partitioning results in a concentration of colloids that are readily mobilized during the next wetting cycle (DeNovio et al., 2004).

30.3.2.6.3 Deposition of Clay

Physical processes that cause a deposition of clays include immobilization by physicochemical interaction with grain surfaces or immobile air–water interfaces and straining of large clays by fine pores or thin films of water on grain surfaces (DeNovio et al., 2004). Deposition of clays may be promoted by chemical factors such as flocculation by high electrolyte concentrations as in saline or calcareous subsoils and adsorption of phyllosilicate clays with permanent negative charge to Fe and Al (hydr)oxides in acid subsoils (Jenny and Smith, 1935). Deposition of clay frequently occurs as the suspension travels into a drier part of the soil where the moving water is imbibed and retained by capillary forces. For example, when a suspension of clay flows through a macropore (Figure 30.8), water is pulled by matric forces into the soil fabric and the clay in suspension is deposited in or on the pore wall. Deposition of this type occurs in tubular, interpedal, and intergranular pores, and results in the characteristic parallel orientation of platy phyllosilicate clays on pore walls. Depositions of clay may be promoted by textural contrasts. Interruption of gravitational water flow as by a fine layer overlying a coarse layer leads to the deposition of colloidal material from the leaching water above the contact (Bartelli and Odell, 1960). The accumulation of clay above the contact then serves to reinforce the textural contrast, creating a positive feedback (Schaetzl and Anderson, 2005).

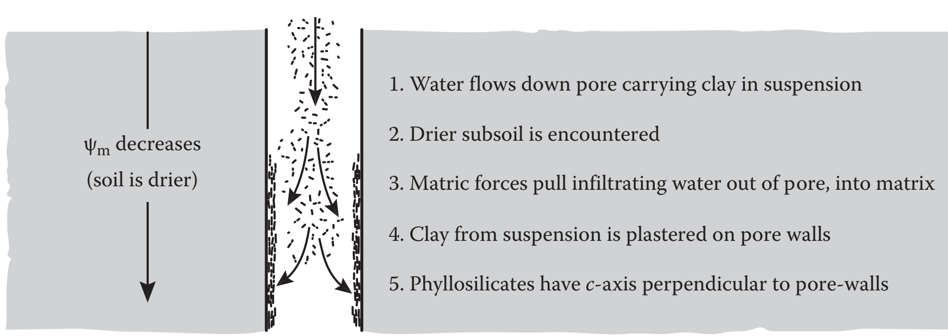


FIGURE 30.8 Illustration of clay deposition from suspension in a tubular pore to form an oriented clay lining (i.e., clay film, channel illuviation argillan).

30.3.2.6.4 Relation to Soil Morphology

Oriented clay coatings, known as clay skins, clay films, clay linings, or argillans, are recognized visually by a discernable thickness and a darker color compared to the inside of the soil peds. Clay skins form rapidly in the laboratory (Dalrymple and Theocharopoulos, 1984) and develop within several decades in the field if pedoturbation is minimal (Graham and Wood, 1991). They are often, but not always, enriched in finer clay sizes than the soil matrix as a whole. Clay coatings are typically most abundant in subsoils (Bt horizons) but are often best preserved and expressed within the fractures of underlying saprolite or bedrock (Graham et al., 1994). They may also occur in surface horizons in which eolian deposition provides clay that is then transported by water along ped and pore surfaces (Sullivan and Koppi, 1991). In sandy, quaternary-aged soils, illuvial clay may accumulate in bands called lamellae (Rawling, 2000). The deposition of clays in lamellae has been attributed to several factors including wetting-front-drying, chemical flocculation, and sieving caused by slight variations in texture of the parent material.

Clay deposition is a strong contributor to soil structure formation by creating distinct differences between the surface and interior of peds. Deposited clays on ped surfaces are strongly oriented and can be discerned easily using a petrographic microscope and cross-polarized light. Clay skins are destroyed in soils having extensive shrink/swell activity; they are replaced by shiny slickensides or pressure faces on ped surfaces, which can be distinguished microscopically from clay skins by a different pattern of clay orientation (Nettleton et al., 1969).

30.3.2.7 Complexation and Redistribution of Fe and Al

In many soils, particularly Spodosols, dissolved organic matter plays a critical role in pedogenic processes by complexing metals, predominantly Fe and Al, in surface horizons, translocating them, and depositing them in subsoils. This process enhances a distinctive style of mineral weathering, in which chelation removes weathering products, and produces a characteristic soil morphology, epitomized by an albic horizon overlying a spodic horizon (Section 33.9).

30.3.2.7.1 Reactive Agents and Sources

Dissolved organic acids act as the carriers of metal cations. Some are simple acids derived directly from leachates of relatively fresh plant material, either from the vegetative canopy or from the leaf litter at the soil surface. Typically, the more important dissolved organic acids in soils are the byproducts of microbial decomposition of organic matter produced in the O or A horizons. They are complex, heterogeneous, relatively low-molecular-weight organic acids referred to as fulvic acids (Chapter 11). Organic acids chelate and remove cations from the surface of mineral grains. This type of weathering is very effective, since it leaves a fresh grain surface with no coatings of secondary minerals to impede solution access to the surface, a condition that inhibits further weathering. Mycorrhizal fungi also promote weathering of primary minerals in the E horizon and may transport Fe and Al to the O horizon, where concentrations of dissolved organic carbon (DOC) are highest (Lundström et al., 2000). Since organic acids keep soil solution pH below the pK_{a1} of H_2CO_3 (6.4), bicarbonate weathering is not involved in these systems (Ugolini and Spaltenstein, 1992). Iron and Al are the cations preferentially removed by organic complexation since, being relatively small cations with high valence, they form the most stable chelates (Schnitzer, 1969). At a given pH, Fe and Al will remain in solution at a much higher concentration if organically complexed.

30.3.2.7.2 Translocation and Accumulation

Chelates, forming primarily in the O and A horizons, move in solution with percolating water. During intense leaching episodes, organometal colloids, as well as dissolved metal chelates, may be flushed downward in the profile (Stoner and Ugolini, 1988). The dissolved chelates precipitate when the metal:organic carbon ratio exceeds a critical value at which all polar-bonding sites are full. The precipitation of dissolved organic matter has been studied in the laboratory, by titrating extracts from organic horizons with solutions of Fe and Al (Nierop et al., 2002). Precipitation of dissolved organic matter by Fe^{3+} occurs at lower metal:organic carbon ratios and shows less pH dependence compared to precipitation by Al^{3+} (Figure 30.9). At higher

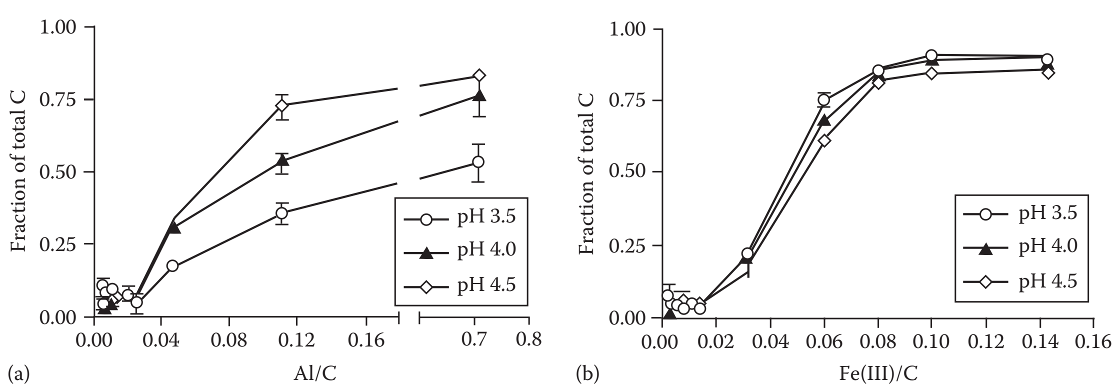


FIGURE 30.9 Fraction of total organic C precipitated form organic horizon extract solutions as a function of: (a) the Al/C molar ratio and (b) the Fe(III)/C molar ratio, at pH 3.5, 4.0, and 4.5, illustrating the precipitation of C in response to high metal-to-C ratios. (Reprinted from Nierop, K.G.J., B. Jansen, and J.A. Verstraten. 2002. Dissolved organic matter, aluminium and iron interactions: Precipitation induced by metal/carbon ratio, pH and competition. Sci. Total Environ. 300:201–211. Copyright (2002), with permission from Elsevier.)

pH, there is less proton competition, which leads to more metal interaction with dissolved organic matter and thus more precipitation of the organic–metal complexes (Nierop et al., 2002). The pH dependence of chelate solubility may explain why translocation of Fe and Al by dissolved organic matter does not occur in calcareous soils until leaching has lowered the soil pH to 5.0 or 5.5 (Schaetzl, 1996). In the field setting, chelates may be arrested even when the metal:carbon ratio is low, by adsorbing on positively charged material, such as Fe, Al (hydr) oxides, or high metal:organic C material already precipitated. Such reactions cause polymerization of soluble low-molecular-weight compounds into insoluble forms. Chelates may also be deposited by desiccation or by aggregation in a zone of relatively high ionic strength, and low H^+ activity, exposed negative charges. Alternatively, precipitation may occur because microbial degradation of the organic ligands releases the Fe and Al from soluble organometal complexes (Lundström et al., 2000). Spodic horizons are often identified by microscopic examination of thin sections that reveal the presence of organic and Fe (hydr)oxide-rich silt-size aggregates, many with cracked coatings indicative of postdepositional dehydration (Deconinck, 1980). Fe and Al accumulate in the spodic horizons as poorly crystalline minerals, ferrihydrite and imogolite (Lundström et al., 2000).

30.3.2.7.3 Relation to Soil Morphology

The O horizon is the major source of dissolved organic acids, which strip Fe and Al from the mineral soil as they move downward (Table 30.3). This produces the bleached E (≈albic) horizon, which exhibits the colors of the fresh primary mineral grains. The zone in which the chelates are deposited is the Bhs (or Bh, or Bs; ≈spodic) horizon. Localized intensified leaching associated with tree-throw pits or macropores (e.g., old root channels) can cause irregularities in horizon boundaries, including overthickened tongues of E horizon.

As time passes, weatherable minerals in the E horizon may become deeply pitted by dissolution while the B horizon is enriched with humus and metals. Microbial activity within the B horizon oxidizes organic C, increasing the metal:organic C ratios and releasing metals from the organic complexes to precipitate as poorly crystalline (hydr)oxides. Thus, the B horizon takes on a dark reddish brown color reflecting the humus and Fe (hydr)oxide components. Metal–humus complexes can accumulate in such a way as to produce cemented horizons. One such feature is the placic horizon, which is a thin (2–10 mm), hard, brittle, and wavy zone cemented by Fe/Mn humus. Quite commonly these features accumulate at a hydrological boundary, such as a change in particle size. A more massive cemented horizon is ortstein, which is essentially a spodic horizon cemented by Fe/Al humus. Both of these features require soil that is relatively free of physical disruption in order to form and persist. A study of carbon storage in relation to Fe and Al in grassland soils has suggested that the processes of chelation and transport of Fe and Al can occur in soils that do not develop typical spodic features (Masiello et al., 2004). In these soils, organic carbon stability was found to be strongly correlated with chelated metal ions in the A horizon and noncrystalline Fe and Al in the B horizon, suggesting a carbon storage mechanism similar to that of Spodosols. However, the morphologic expression of these processes is not visible because of the simultaneous translocation and accumulation of clays in B horizon.

30.3.2.7.4 Relation to Environmental Conditions

The process of Fe and Al translocation and accumulation is associated with specific conditions of climate and vegetation. In general, the effective climate is wet, to provide a strong leaching environment, and cool, to produce a low decomposition rate of organics. In these cool, humid environments, coniferous forests and ericaceous shrubs often prevail and are particularly effective in promoting chelation since they have acidic foliage.

As a consequence of this close relationship between climate, vegetation, and the process of Fe/Al translocation and accumulation in soils, Spodosols and similar soils are found in vast areas north of 45° latitude and at lower latitudes where high precipitation and low temperature prevail, such as coastal regions or

TABLE 30.3 Chemical Variables and Their Roles in the Complexation and Redistribution of Fe and Al in Soils

Variable	Major Role and Interaction with Other Variables	Trend, Interaction with Compartment			
		O	E and/or A	B	C
DOC	Major driving variable mobile anion, acidity source, metal complexing agent	Major source	Minor source	Major sink	No trend
pH	Low pH controlled by DOC; major variable	Lowered greatly	Lowered slightly	Rises greatly	Rises slightly
HCO_3	Controlled by pH, P_{CO_2}	Lowered	Insignificant	Insignificant	Rises significantly
Fe	Complexed and mobilized by DOC, causes DOC immobilization in B	Source	Major source	Major sink	Insignificant
Al	Complexed and mobilized by DOC, causes DOC immobilization in B; may be mobilized inorganically at low pH	Minor source	Major source	Major sink	Insignificant or minor
Basic cations	Leached in association with DOC (upper horizons) and with HCO_3^- (C horizons)	Major source	Source	Sink	Deep leaching (sink)

Source: After Marrett, D.J. 1988. Acid soil processes in the Okpilak Valley, artic Alaska. Ph.D. Dissertation. University of Washington. Seattle, WA.

high elevations. On the other hand, Spodosols are common in warm humid regions where soils are poorly drained. Organic acids in the groundwater chelate Fe and Al, which are concentrated by the fluctuating water table to precipitate in a Bh horizon. The Bh horizons typically contain chelates but little or no free Fe as (hydr)oxides since it is reduced and leached away. Soils formed in this way have very pronounced albic and spodic horizons. They are common in the coastal plains of the southeastern United States and tropical forests such as the Orinoco basin in Brazil. The observation of close ties between organic matter stability and chelated forms of Fe and Al in moist grassland soils, without spodic properties, suggests that redistribution of Fe and Al in association of organic chelates may occur in a wider range of humid environments than previously recognized (Masiello et al., 2004).

Parent material composition is also very influential in the process of metal–humus translocation and accumulation. The process is promoted by relatively low levels of Ca, Fe, and Al. High levels of these cations prevent mobilization because a high metal:organic C complex forms quickly that is not soluble and readily translocated. Furthermore, high Ca contents promote microbial activity that decomposes soluble organics, so they are not available for chelation and leaching. As a rule, translocation of Fe– and Al–humus complexes is favored by silicic or felsic parent materials but not carbonate or mafic materials. It is further favored by coarse-textured materials (sand to coarse loamy) where surface area is low, water infiltration is rapid, and cation release is slower.

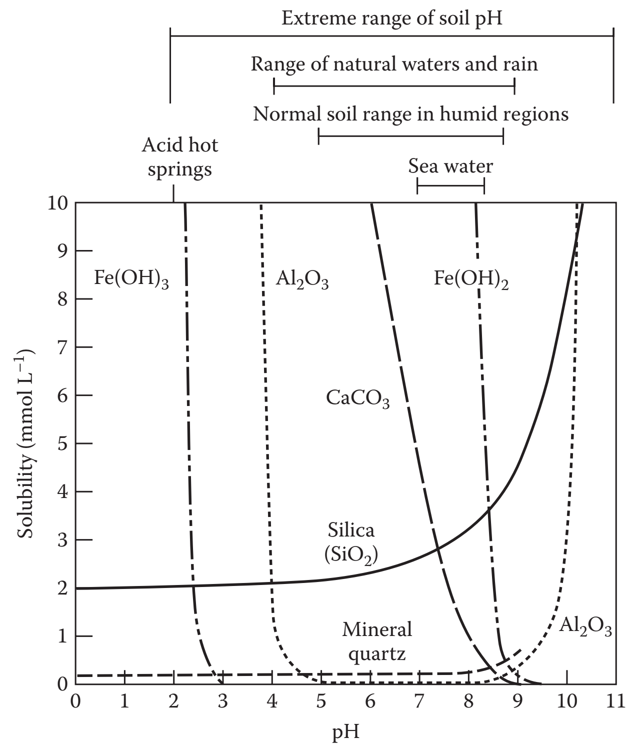


FIGURE 30.10 Mineral solubility as a function of pH.

30.3.2.8 Desilication and Concentration of Resistant Oxides

Weathering of primary minerals occurs to different degrees in all soil environments. It releases highly mobile basic cations, moderately mobile $Si(OH)_4$, and relatively immobile Al^{3+} and Fe^{3+} into soil solution. The fate of these ions depends on soil leaching intensity, organic matter composition, the amount of reactive surface area, and pH. In this section, the cumulative effect of leaching on soil volume change, mineral composition, and soil fabric is discussed.

30.3.2.8.1 Leaching and Mineral Composition

In humid environments, well-drained soils lose many of the mobile constituents in a process that lowers pH and changes solution ionic composition from base cations and $Si(OH)_4$ to Al. At neutral pH, Si is more soluble than Al or Fe (Figure 30.10). In the pH range from 5 to 7, much silica can be lost by leaching. In weathered soils of the North Carolina Piedmont, 39%–75% of Si was lost relative to the parent material, but only 5%–13% of Fe and Al was lost (Oh and Richter, 2005).

Below pH 5, Al is also leached but commonly at slower rates than Si because Al can be strongly sorbed by organic matter. Usually there is enough Al in soil minerals that the pH is buffered near 5 and seldom drops to levels (<4) where Fe is dissolved due to acidity alone (Van Breemen et al., 1983). Reducing conditions are required to solubilize Fe as described in Section 30.3.2.9. Thus, leaching changes soil mineral stability fields in favor of minerals composed of Fe and Al and relatively small amounts of Si.

A typical soil mineral assemblage accumulated after intense weathering in a humid environment is shown in Figure 30.11.

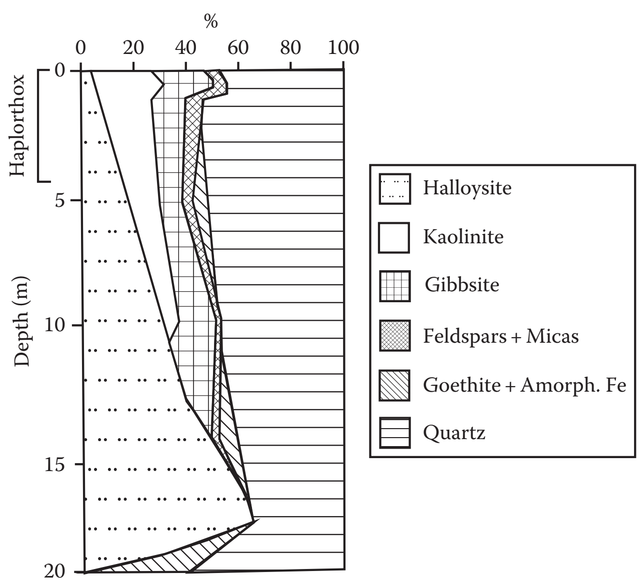


FIGURE 30.11 Mineral distribution in an Oxisol and associated saprolite on granite. (From Eswaran, H., and W.C. Bin. 1978. A study of a deep weathering profile on granite in peninsular Malaysia. I. Physicochemical and micromorphological properties. Soil Sci. Soc. Am. J. 42:144–149. With permission of the Soil Science Society of America.)

The secondary mineral assemblage of highly weathered soils includes kaolin and sometimes hydroxy-interlayered vermiculite (containing Si and Al), gibbsite (containing Al), and hematite and goethite (containing Fe) (Lynn et al., 2002). There is also quartz that is inherited from the parent material or added by dust, and concentrated by the dissolution of other primary minerals. These minerals have low nutrient retention and supply capabilities, which means that nearly all plant nutrients must be derived from breakdown of organic matter or from atmospheric deposition. Because Si has been dramatically depleted and that remaining resides in quartz and kaolinite, these minerals are subject to weathering to a greater extent than in soils that still contain weatherable primary minerals. Even though kaolinite is considered to be a stable end product of weathering, it can decompose (and form) quite rapidly in highly leached soils (Giral-Kacmarcik et al., 1998; Ziegler et al., 2005a). Silicon is conserved by biocycling between rainforest vegetation, where it forms opal phytoliths, and soils, where it is released by weathering of the phytoliths (Lucas et al., 1993; Alexandre et al., 1997; Derry et al., 2005). The preservation of kaolin minerals in the surface horizons of some highly weathered soils may be explained by biocycling of silicon (Kleber et al., 2007).

30.3.2.8.2 Leaching and Collapse

As primary rock minerals are dissolved, they commonly lose volume (Chadwick et al., 1990; Brimhall et al., 1992). This process is countered by any mass addition to soil, which will serve to expand or dilate it. Quantitative studies of leaching losses during soil formation account for these changes through the use of index minerals (or elements), which are extremely resistant to weathering (Brewer, 1964; Smeck and Wilding, 1980; Brimhall and Dietrich, 1987; Brimhall et al., 1992). Dilation is indicated when the quantity of an index element in a soil horizon is less than in the parent material and soil collapse is indicated when the opposite is true. As depicted in Figure 30.12, young soils are often dilated because accumulation of organic matter is more rapid than mineral weathering. Older soils show progressively greater collapse because mineral weathering becomes the dominant control as organic C accumulation is balanced by microbial respiration (Chadwick et al., 1994). Figure 30.12 was constructed by analyzing soils developing in beach sand on a suite of progressively older uplifted marine terraces (Brimhall et al., 1992; Merritts et al., 1992); similar pattern, of early C driven dilation followed by weathering driven collapse occur in soils on lava flows in Hawaii (Vitousek et al., 1997).

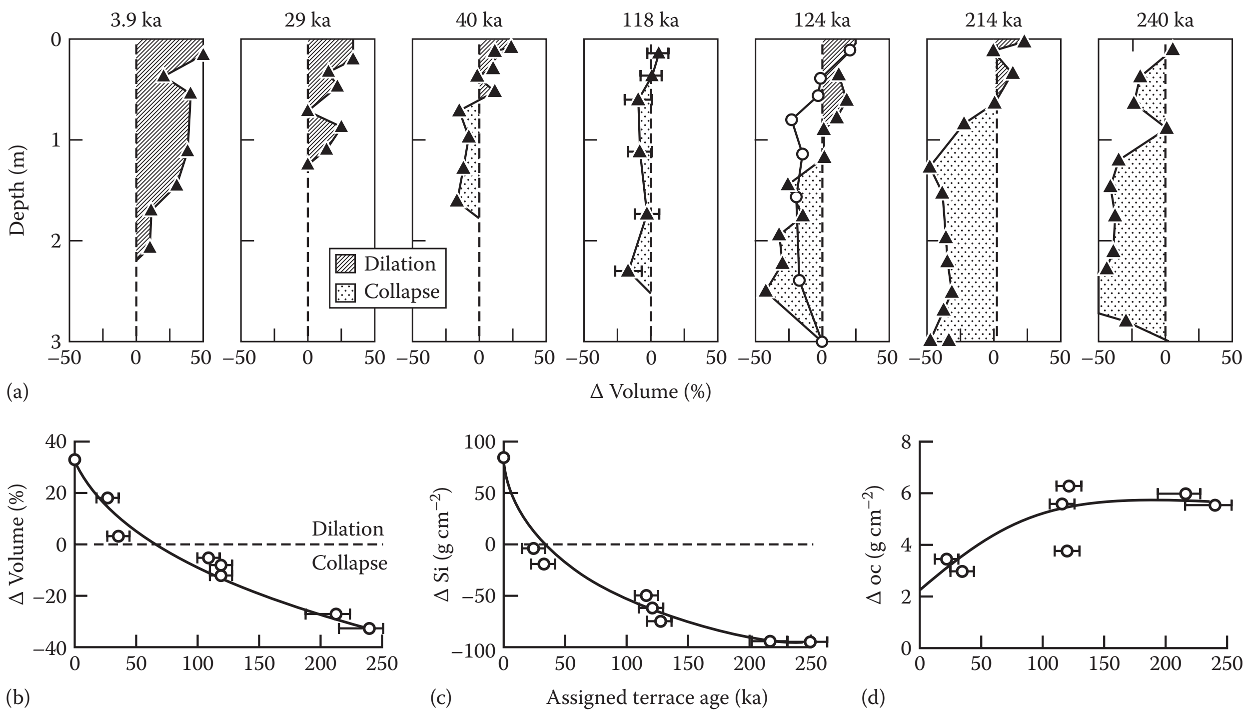


FIGURE 30.12 Parent material and soil values for bulk density and Zr are used to compute volume change that occurs as soils develop from beach sand to 240 ka on uplifted marine terraces. (a) Volume change plotted as a function of depth for each profile sampled; the 124 ka terrace was sampled in two different locations; (b) volume change integrated to the sampling depth (depth-weighted mean) to provide an average value for each profile plotted as a function of terrace age; (c) the quantity of Si leached from each profile as a summation over the soil sampling depth plotted as a function of terrace age. For these soils, base cations are leached more rapidly, and Al is not leached to a significant extent. Given the mass dominance of Si in the arkosic beach sand, Si leaching is the main control on soil collapse; (d) the quantity of organic carbon summed over the sampling depth plotted as a function of terrace age. Organic matter accumulation is the dominant factor driving early dilation and partly offsetting desilication in the older profiles. (Modified from Brimhall, G.H., O.A. Chadwick, C.J. Lewis, W. Compston, I.S. Williams, K.J. Danti, W.E. Dietrich, M.E. Power, D. Hendricks, and J. Bratt. 1992. Deformational mass-transport and invasive processes in soil evolution. Science 255:695–702. Copyright American Association for the Advancement of Science.)

30.3.2.8.3 Accumulation of Atmospherically Derived Constituents

Soils that reside on old stable geomorphic positions accumulate atmospherically transported minerals. For example, dust from Africa augments soils in the Amazon rainforest (Swap et al., 1992; Okin et al., 2004) and dust from Asia can be found in Hawaiian soils (Jackson et al., 1971; Kurtz et al., 2001). In parts of Africa and Australia, deeply weathered bauxite deposits are produced partly by the accumulation of chemically mature Fe and Al compounds that are blown in from other regions (Brimhall et al., 1988). These eolian additions are translocated into the top few meters of a deposit through macropores created by biological activity where they effectively dilate previously collapsed horizons (Brimhall et al., 1992). Below the lower limit of root growth, translocation is no longer accommodated by smaller pores in the saprolite, so leaching has led to collapse. Because these soils have accumulated eolian additions of highly weathered Fe and Al compounds, Si and basic cation losses could be overestimated when comparing soil horizons to underlying material. In contrast, when unweathered primary minerals are added to soils, it is possible that leaching losses will be underestimated. Rate of dust deposition may be similar to weathering rates, depending of mineralogy and climatic conditions (Porder et al., 2007). When weatherable dust is deposited on soils formed from resistant rock types, dust may be the main contributor to the weathering flux. On the other hand, in Hawaii, quartz-rich dust is added to soils formed on mafic bedrock, so that the dust is concentrated in the surface horizons as the bedrock parent material is broken down (Kurtz et al., 2001).

30.3.2.8.4 Relation to Soil Morphology

Horizons in the top 1–2 m of old, highly weathered soils display fine micropeds (<2 mm), which can agglomerate into larger structural units, but they rarely exceed 50 mm in diameter (van Wambeke et al., 1983). The peds are composed of strongly interbonded kaolinite and Al and Fe (hydr)oxides that have low CEC, which defines, in part, the existence of an oxic or kandic diagnostic horizon. In oxic horizons, clay films are rare both because of high aggregate stability and intense bioturbation caused by ants and termites. The soil has high microporosity within peds and relatively large interped pores, which produce excellent permeability.

In Fe-rich soils that are imperfectly drained, localized reduction and oxidation allows Fe to become mobile for short distances. Its redistribution produces localized accumulations of hematite or goethite, which appear as reddish or yellowish mottles. The location of initial deposition can be in small pores where it is thought that precipitation is favored because the chemical potential of water in close association with the solid phase is at a minimum (Tardy and Nahon, 1985), although precipitation can also occur in large pores where high P_{O2} can lead to oxidation (Bouma, 1983). Initially, the accumulated Fe forms a weakly defined glaebule, or segregated body, that includes primary mineral grains and secondary clays as well as Fe (hydr)oxides. Slowly, the glaebule becomes more clearly defined because the engulfed minerals decompose and are replaced by Fe (hydr)oxide (Nahon, 1991). Mineral decomposition is driven by alternating oxidation and reduction of Fe, which creates acidity for enhanced hydrolysis (Brinkman, 1978). Continued growth of the glaebule produces an abrupt boundary between it and the surrounding soil matrix. In time, glaebules grow into each other producing a reticulate pattern of soil material enriched in Fe (hydr)oxides called plinthite, interspersed with less red, kaolin- and gibbsite-rich matrix. Plinthite in perennially moist soils remains soft, but under wetting and drying conditions, it can solidify into iron stone nodules or continuously cemented ferricrete. Wetting and drying can be the result of natural climate, climatic drying, erosional dissection of a plateau, or excavation of a road (Daniels et al., 1971; Nahon, 1991). There is no certain evidence on how long it takes for plinthite to harden when exposed to wetting and drying conditions.

30.3.2.9 Reduction and Oxidation Leading to Depletions and Concentrations

Environmental factors, particularly climate, topography, and the chemical and physical nature of the substratum create the drainage properties of soils, which, in turn, influence the intensity, duration, and spatial occurrence of anoxic conditions. Soils are susceptible to anoxia because O_2 is consumed by belowground microbial respiration but can only be supplied to soil pores by diffusion from the aboveground atmosphere. Soils with few macropores and many water-filled pores often consume O_2 more rapidly than it can be resupplied by diffusion. Many soils will be anoxic for short periods right after intense wetting events and the interiors of peds may be anoxic, even in otherwise well-aerated soils. When O_2 is depleted, microbially induced reduction reactions result in the dissolution of redox sensitive compounds, with the common result being an increase in their solubility, leading to selective elemental loss from the anoxic area and subsequent precipitation during exposure to higher O_2 levels. These processes leave long-lasting visible imprints on soil morphology.

30.3.2.9.1 Reactive Agents in Redox Processes

The reactive agents in redox processes include organic matter, oxygen, Fe and Mn (hydr)oxides, nitrates, sulfides and sulfates, and microbes (Chapter 14). Microbial activity is the key to reduction in soils. Microbes require a C source, supplied by solid or dissolved organic matter, and electron acceptors. In well-aerated soils, O_2 is the electron acceptor, but as it is used up, nitrates, Mn and Fe (hydr)oxides, and sulfates are used by different microbial populations. Such anaerobic conditions are usually associated with saturated or very wet soils in which there is little free pore space for the influx of O_2 from the atmosphere. Each electron acceptor compound is associated with a characteristic range of redox potentials (Chapter 14). Oxygen levels and redox potentials within a soil may show extreme variation at any given time, even on a scale of millimeters, such as from the exterior to the interior of an aggregate (Zausig et al., 1993). The behavior of nutrients that are not redox sensitive may be influenced through their interactions with redox-sensitive species. In oxidized soils rich in Fe, considerable amounts of phosphorus can sorb to iron oxides; however, in reduced soils, the Fe sink for P is lessened and more of it is lost by leaching or remains in organic form in thickened O horizons (Miller et al., 2001).

In addition to Eh, pH also controls the form of redox sensitive species in soils. The theoretical relationship between Eh, pH, and the form of redox sensitive species can be described using stability diagrams (Vepraskas and Faulkner, 2001). However, caution is required in using these thermodynamic relationships to described real world soils, in which the soil solution composition and variation in reaction rates complicate the thermodynamic relationships expressed in a stability diagram. The effect of pH on redox reactions in soils can be observed in certain field settings. This is demonstrated by comparison of two soils with low-lying positions, seasonally high water tables, and reducing conditions on the North Carolina piedmont (McDaniel and Buol, 1991). An Ultisol formed in biotite gneiss with a very strong acid subsoil was observed to be gleyed with very little manganese, while an Alfisol formed in hornblende–epidote gneiss with a moderately acid subsoil contained concentrations of secondary Mn.

30.3.2.9.2 Oxidation of a Reduced Soil

As a wet soil drains, macropores are the first to lose water. Often shrinkage occurs during the drying of soils, so that cracks form between peds. Oxygen penetrates through the interpedal pores, root channels, macrofaunal burrows, and other macropores. In response to the higher Eh that develops within these voids, Fe and Mn (hydr)oxides precipitate from the soil solution within the matrix near the pore wall surface. Precipitation of these (hydr)oxides removes the metal from solution, thereby establishing a diffusion gradient, causing Fe^{2+} and Mn^{2+} to migrate from the still reduced soil matrix to the oxidizing zone adjacent to the macropores (Fanning and Fanning, 1989). Redox sensitive elements such as Fe and Mn often accumulate along macropores or the exterior of peds (Figure 30.13). When a fine-textured horizon overlies a coarse layer, Fe and Mn (hydr)oxides may be concentrated in the upper part of the coarse deposit, reflecting the oxidation of reduced Fe and Mn as they move into the larger pores of the coarse-textured deposit (D'Amore et al., 2004). Oxidized microsites within a reduced matrix, which may be more common in coarse-textured soils (D'Amore et al., 2004), provide the conditions for formation of hard aggregates of Fe and Mn (hydr)oxides, described as nodules or concretions. These features form by an initial precipitation of Fe or Mn (hydr) oxides within the microsite, which acts as a template for further precipitation. When Fe (hydr)oxides precipitate, the color is red, orange, or yellow, depending on the species of Fe (hydr)oxide formed (Chapter 22). Manganese (hydr)oxides are a strong pigment and if they are present, even at levels of tenths of a percent, a pore wall will be blackened.

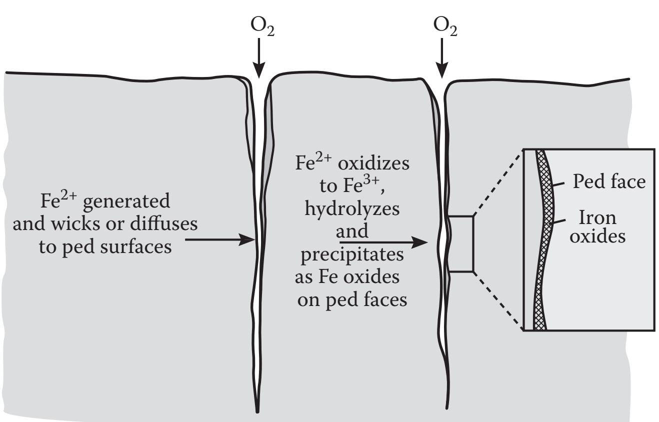


FIGURE 30.13 Macropores (e.g., cracks) facilitate O_2 precipitation of Fe (hydr)oxides on macropore wall surfaces (e.g., ped faces). (Modified from Fanning, D.S., and M.C.B. Fanning. 1989. Soil morphology, genesis, and classification. John Wiley & Sons, New York. Copyright Wiley-VCH Verlag GmbH & Co. KGaA. Reproduced with permission.)

30.3.2.9.3 Reduction of an Oxidized Soil

Soils that are generally well aerated and oxidized periodically become so wet that macropores are filled with water and O_2 is excluded. Spatial and temporal variability in reducing conditions occurs within saturated soils, primarily due to factors that affect the rate of microbial activity, including soil temperature and organic matter content. The length of time required for a saturated soil to develop reducing conditions is variable. On floodplain soils of the Mid-Atlantic Piedmont, it took 20 days for reducing conditions to develop when soil temperatures were <4°C, but only 2 days when the soil temperature was >9°C (Vaughan et al., 2009). This relationship between temperature and development of reducing conditions is attributed to the lower activity of soil microorganisms at low temperatures.

The spatial distribution of reducing conditions is influenced by organic matter from roots, which typically persists in old root channels and along ped faces of strongly structured soils. This organic matter provides an abundant source of C for microbes, and Fe (hydr)oxides in the soil adjacent to the macropore serve as electron acceptors and are reduced to yield Fe^{2+} in solution. At the same time, particularly in highly structured, fine-textured soils, ped interiors and soil matrix at some distance from the macropores may contain sites that are not water saturated and retain relatively high Eh conditions. In this case, Fe^{2+} mobilized from soil adjacent to the pores diffuses into the higher Eh environments within the matrix, where it precipitates as Fe (hydr)oxides (Fanning and Fanning, 1989; Vepraskas, 1996). Soil materials near the pore is the site of redox depletion and is generally a gray or white color, whereas redox concentrations of Fe (red, orange, or yellow) occur within the soil matrix (Figure 30.14). Because Mn reduction and oxidation occur at higher Eh values than Fe, Mn (hydr)oxides should also be dissolved near the water-saturated pores and precipitated within the soil matrix but not as far from the pore as the Fe.

30.3.2.9.4 Redox Soil Systems

A classic expression of the effects of varying redox conditions is found in the catena concept, where upslope soils are usually well

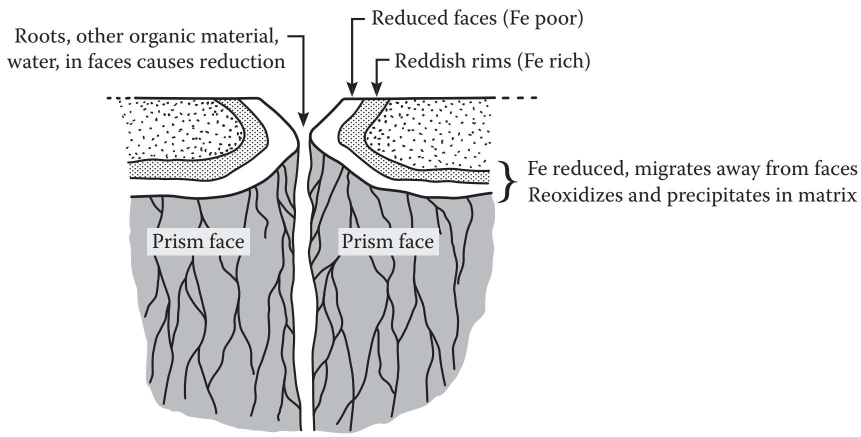


FIGURE 30.14 Occasional, localized reduction occurs in well-drained soils when macropores (e.g., cracks between peds) are temporarily filled with water. Fe (hydr)oxides in rims of peds are dissolved and removed or reprecipitated in more oxidized interior zones of the peds. (After Fanning, D.S., and M.C.B. Fanning. 1989. Soil morphology, genesis, and classification. John Wiley & Sons, New York. Copyright Wiley-VCH Verlag GmbH & Co. KGaA. Reproduced with permission.)

drained and oxidized, while soils at the base of the slope are, at least seasonally, poorly drained and reduced (Fanning and Fanning, 1989). As a result of these topographically induced pedochemical conditions, the upslope soils contain relatively abundant Fe (hydr)oxides and are reddish, whereas soils at the base of the slope contain few Fe (hydr)oxides, are enriched in Mn (hydr) oxides, and generally have low chroma colors (Weitkamp et al., 1996). Redox potentials in the upper slope soils are sufficiently high so that Fe (hydr)oxides are not reduced, although they are occasionally low enough to reduce Mn (hydr)oxides. Mn^{2+} is mobilized and transported to the base of the slope, where Eh is often low enough to cause reduction and loss of Fe (hydr)oxides; thus, Fe and Mn are lost from these soils, but only Mn is replaced by additions from upslope soils.

Soils that contain a slowly permeable layer, such as a dense argillic horizon, fragipan, or permafrost layer, may develop a seasonally perched water table. The resulting epiaquic conditions may not substantially deplete Fe or Mn from the soil as a whole, since there is little downward leaching. The redox sensitive elements are redistributed above and within the restrictive layer, largely by processes described above for the reduction of oxidized soils (Fanning and Fanning, 1989). An E horizon, depleted of Fe and Mn (hydr)oxides by reduction and lateral transport, may develop above the restrictive layer (McDaniel and Falen, 1994).

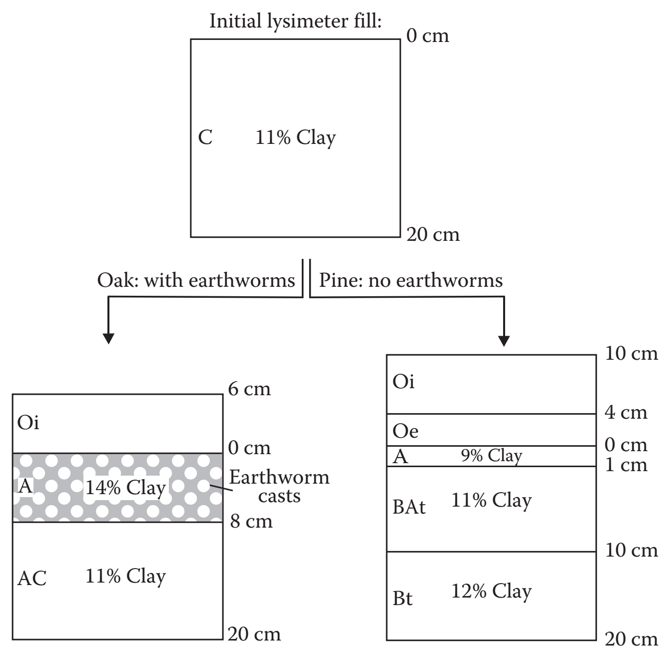


FIGURE 30.15 Soil development after 41 years in lysimeter fill that has been bioturbed by earthworms (oak lysimeter) and in earthworm-free lysimeter fill (pine lysimeter). The bioturbated soils have thinner organic horizons, deeper accumulation of organic C in the mineral soil, and clay accumulation at the surface rather than in the subsoil. (From Graham, R.C., and H.B. Wood. 1991. Morphological development and clay redistribution in lysimeter soils under chaparral and pine. Soil Sci. Soc. Am. J. 55:1638–1646.)

30.3.2.10 Mixing of Soil Materials

Processes that mix soil materials are sometimes viewed as regressive, with the idea that they oppose horizonation. In fact, while intensively mixed soils often display weak horizons, more moderate degrees of mixing can actually play a key role in forming soil horizons (Johnson et al., 1987). For example, bioturbation helps to form A horizons by mixing organic materials into the mineral soil (e.g., Graham and Wood, 1991; Figure 30.15) and various forms of mixing may obliterate rock structure, thus deepening the regolith and promoting soil formation (e.g., Heimsath et al., 1999).

30.3.2.10.1 Agents and Conditions

Many agents of soil mixing can be identified (Hole, 1961); however, the most widely observed include soil biota, seasonally thawed ice, and expansive clays. Mixing by soil biota, referred to as bioturbation, can be caused by both plants and animals. When trees are uprooted, large volumes of soil materials are often moved with the root system. Bioturbation by tree-uprooting is promoted by conditions that prevent deep-rooting (e.g., shallow bedrock, hardpan, shallow water table), making the trees less wind stable (Schaetzl et al., 1989). Animals that are strong agents of bioturbation include burrowing animals (e.g., gophers, earthworms) and mound-building animals (e.g., ants, termites). Mixing by ice, or cryoturbation, is the dominant soil-forming process in permafrost-affected soils (Bockheim and Tarnocai, 1998). Cryoturbation occurs primarily in the seasonally thawed zone above the permafrost, referred to as the active layer. Mixing caused by the shrinking and swelling of clays, referred to as argilliturbation, occurs when clay-rich soils are subject to seasonal wetting and drying cycles, causing vertical cracks to open and close. Argilliturbation is most pronounced in soils rich in expansive clays, such as smectites. Though these agents work to mix the soil by different mechanisms, they are related in that they apply forces that move bulk soil materials and open up voids through which soil material can move.

30.3.2.10.2 Bioturbation

Soil material can be moved upward, downward, or laterally by bioturbation. Burrowing animals move soil upward by depositing excavated soil material at the surface. The material deposited at the surface is size fractionated to different degrees, depending on the animal. Earthworms sort the fine-earth material by ingesting more silt and clay, relative to sand (Graham and Wood, 1991; Figure 30.15), while gophers sort the coarse fragments, by moving gravels, but not cobbles, stones, or boulders (Johnson et al., 1987). Downward movement of soil material by biotic agents can occur due to backfill of macropores left open by soil organisms or by intentional transport of materials by burrowing animals. For example, some earthworms will drag leaves into their burrows, thus mixing organic horizon material into the mineral soil. Lateral transport of soil material occurs when trees are uprooted; leaving an open pit where the tree was standing and creating a mound next to the pit, where soil material slumps off of the rootwad. Mound-building animals can also be responsible for lateral movement of soil materials. For example, when gophers inhabit shallow or poorly drained soils, they redistribute the soil into mounds that are deep enough to build their burrows (Cox and Scheffer, 1991).

30.3.2.10.3 Cryoturbation

Several mechanisms of cryoturbation have been described, which may act individually or together to cause soil mixing. These mechanisms may act during freezing, thawing, or in a multistep process that involves both. Mixing due to cryostatic pressure and the frost-pull mechanism of stone movement are processes associated with freezing. The active layer freezes in fronts moving both down from the surface and up from the permafrost. The volume expansion associated with freezing causes pressure that mixes the unfrozen material between the freezing fronts (Ping et al., 2008). Frost pull occurs when the soil freezes from the surface and the primary direction of expansion is upward. Coarse fragments carried with the upward-expanding soil undergo a net upward displacement if return to their original position upon thawing is prevented by fine material infilling beneath the coarse fragment (Washburn, 1980). Convection cell mixing is a process that is associated with the thawing of frozen soil above the permafrost. Mixing by convection cells occurs over bowl-shaped features of the permafrost table, which cause inward slumping of the thawed soil and upward displacement at the center of the bowl (Mackay, 1980). Frost wedging is a mechanism of cryoturbation that occurs due to the freeze–thaw cycle. When the active layer is frozen, thermal contraction at temperatures below −10°C causes cracks to form, and then as it thaws, melt water fills in the cracks and refreezes. Expansion of the water as it refreezes (between 4°C and 0°C) causes pressure on the surrounding soil and distortion of the land surface (Washburn, 1980).

30.3.2.10.4 Argilliturbation

Mixing by argilliturbation occurs by drying-induced shrinkage of the soil matrix and formation of desiccation cracks, followed by partial infilling of cracks from overlying horizons. The infilling material may be knocked in by biota, washed in by rain, or otherwise carried by gravity. Soils in which shrinkage increases with depth are most susceptible to these transport mechanisms because desiccation cracks become wider with depth, causing the wall to overhang the void (Hallsworth et al., 1955). When the dry, cracked soil is exposed to heavy rains or flooding, the zone beneath the cracks may wet up and expand first, causing subsoil material to be thrust toward the surface. Coarse fragments may also be moved upward, due to the swelling of clays and expansion of the soil volume upward, in the direction of least resistance. If the coarse fragments are smaller than the desiccation cracks, they may be transported back down during subsequent drying cycles. However, if the fragments are larger than the desiccation cracks they are transported upward only and accumulate at the soil surface (Johnson and Hester, 1972).

30.3.2.10.5 Relation to Soil Morphology

Mixing counteracts redistribution processes that form subsurface horizons, transports organic matter to greater depths, and causes irregularities in horizon topography. For example, in a biosequence composed of soils formed under oak, which contained earthworms, and adjacent earthworm-free soils formed under pine, only the pine soils had an argillic horizon (Graham and Wood, 1991; Figure 30.15). Mixing by the earthworms prevented illuvial clay accumulation under the oak. Mixing can also create thicker organic-rich horizons due to the infilling of animal burrows or desiccation cracks with A horizon material. In cryoturbated soils, organic matter is often observed to accumulate on top of or within the permafrost, as a result of downward transport of organic-rich materials from the surface horizons (Bockheim and Tarnocai, 1998; Bockheim, 2007; Ping et al., 2008). Examples of horizon boundaries influenced by mixing processes included swirl-like horizon patterns caused by cryostatic pressure in Gelisols (Bockheim and Tarnocai, 1998) and irregular horizon boundaries caused by up-thrust C horizon material in Vertisols (Hallsworth et al., 1955).

Processes that mix the soil may cause coarse fragments to concentrate at a certain depth in the soil. Coarse fragments too large for transport by burrowing or mound-building animals occur in a layer beneath a mantle of finer-textured material that the animals have brought to the surface (Johnson et al., 1987). Thus, bioturbation can result in a subsurface concentration of coarse fragments, which occurs at the maximum depth of faunal activity. Argilliturbation, on the other hand, concentrates coarse fragments at the surface, in the form of a stone pavement. Stone pavement formation by argilliturbation is exemplified by Vertisols on the Channel Islands, California, which are formed in sediments deposited above bedrock, yet are observed to have an enrichment of bedrock fragments at the soil surface (Johnson and Hester, 1972). Stone pavements may also form through freeze–thaw processes in glacial landscapes (Simón et al., 2000).

Mixing often occurs in localized patches, creating heterogeneity and microtopography in soil landscapes. Microrelief created by bioturbation includes mounds built by soil fauna and pit and mound topography caused by tree throw. Pit and mound topography in Spodosols has been related to variations in horizon thickness due to more intensive leaching of the pits (Schaetzl, 1990). Microhighs and lows, referred to as gilgai relief, are a common feature of argilliturbated soils. The microhighs are often associated with soil material from below the zone of mixing that has been up-thrust toward the surface (Hallsworth et al., 1955). In cryoturbated soils, microrelief and/or sorting of coarse and fine soil materials occurs in repeated shapes, referred to as patterned ground. Examples of patterned ground included polygons, ranging from a few centimeters to a few meters in diameter, formed by frost wedging (Ping et al., 2008) and hummocks, 1–2 m in diameter, formed by slumping of the active layer as it thaws over a curved permafrost table (Mackay, 1980). In between these units, upward injection of C horizon material may occur, resulting in a sorted form of patterned ground (Vandenberghe, 1992; Ugolini et al., 2006).

30.4 From Property to Process

For convenience, specific soil-forming processes have been identified and discussed separately. This leaves the impression that one can easily interpret soil-forming processes from soil properties. To a degree this inversion can be done quite well, but care is required. Many processes occur in the same pedon simultaneously, which imprints a less than clear suite of properties. Furthermore, secondary soil-forming processes may be initiated later in soil development in response to properties developed by the primary processes. Numerous examples of such intrinsic pedogenic thresholds illustrate the complex interactions that can occur between pedogenic processes (Chadwick and Chorover, 2001; Ewing et al., 2006). Development of cemented subsurface horizons can have a strong influence on the pedogenic processes that occur in the surface horizons. For example, by decreasing rooting depth, cemented horizons in forest soils become more susceptible to bioturbation by tree-uprooting (Schaetzl et al., 1989). Altered drainage due to a cemented horizon may also dramatically change the pedogenic processes, as demonstrated in a chronosequence study in southeast Alaska, in which initial processes of Fe and Al translocation form Spodosols with a cemented placic horizon (Ugolini and Mann, 1979). The cemented horizon impedes drainage leading to anaerobic conditions and accumulation of organic material, resulting in the formation of Histosols. Development of surface horizons that impede infiltration, such as desert pavement and vesicular horizons, may also change the pedogenic processes. These horizons cause decreased leaching, leading to accumulation of salts in the subsoil (Young et al., 2004), and increased runoff, which may eventually lead to erosion of the restrictive surface (McDonald et al., 1995). The restriction of vegetation due to stone pavement formation may also reduce the accumulation of organic matter (Simón et al., 2000). The accumulation of clay in an argillic horizon may lead to the secondary process of argilliturbation and formation of a Vertisol, especially if erosion exposes the clay-rich subsoil to surface wetting/drying cycles (Graham and Southard, 1983).

Pedogenic processes may also change over the course of soil development because of changes in the soil-forming factors over time. Soils are often polygenetic because their properties developed under different climatic conditions. Past climate change can superimpose different processes on a pedon at different times during soil formation. Soils formed in glacial deposits are subject to extreme variation in climate between glacial and interglacial periods. Glacial deposits in the Rocky Mountains have undergone phases of carbonate accumulation during interglacial periods, interrupted by cryoturbation during glacial advances (Hall, 1999). Preglacial soils in the Sierra Nevada Mountains of southern Spain include unique features that reflect the transition of pedogenic processes related to changes in climate (Simón et al., 2000). In these soils, clay films in the argillic horizon (formed during the warmer, moisture preglacial climate) have been mixed by cryoturbation during glacial periods, resulting in abundant clay fragments in the matrix. It is important to be alert to the possibility that the present set of soil properties is a composite derived from a series of temporally varying processes. Interpretation of soil processes requires as full an Earth history context as possible. Smart sampling, keen observation, and an open mind are prerequisites to successful understanding.

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31

Soil Taxonomy

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31.1 Conditions Favoring the Development of *Soil Taxonomy*

By 1960, the United States had 60 years of experience with a soil survey program, which mapped and interpreted soils in various parts of the country. Ever since the earliest mapping of areas with specialty croplands and problem saline soils, the primary purpose of the soil survey program has been to predict the consequences of alternative uses of soils.

During that time, significant events had taken place that affected the production and delivery of consistent products and services of the soil survey program. Among these were World War I, new and improved industries, enhanced energy distribution, gas motors and automobiles, mechanization of agriculture, and a shift from mainly family farms to more commercial enterprises. The Dust Bowl devastated the lives of many farmers as land degradation and unfavorable climatic conditions collided. New federal and state parks and forest reserves were established. World War II spanned the globe, followed by the Korean War. Economic reconstruction was promoted, and global markets expanded. Agriculture was also changing from a mode of unbridled production to one of increased emphasis on conservation of soil and water resources in more responsible ways. Many of these changes had impacts on the prediction of the consequences of alternative soil uses.

The model of soil changed during those times. When the soil survey started, it was perceived that soils were derived from the rocks, or from the transported materials, on which they rested, a classic geological conclusion. This concept gradually changed to that proposed by the Russian, Dokuchaev, where soils were considered to be the result of processes that were influenced by the interactions of soil-forming factors, namely, climate, biota, parent material, topography, and time. Soils were recognized as independent natural bodies worthy of study by, and for themselves, and, thus, the course of history of soil science in the United States began to change in the 1920s. Marbut promoted the independence of soils, presented his ideas of soil classification, and helped America become recognized in international affairs of soil science.

Similar climates existed over fairly large areas, as did vegetation groups such as forests and grasses, although both had microvariations. On more local scales, there were differences in parent materials and landform topography that comprised landscapes. The overlapping of the soil-forming factors in space was crucial to applying the model of soil as the result of processes that were influenced by the interactions of the soil-forming factors. Soil mappers discovered empirical relations linking sets of soil properties (generally called soil profiles) to specific landscape features, which represented soil factors. The correlation of soils with landscape segments was found to be consistent enough to be delineated on base maps and to be satisfactory for the purpose of the soil survey.

Soils, therefore, had certain predictability and so did their expected behavior. Where conditions had been the same, the soils would be the same; and where the soils were the same, those responses that depend on soil properties would be the same.

Where similar but not identical conditions or soils occurred, the soil responses would be similar, but not identical to those where the soils were identical.

Soil, which was considered to be a continuum covering the earth's terrestrial surface, could be subdivided into classes in a variety of ways, thereby creating a population or collection of individual soil bodies. Emphasis changed from thinking about the whole with loosely defined and indistinct parts to the concept in which the parts were sharply in focus and the whole was an organized collection of parts.

Locally, the individual soils were called soil types. The members of the same series, that is, soils having a similar sequence of the same kinds of horizons, were separated according to the general texture of the profile. This was a way to recognize different groups of parent material such as glacial till and loess. Initially, the soil types were grouped together into a soil series that had the same horizonation and commonality of properties. Later, soil type referred only to the texture of the surface soil and was considered as a phase of a soil series. The differences in parent materials became a basis for establishing separate series.

Mapping of soil types proceeded rapidly as there were hundreds of survey parties mapping in all parts of the country. Many new soil series were set up; however, an adequate system of correlation and classification lagged behind. It became more and more difficult to keep track of all the information being collected by the soil scientists, to compare soils from one region to another, and to communicate about soil properties and characteristics.

A number of differences in the responses to management and in land uses were found to vary geographically. Some regional variations were related to climate and age, and locally many variations were related to drainage condition and parent material. These observations supported the concept of important geographic differences and similarities and led to the need for some way of recognizing these similarities and differences. The significance of factors and factor interactions varied from region to region, and if reasonable ways to group or separate them could be found, they might serve as a basis for classification.

Over time the question about soil had changed from "How much yield can be expected from this soil with this amount of input?" to "How much input must be used on this soil to produce a given volume of produce?" This recognized that soil was dynamic and capable of modification and manipulation rather than being a static responder to management. Humans and their activities were accepted as major factors of soil formation; in fact, it was recognized that humans were the dominant influence on temporal soil quality. The changing concepts and attitudes about soil and the evolving world economic development were important contributors to the decision to develop a new soil classification system.

31.2 Recognition of Guiding Principles for a Soil Classification System

The leaders of the National Cooperative Soil Survey agreed that a classification system must serve the soil survey program of the United States. The program was undertaken for very practical reasons, namely, to identify and locate soils, and to predict the consequences of alternative uses of soils.

They indicated that a system should provide a basis for developing principles of soil genesis and behavior that would enable them to continue to provide predictions of soil behavior and responses to management and manipulation. Because soil genesis attempts to find cause and effect relationships rather than merely identifying empirical relationships, it was felt that genetic principles should serve well as a means of spatial extrapolation and to assist in interpreting soils for use (Smith, 1963).

It was clear to them that classes of a system must have real counterparts in mappable soil bodies. Some small geographic body was the entity to be classified; thus, the criteria of classification should not be applied independently of the values and variability within natural landscapes. This need to link with identifiable geographic bodies separated the practical from the purely hypothetical organization of soil property information (Cline, 1949).

It was implied that the component soil bodies in nature that were to be classified must be relevant to applied objectives. Geographic size and shape and soil attributes included within those areas were considered important for making relevant interpretations. Experience had demonstrated that this would be a problem in a classification with mutually exclusive classes; however, some of the concerns could be handled with mapping conventions.

Although the system would serve the practical uses of the soil survey, it was not implied that the classes themselves had to serve directly as the technical interpretive units. The leaders realized that valid interpretations of behavior and response likely would involve an additional step, level, or reasoning, which would rely on the known or assumed relationships between soil properties and behavior. This was also true for valid interpretations about soil genesis. It was recognized that no set of classes, or basis for groupings, or basis of subdivision, could provide units homogeneous enough for direct application to multiple objectives. Therefore, the classes had to be capable of being regrouped or subdivided as needed to satisfy different applied purposes. This was most important for the classes of the lowest category, which was the soil series (Smith, 1963).

There was agreement that a classification system must also serve to understand nature by revealing our understanding of soils as natural bodies. The accurate mapping of soils depended on this understanding. This departed from classification schemes of plants and animals where lineage relationships were assumed. This meant that science was to be basic to the process. In as much as such a system would be able to satisfy these ideas, it would be a reflection of the knowledge at a given point in time.

It was desired to have a system that could be applied uniformly by competent soil scientists. This was an important idea because it put soil scientists on an equal footing and allowed mappers to concentrate on other facets of the soil survey program. The new system would have to be objective in the sense that it should be based on properties of soils and not on beliefs of the classifier. The taxa should be defined, or recognized by observable or measurable properties selected to group soils of similar genesis, but also capable of providing groups with similar properties, even if the genesis was uncertain or unknown. Soil genesis would be used to help select those properties that would provide useful groups of soils; however, the properties themselves would be the criteria for actual placement.

31.3 Science and Classification

Several lessons had been learned in trying to modify the 1938 classification. The concepts of some great soil groups were not clear and seemed to overlap with other groups. The methods of determining properties varied from place to place, or from time to time, and the correlations were not always certain. The mapping of soil types and soil series at the lower end of the spectrum, and the division of the pedosphere into regions and provinces and groups at the other end left an undecided middle ground with no good connections. There was great interest in not overly disturbing the thousands of soil series that had been established and used throughout the United States; yet there was a need to be able to link the series to classes in the higher categories of a classification system.

Mapping of soils and detailed laboratory studies of their properties gave rise to many consistent and repeatable correlations of soil features with landscape features. To maintain this consistency, the nature of the operations and methods used to obtain the data need to be specified and used. It is known that explanation is primarily a recognition of familiar correlations among phenomena in nature and that explanations are critical to understanding. It had been suggested that to go beyond empirical correlations to hypothesize the reasons for them will prejudice the future; therefore, there would be an effort to try to minimize using a priori principles, which determine or limit possibilities of new experiences. This meant that concepts of soil genesis would be useful to help select soil properties of interest, but that the operational character of soil facts would be used so as to not limit new experiences in the future.

There already were several thousand series considered to be classes of a lower category. They wanted some broader, more inclusive classes for making generalized maps, and an orderly organization of current understanding. It was also thought that they required flexibility to better serve other needs relying on soil information. Due to the magnitude and complexity of information, a hierarchical system of organization seemed most appropriate. For the most part, the logic of J.S. Mill as explained by Cline (1963) was followed. Some of the requirements follow.

Each category in a multicategoric system contains all members of the population of interest and is defined by abstracting the concepts that group the member classes together. Higher level categories are more abstract than those of the lower level.

The definition of a category suggests criteria that may be appropriate to separate the classes of that category. The criteria are then recognized by properties or features, which are called differentiating characteristics. These characteristics are observable or measurable properties of individuals that are grouped into a class.

Classes separated at a higher categorical level remain separated throughout the lower categories of the system. For ease of operation (placement of objects into the system), the classes are mutually exclusive, that is, without overlap. Operationally, the listing of the categories and the listing of the classes within each category need to be by priority because only a few of the characteristics of individuals are used to obtain placement within the scheme (Smith, 1983). This means that the scheme can be presented as a key or set of keys for consistent application of the definitions and the criteria (differentiating criteria).

The logic of the hierarchy represented by *Soil Taxonomy* is one of its major strengths, yet at the same time it retains the weaknesses of such organizational frameworks. The efficiency with which accessory characteristics are carried along with the differentiating ones is very powerful indeed. On the other hand, the uncertainty of the degree of accessory property relationships and the rigid class structure of the system limit the precision of statements that can be made for soils as they are observed in the field.

Recognition of a soil individual is crucial to the proper functioning of the system. Because the system was to support the soil survey, the individual had to be a recognizable geographic entity and one that could be characterized with limited sampling. The sample unit became the pedon and the soil individual became the polypedon. Their application has not been without some difficulty, which has continued even to the present.

The nomenclature of a hierarchical system is meaningful to the extent that there is consistent composition of the classes. A mnemonic scheme of phrases and combinations seemed a worthy goal. It was thought desirable to have a system of naming that would minimize previous biases and confusion and which might be translatable or usable in other languages. Consequently, roots of words from Latin and Greek and their combinations were proposed and tested. For some soil scientists the "new language" of soil classification was difficult to accept, but in retrospect was a brilliant idea.

And finally, it was agreed that the definitions must be continually tested by the nature of functioning of the soils grouped in a taxon. A taxonomy for the use of the soil survey needed to be tested by the nature of the interpretations that could be made (Smith, 1983). These interpretations included those for genesis and the mapping of soils, as well as for the behavior of soils under multiple uses.

31.4 Definitions of Categories of *Soil Taxonomy*

A category is the aggregate of classes that is formed by differentiation within a population on a single basis. A category includes the entire population; it includes all classes differentiated on one basis; it is distinguished from class that is only one part of a category; and it is definable only in terms of the basis of differentiation. In a hierarchy, the higher categories have fewer classes and they are more inclusive than the classes of the lower categories, which have accumulated attributes from all of the higher categories. Thus, high categorical levels are associated with high-level generalizations or abstractions. These abstractions are used as the bases of differentiation; however, these ideas are expressed in terms of attributes that are assumed to be their consequences. Thus, soil attributes that are thought to be the results of such processes are the criteria used to segregate soils.

The definitions of categories, orders through series, although not stated very clearly, are intended to guide the selection and testing of properties and features used to characterize and classify soils. Soil properties that are thought to reflect processes or control processes are of great importance in this system. Although the processes of formation are influential in the selection of properties, it is the properties themselves that are used to determine the placement of soils into respective classes. The applications of quantified definitions and observations are the tests of adequacy of the theories behind this model of soil. Hence, *Soil Taxonom*y is considered a morphogenetic classification system.

31.4.1 Order

Classes at the order level are separated on the basis of properties resulting from the major processes and pathways of soil formation. Neither the genetic processes nor the courses of development are precisely known but the accepted concepts have influenced the selection of soil properties that are used to recognize and define the 12 classes currently considered. Many of the features are thought to have taken a reasonably long time to develop, are stable in a pedological sense, and are mostly static from a historical perspective.

31.4.2 Suborder

Classes at the suborder level are separated within each order on the basis of soil properties that are major controls, or reflect such controls on the current set of soil-forming processes. Most of the properties selected are dynamic such as soil moisture regime or cold soil temperatures. Other properties relate to materials or processes that retard horizon development, such as sand or alluvial sedimentation.

31.4.3 Great Group

Classes at the great group level are differentiated within each suborder on the basis of properties that constitute subordinate or additional controls, or reflect such controls on the current set of soil-forming processes. The properties selected are generally static, such as layers that retard percolation of water or root extension, but some are dynamic, such as the moisture regime where it was not a criterion at the suborder level.

31.4.4 Subgroup

Classes at the subgroup level are differentiated within each great group on the basis of properties resulting from either (1) a blending or overlapping of sets of processes in space or time that cause one kind of soil to develop from, or toward another kind of soil that has been recognized at the great group, suborder, or order level; or (2) sets of processes or conditions that have not been recognized as criteria for any class at a higher level. A third kind of subgroup fits neither (1) nor (2) but is considered to typify the central concept of the great group or simply represents what is not represented by the other subgroups.

31.4.5 Family

Classes at the family level are separated with the subgroup on the basis of properties that reflect important conditions affecting behavior or the potential for further change. Particle size, mineralogy, and soil depth are mainly capacity factors, whereas soil temperature and exchange activity are mainly intensity factors.

31.4.6 Series

Classes at the series level are separated within the family on the basis of properties that reflect relatively narrow ranges of soil-forming factors and of processes that transform parent materials into soils. Some properties are indicative of parent materials such as coarse fragments, sand or silt content, color, and horizon thickness or expression. Others reflect influences on processes such as differences in intensity or amount of precipitation, and depth to the presence or concentration of soluble compounds.

31.5 Differentiating Characteristics

The differentiating characteristics mentioned above are referred to in *Soil Taxonomy* as diagnostic horizons and characteristics or features. They are the building blocks of *Soil Taxonomy*. The diagnostic horizons and characteristics help define the criteria for the various taxa of *Soil Taxonomy*. They are also terms used by scientists to communicate the language of soil. Many soil scientists consider the concept of diagnostic horizons and features as one of the greatest contributions to soil classification.

31.5.1 Diagnostic Horizons

Diagnostic horizons are defined in the *Keys to Soil Taxonomy* (Soil Survey Staff, 2010). The intent in this chapter is not to reiterate the complete definition of the diagnostic horizons, but to provide a brief summary.

31.5.2 Diagnostic Epipedons

Diagnostic horizons that form at or near the surface are referred to as epipedons. There are eight epipedons recognized in *Soil Taxonomy*, which are briefly described below.

31.5.2.1 Anthropic Epipedon (Gr. *anthropos*, Human Being)

This is formed during long continuous use by humans, either as a place of residence (kitchen middens) or as a site for irrigated crops.

Anthropic epipedons often have the dark color of an umbric or a mollic epipedon, but were formed largely due to the actions of humans. Anthropic epipedons occur in a variety of soil orders, but are not widely recognized. Additional data and research are needed on this epipedon.

31.5.2.2 Folistic Epipedon (L. *folia*, Leaf)

This epipedon contains high amounts of organic C, but is not saturated with water for more than a few days after heavy rains. These epipedons occur mostly in cool, humid regions under forest vegetation. Folistic epipedons occur most commonly in Spodosols (see color insert) and Inceptisols (see color insert), but can occur in other orders.

31.5.2.3 Histic Epipedon (Gr. *histos*, Tissue)

This is 20–60 cm thick and contains high amounts of organic C. It differs from the folistic epipedon in that it is commonly saturated with water for long periods. Histic epipedons can occur in many soil orders.

31.5.2.4 Melanic Epipedon (Gr. *melas*, melan-, Black)

This is a thick, black horizon that contains high concentrations of organic C and short-range order minerals or Al–humus complexes. It commonly occurs in areas associated with a volcanic influence under forest vegetation. Melanic epipedons are associated with Andisols (see color insert for Andisol and Andisol (2)).

31.5.2.5 Mollic Epipedon (L. *mollis*, Soft)

This is a relatively thick, dark-colored, humus-rich surface horizon in which divalent cations are dominant on the exchange complex. It forms from the underground decomposition of organic residues, chiefly grasses, although particularly in ustic soil moisture regimes occur under a variety of vegetation, including shrubs and trees. Mollic epipedons have good structure and porosity and are associated with some of the richest agricultural soils in the world. Mollic epipedons are most commonly associated with Mollisols (see color insert for Mollisol and Mollisol (2)); they very rarely occur in other soil orders.

31.5.2.6 Ochric Epipedon (Gr. *ochros*, Pale)

This fails to meet the definition of any of the other diagnostic epipedons. It can be light colored, thin and dark colored, or even thin with high amounts of organic C. It has few or no accessory characteristics, but was defined to serve as a means of placing a name on the surface horizons that failed the criteria of any of the other seven epipedons. Ochric epipedons can occur in a variety of soil orders.

31.5.2.7 Plaggen Epipedon (Ger. *Plaggen*, Sod)

Plaggen epipedons are rare and occur mostly in Europe where in medieval times sod or other materials were used for bedding livestock, and the manure was spread on cultivated fields. The gradual additions of bedding materials raised the level of the fields. The plaggen epipedon is associated with Inceptisols.

31.5.2.8 Umbric Epipedon (L. *umbra*, Shade, Hence Dark)

Umbric epipedons resemble mollic epipedons but form largely under a forest vegetation and have a lower natural fertility (base saturation) than the mollic epipedon. The umbric epipedon is most commonly associated with Inceptisols, Andisols, Alfisols (see color insert), Ultisols (see color insert for Ultisol and Ultisol (2)), and Gelisols (see color insert).

31.5.3 Diagnostic Subsurface Horizons

The subsurface horizons form below the surface, but can be at or near the surface in eroded soils. Brief descriptions follow.

31.5.3.1 Agric Horizon (L. *ager*, Field)

This is an illuvial horizon, which has formed under cultivation and contains significant amounts of illuvial silt, clay, and humus. This horizon is relatively rare, or perhaps not readily recognized, and is often more associated with old world agriculture. Agric horizons are associated with Alfisols.

31.5.3.2 Albic Horizon (L. *albus*, White)

This is an eluvial horizon 1.0 cm or more thick with a color that is largely determined by the color of the primary sand and silt grains, rather than the color of their coatings. The clay and Fe oxides have been removed by pedogenesis. Albic horizons are most commonly associated with Spodosols, Alfisols, Ultisols, and Mollisols.

31.5.3.3 Argillic Horizon (L. *argilla*, White Clay)

An argillic horizon is commonly a subsurface horizon with a significantly higher content of phyllosilicate clay than the overlying horizon. There is also evidence of clay illuviation. Argillic horizons occur on stable landscapes and are most commonly associated with Alfisols, Ultisols, Aridisols (see color insert), and Mollisols.

31.5.3.4 Calcic Horizon (L. *calcis*, Lime)

This is an illuvial horizon in which secondary $CaCO_3$ has accumulated to a significant extent. In arid environments, precipitation is insufficient to move carbonates to great depths. In soils with water near the surface, capillary rise, evaporation, and transpiration concentrate $CaCO_3$ toward the surface. Calcic horizons occur in Aridisols, Alfisols, Andisols, Mollisols, Inceptisols, Vertisols (see color insert for Vertisol and Vertisol (2)), and Gelisols.

31.5.3.5 Cambic Horizon (L. *cambiare*, to Exchange)

The cambic horizon forms as a result of physical alterations, chemical transformations, accumulations, removals, or a combination of these processes. They are commonly identified by structure and a higher clay content, redder hue, or higher chroma than an overlying horizon. Cambic horizons are most commonly associated with Inceptisols, but they can occur in Aridisols, Mollisols, Vertisols, Andisols, and Gelisols.

31.5.3.6 Duripan (L. *durus*, Hard)

A duripan is cemented by illuvial silica. Duripans limit the downward growth of roots and movement of water. They are most commonly associated with the Aridisols, but are known to occur in Alfisols, Mollisols, Andisols, Inceptisols, Spodosols, and Vertisols.

31.5.3.7 Fragipan (L. *fragilis*, Brittle, and Pan)

This is a subsurface horizon that is noncemented, but restricts the entry of water and roots into the soil matrix. Fragipans most commonly are associated with Ultisols, Alfisols, Spodosols, and Inceptisols.

31.5.3.8 Glossic Horizon (Gr. *glossa*, Tongue)

The glossic horizon develops as a result of the degradation of an argillic, kandic, or natric horizon. Clay and Fe oxides are removed starting from the exterior of the peds. These horizons are transitional from argillic, natric, and kandic horizons to albic horizons. Glossic horizons occur most commonly in the Alfisols and Ultisols.

31.5.3.9 Gypsic Horizon (L. *gypsum*)

This is an illuvial horizon in which secondary gypsum has accumulated to a significant extent. Most gypsic horizons occur in arid environments where the parent materials are rich in gypsum. In soils with groundwater close to the surface, gypsum can accumulate by capillary rise, evaporation, and transpiration. Gypsic horizons most commonly occur in the Aridisols, but a few Inceptisols, Vertisols, and Gelisols also have gypsic horizons.

31.5.3.10 Kandic Horizon (Modified from Kandite)

This is a subsurface horizon with a significantly higher content of clay than an overlying horizon. Kandic horizons are dominated by low activity clays (1:1) and, therefore, have low cation exchange capacity (CEC). Kandic horizons occur in the Ultisols and Oxisols (see color insert for oxisol and oxisol (2)).

31.5.3.11 Natric Horizon (L. *natrium*, Sodium)

The natric horizon has all the characteristics of the argillic horizon (significant clay increase) and, in addition, has an accumulation of Na. Sodium adversely affects the physical properties of a soil. Of the soil orders, natric horizons are most commonly associated with the Aridisols, but they also occur in the Alfisols and Mollisols and to a very limited extent in the dry Gelisols.

31.5.3.12 Ortstein

Ortstein is a cemented horizon that consists of complexes of Al and organic matter with or without Fe (spodic materials). Ortstein limits root growth and the downward movement of water. Ortstein occurs in Spodosols.

31.5.3.13 Oxic Horizon (F. *oxide*)

This is a subsurface horizon of sandy loam or a finer texture with a low CEC and low amount of weatherable minerals. Oxic horizons occur in old, weathered soils or soils derived from highly weathered parent materials. Oxic horizons occur in Oxisols.

31.5.3.14 Petrocalcic Horizon (Gr. *petra*, Rock and Calcic)

This is cemented or indurated by $CaCO_3$ or Ca and Mg carbonates. It is a barrier to roots and the downward movement of water. Petrocalcic horizons are most commonly associated with the Aridisols, but also occur in the Alfisols, Mollisols, Vertisols, and Inceptisols.

31.5.3.15 Petrogypsic Horizon (Gr. *petra*, Rock and Gypsic)

This is cemented or indurated by gypsum. It is a barrier to roots and downward movement of water. Petrogypsic horizons are most commonly associated with Aridisols.

31.5.3.16 Placic Horizon

This is a thin (<25 mm), dark-colored horizon that is cemented by either Mn or Fe and Mn and organic matter. Commonly, they occur in moist, cool climates and are associated with the Spodosols and Inceptisols.

31.5.3.17 Salic Horizon (L. *sal*, Salt)

This is an accumulation of salts that are more soluble than gypsum in water. Many plants are intolerant of high concentrations of salt. Most salic horizons are associated with the Aridisols; a few occur in the drier Gelisols.

31.5.3.18 Sombric Horizon (F. *sombre*, Dark)

This is an illuvial horizon that contains humus that is not associated with Al nor dispersed by Na. It is largely confined to the cool, moist soils of high plateaus or mountains of the tropics and subtropics. Sombric horizons are associated with Inceptisols, Oxisols, and Ultisols.

31.5.3.19 Spodic Horizon (Gr. *spodos*, Wood Ash)

This is an illuvial horizon that contains active amorphous materials composed of organic matter and Al with or without Fe. Spodic horizons are associated with Spodosols.

31.5.3.20 Sulfuric Horizon (L. *sulfur*, Sulfur)

The sulfuric horizon forms when sulfide rich and organic materials are oxidized as a result of drainage, most commonly artificial. Sulfuric horizons are toxic to most plants. They occur in Entisols (see color insert), Histosols (see color insert), and Inceptisols.

31.5.4 Other Diagnostic Soil Characteristics

The diagnostic soil characteristics are features of the soil that are used repeatedly in *Soil Taxonomy*.

Abrupt textural change is a considerable increase in clay within a short distance, a very important feature used to help predict water movement in soils.

Albic materials are light-colored soil materials that reflect the color of the primary sand and silt particles rather than the color of the coatings.

Andic soil properties result from the presence of significant amounts of allophane, imogolite, ferrihydrite, or Al–humus complexes.

Anhydrous conditions refer to the moisture condition of soils in very cold deserts and other areas that have dry permafrost.

Aquic conditions arise in soils that currently experience continuous or periodic saturation and reduction. The presence of these features is indicated by redoximorphic features. There are several types of saturation: Episaturation describes a perched water table, endosaturation describes a groundwater table, and anthraquic saturation describes controlled flooding such as that used to grow rice and cranberries.

Coefficient of linear extensibility (COLE) is the ratio of the difference between the moist and dry lengths of a clod to its dry length. COLE is used to determine the shrink/swell potential of a horizon.

Cryoturbation (frost churning) is the mixing of the soil matrix that results in irregular or broken horizons.

Densic contact occurs between soil and densic materials and is a barrier to roots.

Densic materials are noncemented, relatively unweathered, mostly earthy materials such as till, volcanic mudflows, and noncemented rocks.

Durinodes are cemented to indurated nodules with SiO_2, presumably opal and microcrystalline forms of silica.

Fragic soil properties are similar to a fragipan, but do not have the required thickness or volume for a fragipan.

Gelic materials are soil materials that show evidence of cryoturbation or ice segregation in the active layer.

Glacic layer is massive ice in the form of ice lenses and wedges.

Identifiable secondary carbonates refer to translocated authigenic $CaCO_3$ that has been precipitated in place from the soil solution rather than precipitated in place from the parent material.

Interfingering of albic materials refers to albic materials that penetrate 5 cm or more into an underlying horizon.

Lamellae are thin (<7.5 cm) illuvial horizons that contain evidence of illuvial clay and occur only in coarse-textured soils.

Linear extensibility is the product of the thickness (cm) multiplied by the COLE of the soil layer in question and is used to predict shrink and swell.

Lithic contact occurs between soil and a coherent underlying material that is in a strongly or more cemented class. Lithic contacts occur at the boundary between soil and hard bedrock.

Lithologic discontinuities are significant differences in particle-size distribution or mineralogy that represent differences in lithology of a soil.

N value is used to predict if a soil can support loads and to the degree of subsidence after drainage.

Paralithic contact occurs between soil and paralithic materials (defined below) such as between soil and unconsolidated or weathered bedrock.

Paralithic materials are relatively unaltered (by pedogenesis) materials that are in a moderately or less cemented class.

Permafrost refers to soil material that remains below 0°C for 2 or more years in succession.

Petroferric contact is a boundary between soil and a continuous layer of indurated material in which Fe is an important cement and organic matter is either absent or present in trace amounts.

Plinthite is an Fe-rich, humus-poor material that irreversibly hardens when exposed to wetting and drying, especially if heated by the sun.

Resistant minerals refer to durable minerals in the 0.02–2.0 mm fraction, such as quartz, zircon, etc.

Slickensides are polished and grooved surfaces produced by one mass sliding past another. They are a feature of soils with a high capacity to shrink and swell.

Spodic materials are illuvial, amorphous materials composed of organic material and aluminum with or without iron.

Sulfidic materials contain oxidizable sulfur compounds.

31.5.5 Soil Moisture and Temperature Regimes

Soil Taxonomy is unique among many systems used to classify soils because it recognizes the temperature of a soil and the moisture status of a soil over time. Both temperature and moisture are important properties of the soil and convey essential information about soils. Because complete definitions for the soil moisture and temperature regimes can be found in the *Keys to Soil Taxonomy* (Soil Survey Staff, 2010), only brief descriptions are presented here.

31.5.5.1 Soil Moisture Regimes

For most soil orders, moisture regime is used to determine placement of a soil at the suborder level.

Aquic moisture regime signifies a reducing regime virtually free of dissolved O_2 from saturation with water. Most soils that have an aquic moisture regime are not saturated with water to the soil surface year round.

Aridic (torric) moisture regime applies to soils that commonly occur in arid climates and are dry in all parts more than half the time during the growing season and moist in some or all parts for less than 90 consecutive days during the growing season.

Udic moisture regime applies to soils that occur in climates with well-distributed rainfall or sufficient summer rain so that rainfall plus stored moisture equals or exceeds the amount of evaporation.

Ustic moisture regime is moister than aridic and drier than udic. The concept of ustic is one of limited moisture, but at least some moisture at a time when conditions are suitable for plant growth.

Xeric moisture regime applies to soils that occur in climates with cool, moist winters and warm, dry summers.

Perudic moisture regime applies to soils that occur in areas where precipitation exceeds evaporation in all months when the soil is not frozen.

TABLE 31.1 Example of Naming Systems for Soils Used in *Soil Taxonomy*

Order	Suborder	Great Group	Subgroup	Family	Series
Aridisols	Argids	Calciargids	Ustic Calciargids	Fine-loamy, mixed, superactive, mesic	Barx "Clovis, Flaco".

31.5.5.2 Soil Temperature Regimes

The following soil temperature regimes are based on the mean annual temperature of a soil measured at 50 cm or at a lithic, paralithic, or densic contact, whichever is shallowest.

Cryic soil temperature regime applies to soils that have a mean annual soil temperature of <8°C, but no permafrost and a summer temperature cooler than soils in a frigid regime.

Frigid soil temperature regime applies to soils that have a mean annual soil temperature of <8°C.

Mesic temperature regime applies to soils that have a mean annual soil temperature of 8°C but <15°C.

Thermic temperature regime applies to soils that have a mean annual soil temperature of 15°C but <22°C.

Hyperthermic temperature regime applies to soils that have a mean annual soil temperature of 22°C.

When the difference between mean summer and winter soil temperatures is less than 6°C, *iso* is added to the name. *isofrigid*, *isomesic*, *isothermic*, and *isohyperthermic* are the only classes used. Soil temperature is most commonly used at the family level in *Soil Taxonomy*.

31.6 Categories of *Soil Taxonomy*

The information that follows is largely from the second edition of *Soil Taxonomy* (Soil Survey Staff, 1999). The diagnostic horizons and characteristics help define the categories of *Soil Taxonomy*.

A category of *Soil Taxonomy* is a set of classes that is defined approximately at the same level of generalization or abstraction and includes all soils. There are six categories or levels in *Soil Taxonomy*. In order of decreasing rank and increasing number of differentiae and classes, the categories are order, suborder, great group, subgroup, family, and series.

The nomenclature of *Soil Taxonomy* is based on the following premises: Each taxon requires a name if it is to be used in speech; a good name is short, easy to pronounce, and distinctive in meaning; a name is connotative, that is, capable of mnemonic attachment to the concept of the thing itself (Heller, 1963); it is useful if the name of a taxon indicates its position in the classification, if similarities in important properties are reflected by similarities in names, if the mnemonic attachments hold in many languages, and if the name fits into many languages without translation.

The name of each taxon above the category of series indicates its class in all categories of which it is a member. The name of a soil series indicates only the category of series. Thus, a series name may be recognized as a series, but one cannot tell from the name to what order, suborder, and so on, it belongs.

Table 31.1 gives a few examples of names of taxa in each category from order to series. Because of the system for assigning names and because most formative elements carry the same meaning in any combination, a name can convey a great deal of information about a soil.

31.7 Recognition of the Categories

31.7.1 Orders

The name of each order ends in sol (L. *solum*, soil) with the connecting vowel *o* for Greek roots and *i* for other roots. Each name of an order contains a formative element that begins with the vowel immediately preceding the connecting vowel and ends with the last consonant preceding the connecting vowel. In the order name, Entisol, the formative element is *ent*. In Aridisol, it is *id*. These formative elements are used as endings for the names of suborders, great groups, and subgroups. Thus, the names of all taxa higher than the series that are members of the Entisol order end in *ent* and can be recognized as belonging to the order of Entisol. Names ending in *id* are names of taxa belonging to the order of Aridisols. Table 31.2 lists all the soil orders and their formative elements.

31.7.2 Suborders

Names of suborders have exactly two syllables. The first syllable connotes something of the diagnostic properties of the soils or the soil moisture regime. The second is the formative element from the name of the order. Thirty formative elements

TABLE 31.2 Formative Elements in Names of Soil Orders in *Soil Taxonomy*

Formative Element in Name	Name of Order	Derivation of Formative Element	Pronunciation of Formative Element
Alf	Alfisol	Meaningless syllable	Pedalfer
And	Andisol	Modified from ando	Ando
Id	Aridisol	L. *aridus*, dry	Arid
Ent	Entisol	Meaningless syllable	Recent
El	Gelisol	L. *gelare*, to freeze	Jell
Ist	Histosol	Gr. *histos*, tissue	Histology
Ept	Inceptisol	L. *inceptum*, beginning	Inception
Oll	Mollisol	L. *mollis*, soft	Mollify
Ox	Oxisol	F. *oxide*, oxide	Oxide
Od	Spodosol	Gr. *spodos*, wood ash	Odd
Ult	Ultisol	L. *ultimus*, last	Ultimate
Ert	Vertisol	L. *verto*, turn	Invert

TABLE 31.3 Formative Elements in Names of Suborders in *Soil Taxonomy*

Formative Element	Derivation	Connotation
Alb	L. *albus*, white	Presence of albic horizon
Anth	Gr. *anthropos*, human	Modified by humans
Aqu	L. *aqua*, water	Aquic conditions
Ar	L. *arare*, to plow	Mixed horizon
Arg	L. *argilla*, white clay	Presence of argillic horizon
Cal	L. *calcis*, lime	Presence of calcic horizon
Camb	L. *cambiare*, to exchange	Presence of cambic horizon
Cry	Gr. *kryos*, icy cold	Cold
Dur	L. *durus*, hard	Presence of duripan
Fibr	L. *fibra*, fiber	Least decomposed stage
Fluv	L. *fluvius*, river	Floodplain
Fol	L. *folia*, leaf	Mass of leaves
Gel	L. *gelare*, to freeze	Jell
Gyps	L. *gypsum*, gypsum	Presence of gypsic horizon
Hapl	Gr. *haplos*, simple	Minimum horizon development
Hem	Gr. *hemi*, half	Intermediate stage of decomposition
Hist	Gr. *histos*, tissue	Presence of organic materials
Hum	L. *humus*, earth	Presence of organic matter
Orth	Gr. *orthos*, true	The common ones
Per	L. *per*, throughout time	Perudic moisture regime
Psamm	Gr. *psammos*, sand	Sandy texture
Rend	Modified from rendzina	High carbonate content
Sal	L. *sal*, salt	Presence of salic horizon
Sapr	Gr. *saprose*, rotten	Most decomposed stage
Torr	L. *torridus*, hot and dry	Torric moisture regime
Turb	L. *turbidis*, disturbed	Presence of cryoturbation
Ud	L. *udus*, humid	Udic moisture regime
Ust	L. *ustus*, burnt	Ustic moisture regime
Vitr	L. *vitrum*, glass	Presence of glass
Xer	Gr. *xeros*, dry	Xeric moisture regime

(Table 31.3) are used with the 12 formative elements from names of the orders to make names of over 65 suborders. The suborder of Entisols that has aquic conditions throughout is called Aquents (L. *aqua*, water, plus *ent* from Entisol). The formative element *aqu* is used with this meaning in 9 of the 12 orders. The suborder of Entisols that consists of very young sediments is called Fluvents (L. *fluvius*, river, plus *ent* from Entisol).

31.7.3 Great Groups

The name of a great group consists of the name of a suborder and a prefix that consists of one or two formative elements suggesting something of the diagnostic properties. Names of great groups, therefore, have three or four syllables and end with the name of a suborder. Fluvents that have a cryic temperature regime are called Cryofluvents (Gr. *kryos*, icy cold, plus fluvent). Fluvents that have a torric moisture regime are called Torrifluvents (L. *torridus*, hot and dry). The formative elements for the great groups are listed in Table 31.4.

31.7.4 Subgroups

The name of a subgroup consists of the name of a great group modified by one or more adjectives. The adjective, typic, represents in some instances what is thought to typify the great group and in other instances, typic subgroups simply do not have any of the characteristics used to define the other subgroups in a great group. Each typic subgroup has, in clearly expressed form, all the diagnostic properties of the order, suborder, and great group to which it belongs. Typic subgroups also have no additional properties indicating a transition to another great group. Typic subgroups are not necessarily the most extensive subgroup of a great group.

Intergrade subgroups are those that belong to one great group but that have some properties of another order, suborder, or great group. They are named by using the adjectival form of the name of the appropriate taxon as a modifier of the great group name. Thus, the Torrifluvents that have some of the properties or properties closely associated with Vertisols are called Vertic Torrifluvents. Vertic Torrifluvents have some of the properties of Vertisols superimposed on the complete set of diagnostic properties of Torrifluvents.

Extragrade subgroups are those that have important properties that are not representative of the great group but that do not indicate transitions to any other known kind of soil. They are named by modifying the great group name with an adjective that connotes something of the nature of the aberrant properties. Thus, a Cryorthent that has bedrock that is at least strongly cemented within 50 cm of the mineral soil surface is called a Lithic Cryorthent (lithic, Gr. *lithos*, stone). This subgroup is listed as an example in Table 31.5.

31.7.5 Families

Names of families are polynomial. Each consists of the name of a subgroup and descriptive terms, generally three or more, to indicate the particle-size class (or combinations thereof, if strongly contrasting), the mineralogy (26 classes), the cation exchange activity (4 classes), the calcareous and reaction (4 classes), the temperature (8 classes), and, in a few families, depth of soil (3 classes), rupture resistance (2 classes), and classes of coatings and classes of cracks (3 classes). Names of most families have three to five descriptive terms that modify the subgroup name, but a few have only one or two and a few as many as six. An example given in Table 31.1 is a family of fine-loamy (particle size), mixed (mineralogy), superactive (cation exchange activity), and mesic (soil temperature) Ustic Calciargids.

31.7.6 Series

Names of series as a rule are abstract place names. The name usually is taken from a place near where the series was first recognized. It may be the name of a town, a county, or some local feature. Some series have coined names while many have been carried over from earlier classifications with some having been

TABLE 31.4 Formative Elements in Names of Great Groups in *Soil Taxonomy*

Formative Element	Derivation	Connotation
Acr	Gr. *akros*, at the end	Extreme weathering
Al	Modified from aluminum	High aluminum, low iron
Alb	L. *albus*, white	Presence of an albic horizon
Anhy	Gr. *anhydros*, waterless	Very dry
Anthr	Gr. *anthropos*, human	An anthropic epipedon
Aqu	L. *aqua*, water	Aquic conditions
Arg	L. *argilla*, white clay	Presence of an argillic horizon
Calc	L. *calcis*, lime	A calcic horizon
Cry	Gr. *kryos*, icy cold	Cold
Dur	L. *durus*, hard	A duripan
Dystr (Dys)	Gr. *dys*, ill; dystrophic, infertile	Low base saturation
Endo	Gr. *endo*, within	Implying a groundwater table
Epi	Gr. *epi*, on, above	Implying a perched water table
Eutr (Eu)	Gr. *eu*, good; eutrophic, fertile	High base saturation
Ferr	L. *ferrum*, iron	Presence of iron
Fibr	L. *fibra*, fiber	Least decomposed stage
Fluv	L. *fluvius*, river	Floodplain
Fol	L. *folia*, leaf	Mass of leaves
Frag	L. *fragilis*, brittle	Presence of fragipan
Fragloss	Compound of fra(g) and gloss	See Frag and Gloss
Fulv	H. *fulvus*, dull	Dark brown color, presence of organic C
Gel	L. *gelare*, to freeze	Jell
Glac	L. *glacialis*, icy	Ice lenses or wedges
Gyps	L. *gypsum*, gypsum	Presence of gypsic horizon
Gloss	Gr. *glossa*, tongue	Presence of glossic horizon
Hal	Gr. *hals*, salt	Salty
Hapl	Gr. *haplos*, simple	Minimum horizon
Hem	Gr. *hem*, half	Intermediate stage of decomposition
Hist	Gr. *histos*, tissue	Presence of organic materials
Hum	L. *humus*, earth	Presence of organic matter
Hydr	Gr. *hydor*, water	Presence of water
Kand	Modified from kandite	1:1 layer silicates
Luv	Gr. *louo*, to bath	Illuvial
Melan	Gr. *melasanos*, black	Black, presence of organic C
Molli	L. *mollis*, soft	Presence of a mollic epipedon
Natr	L. *natrium*, sodium	Presence of a natric horizon
Pale	Gr. *paleos*, old	Excessive development
Petro	Gr. *petra*, rock	A cemented horizon
Plac	Gr. *plax*, flat stone	Presence of a thin pan
Plagg	Ger. *plaggen*, sod	Presence of a plaggen epipedon
Plinth	Gr. *plinthos*, brick	Presence of plinthite
Psamm	Gr. *psammos*, sand	Sand texture
Quartz	Ger. *quart*, quartz	High quartz content
Rhod	Gr. *rhodon*, rose	Dark red color
Sal	L. *sal*, salt	Presence of a salic horizon
Sapr	Gr. *saprose*, rotten	Most decomposed stage
Somb	F. *sombre*, dark	Presence of sombric horizon
Sphagn	Gr. *sphagnos*, bog	Presence of sphagnum
Sulf	L. *sulfur*, sulfur	Presence of sulfides or their oxidation products
Torr	L. *torridus*, hot	Tonic moisture regime and dry
Ud	L. *udus*, humid	Udic moisture regime
Umbr	L. *umbra*, shade	Presence of an umbric horizon
Ust	L. *ustus*, burnt	Ustic moisture regime
Verm	L. *vermes*, worm	Wormy or mixed by animals
Vitr	L. *vitrum*, glass	Presence of volcanic glass
Xer	Gr. *xeros*, dry	Xeric moisture regime

TABLE 31.5 Adjectives in Names of Extragrades in *Soil Taxonomy* and Their Meaning

Adjective	Derivation	Connotation
Abruptic	L. *abruptum*, torn off	Abrupt textural change
Aeric[a]	Gr. *aerios*, air	Aeration
Albic	L. *albus*, white	Presence of an albic horizon
Alic	Modified from aluminum	High Al^{3+} status
Anionic	Gr. *anion*	Positively charged colloid
Anthraquic	Gr. *anthropos*, human and L. *aqua*, water	Controlled flooding
Anthropic	Gr. *anthropos*, human	An anthropic epipedon
Arenic	L. *arena*, sand	Sandy between 50 and 100 cm thick
Calcic	L. *calcis*, lime	Presence of a calcic horizon
Chromic	Gr. *chroma*, color	High chroma
Cumulic	L. *cumulus*, heap	Thickened epipedon
Durinodic	L. *durus*, hard	Presence of durinodes
Dystic	Gr. *dys*, ill	Low base status
Eutric	Gr. *eu*, good; eutrophic, fertile	High base status
Fragic	L. *fragilis*, brittle	Presence of fragic properties
Glacic	Gr. *glacialis*, icy	Presence of ice lenses or wedges
Glossic	Gr. *glossa*, tongue	Tongued horizon boundaries
Grossarenic	L. *grossus*, thick; *arena*, sand	Thick, sandy layer
Gypsic	L. *gypsum*, gypsum	Presence of a gypsic horizon
Halic	Gr. *hals*, salt	Salty
Humic	L. *humus*, earthy	Presence of organic matter
Hydric	Gr. *hydor*, water	Presence of water
Kandic	Modified from kandite	Presence of 1:1 layer silicates
Lamellic	L. *lamella*, dim	Presence of lamellae
Leptic	Gr. *leptos*, thin	A thin soil
Limnic	Gr. *limn*, lake	Presence of a limnic layer
Lithic	Gr. *lithos*, stone	Presence of a shallow lithic contact
Natric	L. *natrium*, sodium	Presence of sodium
Nitric	Modified from *nitron*	Presence of nitrate salts
Ombroaquic	Gr. *ombros*, rain; L. *aqua*, water	Surface wetness
Oxyaquic	Oxy representing oxygen and aquic	Aerated
Pachic	Gr. *pachys*, thick	A thick epipedon
Petrocalcic	Gr. *petra*, rock; L. *calcis*, lime	Presence of a petrocalcic horizon
Petroferric	Gr. *petra*, rock; L. *ferrum*,, iron	Presence of a petroferric contact (ironstone)
Petrogypsic	Gr. *petra*, rock; L. *gypsum*, gypsum	Presence of a petrogypsic horizon
Petronodic	Gr. *petra*, rock; L. *nodulus*, a little knot	Presence of concretions and/or nodules
Placic	Gr. *plax*, flat stone	Presence of a thin pan (placic horizon)
Plinthic	Gr. *plinthos*, brick	Presence of plinthite
Rhodic	Gr. *rhodon*, rose	Dark red color
Ruptic[a]	L. *ruptum*, broken	Intermittent or broken horizons
Sodic	Modified from sodium	Presence of sodium
Sombric	F. *sombric*, dark	Presence of sombric horizon
Sulfic	L. *sulfur*, sulfur	Presence of sulfides or their oxidation products
Terric	L. *terra*, earth	A mineral substratum
Thapto(ic)[a]	Gr. *thapto*, buried	A buried soil
Ultic	L. *ultimus*, last	Low base status
Umbric	L. *umbra*, shade	Presence of umbric epipedon
Xanthic	Gr. *xanthos*, yellow	Yellow color

[a] Not strictly an extragrade. Name used to indicate a special departure from the typic subgroup.

in use since 1900. The name of a series carries no meaning to people who have no other source of information about the soils in it. There are over 21,000 soil series currently recognized in the United States.

The Barx, Clovis, and Flaco series (Table 31.1) are three members of the fine-loamy, mixed, superactive, mesic family of Ustic Calciargids. The meaning of each of these terms is defined later, but in a general way the name tells us the following.

Fine-loamy means that from a depth of 25–100 cm there is no marked contrast in particle-size class, the content of clay is between 18% and 35%, 15% or more of the material is coarser than 0.1 mm in diameter (fine sand to very coarse sand plus gravel), but less than 35% by volume of the material is rock fragments 2.0 mm or more in diameter (less than about 50% by weight). The average texture, then, is more likely to be loam, clay loam, or sandy clay loam. *Mixed* means a mixed mineralogy that is less than 40% of any one mineral other than quartz in the fraction between 0.02 and 2.0 mm in diameter; less than 20% (by weight) glauconitic pellets in the fine earth fraction; a total iron plus gibbsite (by weight) in the fine earth fraction of 5% or less; a fine earth fraction that has at least one of the following: free carbonates, the pH of a suspension of 1 g soil in 50 mL 1 M NaF of 8.4 or less after 2 min, or a ratio of −1500 kPa water to measured clay of 0.6 or less. *Superactive* means the CEC divided by the clay content (%) is 0.60 or more, which indicates the soil is relatively high in bases. *Mesic* indicates a mesic temperature regime, that is, the mean annual soil temperature is between 8°C and 15°C (47°F and 59°F) and the soil temperature fluctuates more than 6°C between summer and winter. In other words, the soil is somewhere in the midlatitudes, summer is warm or hot, and winter is cool or cold. *Depth of soil* when no class is used, in Ustic Calciargids, means the soil is 50 cm or more deep.

It is not necessary to know the exact criteria in the *Keys to Soil Taxonomy* in order to communicate useful information about a soil. It is necessary to know the language of *Soil Taxonomy*. For example, from Ustic Calciargids one knows the following.

Ustic at the subgroup level of an Aridisol means that the soil is moister than what is typical for an Aridisol or aridic soil moisture regime and close to having an ustic moisture regime. *Calci* means the soil has a calcic horizon within 150 cm of the soil surface. *Argi* denotes presence of an argillic horizon. The *id* means it is an Aridisol. Therefore, one can say that the soil occurs in an arid, temperate climate. Presence of an argillic horizon implies that the soil occurs on stable, older landscape. It has a calcic horizon and under irrigation, Fe chlorosis may be a problem in sensitive plants. If the soil is not irrigated, it can be used only for limited grazing.

31.8 Forming Names

31.8.1 Names of Orders, Suborders, and Great Groups

The names of the orders, suborders, and great groups that are currently recognized are presented in Chapter 33 under each order.

31.8.1.1 Names of Subgroups

The name of a subgroup consists of the name of a great group modified by one or more adjectives. As explained earlier, the adjective typic is used for the subgroup that is thought to typify the central concept of the great group, or for soils that fail to meet the criteria of the other subgroups defined for a great group.

Intergrade subgroups that have, in addition to the properties of their great group, some properties of another taxon carry the name of the other taxon in the form of an adjective. The names of orders, suborders, or great groups or any of the prior (first) formative elements of those names may be used in the form of an adjective in subgroup names. A few soils may have aberrant properties of two great groups that belong in different orders or suborders. For these, it is necessary to use two names of taxa as adjectives in the subgroup name.

Extragrade subgroups carry the name of one or more special descriptive adjectives (Table 31.5) to modify the name of the great group and connote the nature of the aberrant properties.

31.8.1.2 Names of Intergrades Toward Other Great Groups in the Same Suborder

If the aberrant property of a soil is one that is characteristic of another great group in the same suborder, only the distinctive formative element of the great group name is used to indicate the aberrant properties. Thus, Typic Argidurid is defined, in part, as having an indurated or very strongly cemented duripan. If the only aberrant feature of an Argidic Argidurid is that the duripan is strongly cemented or less cemented throughout, it is considered to intergrade toward the Argids. The name, however, is Argidic Argidurids, not Haplargidic Argidurids. Only the prior (first) formative element is used in adjectival form if the two great groups are in the same suborder.

31.8.1.3 Names of Intergrades Toward a Great Group in the Same Order but Different Suborder

Two kinds of names have been chosen here. If the only aberrant features are color and moisture regime and the hue is too yellow or the chroma is too high or too low for the typic subgroup, the adjectives aeric or aquic are used.

If an Epiaquult has chroma too high for the typic subgroup, but has no other aberrant feature, it is placed in an aeric subgroup. Using an adjective taken from the suborder, udic, would not suggest that the difference is one of aeration alone.

If the only aberrant feature of a Hapludult is redoximorphic features that are too shallow for a Typic Hapludult, the adjective aquic is used in the subgroup name. If redox depletions (accompanied by aquic conditions, unless artificially drained) appear within the upper 60 cm of the argillic horizon, the soil is called an Aquic Hapludult, not an Aquultic Hapludult.

In other instances, the adjective in the subgroup name is made from the first two formative elements of the appropriate great group name in that suborder. For example, if a Paleudult has both shallow redoximorphic features and some plinthite, it is called a Plinthaquic Paleudult, not a Plinthaquultic Paleudult.

Note that the formative element for the order is not repeated in the adjective if the two great groups are in the same order.

31.8.1.4 Names of Subgroups not Intergrading Toward Any Known Kind of Soil (Extragrade)

Some soils have aberrant properties that are not characteristic of a class in a higher category of any order, suborder, or great group. One example might be taken from the concave pedons that are at the base of slopes, in depressions, or in other places where new soil material accumulates slowly on the surface. In these soils, material is added to the A horizon. The presence of an overthickened A horizon is not used to define any great group, but the soils lie outside the range of the typic subgroup and there is no class toward which they intergrade. Hence, a descriptive adjective is required. For this particular situation, the adjective cumulic (L. *cumulus*, heap, plus ic, Gr. *ikos*) is used to form the subgroup names. Pachic is used to indicate an overthickened epipedon if there is no evidence of new material at the surface.

Other soils lie outside the range of typic subgroups in an opposite direction. Such soils are, in effect, truncated by hard rock and are shallow or are intermittent between rock outcrops. They are, in effect, intergrades to nonsoil and are called lithic subgroups. The names of formative elements in groups of this sort, which are called extragrades, are listed together with their derivation in Table 31.5.

31.8.2 Names of Families

Each family requires one or more names. The technical family name consists of a series of descriptive terms modifying the subgroup name. For these terms the class names that are given later for particle-size class, mineralogy, and so on, in family differentiae are used. To have consistent nomenclature, the order of descriptive terms in names of families is particle-size class, mineralogy class, cation exchange activity class, calcareous and reaction class, soil temperature class, soil depth class, rupture resistance class, classes of coatings, and classes of cracks.

Redundancy in names of families is avoided. Particle-size class and temperature classes should not be used in the family name if they are specified above the family. Psamments, by definition, all have a sand or loamy sand texture and are in a sandy particle-size class, unless they are ashy. It is, therefore, redundant to use a particle-size class for Psamments, unless they are ashy.

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32

Other Systems of Soil Classification

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32.1 Introduction

Soil classification is probably as old as farming. The fact that, around 8000 BP, the first farming communities in Europe settled on the better loess soils indicates that, during these times, farmers were already capable of distinguishing between the more and less productive soils. The oldest historical record of soil classification is most likely the Chinese book, *Yugong*, in which the soils of China were classified into three categories and nine classes, based on soil color, texture, and hydrological features (Gong, 1994). Even today, such criteria are still in use by farmers to differentiate soils. Studies on indigenous soil knowledge in northern Ghana, for example, have shown that farmers use texture, color, stoniness, and soil depth to stratify the soils (Asiamah et al., 1997).

This chapter will describe (1) the *Legend of the Soil Map of the World* (FAO–UNESCO, 1974), its revised version (FAO–UNESCO, 1988), and the *World Reference Base for Soil Resources* (WRB) (ISSS–ISRIC–FAO, 1994, 1998; IUSS Working Group WRB, 2006) as international systems of soil classification and correlation; (2) the French system of soil classification (CPCS, 1967), which is still the system in use in large parts of West and Central Africa and the new *Référentiel Pédologique Français* (AFES–INRA, 1990, 1992); (3) the systems of soil classification used in Russia (VASKhNIL, 1986; Shishov et al., 2004); (4) the recently developed soil classification system of China (Gong and Zhang, 2007); and (5) the newly developed system of soil classification for Australia (Isbell, 1996) as classification systems, which are covering a range of ecological regions (from the arctic to the [sub]tropics). In addition, some attention will be paid to the soil classification systems in use in Brazil, Canada, England and Wales, Germany, New Zealand, and South Africa, which are more oriented toward one ecological region.

32.2 The FAO–UNESCO Legend of the Soil Map of the World 1:5,000,000

32.2.1 Introduction

The *Soil Map of the World* is a response to recommendations made by the international soil community in the 1950s to give special attention to developing the classification and correlation of the soils of great regions of the world (FAO–UNESCO, 1974). It was the first attempt to prepare, on the basis of international cooperation, a soil map covering all continents of the world using a uniform legend, thus enabling the correlation of soil units and the comparison of soils on a global scale. It should be borne in mind that the *Soil Map of the World* project produced a legend accompanying the map and not a global soil classification system.

32.2.2 History

The Food and Agriculture Organization (FAO) and United Nations Educational, Scientific, and Cultural Organization (UNESCO), in association with the International Society of Soil Science (ISSS), jointly took up the recommendations made during the Sixth and Seventh ISSS Congresses in 1956 and 1960 to prepare a *Soil Map of the World* at scale 1:5,000,000. The project started in 1961 and was based on the compilation of available soil survey material and field correlation. A scientific advisory panel was convened to study the scientific and methodological problems relative to the preparation of such a soil map of the world.

During a number of meetings, the advisory panel worked out the organization of the field correlation, selected the scale of the map and its topographic base, and prepared the first draft definitions of soil units. These were presented to the Eighth ISSS Congress in 1964. In 1966, a general agreement was reached on the principles for constructing the international legend, on the preparation of the definitions of soil units, and on the adoption of a unified nomenclature. The first draft was presented in 1968 to the Ninth ISSS Congress, which approved the outline of the legend, the definitions, and the nomenclature. Moreover, it recommended that the *Soil Map of the World* be published as soon as possible.

32.2.3 Objectives

The objectives of the *Soil Map of the World* project (FAO–UNESCO, 1974) were to (1) make a first appraisal of the world's soil resources, (2) supply a scientific basis for the transfer of experience between areas with similar environments, (3) promote the establishment of a generally accepted soil classification and nomenclature, (4) establish a common framework for more detailed investigations in developing areas, (5) serve as a basic document for educational, research, and development activities, and (6) strengthen international contacts in the field of soil science.

32.2.4 The Soil Units

The soil units that form the basis of the *FAO–UNESCO Legend* have been defined in terms of measurable and observable properties of the soil itself. They form a monocategorical and not a taxonomic system with different levels of generalization. However, for the sake of logical presentation, they can be grouped on generally accepted principles of soil formation (FAO–UNESCO, 1974). Based on soil development status, material, and major geographical zone, 24 major soils and 106 soil units are distinguished (Table 32.1).

The soil units are characterized by the presence or absence of diagnostic horizons and properties. Key properties have been selected on the basis of generally accepted principles of soil formation so as to correlate with as many other characteristics as possible. Clusters of properties have been combined into so-called diagnostic horizons, which have been adopted to formulate the definitions of the soil units. The definitions and nomenclature of the diagnostic horizons and properties used are drawn from those adopted in *Soil Taxonomy* (Soil Survey Staff, 1975), but the definitions have been summarized and sometimes simplified to serve the purpose of the legend. For full background and details, the user is referred to *Soil Taxonomy* (Soil Survey Staff, 1975, 1996). A brief overview of the diagnostic horizons and properties used in the *FAO–UNESCO Legend* and their meaning are given in Table 32.2.

Volume I of the *Legend of the Soil Map of the World* provides a key to the soil units, which can be used to identify the soil (FAO–UNESCO, 1974). An abbreviated version listing the major soil units is reproduced in Table 32.3.

32.3 The Revised Legend of the FAO–UNESCO Soil Map of the World

32.3.1 Introduction

In 1988, a revised version of the *Legend of the Soil Map of the World* was issued (FAO–UNESCO, 1988), which assessed to what extent the objectives of the original *Legend of the Soil Map of the World* (FAO–UNESCO, 1974; Section 32.2.3) were met and analyzed its present day function. It was realized that, in order to keep the maps and accompanying legend up to date, revisions were necessary. New knowledge on soils, particularly from the developing world, had emerged and more recent soil surveys had yielded better insight into the distribution of soils in the world. The *Revised Legend* did not replace the *1974 Legend*, which is still the reference for the *Soil Map of the World*. The *Revised Legend* served also as a framework for the establishment by the ISSS of the WRB (Section 32.4).

32.3.2 Amendments to the 1974 Legend of the Soil Map of the World

The monocategorical character of the *1974 Legend*, using only soil units, was transformed into a multicategorical system with major soil groupings (MSGs; e.g., Fluvisols), soil units

TABLE 32.1 List of 106 Soil Units of the *FAO–UNESCO Legend of the Soil Map of the World 1:5,000,000*

Non or weakly developed soils									
J	**Fluvisols**	**G**	**Gleysols**	**R**	**Regosols**	**I**	**Lithosols**		
Je	Eutric Fluvisols	Ge	Eutric Gleysols	Re	Eutric Regosols				
Jc	Calcaric Fluvisols	Gc	Calcaric Gleysols	RC	Calcaric Regosols				
Jd	Dystric Fluvisols	Gd	Dystric Gleysols	Rd	Dystric Regosols				
Jt	Thionic Fluvisols	Gm	Mollic Gleysols	Rx	Gelic Regosols				
		Gh	Humic Gleysols						
		Gp	Plinthic Gleysols						
		Gx	Gelic Gleysols						
Soils conditioned by parent material									
Q	**Arenosols**	**E**	**Rendzinas**	**U**	**Rankers**	**T**	**Andosols**	**V**	**Vertisols**
Qc	Cambic Arenosols					To	Ochric Andosols	Vp	Pellic Vertisols
Ql	Luvic Arenosols					Tm	Mollic Andosols	Vc	Chromic Vertisols
Qf	Ferralic Arenosols					Th	Humic Andosols		
Qa	Albic Arenosols					Tv	Vitric Andosols		
Soils from (semi-)arid region									
Z	**Solonchaks**	**S**	**Solonetz**	**Y**	**Yermosols**	**X**	**Xerosols**		
Zo	Orthic Solonchaks	So	Orthic Solonetz	Yh	Haplic Yermosols	Xh	Haplic Xerosols		
Zm	Mollic Solonchaks	Sm	Mollic Solonetz	Yk	Calcic Yermosols	Xk	Calcic Xerosols		
Zt	Takyric Solonchaks	Sg	Gleyic Solonetz	Yy	Gypsic Yermosols	Xy	Gypsic Xerosols		
Zg	Gleyic Solonchaks			Yl	Luvic Yermosols	Xl	Luvic Xerosols		
				Yt	Takyric Yermosols				
Soils from the steppe regions									
K	**Kastanozems**	**C**	**Chernozems**	**H**	**Phaeozems**	**M**	**Greyzems**		
Kh	Haplic Kastanozems	Ch	Haplic Chernozems	Hh	Haplic Phaeozems	Mo	Orthic Greyzems		
Kk	Calcic Kastanozems	Ck	Calcic Chernozems	Hc	Calcaric Phaeozems	Mg	Gleyic Greyzems		
Kl	Luvic Kastanozems	Cl	Luvic Chernozems	Hl	Luvic Phaeozems				
		Cg	Glossic Chernozems	Hg	Gleyic Phaeozems				
Moderately developed soils mainly from temperate regions									
B	**Cambisols**	**L**	**Luvisols**	**D**	**Podzoluvisols**	**P**	**Podzols**	**W**	**Planosols**
Be	Eutric Cambisols	Lo	Orthic Luvisols	De	Eutric Podzoluvisols	Po	Orthic Podzols	We	Eutric Planosols
Bd	Dystric Cambisols	Lc	Chromic Luvisols	Dd	Dystric Podzoluvisols	Pl	Leptic Podzols	Wd	Dystric Planosols
Bh	Humic Cambisols	Lk	Calcic Luvisols	Dg	Gleyic Podzoluvisols	Pf	Ferric Podzols	Wm	Mollic Planosols
Bg	Gleyic Cambisols	Lv	Vertic Luvisols			Ph	Humic Podzols	Wh	Humic Planosols
Bx	Gelic Cambisols	Lf	Ferric Luvisols			Pp	Placic Podzols	Ws	Solodic Planosols
Bk	Calcic Cambisols	La	Albic Luvisols			Pg	Gleyic Podzols	Wx	Gelic Planosols
Bc	Chromic Cambisols	Lp	Plinthic Luvisols						
Bv	Vertic Cambisols	Lg	Gleyic Luvisols						
Bf	Ferralic Cambisols								
Strongly weathered soils mainly from the tropical regions									
A	**Acrisols**	**N**	**Nitosols**	**F**	**Ferralsols**				
Ao	Orthic Acrisols	Ne	Eutric Nitosols	Fo	Orthic Ferralsols				
Af	Ferric Acrisols	Nd	Dystric Nitosols	Fx	Xanthic Ferralsols				
Ah	Humic Acrisols	Nh	Humic Nitosols	Fr	Rhodic Ferralsols				
Ap	Plinthic Acrisols			Fh	Humic Ferralsols				
Ag	Gleyic Acrisols			Fa	Acric Ferralsols				
				Fp	Plinthic Ferralsols				
Organic soils									
O	**Histosols**								
Oe	Eutric Histosols								
Od	Dystric Histosols								
Ox	Gelic Histosols								

Source: FAO–UNESCO. 1974. Soil map of the world 1:5,000,000. Volume I, Legend. United Nations Educational, Scientific and Cultural Organization, Paris, France.

TABLE 32.2 Diagnostic Horizons and Properties in the *FAO–UNESCO Legend of the Soil Map of the World 1:5,000,000*

Horizons	
Albic E	Light colored eluvial horizon generally associated with argic and spodic horizons
Argillic B	Subsurface horizon with distinct clay accumulation
Calcic	Horizon with accumulation of calcium carbonate
Cambic B	Subsurface horizon showing evidence of alteration relative to the underlying horizon(s)
Gypsic	Horizon with accumulation of gypsum
Histic H	Poorly aerated, waterlogged, highly organic surface horizon
Mollic A	Thick, dark-colored surface horizon with high base saturation and moderate to high organic matter content
Natric B	Subsurface horizon with distinct clay accumulation and a high exchangeable sodium percentage
Ochric A	Weakly developed surface horizon, either light colored, or thin, or having a low organic matter content
Oxic B	Strongly weathered subsurface horizon with low cation exchange capacity (CEC)
Spodic B	Dark-colored subsurface horizon with illuvial aluminoorganic complexes, with or without iron
Sulfuric	Extremely acid subsurface horizon with sulfuric acid resulting from oxidation of sulfides
Umbric A	Thick, dark-colored surface horizon with low base saturation and moderate to high organic matter content
Properties and materials	
Abrupt textural change	Sharp increase in clay content within a limited depth range
Albic material	Light colored mineral soil material
Aridic moisture regime	No available water in any part of the moisture control section for as long as 90 consecutive days or more than half the time when soil temperature (at 50 cm depth) is above 5°C
Exchange complex dominated by amorphous material	CEC (pH 8.2) more than 150 $cmol_c \cdot kg^{-1}$ clay; if 15 bar water content is 20% or more, pH of NaF is more than 9.4; ratio of 15 bar water content to clay is more than 1.0; more than 0.6% organic carbon; DTA shows low-temperature endotherm; bulk density is 0.85 $g\ cm^{-3}$ at 1/3 bar tension, low CEC (<24 $cmol_c\ kg^{-1}$ clay by NH_4Cl)
Ferralic ferric	Iron concentrated in large mottles or concretions, or low CEC (<24 $cmol_c\ kg^{-1}$ clay by NH_4Cl)
Gilgai microrelief	Succession of enclosed microbasins and microknolls in level areas or microvalleys and microridges on slopes
High organic matter content	Organic matter content of 1.35% or more averaged to a depth of 100 cm or 1.5% organic matter in the upper part of the B horizon (acrisols only)
High salinity	EC of saturation extract more than 15 $dS\ m^{-1}$ at 25°C at specified depths or more than 4 $dS\ m^{-1}$ within 25 cm of the surface if pH (H_2O, 1:1) exceeds 8.5
Hydromorphic properties	Saturation with groundwater; occurrence of a histic H horizon; dominant neutral (N) hues or hues bluer than 10Y; saturation with water at some period of the year (unless artificially drained) with evidence of reduction or of reduction and segregation
Interfingering	Penetrations of an albic E horizon into an underlying argillic or natric B horizon, not wide enough to constitute tonguing
Permafrost	Perennial temperature at or below 0°C
Plinthite	Iron-rich, humus-poor soil material, which hardens irreversibly upon repeated wetting and drying
Slickensides	Polished and grooved ped surfaces that are produced by sliding past another
Smeary consistence	Presence of thixotropic soil material
Soft powdery lime	Accumulation of translocated calcium carbonate in soft powdery form
Sulfidic materials	Waterlogged deposit containing 0.75% or more sulfur and less than three times as much carbonates ($CaCO_3$ equivalent) as sulfur
Takyric features	Combination of heavy texture, polygonal cracks, and platy or massive surface crust
Thin iron pan	Black to dark reddish layer cemented by iron, by iron and manganese, or by iron–organic matter complexes
Tonguing	Penetrations of an albic E horizon into an argillic B horizon with specified dimensions
Vertic properties	Cracks 1 cm or more wide within 50 cm of the upper boundary of the B horizon, extending to the surface or at least to the upper part of the B horizon
Weatherable minerals	Presence of minerals considered to be unstable relative to other minerals such as quartz and 1:1 lattice clays and which produce plant nutrients upon weathering

Source: FAO–UNESCO. 1974. Soil map of the world 1:5,000,000. Volume I, Legend. United Nations Educational, Scientific and Cultural Organization, Paris, France.

(e.g., Dystric Fluvisols), and soil subunits (e.g., Gleyi-dystric Fluvisols). The introduction of a multicategorical system was necessitated by the increasing use of the *1974 Legend* in more detailed surveys, particularly in Africa, where during the 1980s 1:1,000,000 soil maps were produced of, for example, Kenya, Botswana, and Zambia.

The *Revised Legend of the Soil Map of the World* distinguishes 28 MSGs, 4 more than the *1974 Legend*, and 153 soil units, 47 more than in 1974. Major changes are (1) amalgamation of the Lithosols, Rendzinas, and Rankers into one MSG (Leptosols), since the three had been difficult to show on the map; (2) deletion of the Xerosols and Yermosols, which were characterized by an

TABLE 32.3 Key to Major Soil Units of the 1974 *FAO–UNESCO Legend*

Soils having an H horizon of 40 cm or more (60 cm or more if the organic material consists mainly of sphagnum or moss or has a bulk density of less than 0.1 g cm^{-3}) either extending down from the surface or taken cumulatively within the upper 80 cm of the soil; the thickness of the H horizon maybe less when it rests on rock or on fragmental material of which the interstices are filled with organic matter	**Histosols (O)**
Other soils that are limited in depth by continuous coherent and hard rock within 10 cm of the surface	**Lithosols (I)**
Other soils that, after the upper 20 cm have been mixed, have 30% or more clay in all horizons to at least 50 cm from the surface; at some period in most years have cracks at least 1 cm wide at a depth of 50 cm, unless irrigated, and have one or more of the following characteristics: gilgai microrelief, intersecting slickensides, or wedge-shaped or parallelepiped structural aggregates at some depth between 25 and 100 cm from the surface	**Vertisols (V)**
Other soils developed from recent alluvial deposits, having no diagnostic horizons other than (unless buried by 50 cm or more new material) an ochric or an umbric A horizon, a histic H horizon, or a sulfuric horizon	**Fluvisols (J)**
Other soils having high salinity and having no diagnostic horizons (unless buried by 50 cm or more new material) an A horizon, an H horizon, a cambic B horizon, a calcic or a gypsic horizon	**Solonchaks (Z)**
Other soils showing hydromorphic properties within 50 cm of the surface; having no diagnostic horizons other than (unless buried by 50 cm or more new material) an A horizon, an H horizon, a cambic B horizon, a calcic or gypsic horizon	**Gleysols (G)**
Other soils having either a mollic or an umbric A horizon possibly overlying a cambic B horizon or an ochric A horizon and a cambic B horizon; having no other diagnostic horizons (unless buried by 50 cm or more new material); having to a depth of 35 cm or more one or both of (a) a bulk density (at 1/3 bar water retention) of the fine earth (less than 2 mm) fraction of the soil of less than 0.85 g cm^{-3} and an exchange complex dominated by amorphous material; (b) 60% or more vitric volcanic ash, cinders, or other vitric pyroclastic material in the silt, sand, and gravel fractions	**Andosols (T)**
Other soils of coarse texture consisting of albic material occurring over a depth of at least 50 cm from the surface or showing characteristics of argillic, cambic, or oxic B horizons which, however, do not qualify as diagnostic horizons because of textural requirements; having no diagnostic horizons other than (unless buried by 50 cm or more new material) an ochric A horizon	**Arenosols (Q)**
Other soils having no diagnostic horizons or none other than (unless buried by 50 cm or more new material) an ochric A horizon	**Regosols (R)**
Other soils having an umbric A horizon, which is not more than 25 cm thick; having no other diagnostic horizons (unless buried by 50 cm or more new material)	**Rankers (U)**

TABLE 32.3 (continued) Key to Major Soil Units of the 1974 *FAO–UNESCO Legend*

Other soils having a mollic A horizon that contains or immediately overlies calcareous material with a calcium carbonate equivalent of more than 40% (when the A horizon contains a high amount of finely divided calcium carbonate the color requirements of the mollic A horizon may be waived)	**Rendzinas (E)**
Other soils having a spodic B horizon	**Podzols (P)**
Other soils having an oxic B horizon	**Ferralsols (F)**
Other soils having an albic E horizon overlying a slowly permeable horizon (e.g., an argillic or natric B horizon showing an abrupt textural change, a heavy clay, a fragipan) within 125 cm of the surface; showing hydromorphic properties at least in a part of the E horizon	**Planosols (W)**
Other soils having a natric B horizon	**Solonetz (S)**
Other soils having a mollic A horizon with a moist chroma of 2 or less to a depth of at least 15 cm, showing bleached coatings on structural ped surfaces	**Greyzems (M)**
Other soils having a mollic A horizon with a moist chroma of 2 or less to a depth of at least 15 cm, having one or more of the following: a calcic or a gypsic horizon or concentrations of soft powdery lime within 125 cm of the surface, when the weighted average textural class is coarse, within 90 cm for medium textures, within 75 cm for fine textures	**Chernozems (C)**
Other soils having a mollic A horizon with a moist chroma of more than 2 to a depth of at least 15 cm, having one or more of the following: a calcic or a gypsic horizon or concentrations of soft powdery lime within 125 cm of the surface, when the weighted average textural class is coarse, within 90 cm for medium textures, within 75 cm for fine textures	**Kastanozems (K)**
Other soils having a mollic A horizon	**Phaeozems (H)**
Other soils having an argillic B horizon showing an irregular or broken upper boundary resulting from deep tonguing of the E into the B horizon or from the formation of discrete nodules (ranging from 2 to 5 cm up to 30 cm in diameter) the exteriors of which are enriched and weakly cemented or indurated with iron and having redder hues and stronger chroma than the interiors	**Podzoluvisols (D)**
Other soils having a weak ochric A horizon and an aridic moisture regime; lacking permafrost within 200 cm of the surface	**Xerosols (X)**
Other soils having a very weak ochric A horizon and an aridic moisture regime; lacking permafrost within 200 cm of the surface	**Yermosols (Y)**

(*continued*)

TABLE 32.3 (continued) Key to Major Soil Units of the 1974 *FAO–UNESCO Legend*

Other soils having an argillic B horizon with a clay distribution, where the percentage of clay does not decrease from its maximum amount by as much as 20% within 150 cm of the surface; lacking plinthite within 125 cm of the surface; lacking vertic and ferric properties	
	Nitosols (N)
Other soils having an argillic B horizon; having a base saturation, which is less than 50% (by NH_4OAc) in at least some part of the B horizon within 125 cm of the surface	
	Acrisols (A)
Other soils having an argillic B horizon	
	Luvisols (L)
Other soils having a cambic B horizon or an umbric A horizon, which is more than 25 cm thick	
	Cambisols (B)

Source: FAO–UNESCO. 1974. Soil map of the world 1:5,000,000. Volume I, Legend. United Nations Educational, Scientific and Cultural Organization, Paris, France.

aridic moisture regime (as a general principle, climatic criteria were not to be used in separating the soil units); (3) division of the Acrisols and Luvisols, in 1974 distinguished by base saturation, into four MSGs, introducing clay activity as an additional separating criterion; (4) introduction of new MSGs (Alisols, Calcisols, Gypsisols, Lixisols, Plinthosols, and Anthrosols); and (5) renaming of Nitosols (Nitisols). A listing of the MSGs and the soil units is given in Table 32.4.

The diagnostic horizons and properties have also been largely adapted. In 1974, they were fairly similar to those in *Soil Taxonomy* (Soil Survey Staff, 1975), but in 1988, many were redefined and renamed and new additions made. The argillic and oxic B horizons were renamed argic and ferralic B horizons, respectively. The argic B horizon now includes both the argillic and the kandic horizon as defined in *Soil Taxonomy* (Soil Survey Staff, 1996), while the ferralic B horizon includes additional criteria such as silt/clay ratio and water-dispersible clay content. An addition is the thick, manmade, fimic A horizon, which includes both the anthropic and plaggen epipedons of *Soil Taxonomy*.

Newly defined diagnostic properties include continuous hard rock, fluvic, geric, nitic, and sodic properties. The 1974 hydromorphic properties were split into gleyic (wetness conditioned by groundwater) and stagnic (wetness conditioned by surface water) properties. Definitions of albic material, aridic moisture regime, high organic matter content, and thin iron pan were deleted as they are no longer used in defining soil units. Andic properties have replaced the 1974 exchange complex dominated by amorphous material.

A brief overview of the diagnostic horizons and properties used in the *Revised Legend* and their meaning is given in Table 32.5. A key to the MSGs in the *Revised Legend* is given in Table 32.6.

32.4 The World Reference Base for Soil Resources

32.4.1 Introduction

The first official version of the *World Reference Base for Soil Resources* (WRB) was released at the 16th World Congress of Soil Science at Montpellier in 1998. At the same event, it was also endorsed and adopted as the system for soil correlation and international communication of the International Union of Soil Sciences (IUSS). It is also serving as a correlation scheme for soil maps and databases for the European Union. Although it was not designed for, in several countries it is also applied for different scale of soil mapping. The current official version was released at the 18th World Congress of Soil Science in Philadelphia in 2006, and an updated digital version was published in 2007 on the official, FAO-hosted Web site of the WRB. The main text was translated into 14 languages (Arabic, Chinese, French, German, Hungarian, Italian, Japanese, Latvian, Lithuanian, Polish, Rumanian, Russian, Spanish, and Vietnamese) and most translations are also available online on the Web site http://www.fao.org/nr/land/soils/soil/wrb-documents/jp/

32.4.2 History

The WRB was initiated by the ISSS, FAO, and ISRIC to "provide scientific depth and background to the 1988 FAO–UNESCO–ISRIC *Revised Legend of the Soil Map of the World*, so that it incorporates the latest knowledge relating to global soil resources and interrelationships" (ISSS–ISRIC–FAO, 1994). The initiative was closely related to the *Soil Map of the World 1:5,000,000* project of FAO and UNESCO. After its completion in the early 1980s, it was realized that 20 years had elapsed since the project had started. During this period, numerous new soil surveys, often using local or national soil classification systems, had been carried out both in developing and developed countries, generating new knowledge and insight on the distribution and potential of our soil resources. If the *Soil Map of the World* were to retain its value as a global soil resource inventory, it needed regular updating with the most recent information.

Preliminary discussions were started on the establishment of an International Reference Base for Soil Classification (IRB), an initiative undertaken by FAO and UNESCO, supported by the United Nations Environmental Program (UNEP) and the ISSS. The intention of the IRB project was to work toward the establishment of a framework through which existing soil classification systems could be correlated and ongoing soil classification work could be harmonized (Dudal, 1990). Meetings were organized in Sofia, Bulgaria, to commence such an international program. The outcomes were draft definitions of 16 MSGs occurring globally.

During the ISSS congresses in 1982 and 1986 and expert consultations in 1987 and 1988, the IRB took form, and as a result, 20 MSGs were identified and agreed upon as being representative of the principal components of the world's soil cover.

TABLE 32.4 MSGs and Soil Units, 1988 *Revised Legend of the Soil Map of the World*

AC, **Acrisols**
ACh, haplic Acrisols
ACf, ferric Acrisols
ACu, humic Acrisols
ACp, plinthic Acrisols
ACg, gleyic Acrisols

AL, **Alisols**
ALh, haplic Alisols
ALf, ferric Alisols
ALu, humic Alisols
ALp, plinthic Alisols
ALj, stagnic Alisols
ALg, gleyic Alisols

AR, **Arenosols**
ARh, haplic Arenosols
ARb, cambic Arenosols
ARl, luvic Arenosols
ARo, ferralic Arenosols
ARa, albic Arenosols
ARc, calcaric Arenosols
ARg, gleyic Arenosols

AN, **Andosols**
ANh, haplic Andosols
ANm, mollic Andosols
ANu, umbric Andosols
ANz, vitric Andosols
ANg, gleyic Andosols
ANI, gelic Andosols

AT, **Anthrosols**
ATa, aric Anthrosols
ATc, cumulic Anthrosols
ATf, fimic Anthrosols
ATu, urbic Anthrosols

CL, **Calcisols**
CLh, haplic Calcisols
CLc, luvic Calcisols
CLp, petric Calcisols

CM, **Cambisols**
CMe, eutric Cambisols
CMd, dystric Cambisols
CMu, humic Cambisols
CMc, calcaric Cambisols
CMv, vertic Cambisols
CMo, ferralic Cambisols
CMg, gleyic Cambisols
CMi, gelic Cambisols

CH, **Chernozems**
CHh, haplic Chernozems
CHk, calcic Chernozems
CHl, luvic Chernozems
CHw, glossic Chernozems
CHg, gleyic Chernozems

FL, **Fluvisols**
FLe, eutric Fluvisols
FLc, calcaric Fluvisols
FLd, dystric Fluvisols
FLm, mollic Fluvisols
FLu, umbric Fluvisols
FLt, thionic Fluvisols
FLs, salic Fluvisols

FR, **Ferralsols**
FRh, haplic Ferralsols
FRx, xanthic Ferralsols
FRr, rhodic Ferralsols
FRu, humic Ferralsols
FRg, geric Ferralsols
FRp, plinthic Ferralsols

GL, **Gleysols**
GLe, eutric Gleysols
GLk, calcic Gleysols
GLd, dystric Gleysols
GLa, andic Gleysols
GLm, mollic Gleysols
GLu, umbric Gleysols
GLt, thionic Gleysols
GLi, gelic Gleysols

GR, **Grgreyzems**
GRh, haplic Greyzems
GRg, gleyic Greyzems

GY, **Gypsisols**
GYh, haplic Gypsisols
GYk, calcic Gypsisols
GYl, luvic Gypsisols
GYp, petric Gypsisols

HS, **Histosols**
HSl, folic Histosols
HSs, terric Histosols
HSf, fibric Histosols
HSt, thionic Histosols
HSi, gelic Histosols

KS, **Kastanozems**
KSh, haplic Kastanozems
KSl, luvic Kastanozems
KSk, calcic Kastanozems
KSy, gypsic Kastanozems

LP, **Leptosols**
LPe, eutric Leptosols
LPd, dystric Leptosols
LPk, rendzic Leptosols
LPm, mollic Leptosols
LPu, umbric Leptosols
LPq, lithic Leptosols
LPi, gelic Leptosols

LX, **Lixisols**
LXh, haplic Lixisols
LXr, ferric Lixisols
LXp, plinthic Lixisols
LXa, albic Lixisols
LXj, stagnic Lixisols
LXg, gleyic Lixisols

LV, **Luvisols**
LVh, haplic Luvisols
LVf, ferric Luvisols
LVx, chromic Luvisols
LVk, calcic Luvisols
LVv, vertic Luvisols
LVa, albic Luvisols
LVj, stagnic Luvisols
LVg, gleyic Luvisols

NT, **Nitisols**
NTh, haplic Nitisols
ATr, rhodic Nitisols
NTu, humic Nitisols

PH, **Phaeozems**
PHh, haplic Phaeozems
PHc, calcaric Phaeozems
PHl, luvic Phaeozems
PHj, stagnic Phaeozems
PHg, gleyic Phaeozems

PL, **Planosols**
PLe, eutric Planosols
PLd, dystric Planosols
PLm, mollic Planosols
PLu, umbric Planosols
PLi, gelic Planosols

(*continued*)

TABLE 32.4 (continued) MSGs and Soil Units, 1988 *Revised Legend of the Soil Map of the World*

PT, **Plinthosols**	**PZc**, carbic Podzols	**SCn**, sodic Solonchaks
PTe, eutric Plinthosols	**PZg**, gleyic Podzols	**SCg**, gleyic Solonchaks
PTd, dystric Plinthosols	**PZi**, gelic Podzols	**SCi**, gelic Solonchaks
PTu, humic Plinthosols		
PTa, albic Plinthosols	**RG**, **Regosols**	**SN**, **Solonetz**
	RGe, eutric Regosols	**SNh**, haplic Solonetz
PD, **Podzoluvisols**	**RGc**, calcaric Regosols	**SNm**, mollic Solonetz
PDe, eutric Podzoluvisols	**RGy**, gypsiric Regosols	**SNk**, calcic Solonetz
PLd, dystric Podzoluvisols	**RGd**, dystric Regosols	**SNy**, gypsic Solonetz
PLj, stagnic Podzoluvisols	**RGu**, umbric Regosols	**SNj**, stagnic Solonetz
PDg, gleyic Podzoluvisols	**RGi**, gelic Regosols	**SNg**, gleyic Solonetz
PDi, gelic Podzoluvisols		
	SC, **Solonchaks**	**VR**, **Vertisols**
PZ, **Podzols**	**SCh**, haplic Solonchaks	**VRe**, eutric Vertisols
PZh, haplic Podzols	**SCm**, mollic Solonchak	**VRd**, dystric Vertisols
PZb, cambic Podzols	**SCk**, calcic Solonchaks	**Vrk**, calcic Vertisols
PZf, ferric Podzols	**SCy**, gypsic Solonchaks	**VRy**, gypsic Vertisols

Source: FAO–UNESCO. 1988. FAO–UNESCO soil map of the world, revised legend. World Soil Resources Report No. 60. Food and Agriculture Organization, Rome, Italy.

Subsequently, it became clear that some of the proposed 20 MSGs were too broad to be defined consistently and, consequently, had to be subdivided. By doing so, the list of MSGs became very close to those of the *Revised Legend of the Soil Map of the World* (FAO–UNESCO, 1988). As a result, it was decided in 1992 to adopt the revised legend as the frame for further IRB work. This was also prompted by the fact that both the *Revised Legend* and the IRB were supported by the ISSS and that it would be inappropriate to pursue two programs, which essentially had the same goal, namely, to arrive at a rational inventory of global soil resources (ISSS–ISRIC–FAO, 1994, FAO–ISRIC–ISSS, 1998). The two programs were, therefore, merged under the name WRB. After 8 years of data collection and intensive worldwide testing of the first edition (1998), the second edition (2006 and 2007) was presented after broad discussions. The text of the chapters to follow will be based mostly on the current version (IUSS Working Group WRB, 2006) indicating significant changes from the earlier edition.

32.4.3 Objectives

The specific objectives of the WRB are to (1) develop an internationally acceptable framework for delineating soil resources to which national classifications can be attached and related, using the *FAO Revised Legend* as a guideline; (2) provide this framework with a sound scientific base, so that it can also serve different applications in related fields, such as agriculture, geology, hydrology, and ecology; (3) acknowledge in the framework important lateral aspects of soils and soil horizon distribution as characterized by topo- and chronosequences; and (4) emphasize the morphological characterization of soils rather than to follow a purely analytical approach.

32.4.4 Concepts and Principles

The general principles on which the WRB is based were laid down during the early Sofia meetings in 1980 and 1981 and further elaborated upon by the working groups entrusted with its development. These general principles can be summarized as follows: (1) The classification of soils is based on soil properties defined in terms of diagnostic horizons, properties, and materials, which to the greatest extent possible should be measurable and observable in the field; (2) selection of diagnostic horizons, properties, and materials takes into account their relationship with soil-forming processes; however, at a high level of generalization, it also attempts to select, to the extent possible, diagnostic features, which are of significance for management purposes; (3) climatic parameters are not applied in the classification of soils. It is fully realized that they should be used for interpretation purposes, in dynamic combination with soil properties, but they should not form part of soil definitions; (4) the WRB is a comprehensive classification system that enables people to accommodate their national classification system. It comprises two tiers of categorical detail: *the reference base*, limited to the first level only and having 32 reference soil groups (RSGs) and the *WRB classification system*, consisting of combinations of a set of prefix and suffix qualifiers that are uniquely defined and added to the name of the RSG, allowing very precise characterization and classification of individual soil profiles. The first edition of the *WRB*, published in 1998, comprised 30 RSGs; the second edition, published in 2006, has 32 RSGs. Definitions and descriptions of lower level soil units reflect variations in soil characteristics both vertically and laterally so as to account for spatial linkages within the landscape. The term "Reference Base" is connotative

TABLE 32.5 Diagnostic Horizons and Properties, 1988 *Revised Legend of the Soil Map of the World*

Horizons	
Albic E	Light colored eluvial horizon generally associated with argic and spodic horizons
Argic B	Subsurface horizon with distinct clay accumulation
Calcic	Horizon with accumulation of calcium carbonate
Cambic B	Subsurface horizon showing evidence of alteration relative to the underlying horizon(s)
Ferralic B	Strongly weathered subsurface horizon with low CEC
Fimic A	Surface and subsurface horizons resulting from long-continued cultivation
Gypsic	Horizon with accumulation of gypsum
Histic H	Poorly aerated, waterlogged, highly organic surface horizon
Mollic A	Thick, dark-colored surface horizon with high base saturation and moderate to high organic matter content
Natric B	Subsurface horizon with distinct clay accumulation and a high exchangeable sodium percentage
Ochric A	Weakly developed surface horizon, either light colored, or thin, or having a low organic matter content
Petrocalcic	Continuous cemented or indurated calcic horizon
Petrogypsic	Continuous cemented or indurated gypsic horizon
Spodic B	Dark-colored subsurface horizon with illuvial aluminoorganic complexes, with or without iron
Sulfuric	Extremely acid subsurface horizon with sulfuric acid resulting from oxidation of sulfides
Umbric A	Thick, dark-colored surface horizon with low base saturation and moderate to high organic matter content
Properties	
Abrupt textural change	Sharp increase in clay content within a limited depth range
Andic	High acid oxalate extractable Al and Fe content, low bulk density, high phosphate retention, high amount of coarse volcanoclastic material
Calcareous	Strong effervescence with 10% HCl (more than 2% calcium carbonate)
Calcaric	Presence of calcareous soil material between 20 and 50 cm depth
Continuous hard rock	Presence of coherent rock, practically impermeable for roots
Ferralic	Low CEC (<24 $cmol_c\ kg^{-1}$ clay or 4 $cmol_c\ kg^{-1}$ fine earth)
Ferric	Presence of many coarse mottles or large iron concretions
Fluvic	Presence of fresh fluviatile, lacustrine, or marine sediments at the surface
Geric	Extremely low to negative effective CEC
Gleyic	Wetness producing reduced conditions caused by groundwater
Gypsiferous	Presence ≥5% gypsum
Interfingering	Narrow penetrations of an albic E horizon into an argic or natric B horizon
Nitic	Presence of strongly developed, nut-shaped structure and shiny pedfaces
Organic material	Material containing a very high amount of organic debris
Permafrost	Perennial temperature at or below 0°C
Plinthite	Presence of iron-rich, humus-poor material, which hardens irreversibly upon repeated wetting and drying
Salic	High soluble salt content (electrical conductivity >15 $dS\ m^{-1}$ or >4 if pH exceeds 8.5)
Slickensides	Presence of polished and grooved surfaces produced by one mass sliding past another
Smeary consistence	Presence of thixotropic soil material
Sodic	ESP ≥ 15% or exchangeable Na + Mg percentage ≥50%
Soft powdery lime	Accumulation of translocated calcium carbonate in soft powdery form
Stagnic	Wetness producing reduced conditions caused by stagnating surface water
Strongly humic	High organic matter content
Sulfidic material	Waterlogged deposit containing sulfides and only moderate amounts of calcium carbonate
Tonguing	Wide penetrations of an albic E horizon into an argic B horizon or penetrations of a mollic A horizon into an underlying cambic B horizon or into a C horizon (Chernozems only)
Vertic	Presence of cracks, slickensides, wedge-shaped, or parallelepiped structural aggregates
Weatherable minerals	Presence of minerals unstable in a humid climate relative to other minerals

Source: FAO–UNESCO. 1988. FAO–UNESCO soil map of the world, revised legend. World Soil Resources Report No. 60. Food and Agriculture Organization, Rome, Italy.

TABLE 32.6 Key to MSGs of the 1988 *Revised Legend of the Soil Map of the World*

Criteria	Major soil group
Soils having an H horizon, or an O horizon, of 40 cm or more (60 cm or more if the organic material consists mainly of sphagnum or moss or has a bulk density of less than 0.1 Mg m^{-3}) either extending down from the surface or taken cumulatively within the upper 80 cm of the soil; the thickness of the H or O horizon may be less when it rests on rocks or on fragmental material of which the interstices are filled with organic matter	**Histosols (HS)**
Other soils that are coarser than sandy loam to a depth of at least 100 cm from the surface, having <35% of rock fragments or other coarse fragments in all subhorizons within 100 cm of the surface, having no diagnostic horizons other than an ochric A horizon or an albic E horizon	**Arenosols (AR)**
Other soils in which human activities have resulted in a profound modification or burial of the original soil horizons, through removal or disturbance of surface horizons, cuts and fills, secular additions of organic materials, long-continued irrigation, etc.	**Anthrosols (AT)**
Other soils having no diagnostic horizons other than an ochric or umbric A horizon; lacking soft powdery lime	**Regosols (RG)**
Other soils that are limited in depth by continuous hard rock or highly calcareous materials (calcium carbonate equivalent >40%) or a continuous cemented layer within 30 cm of the surface or having <20% of fine earth over a depth of 75 cm from the surface. Diagnostic horizons may be present	**Leptosols (LP)**
Other soils having a spodic B horizon	**Podzols (PZ)**
Other soils having, after the upper 18 cm have been mixed. Thirty percent or more clay in all horizons to a depth of 50 cm; developing cracks from the soil surface downward which at some period in most years (unless the soil is irrigated) are at least 1 cm wide to a depth of 50 cm; having one or more of the following: Intersecting slickensides or wedge-shaped or parallelepiped structural aggregates at some depth between 25 and 100 cm from the surface	**Vertisols (VR)**
Other soils having ≥25% plinthite by volume in a horizon, which is at least 15 cm thick within 50 cm of the surface or within a depth of 125 cm when underlying an albic E horizon or a horizon that shows stagnic properties within 50 cm of the surface or gleyic properties within 100 cm of the surface	**Plinthosols (PT)**
Other soils showing fluvic properties and having no diagnostic horizons other than an ochric, mollic, an umbric A horizon, or a histic H horizon, or a sulfuric horizon, or sulfidic material within 125 cm of the surface	**Fluvisols (FL)**
Other soils having a ferralic B horizon	**Ferralsols (FR)**
Other soils showing salic properties and having no diagnostic horizons other than an ochric, umbric, or mollic A horizon, a histic H horizon, a cambic B horizon, a calcic, or a gypsic horizon	**Solonchaks (SC)**
Other soils having an E horizon showing stagnic properties at least in part of the horizon and abruptly overlying a slowly permeable horizon within 125 cm of the surface and lacking a natric or a spodic B horizon	**Planosols (PL)**
Other soils, exclusive of coarse textured materials (except when a histic H horizon is present), showing gleyic properties within 50 cm of the surface; having no diagnostic horizons other than an A horizon, a histic H horizon, a cambic B horizon, a sulfuric horizon, a calcic, or a gypsic horizon; lacking plinthite within 125 cm of the surface	**Gleysols (GL)**
Other soils having a natric B horizon	**Solonetz (SN)**
Other soils showing andic properties to a depth of 35 cm or more from the surface and having a mollic or an umbric A horizon possibly overlying a cambic B horizon or an ochric A horizon and a cambic B horizon; having no other diagnostic horizons	**Andosols (AN)**
Other soils having a mollic A horizon with a moist chroma of 2 or less to a depth of at least 15 cm, showing uncoated silt and sand grains on structural pedfaces; having an argic B horizon	**Greyzems (GR)**
Other soils having an argic B horizon showing an irregular or broken upper boundary resulting from deep tonguing of the A into the B horizon or from the formation of discrete nodules larger than 2 cm, the exteriors of which are enriched and weakly cemented or indurated and have redder hues and stronger chromas than the interiors	**Podzoluvisols (PD)**
Other soils having a mollic A horizon with a moist chroma of 2 or less to a depth of at least 15 cm; having a calcic or petrocalcic horizon, or concentrations of soft powdery lime within 125 cm of the surface, or both	**Chernozems (CH)**
Other soils having a gypsic or petrogypsic horizon within 125 cm of the surface; having no diagnostic horizons other than an ochric A horizon, a cambic B horizon or an argic B horizon permeated with gypsum or calcium carbonate, a calcic or petrocalcic horizon	**Gypsisols (GY)**
Other soils having a mollic A horizon with a moist chroma of more than 2 to a depth of at least 15 cm; having one or more of the following: a calcic, petrocalcic, or gypsic horizon or concentrations of soft powdery lime within 125 cm of the surface	**Kastanozems (KS)**
Other soils having a calcic or a petrocalcic horizon, or a concentration of soft powdery lime, within 125 cm of the surface; having no diagnostic horizons other than an ochric A horizon, a cambic B horizon, or an argic B horizon, which is calcareous	**Calcisols (CL)**
Other soils having a mollic A horizon; having a base saturation (by NH_4OAc) of ≥50% throughout the upper 125 cm of the soil	**Phaeozems (PH)**

TABLE 32.6 (continued) Key to MSGs of the 1988 *Revised Legend of the Soil Map of the World*

Other soils having an argic B horizon with a clay distribution, which does not show a relative decrease from its maximum of more than 20% within 150 cm of the surface; showing gradual to diffuse horizon boundaries between the A and B horizons; having nitic properties in some subhorizon within 125 cm of the surface	**Nitisols (NT)**
Other soils having an argic B horizon, which has a CEC equal to or more than 24 $cmol_c\ kg^{-1}$ clay and a base saturation (by NH_4OAc) of less than 50% in at least some part of the B horizon within 125 cm of the surface	**Alisols (AL)**
Other soils having an argic B horizon, which has a CEC of less than 24 $cmol_c\ kg^{-1}$ clay and a base saturation (by NH_4OAc) of less than 50% in at least some part of the B horizon within 125 cm of the surface	**Acrisols (AC)**
Other soils having an argic B horizon, which has a CEC equal to or more than 24 $cmol_c\ kg^{-1}$ clay and a base saturation (by NH_4OAc) of 50% or more throughout the B horizon within 125 cm of the surface	**Luvisols (LV)**
Other soils having an argic B horizon, which has a CEC of less than 24 $cmol_c\ kg^{-1}$ clay and a base saturation (by NH_4OAc) of 50% or more throughout the B horizon within 125 cm of the surface	**Lixisols (LX)**
Other soils having a cambic B horizon	**Cambisols (CM)**

Source: FAO–UNESCO. 1988. FAO–UNESCO soil map of the world, revised legend. World Soil Resources Report No. 60. Food and Agriculture Organization, Rome, Italy.

of the common denominator function that the WRB assumes. Its units have sufficient width to stimulate harmonization and correlation of existing national systems; (5) the reference base is not meant to substitute for national soil classification systems but rather to serve as a common denominator for communication at an international level; (6) in addition to serving as a link between existing classification systems, the WRB also serves as a consistent communication tool for compiling global soil databases and for the inventory and monitoring of the world's soil resources. The nomenclature used to distinguish soil groups retains terms that have been used traditionally or that can be introduced easily in current language.

32.4.5 The Architecture of the WRB

Currently, the WRB comprises two tiers of categorical detail: (1) the RSGs and (2) the combination of RSGs with qualifiers, detailing the properties of the RSGs by adding a set of uniquely defined qualifiers. From the *FAO Revised Legend*, through the 1998 WRB edition, several changes occurred. In 1998, one MSG has been omitted (Greyzems), and three new ones are introduced (Cryosols, Durisols, and Umbrisols). Greyzems were deleted as they constitute the smallest MSG and were amalgamated with the Phaeozems. Cryosols were newly introduced to identify a group of soils, which occur under the unique environmental conditions of thawing and freezing. Durisols have been added to group soils together, which have accumulation of secondary silica, analogous to the Calcisols and Gypsisols. Umbrisols constitute the group of soils that have a thick accumulation of desaturated organic matter at the surface and are the natural counterpart of Chernozems, Kastanozems, and Phaeozems.

The second edition of the *WRB* has undergone a major revision. Technosols and Stagnosols have been introduced, leading to 32 RSGs instead of 30. The Technosols are soils with a certain amount of artifacts, a constructed geomembrane or technic hard rock. The Stagnosols unify the former Epistagnic subunits of many other RSGs. Some rearrangement has taken place in the order of the key, with Anthrosols, Solonetz, Nitisols, and Arenosols moving upward. The definitions of many diagnostic soil horizons, soil properties, and materials have been adjusted. An overview of the diagnostic categories is presented in Table 32.7. The number of qualifiers almost doubled (currently 179), and a significant change was the subdivision of qualifiers into prefix and suffix ones. Prefix qualifiers comprise those that are typically associated with the RSG (in order of their importance) and the intergrades to other RSGs (in order of the key). All other qualifiers are listed as suffix qualifiers.

The RSGs are defined by the key. For each RSG number of possible prefix and suffix qualifiers are listed in priority order. After determining the RSG, the applying prefix qualifier names are put before the RSG name and applying suffix qualifier names are placed between brackets following the RSG name.

The current scheme of the WRB proved to be capable of indicating most of the soil's properties and performed properly for correlation purposes. However, recent applications for mapping purposes indicated that when generalization is required, important information may not show with the current set of the qualifiers. Although WRB was not primarily designed to serve mapping purposes, it is increasingly used for that. Therefore, an addendum has been developed to serve the need for small-scale mapping. The "Guidelines for constructing small-scale map legends using the World Reference Base for Soil Resources" (IUSS Working Group WRB, 2010) is available online at the WRB Web site.

32.4.6 The WRB Reference Soil Groups

The key to the RSGs in the WRB stems from the *Legend of the Soil Map of the World*. The history behind the key to the major soil units of the *Legend of the Soil Map of the World* reveals that it is mainly based on functionality; the key was conceived to derive the correct classification as efficiently as possible. The sequence of the major soil units was such that the central concept of the major soils would come out almost automatically by specifying

TABLE 32.7 An Overview of the Diagnostic Horizons, Properties, and Materials of the WRB

Surface horizons and subsurface diagnostic horizons at shallow depth	
Anthraquic horizon	An anthraquic horizon (from Greek anthropos, human, and Latin aqua, water) is a human-induced surface horizon that comprises a puddled layer and a plow pan
Anthric horizon	A moderately thick, dark-colored surface horizon that is the result of long-term cultivation (plowing, liming, fertilization, etc.)
Folic horizon	Surface horizon or subsurface horizon at shallow depth, consisting of well-aerated organic soil material
Fulvic horizon	Thick, black surface horizon having a low bulk density and high organic carbon content conditioned by short-range-order minerals (usually allophane) and/or organoaluminum complexes
Histic horizon	A surface horizon or a subsurface horizon occurring at shallow depth that consists of poorly aerated organic material
Hortic horizon	A human-induced mineral surface horizon that results from deep cultivation, intensive fertilization, and/or long-continued application of human and animal wastes and other organic residues
Hydragric horizon	A human-induced subsurface horizon associated with wet cultivation
Irragric horizon	Human-induced mineral surface horizon that builds up gradually through continuous application of irrigation water with substantial amounts of sediments
Melanic horizon	Thick, black surface horizon conditioned by short-range-order minerals (usually allophane) and/or organoaluminum complexes
Mollic horizon	Well-structured, dark surface horizons with high base saturation and moderate to high organic carbon content
Plaggic horizon	An organic matter—rich, thick, black or brown human-induced low base mineral surface horizon that has been produced by long-continued manuring
Takyric horizon	Finely textured surface horizon consisting of a dense surface crust and a platy lower part; formed under arid conditions in periodically flooded soils
Terric horizon	A human-induced mineral surface horizon that develops through addition of earthy manures, compost, beach sands, or mud over a long period of time
Umbric horizon	Well-structured, dark surface horizon with low base saturation and moderate to high organic matter content
Yermic horizon	Surface horizon of rock fragments ("desert pavement") usually, but not always, embedded in a vesicular crust and covered by a thin eolian sand or loess layer
Voronic horizon	Deep, well-structured, blackish surface horizon with a high base saturation, high organic matter content, strong biological activity, and well-developed, usually granular, structure. Its carbon content is intermediate between a mollic horizon and a histic horizon
Subsurface diagnostic horizons	
Albic horizon	Bleached eluviation horizon with the color of uncoated soil material, usually overlying an illuviation horizon
Argic horizon	Subsurface horizon having distinctly more clay than the overlying horizon as a result of illuvial accumulation of clay and/or pedogenetic formation of clay in the subsoil and/or destruction or selective erosion of clay in the surface soil
Cambic horizon	Genetically young subsurface horizon showing evidence of alteration relative to underlying horizons: modified color, removal of carbonates, or presence of soil structure
Cryic horizon	Perennially frozen horizon in mineral or organic soil materials
Calcic horizon	Horizon with distinct calcium carbonate enrichment
Duric horizon	Subsurface horizon with weakly cemented to indurated nodules cemented by silica (SiO_2) known as "durinodes"
Ferralic horizon	Strongly weathered horizon in which the clay fraction is dominated by low activity clays and the sand fraction by resistant materials such as iron-, aluminum-, manganese-, and titanium oxides
Ferric horizon	Subsurface horizon in which segregation of iron has taken place to the extent that large mottles or concretions have formed in a matrix that is largely depleted of iron
Fragic horizon	Dense, noncemented subsurface horizon that can only be penetrated by roots and water along natural cracks and streaks
Gypsic horizon	Horizon with distinct calcium sulfate enrichment
Natric horizon	Subsurface horizon with more clay than any overlying horizon(s) and high exchangeable sodium percentage; usually dense, with columnar or prismatic structure
Nitic horizon	Clay-rich subsurface horizon with a moderate to strong polyhedric or nutty structure with shiny pedfaces
Petrocalcic horizon	Continuous, cemented, or indurated calcic horizon
Petroduric horizon	Continuous subsurface horizon cemented mainly by secondary silica (SiO_2), also known as a "duripan"
Petrogypsic horizon	Cemented horizon containing secondary accumulations of gypsum ($CaSO_4 \cdot 2H_2O$)
Petroplinthic horizon	Continuous layer indurated by iron compounds and without more than traces of organic matter
Pisoplinthic horizon	A pisoplinthic horizon contains nodules that are strongly cemented or indurated with Fe (and in some cases also with Mn)

TABLE 32.7 (continued) An Overview of the Diagnostic Horizons, Properties, and Materials of the WRB

Plinthic horizon	Subsurface horizon consisting of an iron-rich, humus-poor mixture of kaolinitic clay with quartz and other constituents and which changes irreversibly to a hardpan or to irregular aggregates on exposure to repeated wetting and drying with free access of oxygen
Salic horizon	Surface or shallow subsurface horizon containing 1% of readily soluble salts or more
Sombric horizon	A dark-colored subsurface horizon containing illuvial humus that is neither associated with Al nor dispersed by Na
Spodic horizon	Dark-colored subsurface horizon with illuvial amorphous substances composed of organic matter and aluminum, with or without iron
Thionic horizon	An extremely acid subsurface horizon in which sulfuric acid is formed through oxidation of sulfides
Vertic horizon	Subsurface horizon rich in expanding clays and having polished and grooved ped surfaces ("slickensides") or wedge- shaped structural aggregates formed upon repeated swelling and shrinking
Diagnostic properties	
Abrupt textural change	Very sharp increase in clay content within a limited vertical distance
Albeluvic tonguing	Iron-depleted material penetrating into an argic horizon along ped surfaces
Andic properties	Result from moderate weathering of mainly pyroclastic deposits. The presence of short-range-order minerals and/or organometallic complexes is characteristic for andic properties
Aridic properties	Refer to soil material low in organic matter, with evidence of eolian activity, light in color and (virtually) base-saturated
Continuous rock	Continuous rock is consolidated material underlying the soil, exclusive of cemented pedogenetic horizons such as petrocalcic, petroduric, petrogypsic, and petroplinthic horizons
Ferralic properties	Indicate that the (mineral) soil material has a "low" CEC or would have qualified for a ferralic horizon if it had been less coarsely textured
Geric properties	Mark soil material of very low-effective CEC or even acting as anion exchanger
Gleyic color pattern	Visible evidence of prolonged waterlogging and reducing conditions by shallow groundwater
Lithological discontinuity	Significant changes in particle-size distribution or mineralogy that represent differences in lithology within a soil
Reducing conditions	Lack of oxygen due the saturation of moisture in some parts of the soil
Secondary carbonates	Significant quantities of translocated lime, soft enough to be readily cut with a finger nail, precipitated from the soil solution rather than being inherited from the soil parent material
Stagnic color pattern	Visible evidence of prolonged waterlogging and reducing conditions by a perched water table
Vertic properties	Vertic properties are characterized with slickensides, or wedge-shaped aggregates, or cracks due to high clay content
Vitric properties	Apply to layers with volcanic glass and other primary minerals derived from volcanic ejecta, which contain a limited amount of short-range-order minerals or organometallic complexes
Diagnostic materials	
Artifacts	Artifacts solid or liquid substances that are created or substantially modified or brought to the surface by human activity from a depth, where they were not influenced by surface processes
Calcaric soil material	Soil material that contains more than 2% calcium carbonate equivalent and shows strong effervescence with 10% HCl in most of the fine earth
Colluvic material	Formed by sedimentation through human-induced erosion
Fluvic soil material	Refers to fluviatile, marine, and lacustrine sediments that receive fresh material at regular intervals or have received it in the recent past
Gypsiric soil material	Mineral soil material, which contains 5% or more gypsum (by volume)
Limnic material	Organic and mineral materials that are deposited in water by precipitation or through action of aquatic organisms, such as diatoms and other algae
Mineral material	The soil properties are dominated by mineral components
Organic soil material	Organic debris that accumulates at the surface and in which the mineral component does not significantly influence soil properties
Ornithogenic material	Material with strong influence of bird excrement
Sulfidic soil material	Waterlogged deposit containing sulfur, mostly sulfides, and not more than moderate amounts of calcium carbonate
Technic hard rock	Consolidated material resulting from an industrial process, with properties substantially different from those of natural materials
Tephric soil material	Unconsolidated, non or only slightly weathered products of volcanic eruptions, with or without admixtures of material from other sources

Source: IUSS Working Group WRB. 2007. World reference base for soil resources 2006, first update 2007. World Soil Resources Reports No. 103. FAO, Rome, Italy. Available online at http://www.fao.org/nr/land/soils/soil/wrb-documents/en/

briefly a limited number of diagnostic horizons, properties, or materials. In the current WRB key, the RSGs are allocated to sets on the basis of dominant identifiers, that is, the soil-forming factors or processes that most clearly condition the soil formation. The sequencing of the groups is done according to the following principles:

1. First, organic soils key out to separate them from mineral soils (*Histosols*).
2. The second major distinction in the WRB is to recognize human activity as a soil-forming factor, hence the position of the *Anthrosols and Technosols* after the Histosols; it also appears logical to key out the newly introduced Technosols close to the beginning of the key, for the following reasons: One can almost immediately key out soils that should not be touched (toxic soils that should be handled by experts); a homogeneous group of soils in strange materials is obtained; politicians and decision-makers, who consult the key will immediately encounter these problematic soils.
3. Next are the soils with a severe limitation to rooting (*Cryosols and Leptosols*).
4. Then comes a group of RSGs that are or have been strongly influenced by water: *Vertisols*, *Fluvisols*, *Solonetz*, *Solonchaks*, and *Gleysols*.
5. The following set of soil groups are the RSGs in which iron (Fe) and/or aluminum (Al) chemistry plays a major role in their formation: *Andosols*, *Podzols*, *Plinthosols*, *Nitisols*, and *Ferralsols*.
6. Next comes the set of soils with perched water: *Planosols* and *Stagnosols*.
7. The next grouping comprises soils that occur predominantly in steppe regions and have humus-rich topsoils and a high base saturation: *Chernozems*, *Kastanozems*, and *Phaeozems*.
8. The next set comprises soils from the drier regions with accumulation of gypsum (*Gypsisols*), silica (*Durisols*), or calcium carbonate (*Calcisols*).
9. Then comes the soils with clay-rich subsoil: *Albeluvisols*, *Alisols*, *Acrisols*, *Luvisols*, and *Lixisols*.
10. Finally, relatively young soils or soils with very little or no profile development, or very homogenous sands, are grouped together: *Umbrisols*, *Arenosols*, *Cambisols*, and *Regosols*.

The key to the RSGs of the WRB is presented in Table 32.8. An example for the classification in the WRB system is presented in Figure 32.1.

The dark, cracking, clayey heavy soil satisfies the criteria of the mollic, the vertic, and the calcic horizons. Therefore, this soil will key out in the Vertisols reference group of the WRB. From the qualifiers, the ones indicated with bold in the figure apply. According to the rules of classification (IUSS Working Group WRB, 2007), the name of the soil is *Calcic Mollic Vertisol* (*Pellic*).

TABLE 32.8 The Key to the RSGs of the WRB

Soils having organic material, either

1. Ten centimeters or more thick starting at the soil surface and immediately overlying ice, continuous rock, or fragmental materials, the interstices of which are filled with organic material or
2. Cumulatively within 100 cm of the soil surface either 60 cm or more thick if 75% (by volume) or more of the material consists of moss fibers or 40 cm or more thick in other materials and starting within 40 cm of the soil surface

Histosols (HS)

Other soils having

1. Either a hortic, irragric, plaggic, or terric horizon 50 cm or more thick or
2. An anthraquic horizon and an underlying hydragric horizon with a combined thickness of 50 cm or more

Anthrosols (AT)

Other soils having

1. Twenty percent or more (by volume, by weighted average) artifacts in the upper 100 cm from the soil surface or to continuous rock or a cemented or indurated layer, whichever is shallower or
2. A continuous, very slowly permeable to impermeable, constructed geomembrane of any thickness starting within 100 cm of the soil surface or
3. Technic hard rock starting within 5 cm of the soil surface and covering 95% or more of the horizontal extent of the soil

Technosols (TC)

Other soils having

1. A cryic horizon starting within 100 cm of the soil surface or
2. A cryic horizon starting within 200 cm of the soil surface and evidence of cryoturbation in some layer within 100 cm of the soil surface

Cryosols (CR)

Other soils having

1. One of the following:
   a. Limitation of depth by continuous rock within 25 cm of the soil surface or
   b. Less than 20% (by volume) fine earth averaged over a depth of 75 cm from the soil surface or to continuous rock, whichever is shallower and
2. No calcic, gypsic, petrocalcic, petrogypsic, or spodic horizon

Leptosols (LP)

Other soils having

1. A vertic horizon starting within 100 cm of the soil surface
2. After the upper 20 cm have been mixed, 30% or more clay between the soil surface and the vertic horizon throughout and
3. Cracks that open and close periodically

Vertisols (VR)

Other soils having

1. Fluvic material starting within 25 cm of the soil surface and continuing to a depth of 50 cm or more or starting at the lower limit of a plow layer and continuing to a depth of 50 cm or more and
2. No argic, cambic, natric, petroplinthic, or plinthic horizon starting within 50 cm of the soil surface and
3. No layers with andic or vitric properties with a combined thickness of 30 cm or more within 100 cm of the soil surface and starting within 25 cm of the soil surface

Fluvisols (FL)

TABLE 32.8 (continued) The Key to the RSGs of the WRB

Other soils having a natric horizon starting within 100 cm of the soil surface

Solonetz (SN)

Other soils having

1. A salic horizon starting within 50 cm of the soil surface and
2. No thionic horizon starting within 50 cm of the soil surface

Solonchaks (SC)

Other soils having

1. Within 50 cm of the mineral soil surface a layer, 25 cm or more thick, that has reducing conditions in some parts and a gleyic color pattern throughout and
2. No layers with andic or vitric properties with a combined thickness of either
   a. Thirty centimeters or more within 100 cm of the soil surface and starting within 25 cm of the soil surface or
   b. Sixty percent or more of the entire thickness of the soil when continuous rock or a cemented or indurated layer is starting between 25 and 50 cm from the soil surface

Gleysols (GL)

Other soils having

1. One or more layers with andic or vitric properties with a combined thickness of either
   a. Thirty centimeters or more within 100 cm of the soil surface and starting within 25 cm of the soil surface or
   b. Sixty percent or more of the entire thickness of the soil when continuous rock or a cemented or indurated layer is starting between 25 and 50 cm from the soil surface and
2. No argic, ferralic, petroplinthic, pisoplinthic, plinthic, or spodic horizon (unless buried deeper than 50 cm)

Andosols (AN)

Other soils having a spodic horizon starting within 200 cm of the mineral soil surface

Podzols (PZ)

Other soils having either

1. A plinthic, petroplinthic, or pisoplinthic horizon starting within 50 cm of the soil surface or
2. A plinthic horizon starting within 100 cm of the soil surface and, directly above, a layer 10 cm or more thick, that has in some parts reducing conditions for some time during the year and in half or more of the soil volume, single or in combination
   a. A stagnic color pattern or
   b. An albic horizon

Plinthosols (PL)

Other soils having

1. A nitic horizon starting within 100 cm of the soil surface
2. Gradual to diffuse horizon boundaries between the soil surface and the nitic horizon
3. No ferric, petroplinthic, pisoplinthic, plinthic, or vertic horizon starting within 100 cm of the soil surface and
4. No gleyic or stagnic color pattern starting within 100 cm of the soil surface

Nitisols (NT)

TABLE 32.8 (continued) The Key to the RSGs of the WRB

Other soils having

1. A ferralic horizon starting within 150 cm of the soil surface
2. No argic horizon that has, in the upper 30 cm, 10% or more water-dispersible clay unless the upper 30 cm of the argic horizon has one or both of the following:
   a. Geric properties or
   b. 1.4 % or more organic carbon

Ferralsols (FR)

Other soils having

1. An abrupt textural change within 100 cm of the soil surface and, directly above or below, a layer 5 cm or more thick that has in some parts reducing conditions for some time during the year and in half or more of the soil volume, single or in combination
   a. A stagnic color pattern or
   b. An albic horizon and
2. No albeluvic tonguing starting within 100 cm of the soil surface

Planosols (PL)

Other soils having

1. Within 50 cm of the mineral soil surface in some parts reducing conditions for some time during the year and in half or more of the soil volume, single or in combination
   a. A stagnic color pattern or
   b. An albic horizon and
2. No albeluvic tonguing starting within 100 cm of the soil surface

Stagnosols (ST)

Other soils having

1. A mollic horizon
2. A Munsell chroma, moist, of 2 or less from the soil surface to a depth of 20 cm or more or having this chroma directly below any plow layer that is 20 cm or more deep
3. A calcic horizon or concentrations of secondary carbonates starting within 50 cm below the lower limit of the mollic horizon and, if present, above a cemented or indurated layer and
4. A base saturation (by 1 M NH_4OAc) of 50% or more from the soil surface to the calcic horizon or the concentrations of secondary carbonates throughout

Chernozems (CH)

Other soils having

1. A mollic horizon
2. A calcic horizon or concentrations of secondary carbonates starting within 50 cm below the lower limit of the mollic horizon and, if present, above a cemented or indurated layer and
3. A base saturation (by 1 M NH_4OAc) of 50% or more from the soil surface to the calcic horizon or the concentrations of secondary carbonates throughout

Kastanozems (KS)

Other soils having

1. A mollic horizon and
2. A base saturation (by 1 M NH_4OAc) of 50% or more throughout to a depth of 100 cm or more from the soil surface or to continuous rock or a cemented or indurated layer, whichever is shallower

Phaeozems (PH)

(*continued*)

TABLE 32.8 (continued) The Key to the RSGs of the WRB

Other soils having

1. A petrogypsic horizon starting within 100 cm of the soil surface or
2. A gypsic horizon starting within 100 cm of the soil surface and no argic horizon unless the argic horizon is permeated with gypsum or calcium carbonate

Gypsisols (GY)

Other soils having a petroduric or duric horizon starting within 100 cm of the soil surface

Durisols (DU)

Other soils having

1. A petrocalcic horizon starting within 100 cm of the soil surface or
2. A calcic horizon starting within 100 cm of the soil surface and no argic horizon unless the argic horizon is permeated with calcium carbonate

Calcisols (CL)

Other soils having an argic horizon starting within 100 cm of the soil surface with albeluvic tonguing at its upper boundary

Albeluvisols (AB)

Other soils having

1. An argic horizon, which has a CEC (by 1 M NH_4OAc) of 24 $cmol_c$ kg^{-1} clay or more throughout or to a depth of 50 cm below its upper limit, whichever is shallower, either starting within 100 cm of the soil surface or within 200 cm of the soil surface if the argic horizon is overlain by loamy sand or coarser textures throughout and
2. A base saturation (by 1 M NH_4OAc) of less than 50% in the major part between 50 and 100 cm

Alisols (AL)

Other soils having

1. An argic horizon that has a CEC (by 1 M NH_4OAc) of less than 24 $cmol_c$ kg^{-1} clay in some part to a maximum depth of 50 cm below its upper limit, either starting within 100 cm of the soil surface or within 200 cm of the soil surface if the argic horizon is overlain by loamy sand or coarser textures throughout, and
2. A base saturation (by 1 M NH_4OAc) of less than 50% in the major part between 50 and 100 cm

Acrisols (AC)

Other soils having an argic horizon with a CEC (by 1 M NH_4OAc) of 24 $cmol_c$ kg^{-1} clay or more throughout or to a depth of 50 cm below its upper limit, whichever is shallower, either starting within 100 cm of the soil surface or within 200 cm of the soil surface if the argic horizon is overlain by loamy sand or coarser textures throughout

Luvisols (LV)

Other soils having an argic horizon, either starting within 100 cm of the soil surface or within 200 cm of the soil surface if the argic horizon is overlain by loamy sand or coarser textures throughout

Lixisols (LX)

Other soils having an umbric or mollic horizon

Umbrisols (UM)

TABLE 32.8 (continued) The Key to the RSGs of the WRB

Other soils having

1. A weighted average texture of loamy sand or coarser, if cumulative layers of finer texture are less than 15 cm thick, either to a depth of 100 cm from the soil surface or to a petroplinthic, pisoplinthic, plinthic, or salic horizon starting between 50 and 100 cm from the soil surface
2. Less than 40% (by volume) of gravels or coarser fragments in all layers within 100 cm of the soil surface or to a petroplinthic, pisoplinthic, plinthic, or salic horizon starting between 50 and 100 cm from the soil surface
3. No fragic, irragric, hortic, plaggic, or terric horizon and
4. No layers with andic or vitric properties with a combined thickness of 15 cm or more

Arenosols (AR)

Other soils having

1. A cambic horizon starting within 50 cm of the soil surface and having its base 25 cm or more below the soil surface or 15 cm or more below any plow layer
2. An anthraquic, hortic, hydragric, irragric, plaggic, or terric horizon
3. A fragic, petroplinthic, pisoplinthic, plinthic, salic, thionic, or vertic horizon starting within 100 cm of the soil surface or
4. One or more layers with andic or vitric properties with a combined thickness of 15 cm or more within 100 cm of the soil surface

Cambisols (CM)

Other soils

Regosols (RG)

Source: IUSS Working Group WRB. 2007. World reference base for soil resources 2006, first update 2007. World Soil Resources Reports No. 103. FAO, Rome, Italy. Available online at http://www.fao.org/nr/land/soils/soil/wrb-documents/en/

32.5 The French Systems of Soil Classification

32.5.1 Introduction

The *Commission de Pédologie et de Cartographie des Sols* (CPCS, 1967) issued the French soil classification, building on previous work published by Aubert and Duchaufour (1956), which has been the basis for many soil surveys during the 1970s and 1980s, not only in France but also in many of the former French colonies, notably in Africa. It was replaced by the pedological reference base (PRB; *Référentiel Pédologique Français*) (AFES–INRA, 1990, 1992).

32.5.2 The 1967 CPCS System

The CPCS soil classification system comprises four main levels: the class (*classe*), the subclass (*sous-classe*), the group (*groupe*), and the subgroup (*sous-groupe*); followed by four minor levels: the family (*famille*), the series (*série*), the type (*type*), and the phase (*phase*). However, due to the limited knowledge at the time of design of the system, the four minor levels have not been developed for all classes.

The class comprises soils that have main characteristics in common, such as a certain degree of profile development, weathering mode, composition and distribution of organic matter, and

Diagnostics of the example profile:

Dark, high base surface horizon with sufficient structure and organic carbon to satisfy the *mollic horizon*

Vertic subsurface horizon with slickensides and wedge-shaped aggregates and deep, wide cracks

Calcic horizon with sufficient amount (≥15%) of secondary $CaCO_3$

Key to the RSGs (according to Table 32.8, Five RSGs Are Not Satisfied before Getting to the Vertisols in the Key)	**Prefix Qualifiers**	**Suffix Qualifiers**
Other soils having 1. A *vertic* horizon starting within 100 cm of the soil surface 2. After the upper 20 cm have been mixed, 30 percent or more clay between the soil surface and the *vertic* horizon throughout 3. Cracks that open and close periodically Vertisols (VR)	Grumic Mazic Technic Endoleptic Salic Gleyic Sodic Stagnic Mollic Gypsic Duric Calcic Haplic	Thionic Albic Manganiferric Ferric Gypsiric Calcaric Humic Hyposalic Hyposodic Mesotrophic Hypereutric Pellic* Chromic Novic

*Pellic: refers to very dark (Munsell value, moist ≤3.5 and a chroma, moist ≤1.5) surface horizon

Full classification and name: *Calcic Mollic Vertisol* (*Pellic*)
(The applying prefix qualifier names are put before the RSG name and applying suffix qualifier names are placed between brackets following the RSG name).

FIGURE 32.1 Example for the classification in the WRB (IUSS WG WRB, 2007).

predominant soil-forming factors (e.g., wetness). The subclass differentiation is related to criteria resulting from climatic factors, which influence, among others, the pedoclimate. The groups are defined according to morphological characteristics corresponding to soil development, while the subgroup is differentiated either on degree of intensity of the fundamental evolutionary characteristics or on the presence of important secondary soil-forming processes.

In total, 12 classes are distinguished, namely, nondeveloped mineral soils (sols minéraux bruts), slightly developed soils (sols peu évolués), Vertisols, Andosols, Ca/Mg-saturated soils (sols calcimagnésiques), humus-rich soils (sols isohumiques), brunified soils (sols brunifiés), podzolized soils (sols podzolisés), soils rich in Fe (sols sesquioxides de fer), ferralitic soils, hydromorphic soils, and sodic soils.

Nondeveloped mineral soils (*sols minéraux bruts*) show very little trace of soil development apart from some accumulation of organic matter at the surface. They are characterized by an (A)C, (A)R, or R horizon sequence. Included in this class are eroded soils (Lithosols, Régosols), alluvial, colluvial and eolian accumulations, volcanic deposits, manmade soils, nondeveloped soils in the arctic regions (Cryosols), and nondeveloped desert soils.

The class of *slightly developed (sols peu évolués)* soils has higher organic matter content than the nondeveloped soils and is characterized by an AC or AR horizon sequence. No B horizon is permitted in this class. Included are soils from the arctic regions, with or without permafrost, soils with high organic matter content directly overlying hard rock (rankers, soils over limestone, and slightly weathered soils on volcanic ashes), slightly developed desert soils, and soils resulting from erosion and deposition.

Vertisols are described as clayey soils, which are homogenized or irregularly differentiated as a result of internal movement and which are dominated by swell/shrink clays. The normal horizon sequence is A(B)C, A(B)gC, or A(B)Cg. Subdivision into subclasses and groups is based on external drainage factors and type of structure in the surface horizon (rounded or angular).

Andosols are defined as soils in which the mineral fraction is dominated by poorly crystalline minerals and/or metal humus complexes, associated with variable, but usually high amounts of organic matter.

The *Ca/Mg-saturated soils* (*sols calcimagnésiques*) have an exchange complex, 90% of which is saturated with Ca and/or Mg, and have a pH above 6.8. They are generally associated with calcareous or basic rocks, and have an AR, AC, A(B)R, or A(B)C horizon sequence. They are subdivided into (1) carbonate-rich soils (rendzinas, cryptorendzinas, and brown calcareous soils), (2) saturated soils (soils containing only traces of primary $CaCO_3$ in the fine earth fraction), and (3) gypsiferous soils.

The class of *humus-rich soils* (*sols isohumiques*) comprises soils characterized by a moderate to high accumulation of well-humified, polymerized organic matter. Normally, base saturation is high with Ca as dominant cation, followed by Mg and, sometimes, Na. If the base saturation is only moderate (50%–80%) in the upper part of the soil, it increases with depth. Profile evolution is slight to moderate, with an A(B)C or ABC, rarely AC horizon sequence. Subdivision of these soils (Brunizems, Chernozems, Chestnut soils, Brown soils Sierozems), is based at subclass level on pedoclimatic characteristics.

Brunified (*sols brunifiés*) soils are well-developed soils with an A(B)C or ABC horizon sequence and are characterized by the presence of dominantly mull-type humus. They may have a structural or textural B horizon. Subdivision is based on climate (humid temperate, continental temperate, boreal, or tropical) and the morphology of the profile, giving rise to the groups of brown soils (with a structural B horizon), eluvial soils (with a textural B horizon), gray wooded soils, dernopodzolic soils, eluvial boreal soils, and eutrophic brown tropical soils.

The class of *podzolic soils* (*sols podzolisés*) is characterized by the processes of alteration and destruction of the silicate minerals by fulvic acids and complexation of liberated Fe and Al. These processes result morphologically in a strongly depleted and light-colored eluvial horizon and an illuvial horizon, which has a higher organic matter content than the eluvial horizon and a sesquioxide content, which is higher than the original material. Division at subclass level is based on climatic or pedoclimatic characteristics.

Soils rich in Fe (*sols sesquioxydes de fer*) have an ABC or A(B)C profile characterized by the presence of Fe and/or Mn (hydr)oxides giving the soils characteristic red, yellow, rusty brown, or even black (in the case of Mn) colors, an SiO_2/Al_2O_3 ratio of >2, a base saturation of >50%, and a low amount of organic matter. Two subclasses are recognized, one in which the role of sesquioxides is dominant (ferruginous tropical soils) and another in which the behavior of the clay fraction predominates (fersiallitic soils).

The ferralitic soils are characterized by complete weathering of primary minerals, residual enrichment of resistant minerals (quartz, rutile, etc.), loss of nutrients, and the presence of neoformations such as kaolinite, gibbsite, goethite, hematite, etc. Subclasses are distinguished on the degree of leaching.

The *class of hydromorphic soils* comprises both organic and mineral soils. The three subclasses are separated on organic matter content. The first subclass (hydromorphic organic soils) is defined as having >30% organic matter over a depth of at least 40 cm if the mineral component is clayey or >20% if the mineral component is sandy. The second subclass (moderately organic hydromorphic soils) has between 8% and 30% organic matter over at least 20 cm depth; while the third subclass, the mineral soils (weakly organic hydromorphic soils) have <8% organic matter over a depth of at least 20 cm. The organic hydromorphic soils are at group level separated on decomposition rate (weak or fibric, moderate or hemic, or strong or sapric), while the two other subclasses are divided on the character of hydromorphism (gley or stagnogley) and accumulation of Fe, $CaCO_3$, or gypsum.

The sodic soils comprise both soils that have a high amount of soluble salts as well as soils with a high exchangeable sodium percentage (ESP). This difference is used to separate the subclasses into sodic soils with a nondegraded structure, comprising the saline soils (solonchak) with an AC horizon sequence, and sodic soils with a degraded structure, having an A(B)C or ABC horizon sequence and comprising alkaline saline soils (high ESP), sodic soils with a textural B horizon (solonetz), and so-called solodized solonetz, which are acid at the surface.

32.5.3 The Pedological Reference Base (*Référentiel Pédologique Français*)

This differs basically from the older CPCS system, which it replaces, in that it is being presented as a reference system, not a hierarchical classification (AFES–INRA, 1992). It considers the soil mantles as objects of study for which three sets of data are required: (1) the composition of the soil mantle (mineral, organic, etc.), (2) the internal arrangement of the individual constituents (e.g., structure), and (3) the soil dynamics (e.g., evolution over time).

At the highest level, the PRB recognizes the pedological system, which comprises several associated horizons grouped in a 3D space pattern. A horizon is defined as a part of the soil mantle, which can be considered homogenous. Because dimensions of horizons and pedological systems are not infinite, vertically and laterally, they merge into other systems (e.g., bedrock or other pedological systems).

The authors of the PRB have tried to design a system, which, at the same time, is both scientific and practical as well as precise but flexible. An example of this is the depth indications used in the descriptions of the horizons; the PRB starts off with the tolerance limits concerning depth or thickness requirements, for example, 10 cm must be considered as 5–15 cm, 40 cm means 30–50 cm, etc. Therefore, only two categories are distinguished—the references and the types, the latter being indicated by one or more qualifiers. The reference horizons form the basis of the system. So far, 50 have been proposed (AFES–INRA, 1992), defined, and described by several of the following: (1) morphological characteristics (constituents, pedological features, etc.), (2) analytical data, (3) pedogenetic significance, (4) major possible variations of the characteristics, and (5) most common positions within soil mantles. A succession or combination of reference horizons identifies a diagnostic solum and permits the association of such a solum with a reference. The PRB has proposed some 90 references (Table 32.9). Several of

TABLE 32.9 Major Groupings of References (MGR) and References of the AFES–INRA Pedological Reference Base

MGR	Reference	Brief Description
Alocrisols	Typic Alocrisols	Very acid, brown or yellow soils with a high amount of exchangeable aluminum (2–8 $cmol_c\ kg^{-1}$ fine earth and Al saturation of 20%–50%)
	Humic Alocrisols	Very acid, brown or yellow soils with a thick, dark colored surface horizon high in organic matter and a high amount of exchangeable aluminum
Alu-Andisols	Humic Alu-Andisols	Non-allophanic soils in weathered volcanic deposits having a thick surface horizon rich in organic matter
	Typic Alu-Andisols	Soils in volcanic deposits or strongly weathered ferralitic material having an aluminum-rich surface horizon and an allophane-dominated subsurface horizon
Andosols	Humic Andosols	Allophane-rich soils having a thick surface horizon rich in organic matter
	Eutric Andosols	Allophane-rich soils having a surface horizon with base saturation > 50%
	Dystric Andosols	Allophane-rich soils having a surface horizon with base saturation < 50%
	Perhydrated Andosols	Allophane-rich soils having a high irreversible water content
Anthroposols	Transformed Anthroposols	Soils modified by intensive or long-continued human activities
	Artificial Anthroposols	Man-made soils consisting of non-soil material (mine refuse, urban debris, etc.)
	Reshaped Anthrosols	Man-made soils consisting of transported soil material
Arenosols	Arenosols	Deep (>120 cm) sandy soils
Brunisols	Saturated Brunisols	Non-calcareous soils with a structural B horizon and 80%–100% base saturation
	Meso-saturated Brunisols	Non-calcareous soils with a structural B horizon and 50%–80% base saturation
	Oligo-saturated Brunisols	Non-calcareous soils with a structural B horizon and 20%–50% base saturation
	Resaturated Brunisols	Non-calcareous soils with a structural B horizon and >80% base saturation due to cultivation
Calcarisols	Calcarisols	Soils with a calcic horizon at least 10 cm thick, starting within 20 cm depth
Calcisols	Calcisols	Soils with non-calcareous, base-saturated (mainly Ca^{2+}) A and B horizons
Calcosols	Calcosols	Soils with calcareous A and B horizons ($CaCO_3$ > 5%)
Castanosols	Castanosols	Soils with a moderately thick to thick, dark colored, base-saturated surface horizon rich in organic matter
Chernosols	Chernosols	Soils with a thick, very dark colored, base-saturated surface horizon rich in organic matter
Colluviosols	Colluviosols	Soils in colluvial deposits
Cryosols	Histic cryosols	Soils with permafrost within 1 m depth and a histic surface horizon
	Mineral cryosols	Soils with permafrost within 2 m depth lacking a histic surface horizon
Dolomitosols	Dolomitosols	Soils with dolomitic A and B horizons (molar ratio of $CaCO_3/MgCO_3$ < 1.5)
Ferrallisols	Soft Ferrallisols	Strongly weathered soils with a ferralitic or oxidic mineralogy
	Nodular Ferrallisols	Strongly weathered soils with a high amount of sesquioxide nodules
	Petroxydic Ferrallisols	Strongly weathered soils with indurated sesquioxide layers (e.g., cuirasses)
Fersialsols	Carbonated Fersialsols	Calcareous soils with significant amounts of 2:1 clays and "free iron"
	Saturated Fersialsols	Base-saturated soils with significant amounts of 2:1 clays and "free Fe"
	Desaturated Fersialsols	Desaturated soils with significant amounts of 2:1 clays and "free Fe"
	Xanthic Fersialsols	Yellowish soils with significant amounts of 2:1 clays and "free Fe"
Fluviosols and thalassosols	Raw Fluviosols	Soils in fluviatile deposits lacking any horizon development
	Typical Fluviosols	Soils in fluviatile deposits with one or more not fully developed reference horizons
	Brunified Fluviosols	Soils in fluviatile deposits with a well-developed structural B horizon
	Thalassosols	Non- or weakly developed soils in marine or fluvio-marine deposits
Gypsosols	Gypsosols	Soils with accumulation of gypsum
Histosols	Leptic Histosols	Shallow organic soils with consolidated or unconsolidated rock within 40 cm
	Fibric Histosols	Organic soils with weakly decomposed organic material more than 60 cm thick
	Mesic Histosols	Organic soils with moderately decomposed organic material more than 40 cm thick
	Sapric Histosols	Organic soils with strongly decomposed organic material more than 40 cm thick
	Composite Histosols	Organic soils without dominance of either fibric, mesic, or sapric materials between 40 and 120 cm
	Covered Histosols	Organic soils with a cover of mineral soil material 10–40 cm thick
	Floating Histosols	Organic soils on water occurring between 40 and 160 cm depth
Lithosols	Lithosols	Shallow soils (<10 cm) over continuous hard rock or indurated layer
Luvisols	Neoluvisols	Soils with a moderately developed eluvial and well developed textural B horizon
	Typic Luvisols	Soils with a well-developed eluvial and textural B horizon
	Degraded Luvisols	Soils with a well-developed, partially light colored and hydromorphic eluvial horizon penetrating a gleyed textural B horizon

(*continued*)

TABLE 32.9 (continued) Major Groupings of References (MGR) and References of the AFES–INRA Pedological Reference Base

MGR	Reference	Brief Description
	Dernic Luvisols	Soils with a well-developed, partially light colored eluvial horizon penetrating a textural B horizon
	Truncated Luvisols	Soils with a textural B horizon but lacking an eluvial horizon
Magnesisols	Magnesisols	Soils with non-calcareous, base-saturated ($Ca^{2+}/Mg^{2+} < 2$) A and B horizons
Organosols	Calcaric Organosols	Well-drained, organic matter rich (>8% organic C), calcareous soils directly overlying an unconsolidated or consolidated substratum
	Saturated Organosols	Well-drained, organic matter rich (>8% organic C), base-saturated ($Ca^{2+}/Mg^{2+} > 5$) soils directly overlying an unconsolidated or consolidated substratum
	Undersaturated Organosols	Well-drained, organic matter rich (>8% organic C), undersaturated (BS < 80%) soils directly overlying an unconsolidated or consolidated substratum
	Tangelic Organosols	Well-drained, organic matter rich (>8% organic C), base-saturated soils with a thick, greasy horizon consisting of soil animal casts ("tangel horizon")
Pelosols	Typic Pelosols	Clay-rich, slightly weathered soils lacking coloration in the B horizon
	Brunified Pelosols	Clay-rich, slightly weathered soils with a brown colored B horizon
	Differentiated Pelosols	Clay-rich, slightly weathered soils with a clear eluvial horizon
Peyrosols	Stony Peyrosols	Soils which have throughout the upper 50 cm 40% or more stones plus 20% or more other coarse fragments
	Gravelly Peyrosols	Soils which have throughout the upper 50 cm 60% or more gravel, stones and boulders, but less than 40% stones
Planosols	Typic Planosols	Soils with abrupt textural change and temporary perched watertable within 50 cm
	Distal Planosols	Soils with abrupt textural change and temporary perched watertable below 50 cm
	Structural Planosols	Soils with a temporary perched watertable within 50 cm of the surface caused an impermeable layer which is not texturally induced (e.g., fragipan, duripan)
Podzosols	Duric Podzosols	Soils with an eluvial horizon and a cemented podzol B horizon
	Humo-Duric Podzosols	Soils with an indurated podzol B horizon, lacking an eluvial horizon
	Soft Podzosols	Soils with an eluvial horizon and a soft podzol B horizon
	Placic Podzosols	Soils with podzol B horizon and a placic horizon
	Ochric Podzosols	Soils with a weakly developed humic podzol B horizon, lacking an eluvial horizon
	Humic Podzosols	Soils with a soft humic podzol B horizon, lacking an eluvial horizon
	Post_Podzosols	Man-modified soils in which (remnants of) the podzol B horizon can be recognized
	Eluvial Podzosols	Soils lacking a podzol B horizon, but having lateral linkage to a podzol B horizon
Rankosols	Rankosols	Soils with a moderately thick A horizon with non-calcareous coarse fragments overlying consolidated or unconsolidated rock
Reductisols and redoxisols	Typic Reductisols	Hydromorphic soils conditioned by saturation of fluctuating groundwater table
	Stagnic Reductisols	Hydromorphic soils conditioned by a perched water table
	Duplex Reductisols	Hydromorphic soils conditioned by groundwater and stagnating surface water
	Redoxisols	Soils with a textural discontinuity and a perched water table
Regosols	Regosols	Shallow soils (<10 cm) over unconsolidated material or only slightly coherent rock
Rendisols	Rendisols	Soils with non-calcareous, base-saturated (mainly Ca^{2+}) A horizon over consolidated or unconsolidated calcareous rock
Rendosols	Rendosols	Soils with deep (>30 cm) calcareous A horizon ($CaCO_3$ > 5%) over consolidated or unconsolidated calcareous rock
Salisols	Chloridi-Sulfatic Salisols	Neutral soils with a high amount of sodium, magnesium or calcium salts
	Carbonatic Salisols	Alkaline soils with a high amount of carbonate/bicarbonates
Sodisols	Undifferentiated Sodisols	Alkaline soils with a high amount of exchangeable sodium
	Solonetzic Sodisols	Soils with clay illuviation and moderate leaching of sodium in the upper part
	Solodic Sodisols	Soils with clay illuviation and strong leaching of sodium in the upper part
Sulfatosols	Sulfatosols	Very acid soils with jarosite within 50 cm depth
Thiosols	Thiosols	Waterlogged soils containing sulfide minerals, rapidly acidifying upon aeration
Veracrisols	Veracrisols	Soils with a thick (50–150 cm), acid, dark colored surface horizon with a high biological activity overlying a slowly permeable horizon (e.g., textural B horizon)
Vertisols	Topovertisols	Deep, clayey soils in level, low-lying positions which crack and show gilgai microrelief, slick
Vitrandisols	Vitrandisols	Soils in slightly weathered pyroclastic material
Yermosols	Yermosols	Hot desert soils

Source: AFES–INRA (Association Française pour l'Etude du Sol) (Institut National de la Recherche Agronomiqie). 1992. Référentiel Pédologique, principaux sols d'Europe. INRA, Paris, France.

these are closely associated with each other because, for example, they may have the same reference horizons. Such references are described together as major groupings of references to avoid duplication and to associate the references with traditional pedological concepts. For example, the Podzosols major grouping of references comprises seven references characterized by a process of podzolization. However, major groupings of references do not form part of the PRB.

32.6 The Soil Classification System of Russia

32.6.1 The Soil Classification System of the Former USSR

32.6.1.1 Introduction

This system was available for soil survey from 1967 as a technical document. An official revised version was published in Russian in 1977 and in English in 1986 (VASKhNIL, 1986). This classification serves as a manual for soil examination and survey in many countries of the former USSR, for governmental assessment of land records and evaluation, as well as a reference book on agricultural and industrial planning for agronomists, land use planners, reclamation specialists, and others. It was amended (Shishov and Sokolov, 1990) to correct obvious errors and to integrate new knowledge and data. The taxonomic levels of the classification system are expanded and new names have been introduced for a number of soils.

32.6.1.2 Structure of the System

The soils are not defined as sequences of diagnostic horizons with fixed frames of properties, like in many other systems. Every soil type has a description of a "central concept," a typical profile. Profound knowledge of the soil-forming processes is required in order to recognize and classify the soils. The higher levels are known as types, subtypes, genera, and species. At subtype level, apart from subtypes with overlapping soil-forming process, the facies modifier may be added to indicate the thermal regime. Twenty-seven of these facies are recognized, ranging from arctic permafrost to subtropical hot nonfreezing.

32.6.1.3 Brief Description of the Types

Some 71 soil types are distinguished at the highest level and are characterized as follows:

Podzolic soils are characterized by either downward movement of organic acids facilitating decomposition of primary and secondary minerals and removal of weathering products or downward movement of clay particles.

Bog-podzolic soils have a combination of downward movement of organic acids or clay particles with either stagnating water conditions in the upper part of the soil or groundwater affecting the bottom part, resulting in a bleached eluvial horizon and hydromorphic properties in both the eluvial and illuvial horizons. Some of these soils have a peaty layer at the surface.

Sod-calcareous soils have a dark-colored, base-saturated, and humus-rich surface horizon overlying calcareous parent material.

Sod-gley soils are poorly drained with a dark, humus-rich surface horizon and hydromorphic features at shallow depth.

Gray forest soils have a dark-colored, humus-rich surface horizon of variable thickness with a bleached horizon or conspicuous white powdery spots overlying a clay-illuviated subsurface horizon.

Gley gray forest soils are similar to those above but with distinct hydromorphic features below the surface horizon.

Brown forest soils (*or Burozems*) have well-developed nut-like structure and intensive brownish and yellow color in B horizon combined with certain enrichment with organic matter in the surface horizon(s).

Gley brown forest soils (*or Gley burozems*) are similar to those above except for clear hydromorphic features below the surface horizon.

Podzolic brown forest soils (*or Podzolic burozems*) have a clearly developed eluvial horizon, clay-illuvial horizon, and weak hydromorphism due to seasonal surface waterlogging.

Gley podzolic brown forest soils (*or Gley podzolic burozems*) are similar to those above, but with seasonal wetness more pronounced. The process of acidic hydrolysis may take place in the upper part of these soils.

Bleached meadow soils (*Podbels*) are seasonally waterlogged with a bleached horizon near the surface, in which segregation of Fe in concretions is the main ongoing process.

Meadow chernozem-like soils have a thick, dark-colored, and humus-rich surface horizon and distinct features of hydromorphism (gray and rusty colors, Fe/Mn concretions, white powdery coatings) in the lower part of the soils, with no evidence of secondary carbonates.

Chernozem-like dark meadow soils are waterlogged with a peaty or mucky surface horizon overlying a dark-colored, humus-rich mineral horizon with rust colored mottles.

Chernozems are well-drained, base-saturated, or only slightly undersaturated soils with a thick, dark-colored surface horizon rich in organic matter and accumulation of illuvial secondary carbonates.

Meadow-chernozem soils are similar to those above, but with some features of wetness in the lower part of the solum.

Chestnut soils are well-drained and base-saturated with a dark-colored surface horizon, which is less thick and less rich in organic matter than in chernozems. The lower part of the solum often contains accumulations of calcium carbonate and/or gypsum.

Meadow chestnut soils are similar as the soils above, but with distinct features of wetness in the lower part of the solum.

Meadow soils are conditioned by a brief period of surface waterlogging and a longer period of saturation by groundwater, resulting in humus-rich surface horizons overlying a gleyed subsoil.

Semidesert brown soils have an accumulation of calcium carbonate, possibly overlying accumulations of gypsum and a crusty surface horizon.

Semidesert meadow brown soils are similar to those above, but with a higher organic matter content, weak signs of hydromorphism, and deeper $CaCO_3$ accumulations.

Desert gray brown soils are calcareous with a low organic matter content and a variable degree of salinization.

Desert takyr-like soils are weakly developed with a friable porous surface crust.

Takyrs have a hard, porous but crusted surface horizon cracking into polygonal patterns.

Desert sandy soils are coarse textured with little horizon differentiation apart from some accumulation of organic matter and segregation of calcium carbonate at depth.

Meadow desert soils are poorly differentiated, characterized by enrichment with organic matter at the surface and signs of hydromorphism in the subsoil.

Serozems have a shallow gray humus-enriched horizon and a calcareous illuvial layer in the subsoil.

Meadow-serozem soils are similar to those above, but signs of wetness occur in the deeper subsoil.

Semidesert and desert meadow soils have periodic or permanent wetness through capillary rise reaching the surface, giving rise to a well-developed, humus-rich surface horizon and a gleyed subsoil, in which carbonate concentrations are linked to the groundwater level.

Irrigated soils include a variety of types, which are all related to the original soil (irrigated serozems, irrigated brown soils, irrigated meadow brown soils, irrigated gray brown soils, irrigated takyr-like soils, irrigated meadow desert soils, irrigated meadow soils, and irrigated bog soils), in which irrigation has caused considerable modification including enhancement of biological activity, leaching, accumulation of sediments from irrigation water, enrichment in carbonates and soluble salts, etc.

Gray-cinnamon brown soils have a well-developed surface horizon with a low amount of organic matter and a clay-enriched subsurface horizon. The soils are calcareous throughout and differ from chestnut soils in that they do not freeze during wintertime.

Meadow gray-cinnamon brown soils are similar to those above, but with clear indications of increased wetness in the subsoil.

Cinnamon brown soils are deep with a high amount of organic matter, a clay-enriched subsurface horizon with a characteristic cinnamon brown color, and secondary carbonates accumulation.

Meadow cinnamon brown soils are similar to those above, but with clear indications of increased wetness in the subsoil.

Gray meadow forest soils have a thick, humus-enriched surface horizon and hydromorphic features starting at or near to the surface.

Zheltozems are leached subtropical soils with no or only weak textural differentiation and rich in sesquioxides.

Gley zheltozems are similar to those above, but with pronounced gleying throughout the profile.

Podzolic–zheltozem soils are leached subtropical soils with clear textural differentiation and a high content in sesquioxides. Gley features are common in the transition zone between the eluvial and illuvial horizons.

Podzolic–zheltozem gley soils are similar to those above, but with pronounced gleying throughout the profile, resulting in the accumulation of Fe in the illuvial horizon.

Krasnozems are strongly weathered subtropical soils with a high amount of sesquioxides. The clay fraction mainly contains kaolinite, halloysite, goethite, and gibbsite.

Peat-bog soils are a group that includes two soil types, high peat-bog soils and low peat-bog-soils. Both are waterlogged organic soils, the first type forming in upland positions and fed with rainwater, the second one—in lowland positions and fed with groundwater.

Reclaimed peat soils include corresponding two types that are drained peat soils with a plow layer.

Meadow-bog soils are waterlogged mineral soils with or without a shallow organic layer at the surface.

Bog soils of the semideserts and deserts have shallow groundwater (usually <50 cm) and an organic matter-rich surface horizon in desert or semidesert conditions.

Solods are degraded solonetzes and solonetzic soils of which the upper horizons are acidified, resulting in a well-differentiated soil with a humus-rich surface horizon, a white eluvial horizon, and a brownish colored, compact illuvial B horizon.

Solonetzes conform a group of types that have a high amount of exchangeable Na, an (near) absence of readily soluble salts, resulting in a well-expressed eluvial horizon and a compact illuvial horizon. There are three types in this group: automorphic (related to the parent material), semihydromorphic, and hydromorphic (related to groundwater influence) solonetzes.

Solonchaks have a high amount of soluble salts. Two soil types, automorphic and hydromorphic solonchaks, are distinguished.

Alluvial soils are divided into three main groups: alluvial soils with deep groundwater and only a brief period of flooding (sod alluvial soils), alluvial soils influenced by both flooding and groundwater at moderate (1–2 m) depth (meadow alluvial soils), and alluvial soils that are conditioned by a long period of flooding or shallow groundwater in combination with surface flooding (bog alluvial soils). In addition, the soil reaction, related to soil zone, is used to further subdivide these groups into types.

Mountain meadow soils occur in cold and moist high mountains, which receive a large excess of moisture resulting in a strongly leaching regime, with an acid soil reaction and a considerable accumulation of organic matter in the surface horizon.

Chernozem-like mountain meadow soils occur in high mountains, which, although receiving an excess in moisture, have only a moderately leaching regime, resulting in a weakly acid to weakly alkaline soil reaction as well as a considerable accumulation of organic matter in the surface horizon.

Mountain meadow-steppe soils develop under similar conditions as above but have a much lower exchange capacity than the chernozem-like mountain meadow soils.

An overview of the types and subtypes distinguished in the 1986 USSR soil classification is given in Table 32.10.

TABLE 32.10 Types and Subtypes of the 1986 USSR Soil Classification

Types	Subtypes
Podzolic soils	Gley-podzolic soils
	True podzolic soils
	Sod-podzolic soils
Bog-podzolic soils	Surface gleyed peaty podzolic soils
	Surface gleyed soddy-podzolic soils
	Surface gleyed mucky-podzolic soils
	Subsoil gleyed peaty-podzolic soils
	Subsoil gleyed soddy-podzolic soils
	Subsoil gleyed mucky-podzolic soils
Sod-calcareous soils	Typical sod-calcareous soils
	Leached sod-calcareous soils
	Podzolized sod-calcareous soils
Sod-gley soils	Sod surface gleyey soils
	Mucky surface gley soils
	Sod subsurface gleyey soils
	Mucky subsurface gley soils
Gray forest soils	Light gray forest soils
	Gray forest soils
	Dark gray forest soils
Gley gray forest soils	Surface gleyey (and surface meadow gray forest) soils
	Subsurface gleyey gray forest soils
	Subsurface gley gray forest soils
Brown forest soils	Acid mor brown forest soils
	Acid mor podzolized brown forest soils
	Acid brown forest soils
	Acid podzolized brown forest soils
	Slightly unsaturated brown forest soils
	Slightly unsaturated podzolized brown forest soils
Gley brown forest soils	Podzolized surface gleyey brown forest soils
	Podzolized surface gley brown forest soils
	Gleyey brown forest soils
	Gley brown forest soils
Podzolic brown forest soils	Unsaturated podzolic brown forest soils
	Slightly unsaturated podzolic brown forest soils
Gley podzolic brown forest soils	Surface gleyey podzolic brown forest soils
	Surface gley podzolic brown forest soils
	Gleyey podzolic brown forest soils
	Gley podzolic brown forest soils
Bleached meadow soils	Podzolized bleached meadow soils
	Podzolized gley bleached meadow soils
Meadow chernozem-like soils	Meadow chernozem-like soils (surface-wet)
Chernozem-like dark meadow soils	Dark meadow prairie soils
	Dark moist-meadow prairie soils
Chernozems	Podzolized chernozems
	Leached chernozems
	Typical chernozems
	Ordinary chernozems

(*continued*)

32.6.2 The Soil Classification System of the Russian Federation

32.6.2.1 Introduction

The 1977 classification and diagnostics of soils of USSR (VASKhNIL, 1986) did not satisfy both scientists and practical experts. Just 5 years after its publication, the Dokuchaev Soil Science Institute initiated work on a new version of soil classification. A complete version was published in 1997; an English translation is also available, in an adapted and improved form (Arnold, 2001). The discussion of the classification resulted in its further revision, and soon after, a new version was published (Shishov et al., 2004; see also Krasilnikov et al. 2009) for detailed description and correlation. In the Russian Federation, this system is currently introduced and used along with the older soil classification of the USSR.

32.6.2.2 Structure of the System

Like most actual soil classifications, the Russian taxonomy uses the concept of diagnostic horizons; a unique sequence of genetic diagnostic horizons comprises a soil type. In the Russian classification, the designated horizons are mutually exclusive, which provides more order to the taxonomy; for example, this classification has a key for diagnostic horizons that is impossible in other classifications. Soil types are grouped in sections and trunks. On the lower level, the types are subdivided on the basis of overlapping pedogenetic processes, particular properties, texture, and origin of parent material. The Russian classification stresses the importance of agricultural transformation of soils: soils having minor agrogenic impact are defined as subtypes of

TABLE 32.10 (continued) Types and Subtypes of the 1986 USSR Soil Classification

Types	Subtypes
	Southern chernozems
	Mountain chernozems
Meadow-chernozem soils	Meadowy chernozemic soils
	Meadow-chernozem soils
Chestnut soils	Dark chestnut soils
	Chestnut soils
	Light chestnut soils
	Mountain chestnut soils
Meadow chestnut soils	Meadowy chestnut soils
	Meadow chestnut soils
Meadow soils	Meadow soils
	Moist-meadow soils
Semi-desert brown soils	Semi-desert brown soils
Semi-desert meadow brown soils	Semi-desert meadowy brown soils
	Semi-desert meadow brown soils
Desert gray brown soils	Desert gray brown soils
Desert takyr-like soils	Desert takyr-like soils
Takyrs	Takyr
Desert sandy soils	Desert sandy soils
Meadow desert soils	Meadow desert (meadowy takyr-like) soils
	Meadow desert (meadow takyr-like) soils
	Meadow desert soils with complementary surface moistening
	Gray brown meadow desert soils
	Sandy meadow desert soils
Serozems light colored	Typical serozems
	Dark serozems
Meadow-serozem soils	Meadowy serozem
	Meadow serozem
Semi-desert and desert meadow soils	Semi-desert and desert meadow
	Semi-desert and desert moist-meadow soils
Irrigated serozems	Irrigated light colored serozem soils
	Irrigated typical serozem soils
	Irrigated dark serozem soils
	Old irrigated serozem soils
Irrigated meadow serozems	Irrigated meadow serozem soils
	Irrigated serozem meadow soils
Irrigated brown soils	
Irrigated meadow brown soils	
Irrigated gray brown soils	
Irrigated takyr-like soils	Irrigated takyr-like soils
	Ancient irrigated takyr-like soils
Irrigated meadow desert soils	Irrigated meadow desert soils
	Ancient irrigated meadow desert soils
Irrigated meadow soils	Irrigated meadow soils
	Irrigated moist meadow soils
	Ancient irrigated meadow soils
Irrigated bog soils	
Gray-cinnamon brown soils	Dark gray-cinnamon brown soils

TABLE 32.10 (continued) Types and Subtypes of the 1986 USSR Soil Classification

Types	Subtypes
	Common gray-cinnamon brown soils
	Light gray-cinnamon brown soils
Meadow gray-cinnamon brown	Surface-meadowy gray-cinnamon brown soils
	Meadowy gray-cinnamon brown soils
	Meadow gray-cinnamon brown soils
Cinnamon brown soils	Leached cinnamon brown soils
	Typical cinnamon brown soils
	Calcareous cinnamon brown soils
Meadow cinnamon brown soils	Surface-meadowy cinnamon brown soils
	Meadowy cinnamon brown soils
	Meadow cinnamon brown soils
Gray meadow-forest soils	Gray meadow-forest soils
	Gray wet meadow-forest soils
Zheltozems	Unsaturated zheltozems
	Weakly unsaturated zheltozems
	Unsaturated podzolized zheltozems
	Weakly unsaturated podzolized zheltozems
Gley zheltozems	Surface gleyey zheltozems
	Gleyey zheltozems
	Gley zheltozems
Podzolic–zheltozem soils	Unsaturated podzolic–zheltozem soils
	Slightly unsaturated podzolic–zheltozem soils
Podzolic–zheltozem gley soils	Podzolic–zheltozem surface-gleyey soils
	Podzolic–zheltozem gleyey soils
	Podzolic–zheltozem gley soils
Krasnozems	Typical krasnozems
	Podzolized krasnozems
Upland peat-bog soils	Upland peat-gley bog soils
	Upland peat-bog soils
Lowland peat-bog soils	Depleted lowland peat-gley bog soils
	Lowland (typical) peat-gley bog soils
	Depleted lowland peat-bog soils
	Lowland (typical) peat-bog soils
Reclaimed upland peat soils	
Reclaimed lowland peat soils	Reclaimed depleted lowland peat-gley soils
	Reclaimed depleted lowland peat soils
	Reclaimed lowland muck-gley soils
	Reclaimed lowland mucky-peat soils
Meadow bog soils	Mucky meadow-bog soils
	Clayey meadow-bog soils
Bog soils (semi-desert/deserts)	Peat-bog soils
	Clayey-bog soils
Solods	Meadow-steppe solods
	Solods meadow
	Meadow-bog solods
Automorphic solonetzes	Chernozemic solonetzes
	Chernozemic solonchak-solonetzes
	Solonchakic chernozemic solonetzes
	Deep-solonchakic chernozemic solonetzes
	Deep-salinized chernozemic solonetzes
	Chestnut solonetzes
	Semi-desert solonetzes

TABLE 32.10 (continued) Types and Subtypes of the 1986 USSR Soil Classification

Types	Subtypes
Semihydromorphic solonetzes	Meadow-chernozemic solonetzes
	Meadow-chernozemic solonchak-solonetzes
	Solonchakic meadow-chernozemic solonetzes
	Deep-solonchakic meadow-chernozemic solonetzes
	Meadow-chestnut solonetzes
	Meadow semi-desert solonetzes
	Semihydromorphic cryogenic solonetzes
Hydromorphic solonetzes	Chernozemic-meadow solonetzes
	Chestnut-meadow solonetzes
	Meadow-bog solonetzes
	Cryogenic-meadow solonetzes
Automorphic solonchaks	Typical automorphic solonchaks
	Takyrized automorphic solonchaks
Hydromorphic solonchaks	Typical solonchaks
	Meadow-solonchaks
	Bog-solonchaks
	Sor-solonchaks
	Mud-volcanic solonchaks
	Hummocky solonchaks
Sod acidic alluvial soils	Primitive stratified sod acidic alluvial soils
	Stratified sod acidic alluvial soils
	Typical sod acidic alluvial soils
	Podzolized sod acidic alluvial soils
Saturated sod alluvial soils	Primitive stratified saturated sod alluvial soils
	Stratified saturated sod alluvial soils
	Typical stratified sod alluvial soils
	Saturated steppe sod alluvial soils
Calcareous sod desertified alluvial soils	Primitive stratified calcareous sod desertified alluvial soils
	Stratified calcareous sod desertified alluvial soils
	Typical calcareous sod desertified alluvial soils
Meadow acid alluvial soils	Primitive stratified acid meadow alluvial soils
	Stratified acid meadow alluvial soils
	Typical acid meadow alluvial soils
Meadow saturated alluvial soils	Primitive stratified saturated meadow alluvial soils
	Stratified saturated meadow alluvial soils
	Typical saturated meadow alluvial soils
	Dark colored saturated meadow alluvial soils
Calcareous meadow alluvial soils	Stratified calcareous meadow alluvial soils
	Tugai calcareous meadow alluvial soils
	Typical calcareous meadow alluvial soils
Meadow-bog alluvial soils	Typical meadow-bog alluvial soils
	Peaty meadow-bog alluvial soils

TABLE 32.10 (continued) Types and Subtypes of the 1986 USSR Soil Classification

Types	Subtypes
Clayey-peat bog alluvial soils	Clayey-peat-gley bog alluvial soils
	Clayey-peat bog alluvial soils
Mountain meadow soils	Alpine mountain meadow soils
	Subalpine mountain meadow soils
Chernozem-like mountain meadow soils	Typical chernozem-like mountain meadow soils
	Leached chernozem-like mountain meadow soils
	Calcareous chernozem-like mountain meadow soils
Mountain meadow-steppe soils	Alpine mountain meadow-steppe soils
	Subalpine mountain meadow-steppe soils

Source: VASKhNIL (V.V. Dokuchaev Institute of Soil Science). 1986. Classification and diagnostics of soils of the USSR. Translated from Russian by S. Viswanathan. Amerind Publishing Co., New Delhi, India.

superficially disturbed soils, those under intensive cultivation enter "agrotypes" within the same section (e.g., *agropodzolic soils* vs. *podzolic soils*), and those completely transformed by agricultural practices are grouped in a special section of Agrozems.

32.6.2.3 Brief Description of the Sections

The classification includes 227 soil types, which are grouped in 27 sections and 3 trunks, defined as follows:

Postlithogenic soils: *a trunk of soils*—bringing together soils where soil-formation processes occur on a previously formed parent material and modern accumulation of matter on the surface is negligible. Includes the following soil sections:

Texture differentiated soils: soils with distinct clay illuviation, expressed both as the difference in clay content between the topsoil and B horizon and as clay coating in the illuvial horizons

Al–Fe–humus soils: soils having illuviation of aluminum and iron in complexes with organic matter

Iron metamorphic soils: soils having a B horizon, altered in situ. The alteration is expressed mainly as iron (hydr) oxides crystallization

Structural metamorphic soils: soils having a B horizon, altered in situ. The alteration is expressed mainly as soil structure modification comparing with that of parent material

Cryometamorphic soils: soils having a B horizon, altered in situ. The alteration is related mainly to the presence of permafrost and the effect of freezing–thawing cycles in the topsoil

Pale metamorphic soils: soils having structural alteration, combined with secondary calcium carbonates and (provisionally) soluble salts accumulation. These soils are geographically related to extracontinental cryoarid regions of Eastern Siberia

Cryoturbated soils: soils formed in the presence of permafrost and affected with soil material cryoturbation due to freezing–thawing cycles in the topsoil

Gleyic soils: soils affected with excessive moisture due to high groundwater level

Humus-accumulating soils: soils having a deep dark, well-structured humus-accumulative topsoil horizon

Light-humus carbonate-accumulating soils: soils having relatively poor and light topsoil, combined with secondary calcium carbonates accumulation

Alkaline clay-differentiated soils: soils with alkaline reaction and with a clay-illuvial horizon

Halomorphic soils: soils with toxic concentrations of soluble salts

Hydrometamorphic soils: soils affected with excessive moisture due to high groundwater level but having no typical "gleyic" (bluish-green) colors due to high organic matter content and/or alkalinity

Organic matter–accumulating soils: soils with no other diagnostic horizons than humus-accumulative one

Eluvial soils: soils with a distinct eluvial (clay and/or iron depleted) horizon but without evidences of clay and iron illuviation beneath

Lithozems: shallow soils (less than 30 cm) with a humus-accumulative horizon

Weakly developed soils: immature soils either in sandy or clayey loose materials or over consolidated rock

Abrazems: strongly eroded soils with exposed B or BC horizon

Agrozems: soils mixed in the course of agricultural management down to the B or BC horizon

Agroabrazems: cultivated soils, previously affected with intensive erosional processes

Turbated soils: soils mixed down to the depth of more than 1 m in the process of land improvement

Sinlithogenic soils: *a trunk of soils*—gathering together the soils, where soil formation occurs in the condition of periodical or continuous accumulation of new material on the surface. It includes the following soil sections:

Alluvial soils: soils formed in actual river valleys in recent alluvial deposits

Volcanic soils: soils formed under influence of recent periodic volcanic ash deposition

Stratozems: include all the soils with periodical eolian or fluvial accumulation of sediments (including long-term irrigation) with no or weak development of soil profile on the surface

Weakly developed soils: soils formed in fresh eolian or fluvial sediments, with incipient pedogenesis

Organogenic soils: *a trunk of soils*—including the soils formed in organic material, mainly in peat. There are two sections within this trunk, namely:

Peat soils: soils formed in organic materials

Torfozems: drained peat soils, in places deeply cultivated and ameliorated with mineral ground

32.7 The Chinese Soil Taxonomic Classification

32.7.1 Introduction

Until 1949, the Chinese soil classification was based on that of the United States but then was replaced by that of the USSR geographical classification (ISS-AS, 1990), and integrating locally important soils, such as long-continued cultivated soils, paddy soils among others. In 1994, a drastically renewed first proposal for a new *Chinese Soil Taxonomic Classification* (CSTC) based largely on *Soil Taxonomy* was issued (Gong, 1994), which is still under review and subject to modification. In developing the CSTC system, many elements of the *Legend of the Soil Map of the World* (FAO–UNESCO, 1974) and other soil classifications were incorporated.

32.7.2 Structure

The CSTC is a hierarchical system with seven categories: order, suborder, group, subgroup, genus, species, and variety. The first four levels are used to construct mapping units for small-scale maps, the lower three for more detailed maps.

The order is based on soil properties, which result from or reflect major soil-forming processes. The suborder is defined according to soil properties, which either control recent soil-forming processes or reflect dominant limiting factors. At group level, intensities of major or secondary soil-forming processes are used to differentiate among the soils, while the subgroup reflects the deviation from the central concept of the group.

The system presently has defined 13 orders, 33 suborders, 78 groups, and 301 subgroups. The nomenclature used is a mixture of older and more recent names, as well as local names for typical Chinese soils. The names of the order and suborder are linked, the group and subgroup nomenclature is different, to avoid names becoming too long. An overview of the orders, suborders, and groups of the Chinese Soil Taxonomic System is given in Table 32.11.

32.7.3 Diagnostic Horizons and Characteristics

Like many other soil classification systems, the Chinese Soil Taxonomic System uses diagnostic horizons and characteristics to identify the soil. A number of the diagnostic horizons are directly taken from other systems, such as the argillic horizon from *Soil Taxonomy*, while others are newly defined to suit Chinese conditions. Thirty diagnostic horizons are defined, 8 surface horizons or epipedons, 10 subsurface horizons, and 12 horizons, which may occupy any position in the soil profile.

Three categories of diagnostic surface horizons are recognized: histic, humus (isohumic, umbrihumic, and ochrihumic epipedons), and anthropic epipedons. Particularly, the latter category has a number of diagnostic epipedons, which do not occur in other systems, namely, the warpic (finely stratified epipedon resulting from long-continued irrigation), cumulic (thick

TABLE 32.11 Order, Suborder, and Groups of the Chinese Soil Taxonomy

Order	Suborder	Group
Histosols	Permagelic Histosols	Foli-Permagelic Histosols
		Fibri-Permagelic Histosols
		Hemi-Permagelic Histosols
	Orthic Histosols	Foli-Orthic Histosols
		Fibri-Orthic Histosols
		Hemi-Orthic Histosols
		Sapri-Orthic Histosols
Anthrosols	Stagnic Anthrosols	Gleyi-Stagnic Anthrosols
		Fe-Leachi-Stagnic Anthrosols
		Fe-Accumuli-Stagnic Anthrosols
		Hapli-Stagnic Anthrosols
	Orthic Anthrosols	Fimi-Orthic Anthrosols
		Siltigi-Orthic Anthrosols
		Mud-Cumuli-Orthic Anthrosols
		Earth-Cumuli-Orthic Anthrosols
Spodosols	Humic Spodosols	Hapli-Humic Spodosols
	Orthic Spodosols	Hapli-Orthic Spodosols
Andosols	Cryic Andosols	Geli-Cryic Andosols
		Hapli-Cryic Andosols
	Vitric Andosols	Usti-Vitric Andosols
		Udi-Vitric Andosols
	Udic Andosols	Humi-Udic Andosols
		Hapli-Udic Andosols
Ferralosols	Udic Ferralosols	Rhodi-Udic Ferralosols
		Xanthi-Udic Ferralosols
		Hapli-Udic Ferralosols
Vertosols	Aquic Vertosols	Calci-Aquic Vertosols
		Hapli-Aquic Vertosols
	Ustic Vertosols	Calci-Ustic Vertosols
		Hapli-Ustic Vertosols
	Udic Vertosols	Humi-Udic Vertosols
		Calci-Udic Vertosols
		Hapli-Udic Vertosols
Aridosols	Cryic Aridosols	Calci-Cryic Aridosols
		Gypsi-Cryic Aridosols
		Argi-Cryic Aridosols
		Hapli-Cryic Aridosols
	Orthic Aridosols	Calci-Orthic Aridosols
		Sali-Orthic Aridosols
		Gypsi-Orthic Aridosols
		Argi-Orthic Aridosols
		Hapli-Orthic Aridosols
Halosols	Alkalic Halosols	Takyri-Alkalic Halosols
		Aqui-Alkalic Halosols
		Hapli-Alkalic Halosols
	Orthic Halosols	Aridi-Orthic Halosols
		Aqui-Orthic Halosols
Gleyosols	Permagelic Gleyosols	Histi-Permagelic Gleyosols
		Hapli-Permagelic Gleyosols
	Stagnic Gleyosols	Histi-Stagnic Gleyosols
		Hapli-Stagnic Gleyosols

TABLE 32.11 (continued) Order, Suborder, and Groups of the Chinese Soil Taxonomy

Order	Suborder	Group
	Orthic Gleyosols	Histi-Orthic Gleyosols
		Molli-Orthic Gleyosols
		Hapli-Orthic Gleyosols
Isohumosols	Lithomorphic Isohumosols	Phosphi-Lithomorphic Isohumosols
		Black-Lithomorphic Isohumosols
	Ustic Isohumosols	Cryi-Ustic Isohumosols
		Cumuli-Ustic Isohumosols
		Pachi-Ustic Isohumosols
		Calci-Ustic Isohumosols
		Hapli-Ustic Isohumosols
	Udic Isohumosols	Stagni-Udic Isohumosols
		Argi-Udic Isohumosols
		Hapli-Udic Isohumosols
Ferrosols	Ustic Ferrosols	Argi-Ustic Ferrosols
		Hapli-Ustic Ferrosols
	Perudic Ferrosols	Carbonati-Perudic Ferrosols
		Alliti-Perudic Ferrosols
		Hapli-Perudic Ferrosols
	Udic Ferrosols	Carbonati-Udic Ferrosols
		Hi-Weatheri-Udic Ferrosols
		Alliti-Udic Ferrosols
		Argi-Udic Ferrosols
		Hapli-Udic Ferrosols
Argosols	Boric Argosols	Albi-Boric Argosols
		Molli-Boric Argosols
		Hapli-Boric Argosols
	Ustic Argosols	Carbonati-Ustic Argosols
		Calci-Ustic Argosols
		Ferri-Ustic Argosols
		Hapli-Ustic Argosols
	Perudic Argosols	Carbonati-Perudic Argosols
		Ali-Perudic Argosols
		Hapli-Perudic Argosols
	Udic Argosols	Albi-Udic Argosols
		Carbonati-Udic Argosols
		Claypani-Udic Argosols
		Ali-Udic Argosols
		Acidi-Udic Argosols
		Ferri-Udic Argosols
		Hapli-Udic Argosols
Cambosol	Gelic Cambosol	Permi-Gelic Cambosol
		Aqui-Gelic Cambosols
		Matti-Gelic Cambosols
		Molli-Gelic Cambosols
		Umbri-Gelic Cambosols
		Hapli-Gelic Cambosols
	Aquic Cambosols	Litteri-Aquic Cambosols
		Shajiang-Aquic Cambosols
		Dark-Aquic Cambosols
		Ochri-Aquic Cambosols

(*continued*)

TABLE 32.11 (continued) Order, Suborder, and Groups of the Chinese Soil Taxonomy

Order	Suborder	Group
	Ustic Cambosols	Siltigi-Ustic Cambosols
		Ferri-Ustic Cambosols
		Endorusti-Ustic Cambosols
		Molli-Ustic Cambosols
		Hapli-Ustic Cambosols
	Perudic Cambosols	Bori-Perudic Cambosols
		Stagni-Perudic Cambosols
		Carbonati-Perudic Cambosols
		Ali-Perudic Cambosols
		Acidi-Perudic Cambosols
		Hapli-Perudic Cambosols
	Udic Cambosols	Bori-Udic Cambosols
		Carbonati-Udic Cambosols
		Purpli-Udic Cambosols
		Ali-Udic Cambosols
		Ferri-Udic Cambosols
		Acidi-Udic Cambosols
		Hapli-Udic Cambosols
Primosols	Anthric primosols	Turbi-Anthric Primosols
		Silti-Anthric Primosols
	Sandic primosols	Geli-Sandic Primosols
		Aqui-Sandic Primosols
		Aridi-Sandic Primosols
		Usti-Sandic Primosols
		Udi-Sandic Primosols
	Alluvic primosols	Geli-Alluvic Primosols
		Aqui-Alluvic Primosols
		Aridi-Alluvic Primosols
		Usti-Alluvic Primosols
		Udi-Alluvic Primosols
	Orthic primosols	Loessi-Orthic Primosols
		Purpli-Orthic Primosols
		Rougi-Orthic Primosols
		Geli-Orthic Primosols
		Aridi-Orthic Primosols
		Usti-Orthic Primosols
		Udi-Orthic Primosols

Source: Gong, Z.T., and G.L. Zhang. 2007. Pedogenesis and soil taxonomy. Science Press, Beijing, China (in Chinese).

epipedon resulting from additions of manure and organic-rich soil material), mellowic (epipedon resulting from long and intensive cultivation and applying night soil, organic trash, and manure), and hydragric (epipedon resulting from cultivating the soil under wet conditions) epipedons.

The 10 diagnostic subsurface horizons are the albic (white), weathering B, humilluvic (illuvial organic matter), argillic (clay illuviated), clayific (in situ clay accumulation through weathering of primary minerals), claypan (slowly permeable, clay-rich horizon), alkalic (clay-illuviated horizon with high Na percentage), spodic (illuviation of amorphous organic compounds in combination with Al and/or Fe), agric (illuviated humus clay or humus silt clay under conditions of cultivation), and hydragric–redoxic (oxidation–reduction horizon related to wet cultivation) horizons.

The other diagnostic horizons comprise the calcic, hypercalcic, gypsic, hypergypsic, salic, hypersalic, sulfuric, phosphic, and gleyic horizons and the calci-, gypsi-, and salipans.

Twenty-three diagnostic characteristics (characters, properties, features, contacts, saturation, materials, regimes, or layers) are defined. These are lithologic characters (features inherited from the parent material), vertic features (shrink/swell phenomena), desertic features (surface and topsoil features related to arid conditions), takyric features (cracking of surface soil in arid conditions induced by wetting and drying), redoxic features (hydromorphism), frost/thaw features, regosolic features (<35% coarse fragments), skeletic features (between 35% and 70% coarse fragments), lithic features (>70% coarse fragments), lithic contact, paralithic contact, siallic properties (weathering B horizon of which the clay fraction is dominated, 2:1 or 2:1:1 clays), fersiallic properties (weathering B horizon of which the clay fraction is dominated 2:1 or 2:1:1 clays and which has ≥2% free Fe oxides), ferralic properties (weathering B horizon of which the clay fraction is dominated 1:1 clays and which has ≥2% free Fe oxides), andic properties (related to weathering of pyroclastic deposits), base saturation, Al saturation, calcaric properties (≥1% $CaCO_3$), humic properties (high organic matter content), organic soil materials, soil moisture regimes, soil temperature regimes, and permafrost layer (perennial temperature at or below 0°C).

When important and obvious properties of soil horizon do not fulfill the combination required for a diagnostic horizon, the CSTC uses the term evidence. They are used to identify soil taxa, particularly at subgroup level. Twelve diagnostic evidences are defined so far, namely, histic, warpic, cumulic, mellowic, clayific (all meeting the requirements of the related diagnostic horizons apart from thickness), alkalic (lower ESP), spodic (lower ratio of pyrophosphate extractable Al + Fe to clay), calcic (either shallower thickness or lower $CaCO_3$ content), gypsic (either shallower thickness or lower gypsum content), salic (lower salt content), gleyic (discontinuous gleyic parts), and vertic (less wider cracks and a clayey subhorizon within 50 cm depth) evidences.

32.8 The Australian Soil Classification

32.8.1 Introduction

In 1996, a new soil classification system for Australia was issued (Isbell, 1996). Until then, two different soil classification systems were widely used (Stace et al., 1968; Northcote, 1979). A soil classification committee has worked over 15 years to develop the new system, taking advantage of the previously used systems, the numerous soil surveys carried out all over the country, and the large database containing data of some 14,000 soil profiles, which was constructed during the time of the committee's work.

32.8.2 Structure and Nomenclature

The new system is multicategorical, comprising orders, suborders, great groups, subgroups, and families. In total, 14 orders are recognized (Anthroposols, Calcarosols, Chromosols, Dermosols, Ferrosols, Hydrosols, Kandosols, Kurosols, Organosols, Podosols, Rudosols, Sodosols, Tenosols, and Vertosols), distinguished by fairly straightforward criteria, which can easily be recognized in the field. Nomenclature is often based on Latin or Greek roots, as with many other modern soil classification systems, but the names are clearly different from *Soil Taxonomy*, the *FAO–UNESCO Legend*, or the *World Reference Base for Soil Resources*, in order to avoid confusion.

Unlike many other soil classification systems, there is no fixed soil depth for classification purposes in the new Australian system. The concept of pedologic organization (McDonald et al., 1990) is used to define the soil to be classified. Isbell (1996) describes this as "… a broad concept used to include all changes in soil material resulting from the effect of the physical, chemical and biological processes that are involved in soil formation." Consequently, soil studies for classification purposes may need to go to considerable depth. At family level, this is recognized since one of the family criteria, soil depth, includes a class Giant for those soils that are deeper than 5 m.

Division of seven orders (Chromosols, Dermosols, Ferrosols, Kandosols, Kurosols, Sodosols, and Vertosols) into suborders is based on color criteria (e.g., red, brown, yellow, gray, and black). Suborders of Anthroposols are distinguished on the kind and nature of the human activity, in Calcarosols, the degree and kind (shells, gypsum, $CaCO_3$) of the accumulation are used for the subdivision, suborders of the Hydrosols reflect the type of tidal inundation, the degree of salinity, and the reduction–oxidation regime, Organosols are classically divided according to the degree of decomposition of the organic material, suborders of the Podosols reflect the degree of wetness, Rudosols are divided according the nature of the material, while, in Tenosols, the type of surface horizon, the occurrence of a bleached horizon, and the presence or absence of a weakly developed B horizon are used to characterize the suborders.

Subdivision of the great groups and subgroups is still incomplete for a number of suborders. For those suborders that are developed into great groups and subgroups, division is generally made on the presence or absence of important diagnostic horizons and materials and on the degree of importance of certain characteristics.

Criteria considered at family level are A horizon thickness (not in Organosols, Rudosols, and Vertosols), gravel content of the surface and A1 horizon (not in Organosols), A1 horizon texture (not in Organosols and Vertosols), B horizon maximum texture (not in Organosols and Rudosols), soil depth (not in Organosols), thickness of the soils above the upper boundary of the Bk horizon (in Calcarosols only), nature of the uppermost organic materials and cumulative thickness of the organic materials (in Organosols only), and clay content of the upper 0.1 m (in Vertosols only).

The full soil name is constructed as follows: subgroup, great group, suborder, order, and family. A unique coding system for all levels is provided for recording the classification of soil profiles. Included in this coding system is the opportunity to indicate a confidence level to the soil classification.

32.8.3 Description of the Orders

The 14 orders now recognized in the Australian soil classification, and their characteristics are as follows, in alphabetical order:

Anthroposols (Gr. anthropos, man): soils resulting from human activities

Calcarosols (L. calcis, lime): soils having pedogenetic carbonate accumulations throughout the solum or at least directly below the A1 or Ap horizon

Chromosols (Gr. chroma, color): soils having a clear or abrupt textural B horizon in which the major part of the upper 0.2 m of the B2 horizon is not strongly acid [pH_{H_2O} (1:5) ≥ 5.5 and pH_{CaCl_2} (1:5) ≥ 4.6]

Dermosols (L. dermis, skin): soils having a B2 horizon, which has a structure more developed than weak throughout the major part of the horizon

Ferrosols (L. ferrum, iron): soils having a B2 horizon in which the major part contains >5% free Fe oxide in the fine earth fraction

Hydrosols (Gr. hydor, water): soils that are saturated with water in the major part of the solum for at least 2–3 months in most years. Reducing conditions are not an essential requirement

Kandosols (kandite, 1:1 clay minerals): soils having a well-developed B2 horizon of which the major part is massive or has only a weak grade of structure, and having a maximum clay content in some part of the B2 horizon >15%

Kurosols (no root): soils having a clear or abrupt textural B horizon in which the major part of the upper 0.2 m of the B2 horizon is strongly acid [pH_{H_2O} (1:5) < 5.5 and pH_{CaCl_2} (1:5) < 4.6]

Organosols (no root): soils having more than 0.4 m of organic materials within the upper 0.8 m or having organic materials from the surface to a minimum depth of 0.1 m overlying either hard rock or other hard layers, partially weathered or decomposed rock or fragmental material with interstices (partially) filled

Podosols (R. pod, under, and zola, ash): soils having a Bs, Bhs, or Bh horizon (horizons with illuvial accumulations of amorphous organic matter Al and Al silica complexes, with or without Fe)

Rudosols (L. rudimentum, a beginning): soils having negligible (or rudimentary) pedological organization apart from the minimal development of an A1 horizon or the presence of less than 10% of B horizon material in fissures in the parent rock or saprolite

Sodosols (E. sodium): soils having a clear or abrupt textural B horizon, in which the major part of the upper

0.2 m of the B2 horizon is sodic (ESP of the fine earth soil material ≥6)

Tenosols (L. tenius, weak, slight): soils having weak expression of pedological organization

Vertosols (L. vertere, to turn): soils having ≥35% clay and developing open cracks, which are at least 5 mm wide and extend upward to the surface or to the base plow layer at some time in most years self-mulching horizon, or a thin, surface crust, and having slickensides or lenticular peds

32.9 The German Soil Classification

32.9.1 Introduction

The latest version of the German soil classification system was published in 1998 (Bodensystematik, 1998) and updated in 2005 (Ad-hoc-AG Boden, 2005). It is based on the system laid down by Kubiëna (1953) and Mückenhausen (1962). Older concepts from the nineteenth and the beginning of the twentieth century using different approaches will not be discussed in this book. During recent history, West Germany and East Germany had slightly differing systems, which were brought together again in 1994. Responsible for the German classification system is a Working Group of the German Soil Science Society.

32.9.2 Structure and Nomenclature

The German system is hierarchical with 6 categorical levels (English terms according to Wittmann, 1997): 4 orders, 21 suborders, 56 types, and 230 subtypes are defined. For the lowest levels, the varieties and subvarieties, additional (often quantitative) attributes are provided that have to be added to the subtype name according to specific rules. For common discussions among soil scientists, the type or subtype level is used. A key is available to classify soils down to the type level. The German system is designed as a natural system, whose structure reflects the mutual relationships of the units and the inherent architecture of the whole (Kubiëna, 1953).

In addition to classify soils, a system to classify substrates is provided characterizing texture, coarse fraction, carbonates, coal, lithogenic carbon, and consolidated and unconsolidated rock. Usually, soil and substrate classification are given together.

The orders are defined by the water regime: The Terrestrial soils are dominated by a descending water regime and form by far the largest order. The Semiterrestrial soils are characterized by groundwater influence starting within the first 40 cm or by groundwater fluctuations over greater depth ranges. The Semisubhydric and Subhydric soils comprise the soils in tidal regions and the soils covered by water not exceeding 2 m. The last order is formed by fens and bogs.

The suborders reflect the stage of soil formation and the degree of horizon differentiation. They will be described later.

The types are differentiated according to horizons and horizon sequences reflecting specific pedogenic processes and the properties resulting from these processes (morphogenetic approach). The allocation of soils requires the knowledge of all major soil characteristics and the full horizon sequence. Specific rules apply to fens and bogs and to poorly developed soils, where characteristics of the parent material are included in the definition.

Three kinds of subtypes are recognized: The norm subtypes, the deviation subtypes, and the transition subtypes. A transition subtype shows some properties of another type. The names of both types are combined using the name of the dominant type last. A deviation subtype has some additional characteristics that are not part of the definition of the type but fit well within the type. The additional characteristics are indicated by a prefix added to the name of the soil type. Finally, the norm subtype lacks both, transition features and additional features.

Types and subtypes are defined by horizon sequences. A horizon denomination consists of a capital letter as master symbol followed by one or more lower case letters indicating properties resulting from soil-forming processes. Lower case letters set before the master symbol describe geogenic and anthropogenic characteristics. Most soil characteristics required for the horizon denomination can be recognized in the field.

32.9.3 Description of the Suborders

The 21 suborders recognized in the German soil classification system are described briefly:

Order Terrestrial Soils

O/C-Böden: soils having an organic horizon overlying directly continuous rock or coarse material, the interstices of which are filled with organic material

Terrestrische Rohböden: soils having an initial accumulation of organic matter and an initial chemical weathering of the geological parent material

Ah/C-Böden: soils having a fully developed mineral topsoil horizon enriched with organic matter and some chemical weathering of the geological parent material

Schwarzerden: soils having a thick dark mineral topsoil horizon rich in organic matter with high base saturation and intensive bioturbation and not having a pronounced subsoil horizon

Pelosole: soils originating from parent materials rich in swelling clays and having a well-developed aggregate structure without pronounced chemical weathering

Braunerden: soils having a well-developed subsoil horizon with neoformation of oxides and clay minerals

Lessivés: soils having an accumulation of illuviated clay in the subsoil

Podsole: soils having an accumulation of illuviated Fe, Al, and organic matter in the subsoil

Terrae Calcis: soils having a residual accumulation of large amounts of clay and oxides after limestone weathering

Fersiallitische und Ferrallitische Paläoböden: soils having a kaolinite-rich subsoil; in Germany only as relics from old (mostly Tertiary) weathering

Stauwasserböden: soils having a bleached topsoil or bleached aggregate surfaces and an Fe oxide accumulation in the interiors of the aggregates (if aggregates are present) due to redox processes caused by temporal water stagnation

Reduktosole: soils having a redistribution of Fe oxides due to redox processes caused by reducing gases like CO_2 or CH_4 (which maybe natural or anthropogenic)

Terrestrische Anthropogene Böden: soils having been changed by human activity (like deep cultivation or transportation of natural soil material) to such an extent that the natural horizon sequence is no longer present

Order Semiterrestrial Soils

Auenböden: soils originating from alluvial sediments that are periodically or episodically flooded and that have groundwater fluctuations over greater depth ranges

Gleye: soils having bleached horizons with Fe oxide accumulations on aggregate surfaces (if aggregates are present) or at root channels due to redox processes caused by groundwater that at least temporally reaches a depth above 40 cm below the mineral soil surface

Marschen: soils originating from tidal sediments and having groundwater influence

Strandböden: soils originating from sandy tidal sediments that are periodically or episodically flooded and show an initial accumulation of organic matter

Order Semisubhydric and Subhydric Soils

Semisubhydrische Böden: soils affected by regular tidal fluctuation

Subhydrische Böden: underwater soils covered by water not exceeding 2 m

Order Fens and Bogs

Naturnahe Moore: natural or only slightly altered fens and bogs

Erd- und Mulmmoore: drained and used fens and bogs showing (especially structural) changes toward a terrestrial soil formation

32.10 Classification Systems of Brazil, Canada, England and Wales, New Zealand, and South Africa

32.10.1 Introduction

The soil classification systems of Brazil, Canada, England and Wales, New Zealand, and South Africa have in common that they are developed in one, or at the most, two major world ecological regions, and therefore, focus on the characteristics important in that particular region. The Brazilian soil classification specializes in tropical soils, the Canadian system of soil classification focuses mainly on soils of the (sub)arctic and drier boreal regions, that of England and Wales aims at classifying soil conditions characteristic of the humid temperate regions, as does that of New Zealand, but with focus on volcanic regions, while the South African system is designed to classify soils in dry subtropical areas with emphasis on land use and management aspects.

32.10.2 The Brazilian System

Work on soil inventory of Brazil started in 1947 (Jacomine and Camargo, 1996). However, detailed descriptions and building of classification system started after 1964. The construction started "from top to bottom," filling the taxa with empirical content and developing lower taxonomic levels. Since the 1960s, the classification has gradually improved and became more descriptive and detailed (Jacomine and Camargo, 1996). The latest available version was published in 1999 (EMBRAPA, 1999). The structure and some diagnostic elements are very similar to the U.S. *Soil Taxonomy* (Soil Survey Staff, 1996); however, moisture and temperature regimes are not included as diagnostic criteria in the upper levels of the Brazilian taxonomy. The nomenclature, especially the names of higher categories, is closer to that of the WRB (IUSS Working Group WRB, 2006).

Brazilian soils are divided into 14 orders, at the first level. There are 150 great groups divided into 580 subgroups, which are defined according to qualitative variations in soil properties. The orders are defined as follows:

Alissolos: soils with evident active clay-enriched B horizon with high concentration of exchangeable aluminum

Argissolos: soils with B horizon enriched with low-activity (kaolinitic) clay

Cambissolos: soils with moderate alteration of the structure or color of the B horizon

Chernossolos: soils with dark humus-enriched topsoil horizon

Espodossolos: soils having evidences of iron and aluminum migration in organic complexes

Gleissollos: excessively humid soils affected with high groundwater level

Latossolos: deeply weathered soils

Luvissolos: soils with evident active clay-enriched B horizon with high base saturation

Neossolos: soils with moderate development

Nitososolos: deeply weathered clayey soils with well-developed aggregates with shiny surfaces

Organossolos: soils formed in organic parent material (peat)

Planossolos: soils with clay-depleted topsoil, where water stagnation causes prolonged reductive conditions

Plintossolos: soils having a discrete or continuous layer cemented with iron (hydr)oxides, formed by alternating reductive and oxidative conditions around the groundwater limit

Vertissolos: clayey soils with predominantly swelling clays, which expand in humid periods and shrink in dry seasons

32.10.3 The Canadian System

The Canadian system (Canada Soil Survey Committee, Subcommittee on Soil Classification, 1978) comprises a number of taxa, which are defined on the basis of soil properties; however, the system is genetically biased with respect to the definition of the higher categories. The system is differentiated as follows: orders, suborders, great groups, subgroups, families, and series. The following nine orders are recognized as follows:

Brunosolic: weak to moderately developed soils

Chernozemic: soils with a dark-colored, high base-saturated (dominantly Ca) A horizon, which is at least 10 cm thick and which have an organic carbon content between 1% and 17% (depending on clay content)

Cryosolic: soils that have permafrost within 1 m of the surface, or 2 m if they are strongly cryoturbated

Gleysolic: soils permanently or temporarily saturated with water and which experience reducing conditions

Luvisolic: soils with a subsurface horizon showing evidence of clay accumulation

Organic: soils dominated by organic horizons (horizons >17% organic C)

Podzolic: soils having a podzolic B horizon (horizon with accumulation of amorphous materials)

Regosolic: soils showing little or no soil development

Solonetzic: soils having a solonetzic B horizon (subsurface horizon with prismatic or columnar structures and a Ca/Na ratio ≥ 10)

32.10.4 The Soil Classification System of England and Wales

The differentiating criteria, which build the soil classification for England and Wales (Avery, 1980), comprise the composition of the soil material, the presence or absence of particular horizons or sequences of horizons, or other specified differentiating criteria. Ten MSGs are recognized, which are further subdivided in soil groups and soil subgroups. The 10 MSGs are as follows:

Brown soils: soils having a weathered (cambic) or argillic B horizon

Groundwater gley soils: soils influenced by the presence of a shallow fluctuating groundwater table

Lithomorphic soils: shallow soils (mineral substratum starting within 40 cm depth) with a distinct, humose, or peaty topsoil

Manmade soils: soils having a thick manmade A horizon or which are profoundly disturbed to depths exceeding 40 cm

Peat soils: soils with >40 cm organic material, or >30 cm if directly overlying bedrock, and lacking overlying nonhumose mineral horizons extending below 30 cm

Pelosols: slowly permeable clayey soils

Podzolic soils: soils having a podzolic B horizon (subsurface horizon with accumulations of organic matter and Al, Fe, or both)

Raw gley soils: soils with a gleyed subsurface horizon lacking a distinct topsoil

Surface-water gley soils: soils having a gleyed subsurface horizon, which can be attributed to saturation by surface water

Terrestrial raw soils: soils consisting of little altered mineral material having no diagnostic surface or subsurface horizons

32.10.5 The New Zealand System

The newly published New Zealand soil classification (Hewitt, 1998) comprises 15 orders, which are subdivided into groups and subgroups. They are distinguished from each other by the presence or absence of diagnostic horizons and other differentia, some of which are unique to New Zealand. An important element in the classification is a listing of accessory properties of the order, some of which are directly related to soil management and conservation. The 15 orders are as follows:

Allophanic soils: soils strongly influenced by minerals with short-range order (especially, allophane, imogolite, and ferrihydrite)

Anthropic soils: soils formed by the direct action of people

Brown soils: soils having a weathered B, argillic, or cutanic horizon (horizon containing translocated material, which, however, fails to meet the requirements for argillic or Bh horizon), generally having a low base status

Gley soils: poorly and very poorly drained soils

Granular soils: clayey soils dominated by kaolinite-group minerals

Melanic soils: soils with highly base-saturated, well-structured, very dark A horizons

Organic soils: soils occurring in partly decomposed remains of wetland plants or forest litter

Oxidic soils: soils containing low-activity clays and secondary oxides, which give rise to variable charge properties

Pallic soils: soils with moderate to high base status and low contents of secondary Fe oxides

Podzols: acid soils with low base saturation and subsurface accumulation of organo-Al complexes, with or without Fe

Pumice soils: soils having properties dominated by a pumiceous and glassy skeleton with a low content of clay

Raw soils: soils lacking distinct topsoil development

Recent soils: soils showing only incipient marks of soil-forming processes

Semiarid soils: soils having a high base status and soil water deficit during the growing season

Ultic soils: acid soils with clayey and/or organic illuvial features

32.10.6 The South African System

The South African taxonomic system of soil classification (Soil Classification Working Group, 1991) comprises only two categories, soil forms and soil families. Soil forms are defined by a unique vertical sequence of diagnostic horizons and materials, while the soil families are separated on the basis of other properties. Nomenclature is South African in the sense that both soil forms and soil families bear local, geographical names (e.g., Escourt form, Haarlem family). So far, 73 soil forms are defined, divided into 406 soil families. The definitions of the diagnostic horizons and materials fit the unique South African conditions and are very difficult to correlate with other existing classification systems. Important elements for classification of the soil forms are the various topsoil horizons (organic O, humic A, vertic A, melanic A, and orthic A horizons), the pedality (apedal, structural grade weaker than moderate, vs. structured B horizons), various clay-enriched horizons (prismacutanic, pedocutanic, lithocutanic, and neocutanic B horizons), signs of wetness (e.g., soft and hard plinthic B horizons), carbonate accumulation (neocarbonate B horizon, soft carbonate horizon, and hardpan carbonate horizon), and nature of the underlying material (regic sand, stratified alluvium, saprolite, hard rock). Distinguishing criteria at soil family level comprise the thickness, colors, base status structure, consistency, and wetness of horizons.

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33

Classification of Soils

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33.1 Introduction: General Characteristics of Soil Orders and Global Distributions

Larry P. Wilding

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The purpose of this chapter is to present a discussion of the distribution, attributes, pedogenesis, classification, use, and management of the 12 soil orders recognized by *Soil Taxonomy* (Soil Survey Staff, "1999"). In this chapter, soil orders have been arrayed in a sequence to consider the organic soils first, followed next by soils that are weakly expressed or lack subsoil diagnostic horizons, and closing with soils exhibiting the greatest evidence of pedogenic weathering and/or accumulation of sesquioxides and translocated phyllosilicates. Only general statements can be made at this categorical level, but the information content becomes more precise as one descends to lower taxa (Sections 33.2 through 33.13). Because *Soil Taxonomy* is a dynamic classification system, the content is subject to change. For example, the most recent edition of the *Keys to Soil Taxonomy* (Soil Survey Staff, 2010) includes newly defined taxa for subaqueous soils (soils permanently inundated by shallow water) in the Histosols and Entisols (Sections 33.2 and 33.4).

Because of space limitations, data sets verifying the physical, chemical, mineralogical, and biological attributes of given orders are abridged. However, access to complete soil characterization databases, engineering properties, pedon descriptions, soil interpretations, and other information for different soil taxa are available online at the following address: http://soils.usda.gov/. For example, analytical data for more than 27,000 pedons in the United States and 1,100 pedons from other countries can be accessed from this Web site. Standard morphological pedon descriptions are also available in most cases. Further, this Web site includes databases with listings of published soil surveys in the United States, Puerto Rico, Virgin Islands, Trust Territories, and some foreign countries; areal extent and classification of soil series; descriptions of ongoing soil survey investigations; various documents that define the standards used for soil surveys in the United States including field and laboratory methods; and links to Web Soil Survey and other soil surveys available in digital format.

The global distribution of soil orders at 1:1 million scale is illustrated in Figure 33.1—while Table 33.1 provides the areal extent of suborders associated with these orders. Tables 33.2 and 33.3 present proportional percentages of soil orders occurring in different soil temperature (Table 33.2) and soil moisture (Table 33.3) regimes as defined in *Soil Taxonomy* (Soil Survey Staff, 1999). Broad geographical distribution patterns (Figure 33.1) reflect regional conditions of climate, vegetation, geology, and topography, which function interactively over time. Soils found at any point on the land surface are considered products of multiple sets of pedogenic processes with state factors serving as controls (Chapter 30). Generally, soils are developed along polygenetic pathways, on dynamically evolving landforms under the influence of paleoclimates, in nonuniform parent materials. Hence, rarely can bio-, litho-, climo-, chrono-, and toposequence functions be developed with scientific rigor. Clearly, the distribution of soil orders in which local conditions strongly govern pedogenic dynamics over short-range distances is not well illustrated in Figure 33.1. In particular, such comments are germane to distributions of Histosols, Spodosols, Vertisols, and Gelisols. In these cases, the distributions presented in sections of this chapter may differ somewhat from those presented herein (Figure 33.1). This uncertainty will also impact the accuracy of

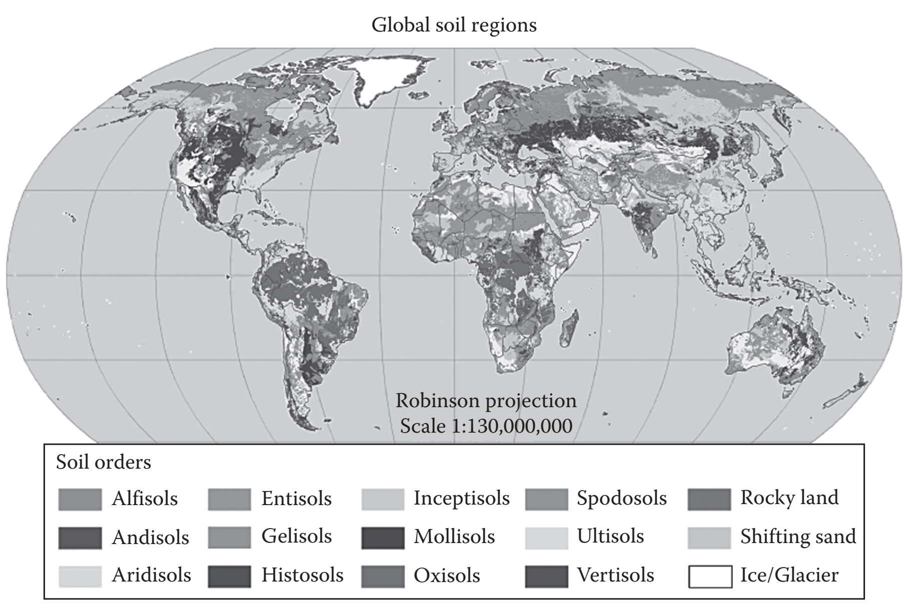

FIGURE 33.1 Global distribution of soil orders. (Courtesy of the USDA-NRCS, Soil Survey Division, World Soil Resources, Washington, DC, 2006.)

TABLE 33.1 Global Areas and Percentages of Suborders and Miscellaneous Land Units Based on Ice-Free Land Area

Order	Suborder		Area ($km^2 \times 10^3$)	Proportion (%)
Alfisols	Aqualfs		1,029	0.8
	Cryalfs		2,531	1.9
	Ustalfs		6,024	4.6
	Xeralfs		893	0.7
	Udalfs		2,678	2.0
		Subtotal	**13,156**	**10.1**
Andisols	Cryands		251	0.2
	Torrands		1	<0.1
	Xerands		32	<0.1
	Vitrands		281	0.2
	Ustands		62	<0.1
	Udands		275	0.2
	Gelands		62	<0.1
		Subtotal	**963**	**0.7**
Aridisols	Cryids		1,036	0.8
	Salids		1,287	1.0
	Gypsids		680	0.5
	Argids		4,678	3.6
	Calcids		4,887	3.7
	Cambids		2,919	2.2
		Subtotal	**15,487**	**11.9**
Entisols	Aquents		109	0.1
	Psamments		4,447	3.4
	Fluvents		3,056	2.3
	Orthents		15,834	12.1
		Subtotal	**23,446**	**17.9**
Gelisols	Histels		1,006	0.8
	Turbels		5,065	3.9
	Orthels		5,692	4.4
		Subtotal	**11,764**	**9.0**
Histosols	Folists		—	<0.1
	Fibrists		196	0.1
	Hemists		981	0.8
	Saprists		331	0.3
		Subtotal	**1,507**	**1.2**
Inceptisols	Aquepts		3,657	2.8
	Anthrepts		450	0.3
	Cryepts		2,598	2.0
	Ustepts		2,230	1.7
	Xerepts		683	0.5
	Udepts		4,102	3.1
	Gelepts		6,043	4.6
		Subtotal	**19,764**	**15.1**
Mollisols	Albolls		7	<0.1
	Aquolls		118	0.1
	Rendolls		261	0.2
	Xerolls		923	0.7

TABLE 33.1 (continued) Global Areas and Percentages of Suborders and Miscellaneous Land Units Based on Ice-Free Land Area

Order	Suborder		Area ($km^2 \times 10^3$)	Proportion (%)
	Cryolls		2,464	1.9
	Ustolls		3,937	3.0
	Udolls		1,263	1.0
	Gelolls		156	0.1
		Subtotal	**9,128**	**7.0**
Oxisols	Aquox		322	0.2
	Torrox		31	<0.1
	Ustox		3,115	2.4
	Perox		1,167	0.9
	Udox		5,233	4.0
		Subtotal	**9,868**	**7.6**
Spodosols	Aquods		167	0.1
	Cryods		2,574	2.0
	Humods		57	<0.1
	Orthods		649	0.5
	Gelods		1,111	0.8
		Subtotal	**4,558**	**3.5**
Ultisols	Aquults		1,285	1.0
	Humults		380	0.3
	Udults		5,540	4.2
	Ustults		3,340	2.6
	Xerults		19	<0.1
		Subtotal	**10,563**	**8.1**
Vertisols	Aquerts		5	<0.1
	Cryerts		17	<0.1
	Xererts		98	0.1
	Torrerts		894	0.7
	Usterts		1,770	1.4
	Uderts		383	0.3
		Subtotal	**3,167**	**2.4**
Miscellaneous	Salt		145	0.1
	Shifting sands		5,341	4.1
	Rock		1,803	1.4
	Ice		2,210	
		Subtotal	**7,288**	**5.6**
Total			**130,658**	**100.0**

Source: Courtesy of USDA-NRCS Soil Survey Division, Washington, DC, 2010.

global distributions and areal data given for Histosols, Andisols, and Inceptisols found interspersed with Gelisols.

The human impact on soils is recognized in some taxa of Entisols and Inceptisols, but, in general, *Soil Taxonomy* does not adequately handle the anthropogenic effects on soils. For example, Mollisols can be transformed into Alfisols when subject to severe erosion; paddy soils can undergo secondary salinization and waterlogging under long-term rice culture; soils in urban/industrial environments can be markedly modified by landfills,

TABLE 33.2 Extent of Soil Orders Found in Different Temperature Regions

	Soil Temperature Regions[a]			
Soil Order	Tropical	Temperate	Boreal	Tundra
Alfisols	38	39	23	0
Andisols	49	23	28	0
Aridisols	12	74	14	<1
Entisols	28	68	4	0
Gelisols	0	0	0	100
Histosols	21	8	71	0
Inceptisols	47	42	11	0
Mollisols	4	50	46	0
Oxisols	98	2	0	0
Spodosols	<1	18	82	0
Ultisols	69	31	<1	0
Vertisols	47	52	1	0

Source: Modified from data supplied by USDA-NRCS, Soil Survey Division, 1998.

[a] Derived from soil temperature regimes defined in *Keys to Soil Taxonomy* (Soil Survey Staff, 2010).

Tropical: isomesic, isothermic, and isohyperthermic (MAST ≥ 8°C in which the difference in mean summer and mean winter soil temperatures is <6°C at 50 cm depth).

Temperate: mesic, termic, and hyperthermic (MAST ≥ 8°C at 50 cm depth).

Boreal: frigid, isofrigid, and cryic (MAST is 0°C–8°C at 50 cm depth).

Tundra: Gelic (MAST <0°C or <1°C in Gelisols).

TABLE 33.3 Extent of Soil Orders Found in Different Soil Moisture Regimes

	Soil Moisture Regimes[a]			
Soil Order	Aridic	Xeric	Ustic	Udic[b]
Alfisols	0	8	56	36
Andisols	4	3	30	63
Aridisols	95	1	3	1
Entisols	60	4	22	14
Gelisols	N/A[c]	N/A	N/A	N/A
Histosols	8	1	23	68
Inceptisols	0	6	42	52
Mollisols	0	13	66[d]	22
Oxisols	<1	0	35	65
Spodosols	<1	1	9	90
Ultisols	0	<1	42	58
Vertisols	28	3	56	13

Source: Modified from data supplied by USDA-NRCS, Soil Survey Division, 1998.

[a] Soil moisture regimes defined in *Keys to Soil Taxonomy* (Soil Survey Staff, 2010).

[b] Udic soil moisture regime includes perudic moisture regime and soils with aquic conditions at a shallow depth (soil horizons saturated, reduced, and with redoximorphic features).

[c] N/A—not appropriate for the Gelisol order.

[d] A small percentage of Mollisols in the Ustic moisture regime may in fact have an Aridic soil moisture regime, especially in steppe regions of Eurasia that lie along the border between Ustic and Aridic soil moisture regimes.

landfarming, earth movement, and heavy metal contamination; and drastically, disturbed soils are common in regions where precious metals, rock aggregate, and fossil fuels have been mined. The International Committee on Anthropic Soils (ICOMANTH) where ANTH stands for anthropogenic is currently addressing this matter. For information on this activity, refer to the ICOMANTH Web site, http://clic.cses.vt.edu/icomanth/

Soil Taxonomy currently does not accommodate systematic cataloging of paleosols formed under remarkably different paleoenvironments. With increasing geomorphic age, properties of soils commonly reflect a welding of contemporaneous and paleoenvironments. In particular, when paleosols are well preserved, they are valuable proxies for biological and physicochemical evolution of the Earth. This is an area for *Soil Taxonomy* development that will likely gain greater attention in the future. With multiple uses of soil surveys for connectivity between the land surface and geological substrata (e.g., for environmental, hydrological, archaeological, and biogeoscience synergisms), there are increasing driving forces to observe soils beyond the 2 m depth limit currently in vogue. This is likely to enhance interest in revising *Soil Taxonomy* to better accommodate recognition and cataloging of paleosols as important morphogenetic markers of paleoenvironments (Chapter 29).

In perusing the global distributions of soil orders (Figure 33.1) and aerial extents given in Tables 33.1 through 33.3, several noteworthy relationships become apparent. These are summarized in a cursory overview in the following paragraphs. The organic-rich soils [Histosols (Section 33.2)], which comprises about 1.2% of the Earth's ice-free land surface (Table 33.1), generally occur in cool, humid, high latitude, boreal regions or in microhabitats such as seasonally ponded areas that favor the balance of organic matter accumulation over decomposition (Tables 33.2 and 33.3). This is because of saturated and reduced conditions, biological inhibitors, and/or cool, short summer periods with low evapotranspiration rates. Histosols are important wildlife and wetland habitats. Because they contain a disproportionally high level of soil carbon in relation to their aerial extent, Histosols are an important component of the global C cycle. These soils can be important sources of methane in the atmosphere, and accelerated organic matter decomposition resulting from more frequent drought and/or elevated temperatures associated with climate change not only potentially reduces their global abundance but also makes them a potential source of atmospheric carbon dioxide. Histosols are used extensively for cranberry, citrus, vegetable, rice, and sugarcane production and are a major fuel source in many regions. Major constraints to their use are subsidence, nutrient deficiencies, and wind erosion. Effective water-table management is critical in minimizing subsidence. These land resources are also critically limiting to construction activities because of their low soil strength and load bearing capacities.

Entisols (Section 33.4) comprise about 17.9% of the Earth's ice-free land surface (Figure 33.1). Over two-thirds of the Entisols are in the temperate region while the remainder mostly occur in the tropics (Table 33.2). Sixty percent of the Entisols have aridic soil moisture regimes with ustic and udic being the next most extensive (Table 33.3). These are mineral soils, which do not have expression of diagnostic subsurface horizons within a specified depth of the soil surface, generally 1–2 m. Subsoil materials of these soils reflect the nature of the geological materials from which they were derived. These weakly expressed soils are commonly found on geomorphic surfaces, which are unstable because of frequent flooding, erosion/truncation, or human impact (drastically disturbed lands). They are formed also in coarse-textured, resistant mineral parent materials (e.g., quartzose sands) that are subject to little pedogenic development over time. Entisols are common along flood plains of river and stream valley systems, sand dunes in desert regions, on high-gradient mountainous terrain, and associated with recently mined or disturbed lands.

Andisols (Section 33.3) comprise <0.7% of the Earth's ice-free land surface (Table 33.1), but these soils are so unique physically, chemically, and mineralogically that a separate soil order was established for them. They are found along the tectonically active Pacific Ring of Fire, Central Atlantic Ridge, North Atlantic rift, the Caribbean, and the Mediterranean regions where volcanic or pyroclastic deposits are common (Figure 33.1). About half of them are located in tropical regions with the remainder split between temperate and boreal climates (Table 33.2). About two-thirds occur under udic soil moisture regimes and one-third in ustic regions (Table 33.3). These weakly developed soils are texturally undifferentiated and characterized by short-range order (amorphous) aluminosilicates that have not been translocated from upper to lower horizons. Andisol land resources are used extensively for crop production in developing countries. They commonly have favorable physical properties for plant growth while in many areas, high native fertility is present because of frequent additions of tephra that renew soil nutrients.

Inceptisols (Section 33.5), which comprise about 15.1% of the Earth's ice-free land surface (Figure 33.1), occur indiscriminately on a global basis because they lack a sharply defined central concept. Over 90% occur in tropical and temperate climates (Table 33.2) under ustic and udic soil moisture regimes (Table 33.3). Inceptisols serve to make *Soil Taxonomy* fully bifurcated. Soils classified as Inceptisols generally contain light-colored surface horizons with weakly expressed diagnostic subsoil horizons (cambic) but, under certain circumstances, may contain more strongly developed subsoil diagnostic horizons (e.g., petrocalcic, petrogypsic, duripan, and fragipan). Where these soils reflect youthfulness, they occur in high-gradient mountainous regions, along major river systems as terraces and deltaic/fluvial plains, and as soils developed from carbonate-rich bedrocks or sediments in Mediterranean environments. They also occur interspersed among more strongly developed Alfisols, Ultisols, and Oxisols in tropical and subtropical regions. Properties of Inceptisols vary widely as does their potential for plant production and engineering use.

Gelisols, the newest of the soil orders (Section 33.6), comprise about 9.0% of the Earth's ice-free land surface (Figure 33.1). These are soils of the high-latitude polar regions underlain by permafrost (materials wet/dry that remain below 0°C for 2 consecutive years) at depths of 1–2 m depending on cryoturbation (frost churning) activity (Table 33.2). Cryoturbation in Gelisols is driven by the physical volume change from water to ice and subsequent moisture migration along thermal, hydrostatic, chemical, and electrical gradients. Cryopedogenic processes include freezing/thawing, cryoturbation, frost heaving, cryogenic sorting, thermal cracking, and ice segregation. Remarkable spatial variability occurs in soil properties over distances of a few meters or less giving rise to pattern ground in the form of stone stripes, stone circles, high and low center polygons, and barren frost boils. Because Gelisols are large C sinks, warming in polar regions may result in melting of permafrost, increased organic matter decomposition, and increased CO_2 release to the atmosphere. Gelisols present management challenges because of volume changes associated with cryopedogenic processes. In addition, melting of ice following disturbance associated with construction or agricultural production leads to subsidence or thermokarst. Only through careful management of subsurface ice and maintenance of a negative thermal balance can the integrity of structures be preserved.

Vertisols (Section 33.7) comprise about 2.4% of the Earth's ice-free land surface (Figure 33.1). About half the Vertisols are found in tropical and about half in temperate environments (Table 33.2). Over 50% of the Vertisols occur in regions with ustic soil moisture regimes where seasonal desiccation is common. The remainder are found primarily under aridic and xeric environments; here, the oscillation from wet to dry soil conditions is less frequent (Table 33.3) and gilgai expression less extreme. These clayey, shrink/swell soils are highly diverse in physical, chemical, biological, and mineralogical properties. Although concentrated in subtropical and tropical environments, they do occur across broad climatic, vegetative, topographic, and geologic regions of Africa, India, Australia, China, and North and South America (Figure 33.1). Swelling caused by wetting dry Vertisols often results in soil failure when swelling pressures exceed shear strength. Soil failure is usually expressed as slickensides occurring as diagonal, polished, and grooved slip surfaces, wedge-shaped structural units, and microtopography in the form of gilgai. Commonly, these soils have self-swallowing and self-mulching surface features where strongly aggregated surface materials fill in vertical structural cracks. While inversion by turbation is commonly invoked as a major pedogenic processes, most movement in these soils is believed to be diagonally along inclined major slickenside fault planes. Major limitations for using these highly productive soils for agriculture, especially in developing countries, are their high energy requirements for tillage and narrow favorable soil moisture range for workability. Construction activities are constrained by their propensity for soil failure and high shrink/swell activity.

Mollisols (Section 33.8) comprise about 7% of Earth's ice-free land surface (Figure 33.1). Essentially, all the Mollisols are approximately equally split between temperate and boreal regions (Table 33.2). About two-thirds of the Mollisols are in ustic soil moisture regimes (Ustolls) with most of the remainder occurring in regions with a udic moisture regime (Udolls) and the rest in climates with winter precipitation and summer periods of soil moisture deficit (Xerolls) (Table 33.3). A few Mollisols may occur under aridic soil moisture regimes in the southern steppe zones of Eurasia but their extent is uncertain. Mollisols are the dark-colored, high base status soils commonly found under prairie grass vegetation in mid-latitudes of North and South America and Eurasia. The distribution pattern of these soils in Eurasia is in east-west trending belts between Alfisols to the north and Aridisols and Inceptisols to the south. This distribution reflects decreasing precipitation and increasing temperature gradients from north to south. In contrast, in North America, the precipitation isohyets and isotherms are set normal to each other with precipitation decreasing from east to west and temperature increasing from north to south. These climatic controls result in Mollisols being bordered on the east by Alfisols and Ultisols, and to the west by Aridisols, Entisols, and drier Alfisols. Mollisols are commonly associated with calcareous or high base status glacial drift, limestones, and loess. Mollisols (especially Udolls) are the "bread basket" soils of the Americas and Eurasia because their high native fertility, occurrence in favorable climates, and excellent physical and chemical properties result in high crop production potentials.

Spodosols (Section 33.9) comprise about 3.5% of Earth's ice-free land area (Figure 33.1) with over 80% occurring in boreal environments (Table 33.2) with udic soil moisture regimes (Table 33.3) between Mollisols of mid-to-high latitudes and Alfisols in lower latitudes. Spodosols are also commonly associated with imperfectly drained sediments in coastal regions such as the Atlantic Coast of the southeast United States. Spodosols commonly have prominent eluvial E horizons overlying black to red spodic and placic horizons. The spodic horizons have formed from translocated illuvial organo-metal (Al and Fe) complexes. These soils have many properties in common with Andisols. The major difference is that Spodosols have formed for a sufficient period of time for the weathered organic–metal complexes to be translocated to subsoils. In Andisols, little or no translocation of these complexes has taken place. Formation of Spodosols is favored by environments with abundant acid litter, coarse-textured resistant parent materials, and strong leaching potentials. These soils are very strongly acid and extremely phosphorus deficient. Native vegetation is composed largely of acid tolerant plants. They are little used for agricultural production because of low nutrient and water retention and high management requirements. The major land use is forests for lumbering and recreation.

Aridisols (Section 33.10) which comprise about 11.9% of the Earth's ice-free land area (Figure 33.1) occur mostly in hot temperate deserts with aridic soil moisture regimes, dry coastal regions, or on rain-shadow plains leeward of high mountains (Tables 33.2 and 33.3). Locally, they also occur in more humid regions where salts have concentrated at or near the soil surface through evaporative pumping from shallow saline ground waters. Aridisols range from base-rich to base-poor soils and exhibit a wide diversity in physical, chemical, and mineralogical properties. Commonly, they represent contemporaneous xerophytic environments, but some Aridisols have well-expressed subsoil diagnostic properties such as argillic, natric, calcic, petrocalcic horizons. These likely reflect more pluvial paleoclimates, and the diagnostic horizons would be, in part, relict. Aridisols are used extensively for rangelands, military bases, and nomadic agrarian activities. Although these soils are not generally considered arable land resources without irrigation, where an adequate quantity of high-quality subsurface water is available they become suitable for intensive agriculture and urban development. Irrigation and drainage must be managed for a favorable salt balance in Aridisols.

Alfisols (Section 33.11) and Ultisols (Section 33.12) are closely related soil orders that comprise, respectively, about 10.1% and 8.1% of the Earth's ice-free land area (Figure 33.1). These soils, developed mostly under deciduous forested or savannah environments, occur extensively in mid-to-lower latitudes. Alfisols are distributed approximately equally across tropical, temperate, and boreal environments whereas the distribution of Ultisols is skewed toward tropical conditions (Table 33.2). For both Alfisols and Ultisols, most of their extent is associated with ustic and udic soil moisture regimes (Table 33.3). Alfisols and Ultisols are differentiated from other orders on the basis of textural differentiation between surface and subsoil horizons. The textural differentiation partly defines argillic and kandic horizons and has often resulted from translocation and/or in situ neoformation of clays. Alfisols differ from Ultisols in having higher base saturation (>35%) at a specified depth in the subsoil. Rationale for this differentiation was that Alfisols could sustain traditional slash/burn subsistence agriculture without significant external fertilizer inputs, while Ultisols would require applied nutrients for crop production. Both orders are used extensively for forest and crop production. Major limitations are nutrient deficiencies, wind and water erosivity, and seasonal soil moisture deficits, especially for those in xeric and ustic regions.

Sesquioxide-rich Oxisols (Section 33.13) comprise about 7.6% of the Earth's ice-free land surface occurring in equatorial regions intermixed with Ultisols and Inceptisols (Figure 33.1). Nearly all the Oxisols occur in tropical environments (Table 33.2) with two-thirds under udic and one-third under ustic soil moisture regimes (Table 33.3). Oxisols contain low activity clays and are nutrient poor, weakly buffered systems that have little or no horizon differentiation. This reflects (1) their high weathering intensities in low latitude, humid environments, (2) their geomorphic stability and age associated with mafic-rich, Precambrian platforms, or (3) development from polcycled, preweathered parent materials derived from highly weathered terrestrial source areas. These soils are often considered residual carcasses, but many are very productive for agriculture and forestry if properly managed for water conservation, soil erosion, and nutrient inputs. With high nutrient inputs and management to conserve soil and water, Oxisols can be very productive agricultural soil systems.

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33.2 Histosols

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33.2.1 Introduction

While most soils of the world comprise primarily mineral materials, a small but important group of soils are formed from organic materials derived from plants, or less frequently, from animals. *Organic soil materials* contain a minimum of 12%–18% organic carbon, depending on the particle size of the mineral component (Soil Survey Staff, 2010). Generally speaking, soils with at least 40 cm of the upper 80 cm that are organic materials, and which do not have permafrost within 1 m of the soil surface, are Histosols. Prior to 1997, organic soils with permafrost were included in the Histosol order; they are now placed in the Histel suborder of Gelisols (Soil Survey Staff, 1998). Following the separation of Histosols and permafrost soils, Histosols occupy about 1% of the global land area while the Histel suborder of Gelisols occupies about 0.8% (Buol et al., 2003). Organic soil materials are commonly referred to as *peat*, and land covered by Histosols or Histels is known as *peatland*. The term *mire* is a synonym of peatland that is more commonly used in Europe. Histosols also include a narrowly distributed group of soils, the Folists, that consist of well-drained organic soil materials that directly overly bedrock or coarse fragments with little or no intervening fine soil. The peat layer in Folists may be (and usually is) thinner than the 40 cm required for other Histosols.

Because of their high organic C content, many Histosols have been utilized as a combustible energy resource. Mankind has mined and burned peat since prehistoric times, and peat is still an important fuel in a number of northern countries, although it has a lower energy rating than oil or coal. In Russia, Germany, and Ireland, peat is not only utilized for domestic heating but is also used on a large scale in electricity generation. In 2009, peat burning was the source of 10% of Ireland's electricity (Public communication, 2009a), and in 2003, about 7% of Finland's electricity was a result of burning peat (Kirkinen et al., 2007). Sweden, Belarus, Latvia, Estonia, and Lithuania also utilize peat for energy; and Canada is currently investigating the possibility of using peat for power production in Ontario and Labrador (Public communication, 2009b, 2009c). In addition to being mined as an energy source, peat is mined for use as a soil amendment in agriculture and horticulture. Moreover, the agricultural value of Histosols has long been recognized. Provided that the water tables can be effectively managed, high yields of vegetables and other specialty crops can be produced on Histosols in such different climatic regions as Michigan and Florida (Lucas, 1982).

Because most Histosols occur in wetlands,* their utilization as agricultural and energy resources has come under intense scrutiny. Histosols perform many of the beneficial functions of wetlands, and they are negatively impacted by mining, drainage, or other practices associated with agriculture; thus, there are benefits derived from preservation of Histosols in their natural state. Histosols are perhaps uniquely fragile and are highly vulnerable to degradation. When drained or dry, organic soil material is highly susceptible to wind and water erosion (Lucas, 1982; Parent et al., 1982). Organic soils also have very low strength, are highly compressible (MacFarlane, 1969; MacFarlane and Williams, 1974), and gradually subside by decomposition if drained (Gesch et al., 2007). Furthermore, within the framework of current discussions of global climate change and carbon budgeting, Histosols contribute significantly to the terrestrial carbon pool. On an areal basis, C storage in Histosols is often greater than mineral soils by an order of magnitude (Rabenhorst, 1995).

33.2.2 Distribution

Histosols occur at all latitudes, but are most prevalent in the boreal forest regions of northern North America, Europe, and Asia (Figure 33.2). The world's largest expanses of Histosols occur in the West Siberian lowland (Walter, 1977) and the Hudson Bay lowland of central Canada (Sjors, 1963; Canada Committee on Ecological (Biophysical) Land Classification, National Wetlands Working Group, 1988). At lower latitudes, Histosols occur locally on humid coastal plains, notably southeast Asia and Indonesia (Anderson, 1983).

Histosols in the United States are most widespread in lowlands of the Great Lakes region, the northeast, the Atlantic Coastal plain and Florida, the Pacific Northwest, and Alaska (Figure 33.2). The largest expanses of Histosols in the continental United States are on the Lake Agassiz plain in north-central Minnesota (Wright et al., 1992). Coastal and estuarine areas inundated by tidal water are also sites for Histosol formation, most notably along the Atlantic and Gulf coastlines. Drained Histosols are widely used in agriculture in the Great Lakes region, southern Florida, and the Sacramento–San Joaquin delta region of California. In the semiarid Great Plains and mountainous west, Histosols are very rare and occur only in areas of steady groundwater discharge (Mausbach and Richardson, 1994) or humid areas at high elevations (Cooper and Andrus, 1994). Organic soils are widespread in lowlands throughout Alaska, although most of the organic soils in central and northern Alaska have permafrost and hence are classified as Histels in the Gelisol soil order rather than Histosols.

* The National Technical Committee on Hydric Soils (USA) has included "All Histosols except Folists" within the criteria for hydric soils (USDA-NRCS, 1996).

FIGURE 33.2 Worldwide distribution of Histosols. (Courtesy of USDA-NRCS, Soil Survey Division, World Soil Resources, Washington, DC, 2010.)

33.2.3 Formation of Histosols

33.2.3.1 Parent Material

In contrast to the wide variety of mineral materials, which may serve as parent materials for other soils, the parent materials from which Histosols are formed are organic in nature. The conditions that cause the accumulation of organic parent materials are very closely tied to the processes, which form various organic soil horizons. The unique properties of Histosols result from the nature of the organic parent material.

Some of the factors, which affect the nature of organic parent materials, include hydroperiod, water chemistry, and vegetation type and will be discussed in more detail in the following sections. The net accumulation of organic materials occurs when rates of additions (usually as primary plant production) exceed rates of decomposition. In natural soils of most ecosystems, a steady state exists between these two processes, which maintains the quantity of organic C in the surface horizons somewhere between 0.5% and 10%, although some forest soils have relatively thin layers of organic soil materials (O horizons). In Histosols, the rates of decomposition are slowed and organic matter accumulates to the degree that the organic materials amass to a significant depth. In most cases, this is caused by saturation in soils leading to anaerobic conditions, which causes organic matter decomposition to be less efficient than under aerobic conditions. Occasionally, usually under cool and moist conditions, organic parent materials may accumulate without prolonged saturation, producing soils in the unique Folist suborder of Histosols. Under certain conditions and landscapes, mineral materials can be added to accumulating organic parent materials, typically by alluviation or eolian deposition. In such cases, the balance between mineral inputs and organic accumulation will determine whether a Histosol or mineral soil will form.

The accumulation of organic parent materials often occurs over long periods of time and under changing conditions. Thus, the stratigraphy of a bog may reflect many thousands of years of organic matter accumulation. Rarely does vegetation remain constant over such long periods. Microscopic examination of the partially decomposed peat or evaluation of pollen or plant microfossils can provide information concerning the types of plants that have contributed to the organic parent material during various stages of accumulation (e.g., Wieder et al., 1994).

The organic parent material of Histosols is a major source of acidity. Acids produced by the partial decomposition of organic matter cause the organic horizons of Histosols to be highly acidic unless the acids are neutralized by bases that were dissolved from mineral soils or rocks and transported into the peat by groundwater. Highly acidic peatlands are called *bogs* and commonly described as *ombrotrophic* (rain-fed, because all nutrients are derived from atmospheric sources). The less acid Histosols that receive base-rich groundwater or runoff are called *fens* and usually described as *minerotrophic* (fed with mineral-derived nutrients). The term *swamp* is often applied to forested systems on both fens and on mineral soil wetlands. Peatlands are also sometimes divided by their acidity into *oligotrophic* (very acid and mineral-poor), *mesotrophic* (intermediate), and *eutrophic* (weakly acidic to neutral and mineral-rich) classes (Moore and Bellamy, 1974; Gore, 1983).

33.2.3.2 Climate

Histosol formation is favored by wet or cold climates. Wetness and cold favor Histosol formation by hindering decomposition of organic matter. The northern Histosol-dominated regions of North America and Eurasia (Figure 33.2) have temperate and boreal climates in which average annual precipitation exceeds annual potential evapotranspiration (Trewartha, 1968). Further to the north, in Greenland and the islands bordering the Arctic Ocean, for example, Histosols are rare because the growing season is so short that there is little production of organic matter. To the south of the boreal zone, Histosol formation is apparently constrained by rapid decomposition of soil organic matter. At lower latitudes, Histosols are restricted mostly to coastal plains with very flat topography, high annual precipitation, and no dry season, or to those areas where a high water table is maintained by tidal waters, leading to coastal marshes and mangrove systems (Anderson, 1983).

Climate affects not only where Histosols occur, but also the chemistry of the resulting soils. Where average annual precipitation exceeds annual potential evapotranspiration, Histosols in suitable settings can remain saturated by rainfall alone (Ivanov, 1981). In such regions, both the highly acidic Histosols of bogs and the less acid Histosols of groundwater-fed fens may form. In less humid regions, where groundwater is required to maintain saturation of soils, only the less acid, minerotrophic Histosols can form.

Climate change scenarios suggest more frequent extreme events such as heat waves, droughts, and high precipitation events (IPCC, 2007). Increases in drought and heat will likely lead to a higher frequency of fire in the boreal zone resulting in significant losses of C from Histosols. Currently, disturbances, mainly fire, reduce the net C uptake of continental boreal peatlands by about 85% (Turetsky et al., 2002). Predicted increases in fire frequency and intensity could lead to an even greater proportion of C uptake being balanced by fire, and boreal peatlands may actually become a net C source to the atmosphere (Turetsky et al., 2002).

33.2.3.3 Topography

While climate controls the occurrence of Histosols on a regional scale, topography controls where they occur on a given landscape. The classical soil-forming factor of topography is considered broadly here to incorporate relief, geomorphic setting, and hydrologic setting. Histosols occur where the setting facilitates concentration of runoff, discharge of groundwater, or retention of precipitation (Figure 33.3a through c). These conditions are most often satisfied in topographic depressions or very flat areas. On plains with low-permeability substrates in the boreal zone, Histosols may cover entire interfluves (Sjors, 1963; Walter, 1977), and in extremely humid climates, such as the British Isles and southeastern Alaska, Histosols may also occur on gentle slopes (Moore and Bellamy, 1974; Sjors, 1985; Ward et al., 2007). Histosols of floodplains receive suspended mineral matter during floods, while other Histosols obtain only dissolved material from source waters or inputs of eolian materials.

In lower latitude coastal areas, rising sea level has caused brackish or saline waters to engulf drowned river valleys or to extend over formerly upland soils. This has led to the formation of coastal marsh Histosols in several geomorphic settings (Darmody and Foss, 1979). Figure 33.4 shows a schematic of a cross section of a submerged upland type marsh where organic soil materials are accumulating over what were previously upland soils, forming Histosols. As sea level continues to rise, the margin of the marsh is pushed landward, and the organic materials continue to thicken such that older and deeper Histosols generally exist closer to the open water (Rabenhorst, 1997).

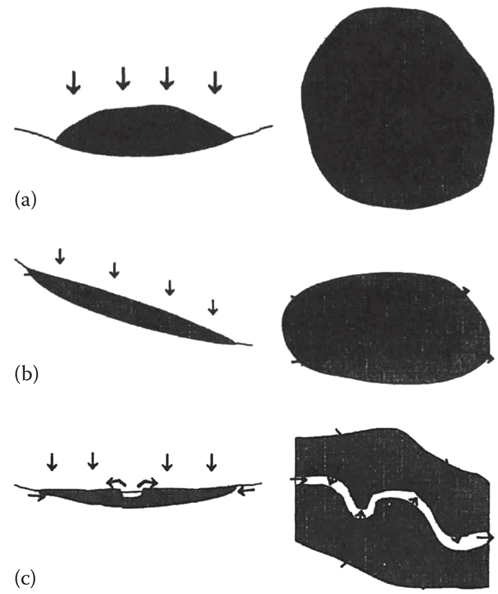

FIGURE 33.3 Settings of peatlands. Peat is shown in gray, and arrows indicate the direction of water movement (a) Bog. The domed surface of the bog precludes input of runoff or groundwater containing bases dissolved from minerals. All water is derived from the atmosphere and evaporates or runs radially off the bog. (b) Fen or swamp. Runoff and groundwater from mineral soils surrounding the peatland supplement precipitation on the peatland. (c) Floodplain fen or swamp. The peatland receives precipitation and runoff or groundwater from mineral soils adjacent to the peatland. Water seeps from the peatland into the stream during periods of low water, while during floods suspended mineral matter is deposited on the peatland, increasing the content of mineral matter in the peat.

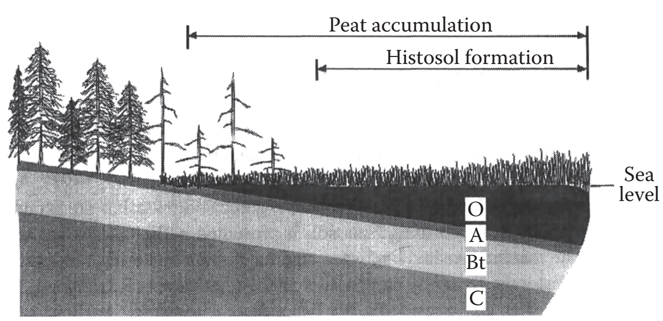

FIGURE 33.4 A schematized cross section of a submerged-upland type marsh showing development of O horizons and a Histosol (probably a Sulfihemist) over what was formerly an upland soil with an argillic (Bt) horizon.

The formation of Histosols, like other soils, is a function of topography, but Histosols are unique among soils in that their formation also modifies the topography. Accumulation of organic soil material can fill depressions and create gentle topographic highs where the topography was once level or depressional (Figure 33.5). The development of a topographic high due to peat accumulation can prevent base-rich water from moving onto the peatland and thereby facilitate formation of a highly

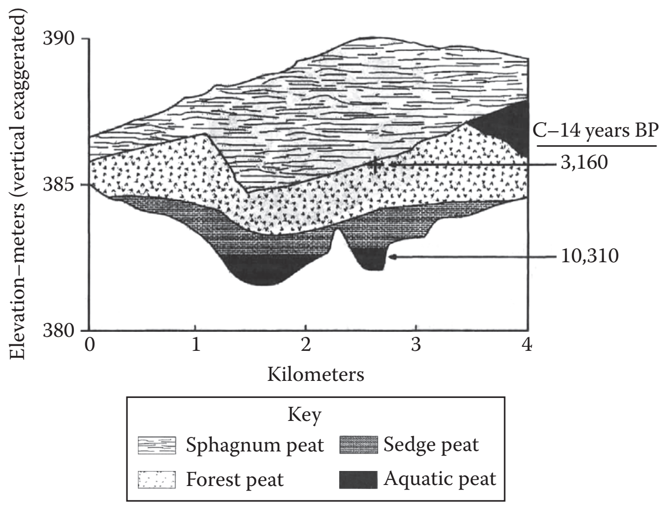

FIGURE 33.5 Stratigraphic cross section of a bog in the Myrtle Lake peatland, northern Minnesota. (Modified from Heinselman, M.L. 1970. Landscape evolution, peatland types and the environment in the Lake Agassiz Peatlands Natural Area, Minnesota. Ecol. Monogr. 40:235–261.)

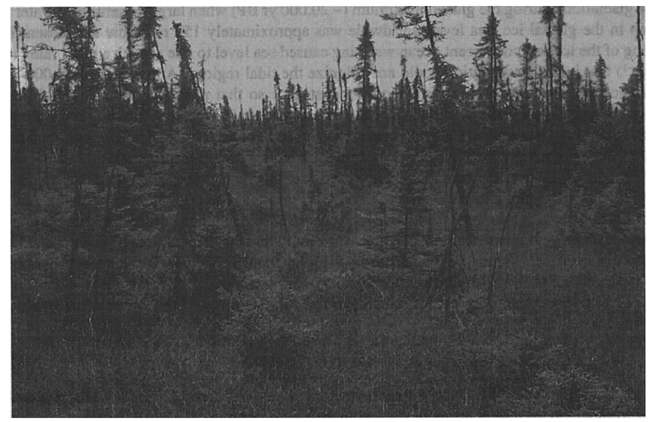

FIGURE 33.6 Typical bog vegetation on a highly acidic Histosol in boreal North America. Trees are stunted black spruce (*Picea mariana*); the largest trees visible are about 4 m tall. Understory plants include ericaceous dwarf shrubs, sedges (*Carex* sp.), and a continuous cover of *Sphagnum* sp. moss. The soil is a Typic Borohemist. Toivola peatland, northeastern Minnesota.

acidic bog. Peat accumulation also produces intriguing microtopography on some peatlands, notably a pattern of ridges (or *strings*) and pools (Foster et al., 1983; Seppälä and Koutaniemi, 1985; Swanson and Grigal, 1988) (or *flarks*), the cause of which is still debated (Washburn, 1980).

Histosols frequently form by accumulation of organic matter in basins of lakes or ponds, a process known as *lakefill* or *terrestrialization*. Histosols may also form by *paludification*, the expansion of wetland onto what was originally drier soils. Both processes operated in the formation of the peatland depicted in Figure 33.5. Folists, unlike other Histosols, generally occur in mountainous regions (Reiger, 1983; Wakeley et al., 1996). Mountainous settings provide the high precipitation and underlying bedrock or fragmental material required for Folist formation.

33.2.3.4 Vegetation

The flora of Histosols varies widely as a result of the great range in climates over which they occur. Moreover, while most Histosols owe their existence to saturation of the soil by water, the depth at which saturation occurs is variable and this has a major effect on the vegetation. The vegetation of bogs in the worldwide circumboreal zone is remarkably uniform, apparently due to the limited number of plants that can tolerate the poor nutrient conditions, cold climate, and high water table of these soils (Figure 33.6). *Sphagnum* mosses cover the ground, along with scattered sedges (*Carex*) and cotton sedges (*Eriophorum*). Low shrubs from the family Ericaceae are common, and trees are usually present but stunted; black spruce (*Picea mariana*) is most widespread in North America and Scot's pine (*Pinus sylvestris*) in Eurasia. The vegetation of bogs in more southerly climates includes different species, but shares with northern bogs the *Sphagnum* moss and prevalence of nutrient-conserving evergreen plants (Anderson, 1983; Hofstetter, 1983).

The less-acid conditions of groundwater-fed Histosols permit a greater diversity of plants, many of which also occur on wet mineral soils. On minerotrophic peatlands with the water table continuously near the surface (fens), most vascular plants are *aerenchymous* grass-like plants (mainly sedges, family Cyperaceae) that can transport oxygen downward to their roots in hollow stems, allowing root metabolism in anaerobic soils (Crawford, 1983). Fens that exhibit an aerobic surface horizon present during at least part of the growing season, allow growth of nonaerenchymous plants, including woody plants (Gill, 1970; Kozlowski, 1984). Though nutrient conditions for tree growth on these Histosols are superior to those of bogs, trees may be stunted due to short duration of aerobic conditions or thinness of the aerobic rooting zone.

Many of the plants that occur on Histosols are highly adapted to specific conditions of pH/nutrients and high water table; thus, plants are useful indicators of these conditions (Heinselman, 1963, 1970; Sjors, 1963; Jeglum, 1971; Vitt and Slack, 1975; Vitt and Bayley, 1984; Andrus, 1986; Janssens and Glaser, 1986; Swanson and Grigal, 1989, 1991; Glaser, 1992; Janssens, 1992). Because the high water table restricts rooting of most plants to near-surface soil horizons, vegetation is a useful indicator of pH/nutrient conditions near the surface but not at depth.

Because the vegetation actually creates most of the soil material in Histosols, composition of the vegetation that formed the soil exerts strong control over its physical and chemical properties. Peats are commonly divided into three broad groups on the basis of botanical composition: moss, herbaceous (sedge), and woody peat (Kivinen, 1977). The botanical origin of organic soil material may be determined by examination of plant remains (Birks and Birks, 1980; Janssens, 1983; Levesque et al., 1988).

The changes in soil drainage and trophic conditions on peatlands that accompany peat accumulation affect the vegetation and hence botanical composition of the peat. As peat

TABLE 33.4 Rates of Peat Accretion/Accumulation in Various Organic Soils

Location	Site Characteristics	Peat Accretion–Accumulation Rate (mm year^{-1})	References
Alberta, Canada	Fen, last 50 years	3.2–6.4	Vitt et al. (2009)
Northwest Territories, Canada	Rich fen	0.38	Robinson (2006)
	Poor fen	0.41	
	Bog	0.47	
West Siberia, Russia	Forested bog	0.40–0.70	Peregon et al. (2007)
	Shrub bog	0.25–0.60	
Western Canada	Bog	3.0	Turetsky et al. (2007)
British Columbia, Canada	Sloping, open peatland	0.94–1.7	Asada and Warner (2004)
Southern Sweden	Bog	5.2–5.9	Belyea and Malmer (2004)
West Siberia, Russia	River valley fen	0.84	Borren et al. (2004)
	Raised bog	0.67	
	Forested fen	0.42	
	Bog	0.37	
	Flow-through fen	0.38	
	Flow-through fen	0.35	
	Bog	0.42	
	Bog	1.13	
North America		0.58	Gorham et al. (2003)
North Dakota, USA			
Minnesota, USA		0.41–0.79	
Quebec, Canada		0.59–0.83	
Maine, USA		0.47–0.56	
New Brunswick, Canada		0.49–0.79	
Nova Scotia, Canada		0.40–0.48	
Newfoundland, Canada		0.39–1.05	
Alaska, USA		0.18–0.38	
West Siberia, Russia	Bog	0.57	Glebov et al. (2002)
Southeastern Norway	Bog	3–28	Ohlson and Okland (1998)
Bergslagen, Sweden	Raised bog	0.3–1.0	Derived from Almquist-Jacobson and Foster (1995)
Slave Lake, Canada	*Sphagnum* peat	0.3–0.6	Kuhry and Vitt (1996)
Eastern United States	Last 100 years		Wieder et al. (1994)
	Minnesota bog	2.4	
	Pennsylvania fen	1.4	
	Maryland fen	1.9	
	West Virginia bog 1	3.1	
	West Virginia bog 2	2.3	
West Greenland	Nearshore peat	0.43	Bennike (1992)
Subarctic and boreal Canada	Based on 138 basal ^{14}C dates	0.31–0.54	Gorham (1991)
S. Sweden and N. Germany		0.70	Tolonen (1979) (after Gorham, 1991)
S. and Central Finland		0.75	Tolonen (1979) (after Gorham, 1991)
N. Europe		0.60	Aaby (1986) (after Gorham, 1991)
Boreal USSR	Raised bogs	0.6–0.8	Botch et al. (1983)
Siberian USSR	Palsa province	0.2–0.4	Botch and Masing (1983)
Eurasia		0.52	Zurek (1976) (after Gorham, 1991)
Maine, USA		0.35–0.75	Tolonen et al. (1988) (after Gorham, 1987)
Minnesota, USA	Red Lake, Minnesota	0.85–1.15	Gorham (1987)
Los Angeles, USA	Coastal marsh		Derived from Nyman and DeLaune (1991)
	Fresh	6.5–8.5	
	Brackish	5.9–9.5	
	Saline	7.5–7.6	

TABLE 33.4 (continued) Rates of Peat Accretion/Accumulation in Various Organic Soils

Location	Site Characteristics	Peat Accretion–Accumulation Rate (mm $year^{-1}$)	References
Chesapeake Bay, MD	Coastal marsh, Brackish	3.3–7.8	Derived from Kearney and Stevenson (1991)
Chesapeake Bay, MD	Coastal marsh, Brackish	3.5–7.5	Griffin and Rabenhorst (1989)
Chesapeake Bay, MD	Coastal marsh, Brackish	1.4–3.2	Hussein (1996)
	Current (^{210}Pb)	0.5–1.1	
	Long term (^{14}C)		
Louisiana, USA	Barataria basin, Coastal marsh	7–13	Hatton et al. (1983)
Massachusetts, USA	Barnstable coastal Marsh	1.1–2.6	Redfield and Rubin (1962)

accumulates, the rooting zone often becomes more and more isolated from mineral nutrient sources, and portions of the peatland may become drier as the ground surfaces rises. A cross section of a bog in Minnesota shows how a lake was filled in with aquatic peat (i.e., limnic material; Figure 33.4). Then, the peatland expanded onto what was originally dry land of the lakeshore as minerotrophic sedge peat was followed by minerotrophic forest peat as the surface rose and became drier. By about 3000 years ago, the center of the peatland became isolated from minerotrophic water, allowing accumulation of highly acidic *Sphagnum* moss peat, which subsequently expanded over the entire peatland and continues to accumulate today. Such changes in vegetation over time in peatlands make it difficult to predict subsurface peat properties from surface vegetation (Swanson and Grigal, 1989).

33.2.3.5 Time

Essentially all extant Histosols have formed since the end of the Pleistocene epoch. Most northern latitude Histosols occupy regions which were covered by glaciers during the last ice age and have formed following the glacial retreat. Reported average rates of peat accumulation in northern bogs and fens have been as high as >3 mm $year^{-1}$, but more typically fall in the range of 0.2–0.7 mm $year^{-1}$ (Table 33.4). These average rates usually are based on basal ^{14}C dates, and actual rates may have been higher or lower during particular periods.

Although distant from the glacial activity, coastal Histosols at lower latitudes were also impacted by the glaciation. During the glacial maximum (approximately 20,000 years ago), sea level worldwide was approximately 150 m below the present level when large quantities of water were tied up in the glacial ice. Melting of the ice and concurrent ocean warming caused sea level to rise at such a rapid rate (10–20 mm $year^{-1}$) that initially vegetation could not colonize the tidal regions. Approximately 3000–5000 years ago, sea level rise slowed to a more modest pace such that marsh vegetation could become established and organic parent materials began to accumulate (Bloom and Stuvier, 1963; Redfield, 1972). As sea level has continued to rise, organic materials have accumulated in Histosols, and coastal marshes and mangrove systems generally have been thought to have accreted at approximately the rate of sea level rise. In addition to the eustatic sea level rise, sediment in transgressing coastal regions is subsiding (e.g., along the Atlantic and Gulf coasts of the United States). The combination of rising sea level (presently estimated at 1 mm $year^{-1}$ worldwide) and coastal subsidence can be joined to yield an apparent sea level rise, which is substantially greater. Estimates of peat accretion in coastal areas generally range from 3 to 8 mm $year^{-1}$, which are much higher than in noncoastal regions, with even higher rates reported in rapidly subsiding areas (Table 33.4). Current evidence suggests that the highest rates of sea level rise may be too great for marsh systems to maintain, and that some of these areas are suffering marsh loss (Kearney et al., 2002).

33.2.4 Morphological Properties of Histosols

Most organic soil material is derived from terrestrial plants, and soil particles initially resemble the plants from which they were derived. As decomposition progresses, the organic matter is converted into a homogenous, dark-colored mass. Some organic soil materials are derived from aquatic plants and animals that accumulate on the bottom of water bodies, producing *limnic materials* (Finney et al., 1974; Soil Survey Staff, 2010).

The most obvious morphological properties of organic soil horizons are related to their degree of decomposition. Master horizons Oi, Oe, and Oa are used to designate fibric, hemic, and sapric horizons, respectively (Table 33.5) and are defined by the portion of the soil material, which retains discernable plant fibers after rubbing and by the color of a Na pyrophosphate extracting solution (Soil Survey Staff, 2010). Other methodologies and rating scales have been developed and utilized for evaluating degree of decomposition; the one most broadly used in Europe is the Von Post scale, which ranks organic materials on a scale of decomposition from 1 to 10 based on soil color, quantity of recognizable fibers, and the proportion of material remaining in one's hand when the sample is squeezed. Von Post scale numbers 1–4 correspond approximately to fibric material, 5–7 to hemic, and 8–10 sapric (Von Post and Granlund, 1926; Clymo, 1983).

There are a number of soil properties which are related to the degree of decomposition of the organic materials, including color and a variety of physical and chemical properties. Field moist colors of sapric organic soil material are usually

TABLE 33.5 Defining Morphological Criteria for Histosol Organic Horizons

Horizon Designation	Type of Material	Common Descriptor	Volumetric Rubbed Fiber (RF) Content	Color of Pyrophosphate Extract (Value/Chroma)
Oi	Fibric	Peat	RF > 3/4 or RF > 2/5	7/1, 7/2, 8/1, 8/2, 8/3
Oe	Hemic	Mucky peat	2/5 > RF > 1/6	Does not otherwise qualify for either fibric or sapric materials
Oa	Sapric	Muck	RF < 1/6	Below or right of line drawn to exclude blocks 5/1, 6/2, 7/3

Source: Soil Survey Staff. 2010 Keys to soil taxonomy. 11th edn. USDA-NRCS, U.S. Government Printing Office, Washington, DC.

nearly black. Fibric peat is lighter colored and often reddish in hue, while hemic peat has an intermediate color (Table 33.6). The relation between the degree of decomposition and various physical properties is discussed later under the section on physical properties.

Undrained Histosols typically lack pedogenic structure as it is usually defined, although flattened plant remains and stratification commonly produces a plate-like structure (Lee and Manoch, 1974). This sedimentary structure often results in Histosols having different shear strength and hydraulic conductivity in horizontal and vertical directions (MacFarlane and Williams, 1974; Rycroft et al., 1975). Drained Histosols may develop pedogenic structure in the man-made aerobic zone, such as granular structure due to earthworm casts and blocks or prisms due to cycles of wetting and drying (Lee and Manoch, 1974).

The mineral soil material that occurs under the peat in Histosols is typically chemically reduced as a result of saturation by water and the abundance of organic matter above it. Where paludification has occurred, soil horizons of prior mineral soils may be buried beneath the peat. In some cases, the preceding mineral soil pedogenesis may facilitate paludification by producing low-permeability horizons such as placic horizons (Ugolini and Mann, 1979; Klinger, 1996). The depth to mineral soil underlying drained Histosols has been measured with some success by ground-penetrating radar (Shih and Doolittle, 1984; Collins et al., 1986; Sheng et al., 2004; Kettridge et al., 2008).

The morphology of Folists differs from that of other Histosols in that the peat layers are thinner and underlain by fragmental material or bedrock (Witty and Arnold, 1970; Everett, 1971; Lewis and Lavkulich, 1972). Folists drain freely and as a result are less saturated than other Histosols (not saturated for >30 cumulative days per year; Soil Survey Staff, 2010).

TABLE 33.6 Color of Organic Soil Material in Relation to Degree of Decomposition[a]

Degree of Decomposition	Median Munsell Color (Hue Value/Chroma)	*N*
Sapric	10 YR (2/1)	69
Hemic	10 YR (3/2)	49
Fibric	7.5 YR (3/2)	18

[a] Data include organic horizons of Histosols in the USDA Natural Resources Conservation Service, National Soil Survey Laboratory characterization database.

33.2.5 Micromorphology of Histosols

Micromorphological observations can provide a direct examination of the structural integrity of plant fragments and components in organic soil materials. Levesque and Dinel (1982) and Fox (1985) have summarized the characteristics of organic soil materials at various stages of decomposition (fibric, hemic, and sapric materials). Fibric materials mainly show unaltered or slightly altered plant tissues without appreciable darkening and with little organic fine material. The plant fragments, which are only slightly decomposed, appear to be loosely arranged with a porous and open structure. Partially decomposed (hemic) materials also possess a fibrous appearance, and most fragments show incomplete degradation. The development of brown or black colors in the plant tissues is typical. Fine organic material is also present intermixed with, or adhering to, the coarser fragments of plant tissue (Figure 33.7). Fecal pellets, which are evidence of faunal activity, can also be common. In the most highly decomposed (sapric) materials, organic fragments are sufficiently darkened and decomposed, so that identification of botanical origin is not possible. Fine organic material is usually the dominant component although faunal excrement is also common.

The effects of draining organic soils can sometimes be seen during microfabric examination. When an undrained sphagnum

FIGURE 33.7 Thin section showing organic material from hemic Oe horizon (80–88 cm) of a Typic Sulfihemist in a coastal marsh of Chesapeake Bay; the organic material reflects an intermediate degree of decomposition with some discernable plant structures and cell components intermixed with decomposed organic material; frame length = 5 mm; PPL.

peat profile in Ireland was compared with those which had been drained for between 10 and 100 years, the drained profiles had undergone substantial alteration and decomposition leaving little of the original tissue structures (Hammond and Collins, 1983). The fine organic material was dominant, showing some biological granulation (Pons, 1960; Lee, 1983). The change in microfabric materials directly corresponded to an increase in density of the material.

In another study, Lee and Manoch (1974) concluded that 50 years of drainage and cultivation of organic soils led to significant decomposition and the development of pedogenic structure in the subsoil, whereas in the lower portion of the profile where the soil remained saturated, sedimentary structure persisted and the peat was more highly fibrous and less decomposed. The activity of soil fauna in the drained portions of the soil contributed to biological granulation and the formation of two distinct types of surface horizons. The *moder* mostly consists of faunal excrement and usually forms in oligotrophic peats, while the *mull* is formed by an intense mixing and binding of organic with mineral particles by larger organisms such as earthworms, and usually forms in mesotrophic or eutrophic peats.

While not widely reported, following drainage and cultivation of organic soils, illuvial humus termed *humilluvic material* (Soil Survey Staff, 2011) may accumulate in the lower parts of acid organic soils (Van Heuveln et al., 1960). Both the lower pH of the oligotrophic peat and the disturbance by cultivation apparently contribute to the dispersion of the organic fraction, which can then be translocated within the soil, and accumulate within the lower horizons of the peat, at the peat–mineral soil contact, or within the underlying mineral soil material.

33.2.6 Classification of Histosols

The definition of *organic materials* for saturated soils requires a minimum of 12% OC if there is no clay, and a minimum of 18% OC if the soil contains 60% or more clay, with a sliding scale for intermediate textures. Those soils which are not saturated must contain at least 20% OC to be considered organic soil materials (Soil Survey Staff, 1998, 2011). For a soil to be classified as a Histosol, at least 40 cm of the upper 80 cm must comprise organic materials, and it must not have permafrost within 1 m of the surface. However, if the organic materials are especially low in density ($<0.1\ \mathrm{g\ cm^{-1}}$), then at least 60 cm of the upper 100 cm must be organic materials. Histosols may be buried by as much as 40 cm of overlying mineral soil materials and still be considered Histosols.

The types of differentiating characteristics used to discriminate between classes of soils at the various categorical levels are presented in Table 33.7. Basically, organic soils that are not saturated for extended periods are classified as Folists, while the saturated organic soils are differentiated according to the degree of decomposition of the organic materials in the subsurface tier (the zone approximating 40–100 cm) into Fibrists, Hemists, or Saprists. Within the United States, some 314 soil series have been established for Histosols. Table 33.8 shows the distribution of those 314 series among the classes of the various

TABLE 33.7 Criteria Utilized in the Classification of Histosols

Suborder	Great Group	Subgroup	Family
Degree of saturation with water	Soil temperature regime	Thickness of organic materials (Terric vs. Typic)	Particle size and mineralogy (used only for Terric subgroups or for those containing Limnic materials)
Degree of decomposition of the subsurface tier	Special components (sphagnum fibers, sulfidic materials or sulfuric horizon, humilluvic materials)	Underlying materials Special materials contained (Limnic) Intergrades to other great groups (Cryic and Sphagnic)	Reaction (pH in 0.01 M $CaCl_2$) Temperature regime Soil depth (only used if <50 cm deep)

TABLE 33.8 Number of Soil Series in the United States That Are Classified into Particular Taxonomic Groups

Suborders		Great Groups		Subgroups	
Formative Element	Number of Series	Formative Element	Number of Series	Formative Element	Number of Series
Fibrists	26	Cryo	58	Fluvaquentic	14
Folists	35	Haplo	198	Limnic	19
Hemists	82	Sphagno	7	Lithic	37
Saprists	171	Sulfi	17	Terric	124
		Udi	19	Typic	102
		Usti	7	Hemic	7
		Others	8	Others	11
Total	314	Total	314	Total	314

taxonomic categories. The number of series that exist within a particular class may result from many factors and should not be taken to represent the areal extent of those soils. Histosols classified at the family level are differentiated into classes based upon particle size and mineralogy (used only for terric subgroups or for those containing limnic materials), reaction (pH in 0.01 M $CaCl_2$), temperature regime, and soil depth (only used if <50 cm deep).

33.2.7 Biological and Chemical Properties

Organic carbon contents are generally higher and nitrogen contents lower in less-decomposed peats than highly decomposed peats (Table 33.9). As a result, C:N ratios are generally higher for less-decomposed peats (Table 33.9; Lee et al., 1988). The ash content (mineral component) is also higher for more highly decomposed peats (Table 33.9; Lévesque et al., 1980; Lee et al., 1988). Thus, more highly decomposed peats are generally more fertile than less-decomposed peats. Some drained, sapric peats may supply nitrogen in excess of crop requirements without fertilization. However, nutrients derived from minerals, such as phosphorus, potassium, and most micronutrients, are frequently deficient in Histosols (Lucas, 1982; Yefimov, 1986).

The chemical and physical properties of peats are also related to their botanical composition. Peats derived mostly from *Sphagnum* mosses tend to be more acid, less decomposed, contain less ash, have lower cation exchange capacity (CEC), and have lower bulk density than woody peats; sedge peats typically have intermediate properties (Farnham and Finney, 1965; Lévesque et al., 1980).

33.2.7.1 Soil Carbon

Interest in the global C cycle has focused attention on the high proportion (3/4) of terrestrial C stored in soils (Lal et al., 1995). As a group, wetland soils maintain a disproportionately high level of soil carbon, and Histosols, which are composed largely of organic matter, clearly store the largest quantities of soil carbon. Although peatlands only occupy approximately 4% of the global land surface, they store about 30% of the global soil C (Lavoie et al., 2005). While typical agricultural soils may contain between 2 and 10 kg C m^{-2}, reported values for Histosols typically are an order of magnitude greater, with some values >200 kg C m^{-2} (Table 33.10). The quantity of C stored in some very deep Histosols is undoubtedly even higher.

Histosols are very dynamic and may be particularly significant in the overall terrestrial C budget. Many Histosols continue to sequester C at significant rates. This is particularly true for soils of coastal marshes, where rising sea level continues to power the engines of marsh accretion and C storage. Therefore, Histosols are generally viewed as a significant C sink although some studies indicate that climate change either through increased frequency and intensity of fire (Turetsky et al., 2002) or through elevated decomposition as a result of rising temperatures (Billett et al., 2004) is possibly switching some peatlands to sources of C to the atmosphere. In addition, if Histosols are drained or in some other way exposed to an aerobic environment, they may begin to oxidize and yield large quantities of C to the atmosphere (e.g., Nykanen et al., 1997), although other studies indicate that drainage can increase C sequestration as a result of greater aboveground and belowground productivity (e.g., Minkkinen and Laine, 1998).

Most of the discussion of possible global warming and greenhouse gas emission has focused on rising levels of CO_2 in the atmosphere. Methane (CH_4), however, is 32 times more efficient than CO_2 in trapping infrared radiation. Because many Histosols are strongly reducing (low Eh), they represent an ideal environment for the formation of CH_4. In systems where SO_4^{2-} is more abundant in the soil solution, such as in coastal or estuarine environments, sulfate reduction is favored over methanogenesis and methane production may be more limited (Bartlett et al., 1987; Widdell, 1988; Dise and Verry, 2001). However, in many freshwater or inland areas, Histosols may be the locus of significant methane emission to the atmosphere (Table 33.11). Minerotrophic fens have higher methane

TABLE 33.9 Chemical Properties of Organic Soil Material as Related to Degree of Decomposition: Mean (Standard Deviation, N)[a]

Property	Sapric	Hemic	Fibric	AOV Probability[b]
Organic carbon, g kg^{-1}	313 (128, 129)	347 (135, 61)	372 (130, 26)	0.055
Total nitrogen, g kg^{-1}	18 (9, 131)	16 (6, 54)	14 (5, 23)	0.058
C:N ratio	21 (10, 113)	25 (11, 48)	27.5 (10, 20)	0.007
CEC, cmol kg^{-1}	101 (44, 129)	88 (41, 61)	83 (33, 28)	0.046
CEC, cmol L^{-1}	76 (42, 28)	44 (25, 25)	21 (2, 5)	0.000
Ash, g kg^{-1}%	250 (nd)	178 (110, nd)	100 (50, nd)	nd
pH	5.1 (1.2, 143)	4.9 (1.2, 59)	4.5 (1.2, 27)	0.024
Al, mol Al mol^{-1} TEA[c]	0.074 (0.076, 49)	0.022 (0.024, 23)	0.038 (0.050, 16)	0.004

[a] Data (except for ash) is for all organic horizons of Histosols in the USDA Natural Resources Conservation Service, National Soil Survey Laboratory characterization database. Ash content is taken from Lee et al. (1988; data for 1300 samples of Wisconsin Histosols, nd—no data).

[b] F-test probability from one-way analysis of variance (AOV) between fibric, hemic, and sapric peats.

[c] KCl extractable Al divided by total NH_4 acetate extractable acidity.

TABLE 33.10 Carbon Storage Values for Organic Soils

Site Characteristics	Location	Carbon Accumulation Rate (kg m^{-2} year^{-1})	Quantity of Stored C (kg m^{-2})	References
10 year averages	West Siberia			Golovatskaya and Dyukarev (2009)
Pine peatland		0.021		
Stunted pine peatland		0.11		
Sedge fen		0.10		
Fen, last 50 years	Alberta, Canada	0.14–0.25		Vitt et al. (2009)
Various peatland types	Alberta, Canada		53–165, mean = 129	Beilman et al. (2008)
Bog, last 100 years	Western Canada	0.09		Turetsky et al. (2007)
Fen 1	Saskatchewan, Canada		29–210	Robinson (2006)
Fen 2			20–120	
Rich fen	Northwest territories, Canada	0.014		Robinson (2006)
Poor fen		0.018		
Bog		0.019		
Rich fen	Western Canada	0.025		Yu (2006)
Tropical peatlands	Micronesia	0.3		Chimner and Ewel (2005)
Sloping, open peatland	British Columbia, Canada	0.007–0.039		Asada and Warner (2004)
Bog	Southern Sweden	0.060–0.072		Belyea and Malmer (2004)
River valley fen	West Siberia	0.069		Borren et al. (2004)
Raised bog		0.033		
Forested fen		0.034		
Bog		0.027		
Flow-through fen		0.021		
Flow-through fen		0.020		
Bog		0.019		
Bog		0.040		
Variety peatland types	West Siberia		<30 to >300	Sheng et al. (2004)
Various peatland types	North Dakota, USA	0.038		Gorham et al. (2003)
	Minnesota, USA	0.021–0.038		
	Quebec, Canada	0.022–0.028		
	Maine, USA	0.022–0.030		
	New Brunswick, Canada	0.017–0.019		
	Nova Scotia, Canada	0.019–0.041		
	Newfoundland, Canada	0.008–0.016		
	Alaska, USA			
Raised bog	Southern Sweden			Malmer and Wallen (1999)
Hummock		0.05–0.18		
Hollow		0.03–0.15		
Last 100 years	Eastern USA			Wieder et al. (1994)
Minnesota bog		0.16		
Pennsylvania fen		0.14		
Maryland fen		0.15		
West Virginia bog 1		0.15		
West Virginia bog 2		0.18		
Coastal marsh	Los Angeles, USA			Derived from Nyman and DeLaune (1991)
Fresh		0.17–0.22		
Brackish		0.17–0.27		
Saline		0.21–0.22		
Coastal marsh Brackish	Chesapeake Bay, MD	0.12–0.42		Derived from Kearney and Stevenson (1991)
Coastal marsh Brackish	Chesapeake Bay, MD		59 (range 18–166)	Derived from Griffin and Rabenhorst (1989)
Barataria Basin, coastal marsh	Louisiana, USA	0.18–0.30		Smith et al. (1983)
Coastal marshes	Atlantic and Gulf Coasts, USA		64 (range 9–191)	Rabenhorst (1995)
Sphagnum peat	Slave Lake, Canada	0.014–0.035		Kuhry and Vitt (1996)
Based on 138 basal ^{14}C dates	Subarctic and Boreal Canada	0.023–0.029		Gorham (1991)

TABLE 33.11 Reported Fluxes of CO_2 and CH_4 Emissions from Histosols

			CO_2 Emission Rate (mmol m^{-2} day^{-1})		CH_4 Emission Rate (mmol m^{-2} day^{-1})		
Location	Site Details	Notes	Mean	Range	Mean	Range	References
West Siberia	10 year averages	Growing season					Golovatskaya and Dyukarev (2009)
	Pine peatland		45.0		0.030		
	Stunted pine peatland		28.5		0.045		
	Sedge fen		29.9		3.30		
Quebec, Canada	Poor fen—control	Growing season		42–250		0.95–3.55	Strack and Waddington (2007)
	Poor fen—w/water table drawdown			46–242		0.63–6.40	
Western Canada	Bog	Transplant experiment	55		0.4		Turetsky et al. (2007)
Northern England	Acidic blanket peat	Growing season				0.15–4.2 (range for all treatments)	Ward et al. (2007)
	Control			5.4–327			
	Burned			10.9–491			
	Grazed			5.4–436			
Micronesia	Forested peatland	Annual	198				Chimner and Ewel (2004)
	Cultivated for taro		110				
New York, USA	Conifer/maple peatland	Annual	103	8.6–216	0.05	−0.17–0.69	Coles and Yavitt (2004)
Ontario, Canada	Mesocosms—controlled temperature, water table, and humidity	Net production reported		6.1–602		−0.25–1.1	Blodau and Moore (2003)
Quebec, Canada	Poor fen—control	Growing season			8.9		Strack et al. (2004)
	Poor fen—w/water table drawdown				4.0		
Minnesota, USA	Poor fen—control	Growing season			15.1	6.2–31.9	Dise and Verry (2001)
	Ammonium nitrate added				16.0	7.5–30.0	
	Ammonium sulfate added				10.2	5.0–15.6	
Minnesota, USA	Bog	Mesocosms		275–2500		1.3–41.3	Updegraff et al. (2001)
	Fen	Growing season					
Quebec, Canada	Gatineau Park			1.97–7.24		1.15–2.18	Buttler et al. (1994)
Wales	Peat monoliths	Laboratory study	14.7	9.6–21.0	14.4	4.7–34.4	Freeman et al. (1993)
Finland	Natural fen	Annual	11.3		3.49		Nykanen et al. (1995)
	Drained fen		30.8		0.03		
Finland	Ombrotrophic	12 C	88.2	42.5–141.3			Silvola et al. (1996)
Alaska, USA		Summer measurements			9.2		After Gorham (1991), after Crill et al. (1988)
Boreal Canada	Swamp (n = 20)	Annual averages			0.21		Derived from Moore and Roulet (1995)
	Fen (n = 6)				0.57		
	Bog (n = 13)				0.39		
Canada (lab study)	Bog flooded	19–23 C	0.005		0.012		Derived from Moore and Knowles (1989)
	Bog drained		0.19		0.006		
	Fen flooded		0.009		0.58		
	Fen drained		0.14		0.025		
Alaska	Moist tundra	August			0.3		Derived from Sebacher et al. (1986)
	Waterlogged tundra				7.4		
	Wet meadows				2.5		
	Alpine fen				18		
Northern Sweden	Ombrotrophic bog	Summer					Svensson and Rosswall (1984)
	Hummocks		10.12		0.05		
	Between hummocks		14.92		0.14		
	Shallow depressions		11.61		0.77		
	Deeper depressions		12.45		1.21		
	Ombrominerotrophic		12.49		2.73		
	Minerotrophic fen		11.48		16.89		

TABLE 33.11 (continued) Reported Fluxes of CO_2 and CH_4 Emissions from Histosols

			CO_2 Emission Rate (mmol m^{-2} day^{-1})		CH_4 Emission Rate (mmol m^{-2} day^{-1})		
Location	Site Details	Notes	Mean	Range	Mean	Range	References
Minnesota	Bog Fen	Sampled during August			8.2 0.25	1.2–29.2 0.19–0.31	After Harriss et al. (1985)
Georgia, USA		During midsummer			6.6		After Gorham (1991), after Crill et al. (1988)
West Virginia, USA	Mountain bog	During midsummer			11.7		After Gorham (1991), after Crill et al. (1988)
Minnesota, USA	Forest bog Forest fen Open bog Neutral fen Acid fen	Summer measurements			3.6 6.7 14 15 4.8	0.5–33 3.2–12 0.9–41 7.1–33	After Crill et al. (1988), after Mitsch and Wu (1995)
Minnesota, USA	Open poor fen Open bog Forest bog hollow Hummock	Winter measurements			3.0 0.7 0.8 0.3		Dise (1992)
Virginia	Coastal marsh York river Chesapeake Bay estuary	Summer Creek Bank Short Spartina High marsh			0.45 0.29 0.09	0.13–0.82 0.05–0.87 0–0.36	Derived from Bartlett et al. (1985)
Virginia	Coastal marsh Tidal Creek in Chesapeake Bay estuary	Summer Low salinity Moderate salinity High salinity			 11 7.5 2	 5–16 4–11 0.5–2.5	Derived from Bartlett et al. (1987)
West Virginia	Appalachian bog		127	75–250	17.0	0–53	Wieder et al. (1990)
Maryland	Appalachian bog		152	100–250	4.4	0–12	Wieder et al. (1990)
Louisiana, USA	Barataria basin and Coastal marsh	Annual averages		41–141			Smith et al. (1983)
Florida, USA		During midsummer			6.0		After Gorham (1991), after Crill et al. (1988)
Malaysia	Ombrotrophic bog		170				Wosten et al. (1997)
Malaysia	Drained and cultivated peatland			139–727			Murayama and Bakar (1996)

emissions than ombrotrophic bogs (Dise, 1993). Methane emission from Histosols seems to be directly related to the location of the water table, with greater generation when soils are saturated to or above the surface (Updegraff et al., 2001; Strack et al., 2004). Greater methane emissions occur when soils are saturated. If aerobic zone exists in the profile, soil microbes will utilize the methane as it passes through on its way to the soil surface. Carbon dioxide emissions, which are produced by oxidation of soil organic matter, have been found to be both greater (Nykanen et al., 1997) and unchanged (Updegraff et al., 2001; Strack and Waddington, 2007) following the lowering of water tables.

33.2.7.2 Sulfides

The biogeochemical environment in which Histosols form can also be conducive to the formation of sulfides. The occurrence of iron sulfides can, in some circumstances, lead to the generation of extreme acidity and acid sulfate soils (Van Breemen, 1982). Sulfate reduction generally requires the presence of organic matter, which serves as an energy source, low redox potentials, sulfate that functions as an electron acceptor, and sulfate reducing bacteria (Rickard, 1973). If sulfide is formed in the presence of a reactive iron source, then such minerals as pyrite (FeS_2) can form. The C source and anaerobic conditions are almost always present in Histosols, but sulfate (SO_4^{2-}) levels may vary dramatically among environments. Many inland Histosols receive SO_4^{2-} only in small amounts as atmospheric deposition, and under these circumstances, sulfidization (Rabenhorst and James, 1992) is insignificant, and most of the S in those soils is bound in organic S forms (Novak and Wieder, 1992). Some inland peats have developed acid sulfate conditions, although usually they are associated with deposits of coprogeneous earth (Lucas, 1982). Sulfate reduction is common in coastal Histosols, which contain an abundance of SO_4^{2-} from sea water. Extensive areas of these Sulfihemist soils have been identified along the Atlantic coast of the United States. The distribution of pyrite within coastal Histosols can be highly variable and is often related to microsite differences in the availability of either reactive iron or sulfide (Rabenhorst and Haering, 1989).

33.2.7.3 Acidity and Base Saturation

In bog Histosols, where inputs of bases are minimal because soil water is derived from rainfall that has never contacted mineral soil, base saturation is low (at least in the rooting zone) and the organic acids produced by partial decomposition of organic matter typically buffer soil water pH near 4; the soil pH in $CaCl_2$ is typically 3–4 (Figures 33.8 and 33.9; Gorham et al., 1985; Urban, 1987). Minerotrophic (fen) peats have higher base saturation,

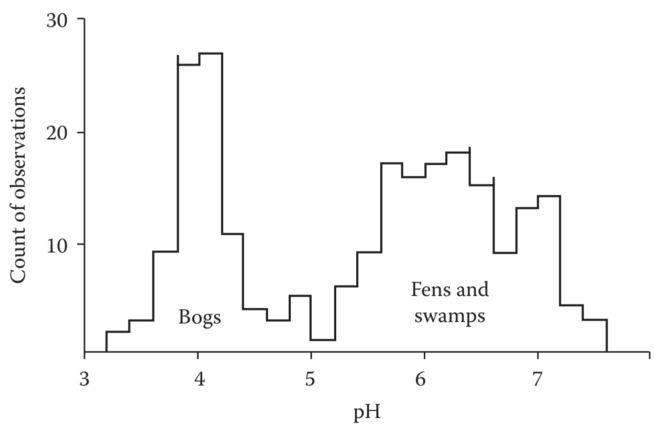

FIGURE 33.8 Frequency distribution of pH in peatland surface water for 232 sites in Minnesota. Samples with surface water pH near 4 are from bogs and those with pH above 5 are from fens (unforested) and swamps (forested). (From Swanson, D.K., and D.F. Grigal. 1989. Vegetation indicators of organic soil properties in Minnesota. Soil Sci. Soc. Am. J. 53:491–495.)

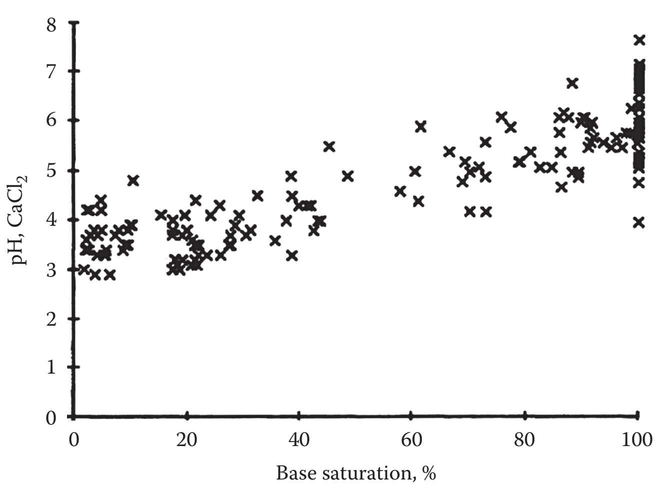

FIGURE 33.9 Relationship between base saturation and pH of organic soil horizons of Histosols. Peat samples with pH in $CaCl_2$ of less than 4.5 (presumably deposited in bogs) generally have less than 50% base saturation. Peat samples with pH greater than 4.5 (deposited in fens and swamps) typically have high base saturation. Data are for all organic horizons of Histosols in the USDA Natural Resources Conservation Service, National Soil Survey Laboratory characterization database.

and pH in water is generally above 5 ($CaCl_2$ pH above 4.5). The pH of minerotrophic peats is buffered by cation exchange within the soil (Bloom et al., 1983) or by carbonates if they are present. The aluminum ion, which is derived from silicate minerals, comprises a small proportion of the total acidity in most Histosols because of its low solubility at pH > 5.5 (Table 33.9). Less-decomposed peats generally have lower pH than more highly decomposed peats (Table 33.9; Lee et al., 1988).

The distinction between euic and dysic reaction classes at the family level in *Soil Taxonomy* separates the highly acid, bog peats (dysic) from less acid fen peats (euic) (Farnham and Finney, 1965). Because a pH (in $CaCl_2$) greater than 4.5 anywhere in the control section (i.e., anywhere within 130 or 160 cm of the surface) places the soil into the euic class, a bog with a highly acidic near-surface rooting zone may classify as euic rather than dysic due to the presence of higher pH horizons at depth in the soil.

Histosols which contain sulfide minerals such as pyrite have the potential to develop extreme acidity. Under saturated and anaerobic conditions, such sulfide bearing soils have circumneutral pH. If it is drained, dredged, or in some other way exposed to oxidizing conditions, the soil will undergo acid sulfate weathering, a process that is the result of sulfide mineral oxidation and the concomitant production of sulfuric acid (Van Breemen, 1982). The oxidation of pyrite can lead both to the extreme acidification of the soil (pH < 3.5) and also to the generation of acidity, which can be moved offsite through mobilization of acid generating soluble salts such as $FeSO_4$.

33.2.7.4 Cation Exchange Capacity

The CEC of Histosols is quite high as a result of the high cation exchange of organic matter (Table 33.9). The CEC is higher for more highly decomposed peats than fibric peats (Table 33.9; Lévesque et al., 1980). The difference between the CEC of sapric and fibric peats is even more dramatic if the CEC is expressed on a volume rather than mass basis, as a result of the very low bulk densities of fibric materials (Table 33.9). The CEC per unit soil volume is a more useful measure than CEC per unit soil mass in Histosols, because their bulk densities are very low and because roots occupy a volume rather than mass of soil. The CEC per unit volume of fibric peats, near 20 $cmol_c$ L^{-1}, is comparable to that of most mineral soils (5–20 $cmol_c$ L^{-1}) assuming mineral soil bulk density of 1.0–1.5 kg L^{-1} (Holmgren et al., 1993). Even when expressed per unit volume, the average CEC of sapric and hemic peats (Table 33.9) is much higher than the 5–20 $cmol_c$ L^{-1} of most mineral soils.

33.2.8 Physical Properties

33.2.8.1 Bulk Density

The physical properties of Histosols differ greatly from those of mineral soils. Bulk densities for organic soil materials generally are quite low, ranging from as little as 0.02 up to 0.8 g cm^{-3} (Table 33.12). Bulk density is related to the degree of decomposition

TABLE 33.12 Physical and Hydraulic Properties of Organic Soils

Location	Site Characteristics	Bulk Density (g cc^{-1})	Hydraulic Conductivity (10^{-5} cm s^{-1})	References
Alberta Canada	Rich fens	0.04–0.12		Vitt et al. (2009)
Western Canada	Bog	0.056		Turetsky et al. (2007)
Manitoba, Saskatchewan and Alberta, Canada	Various peatland types	0.07–0.26		Bauer et al. (2006)
Northwest Territories, Canada	Rich fen	0.03–0.14		Robinson (2006)
	Bog	0.08–0.20		
Western Canada	Rich fen	0.18		Yu (2006)
Micronesia	Tropical peatlands	0.11–0.13		Chimner and Ewel (2005)
West Siberia	Various peatland types	0.05–0.41		Sheng et al. (2004)
England	Blanket peats (fens)	0.15–0.27	0.01–1.04, mean = 0.24	Holden and Burt (2003)
New Zealand	Bog	0.06		Schipper and McLeod (2002)
Poland	Various peatland types	0.07–0.60		Bogacz (2000)
Finland	Drained and harvested fens		0.4–60	Klove (2000)
Southern Sweden	Bog			Malmer and Wallen (1999)
	Lichen hummock	0.18		
	Sphagnum hummock	0.27		
	Sphagnum lawn	0.27		
Finland	Various peatland types	0.12–0.16		Minkkinen and Laine (1998)
Southeastern Norway	Bog	0.2–0.8		Ohlson and Okland (1998)
Northern Minnesota	Undecomposed		3,810–15,000, mean = 8,650	Boelter (1965)
	Partially decomposed		13.9–132, mean = 63	
	Decomposed		0.9–15 mean = 5.1	
Wyoming	Mountain bog 46 cm	0.16–0.22	0.0277	Sturges (1968)
	91 cm		0.0185	
Ottawa and St. Lawrence River Valleys, Canada	Swamp and bog			Mathur and Levesque (1985)
	0–60 cm		624	
	0–100		366	
	0–125		269	
Northern Minnesota	Lost river peatland	0.06–0.14		Chason and Siegel (1986)
	Bog		25–560	
	Fen margin		150–2,600	
	Spring fen		67–1,600	
Quebec, Canada	Gatineau Park	0.03–0.10		Buttler et al. (1994)
Eastern New Brunswick, Canada	Sphagnum peat from raised bogs			Korpijaako and Radforth (1972)
	Van post scale 1–2	0.02–0.08	90–175	
	3–4	0.03–0.10	6.9–56	
	5–6	0.06–0.11	1.4–17	
	7–8	0.09–0.13	0.14–2.8	
Wisconsin, USA	Fibric	0.13		Lee et al. (1988)
	Femic	0.17		
	Sapric	0.20, 0.24		
Minnesota	Bog			Gafni and Brooks (1990)
	0–10 cm Von Post 1		23,495	
	10–20 cm Von Post 2–3		7,697	
	20–30 cm Von Post 4–5		5,498	
	30–40 cm Von Post 5–6		799	
	40–50 cm Von Post 5–7		995	
	Fen			
	0–10 cm		31,597	
	10–20 cm		5,000	
	20–30 cm		1,400	

of the organic materials with bulk density generally increasing as the materials become more highly decomposed (Figure 33.10). Because organic materials contain varying amounts of mineral matter, this also can affect the bulk density of organic soil horizons. For coastal marsh peat, the organic matter content is generally about twice the content of organic C, with the remainder representing the mineral fraction. Figure 33.11 illustrates the relationship between the mineral content (roughly the difference remaining from twice the OC content) and the bulk density for Oe and Oa horizons from Sulfihemists along the Atlantic Coast (Griffin and Rabenhorst, 1989).

33.2.8.2 Porosity, Hydraulic Conductivity, and Water Retention

Histosols have very high porosity levels, which can reach over 80% as shown by the water content at saturation (Figure 33.12; Boelter, 1969). The high porosity and low bulk density of organic soils would lead one to expect high rates of water transmission through Histosols. Weakly or undecomposed peat often has a fairly high hydraulic conductivity; however, as the material becomes more decomposed, the hydraulic conductivity decreases (Table 33.12). The hydraulic conductivity of sapric and some hemic peats is quite low, comparable to that of clay (i.e., 10^{-5} m s^{-1} or less). Because the peat in the lower part of Histosol profiles is often sapric or hemic, little deep percolation occurs and water tends to evaporate or move laterally through the less decomposed, surface horizons (Ivanov, 1981).

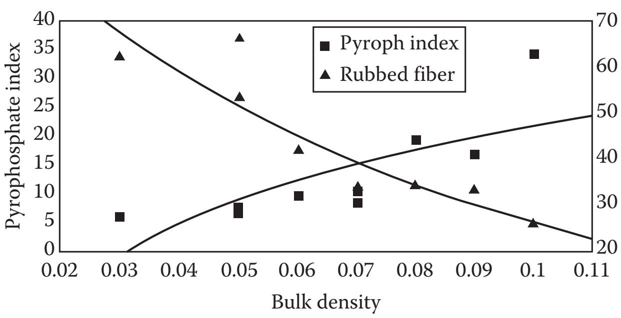

FIGURE 33.10 Relationship between bulk density of organic materials and indices of decomposition such as rubbed fiber content or pyrophosphate index. (Derived from Buttler, A., H. Dinel, and P.E.M. Levesque. 1994. Effects of physical, chemical and botanical characteristics of peat on carbon gas fluxes. Soil Sci. 158:365–374.)

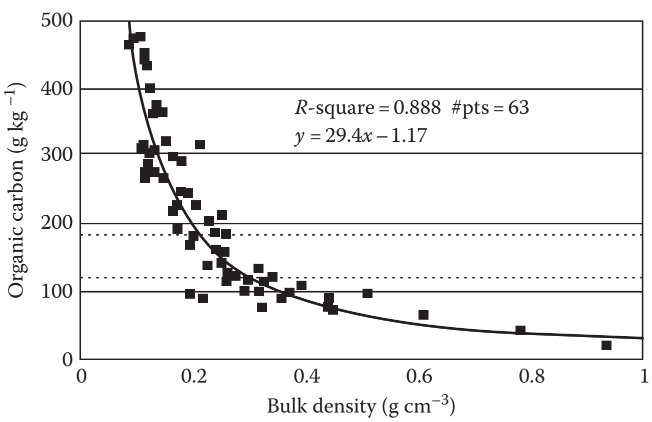

FIGURE 33.11 The effect of mineral content (roughly the difference remaining from 2× the OC content) on the bulk density of Oe and Oa horizons from Sulfihemists along the Atlantic Coast. Dashed lines represent 12% and 18% OC, which is necessary for soil materials to be considered organic. (Derived from Griffin, T.M., and M.C. Rabenhorst. 1989. Processes and rates of pedogenesis in some Maryland tidal marsh soils. Soil Sci. Soc. Am. J. 53:862–870.)

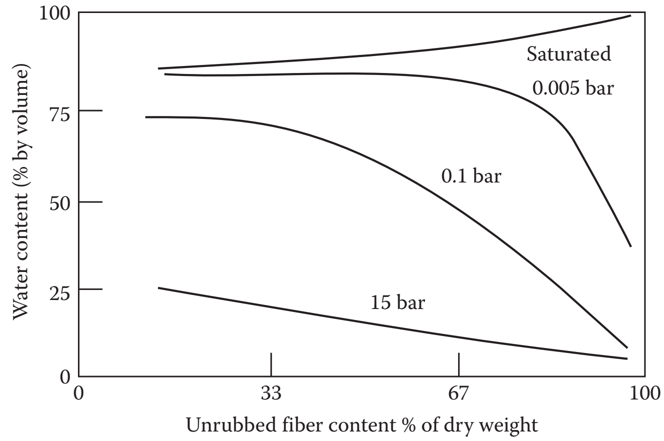

FIGURE 33.12 Volumetric soil water content as a function of unrubbed fiber content at various moisture tensions. (After Boelter, D.H. 1969. Physical properties of peats as related to degree of decomposition. Soil Sci. Soc. Am. J. 33:606–609.)

Water retention in Histosols is also closely related to the degree of decomposition (Figure 33.12). The large pores of slightly decomposed peats drain readily at low suction. In contrast, the fine pores of more well-decomposed peats retain considerable water at low suctions (Boelter, 1974). The water retention difference (water content at 0.01 MPa minus that at 1.5 MPa suction; approximates plant-available soil water capacity) of most Histosols is very high, approaching 50% in some peats and exceeding 25% in all others except the least-decomposed peats (Figure 33.12).

33.2.9 Utilization and Management of Histosols

33.2.9.1 Interest in Preservation as Coastal and Nontidal Wetlands

Apart from any benefits which may be achieved from managed systems, such as commercial forestry, grazing, or intensive agriculture, Histosols perform a number of environmental and ecological functions. Because most Histosols occur in wetland environments, they typically provide wetland functions including wildlife habitat, floodwater control, groundwater recharge, nutrient and biogeochemical cycling, ion sorption, purification of surface and shallow groundwater, as well as functioning as an important sink for terrestrial C. The importance of peatlands as paleoenvironmental and archaeological archives has also been documented (Godwin, 1981). Throughout the years, the value society has placed on these functions has been minimal, and the Histosols of peatlands and coastal marshes have been exploited and extensive areas have been destroyed. More recently, people have recognized that the functions which Histosols perform in a natural setting have significant benefit for society. Thus, legislation has been passed in the United States and elsewhere to preserve Histosols and other wetlands.

33.2.9.2 Histosols as Agricultural Resources

As was mentioned earlier, Histosols are utilized as important agricultural soils in many areas, so long as water tables can be effectively managed. For example, in Japan, there are over 70,000 ha of rice that is grown on peatlands. Peatlands in a number of tropical countries such as Cuba, Guyana, Malaya, and Indonesia have been reclaimed from mangroves and are used for the production of sugarcane (Moore and Bellamy, 1974). Within the United States, there has also been extensive growth of vegetables on peatlands in the Northern United States (such as Michigan), and there has also been extensive agricultural development in the Everglades region of Florida. The agricultural use of Histosols presents some special challenges regarding nutrient management and fertility, but the most significant problem is the high water tables requiring drainage.

33.2.9.3 Impacts of Drainage

Where Histosols have been converted for higher intensity land uses, such as in agriculture, horticulture, or silviculture, they are almost always drained to lower the water table. Such drainage results in a number of short-term effects, such as shrinkage and consolidation due to desiccation, the loss of the buoyant force of groundwater, and compaction. There is also ongoing consolidation and soil alteration caused by the enhanced decomposition of the organic materials following the shift from an anaerobic to an oxidizing regime (Stephens and Speir, 1969; Minkkinen and Laine, 1998). Reported rates of long-term peat subsidence range up to 10 cm year^{-1} but most reports are in the range of 2–5 cm year^{-1} (Table 33.13). This consolidation is accompanied by changes in the physical properties of the peat, including higher bulk densities and lower moisture contents.

33.2.9.4 Histosols as Energy Resources

Peat is mined mainly in northern Europe and used as fuel (Table 33.14). In addition, there is extensive mining and export of peat as a horticultural amendment. Important examples of the latter uses include ingredients for potting soils, and mixed fertilizers, components of mushroom beds, as a seed inoculant, as a material for packing of flowers and other plants as well as a general soil amendment to increase organic matter content in gardens, golf courses, etc. Approximately 635,000 metric tons of peat are utilized annually for these types of uses in the United States (Public communication, 2009c).

33.2.9.5 Engineering Properties

Histosols are notorious for their low strength and great compressibility, which make them poor foundation materials for

TABLE 33.13 Reported Rates of Subsidence after Drainage of Organic Soils

Location	Site Characteristics	Subsidence Rate (cm year^{-1})	Length of Record (year)	References
New Zealand	Bog	3.4	40	Shipper and McLeod (2002)
Finland	Various peatland types	0.4	60	Minkkinen and Laine (1998)
Florida, USA	Everglades, muck, and peat	3.2	41	Thomas (1965)
Hunts, England	Holme marsh	3.4	103	Nickolson (1951)
California, USA	Sacramento–San Joaquin delta, muck, and peat 1–3 m deep	6.4–9.8	26	Weir (1950)
Indiana, USA		3.0	30+	Ellis and Morris (1945)
Northern Indiana, USA	Muck	1.1–3.0 dependent on WT level	6	Jongedyk et al. (1950)
Michigan, USA		0–3.6	5	Davis and Engberg (1955) (after Thomas, 1965)
Southern Ontario, Canada	Holland marsh and deep loose muck	3.3	19	Mirza and Irwin (1964)
Minnesota, USA		5.1	3	Row (1940)
Florida, USA	Everglades	3	55	Stephens and Speir (1969)
Florida, USA	Everglades	2.3–1.8	20	Shih et al. (1981)
California USA	Sacramento–San Joaquin delta	2.3	78	Rojstaczer and Deverel (1995)
Quebec, Canada		2.1	38	Parent et al. (1982)
Malaysia	Ombrotrophic bog	2	21	Wosten et al. (1997)

TABLE 33.14 Peat Mining by Country, 2007 (in Thousands of Metric Tons)[a]

Country	Horticultural Use	Fuel Use	Total
Argentina	15	0	15
Australia	nd	nd	7
Belarus	100	2400	2500
Burundi	0	10	10
Canada	1250	0	1250
Denmark	300	0	300
Estonia	1300	600	1900
Finland	900	8200	9100
France	200	0	200
Germany	120	0	120
Ireland	500	3800	4300
Latvia	nd[b]	nd	1000
Lithuania	nd	nd	307
Moldova	0	475	475
New Zealand	27	0	27
Norway	30	0	30
Poland	500	0	500
Russia	nd	nd	1300
Spain	nd	nd	60
Sweden	380	900	1280
Ukraine	nd	nd	395
United States	635	0	635

[a] Data from Public communication (2009c).
[b] No data.

roads, buildings, and other structures. Compression and settlement of peats may continue for years after loading (MacFarlane, 1969; MacFarlane and Williams, 1974; Dhowian and Edil, 1980). Special engineering techniques, such as removal of the peat or precompression before construction, are thus required. The high water content and acidity of most Histosols also make corrosion of concrete and metal structures a potential problem (MacFarlane, 1969; MacFarlane and Williams, 1974). While Histosols are poor foundation materials, their high porosity and great adsorption capacity make them very useful for treatment of wastewater. Peats have potential for treatment of municipal effluent and removal of heavy metals and hydrocarbon pollutants from wastewater (Malterer et al., 1996).

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33.3 Andisols

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33.3.1 Introduction

Andisols are soils that typically form in loose volcanic ejecta (tephra) such as volcanic ash, cinders, or pumice. They are characterized by andic properties that include physical, chemical, and mineralogical properties that are fundamentally different from those of soils of other orders. These differences resulted in a proposal to recognize these soils at the highest level in the USDA soil classification system (Smith, 1978). In 1990, Andisols were added to *Soil Taxonomy* as the 11th soil order (Soil Survey Staff, 1990; Parfitt and Clayden, 1991). A very similar taxonomic grouping, Andosols, is 1 of the 32 soil reference groups recognized in the World Reference Base for Soil Resources (IUSS Working Group WRB, 2006). Andisols (and Andosols) are classified on the basis of selected chemical, physical, and mineralogical properties acquired through weathering and not on parent material alone. Both soil names relate to two Japanese words, *anshokudo* meaning "dark colored soil" (*an*, dark; *shoku*, color or tint; *do*, soil) and *ando* meaning "dark soil." *Ando* was adopted into western soil science literature in 1947 (Simonson, 1979).

The central concept of Andisols is one of deep soils commonly with depositional stratification developing mainly from ash, pumice, cinders (scoria), or other explosively erupted, fragmental volcanic material (referred to collectively as tephra) and volcaniclastic or reworked materials. Andisols occur much less commonly on lavas. Unlike many other soils, Andisol profiles commonly undergo "upbuilding pedogenesis" as younger tephra materials are deposited on top of older ones. The resulting profile character is determined by the interplay between the rate at which tephras are added to the land surface and classical "top-down" processes that form soil horizons. Understanding Andisol genesis in many instances thus requires a stratigraphic approach combined with an appreciation of buried soil horizons and polygenesis.

The coarser fractions of Andisols are often dominated by volcanic glass. This glass weathers relatively quickly to yield a fine colloidal or nanoscale fraction (1–100 nm) dominated by short-range order materials composed of “active” Al, Si, Fe, and organic matter, especially humus. Previously described erroneously as “amorphous,” short-range order materials comprise extremely tiny but structured nanominerals, the main ones being allophane and ferrihydrite (Hochella, 2008). A useful collective descriptor for them is “nanocrystalline” (Michel et al., 2007). Another colloidal constituent, imogolite, comprises long filamental tubes and therefore has both short- and long-range order (Churchman, 2000). The nanominerals, chiefly allophane, ferrihydrite, and metal–humus complexes, are responsible for many of the unique properties exhibited by Andisols.

Despite covering less of the global ice-free land area than any other soil order (~1%), Andisols generally support high population densities, about 10% of the world’s population (Ping, 2000). This is because they typically have exceptional physical properties for plant growth and, in many localities, high native fertility because relatively frequent additions of tephra can renew potential nutrient sources (Ugolini and Dahlgren, 2002; Dahlgren et al., 2004). The majority of Andisols occur in humid regions where there is adequate rainfall. Andisols often have high organic carbon contents. These and other factors make Andisols generally well suited for agriculture production and historically allowed establishment of nonshifting agricultural practices. Despite their generally favorable properties for plant growth, Andisols do pose some engineering and fertility challenges. These soils have low bulk densities, resulting in low weight-bearing capacity. Andisols also exhibit thixotropy and sensitivity, properties that cause them to behave in a fluid-like manner when loading pressures are applied (Neall, 2006; Arnalds, 2008). Andisols may exhibit substantial fertility limitations, including P fixation, low contents of exchangeable bases (especially K) and other nutrients, and strong acidity and Al toxicity (Shoji and Takahashi, 2002; Dahlgren et al., 2004; Lowe and Palmer, 2005).

33.3.2 Geographic Distribution

Andisols cover approximately 124 million ha or about 0.84% of the Earth’s ice-free surface (Soil Survey Staff, 1999). They are closely associated with areas of active and recently active volcanism, and their global distribution is depicted in Figure 33.13. The greatest concentration of Andisols is found along the Pacific Ring of Fire, a zone of tectonic activity and volcanoes stretching from South through Central and North America via the Aleutian Islands to the Kamchatka Peninsula of Russia through Japan, Taiwan, the Philippines, and Indonesia to Papua New Guinea and New Zealand. Other areas include the Caribbean, central Atlantic ridge, northern Atlantic rift, the Mediterranean, parts of China, Cameroon, the Rift Valley of east Africa, and southern Australia (Soil Survey Staff, 1999). There are numerous volcanic islands where Andisols are common, including Iceland, the Canary Islands, Azores, Galapagos Islands, Hawaiian Islands, the West Indies, and various small islands in the Pacific.

The global distribution of Andisols encompasses a wide variety of climatic conditions—cold-to-hot and wet-to-dry. This suggests that climate is less important to the formation of Andisols than is proximity to volcanic or pyroclastic parent materials. Nevertheless, the majority of Andisols are found in higher-rainfall regions of the world. Almost two-thirds of Andisols occur in humid regions (udic soil moisture regimes) while fewer than 5% occur in aridic moisture regimes (Mizota and van Reeuwijk, 1989; Wilding, 2000). Approximately half of the world’s Andisols occur in the tropics, with the remaining half being split between boreal and temperate regions (Wilding, 2000; IUSS Working Group WRB, 2006).

There are almost 15.6 million ha of Andisols in the United States (Soil Survey Staff, 1999). The largest areas occur in Alaska

FIGURE 33.13 Global distribution of Andisols. (Courtesy of USDA-NRCS, Soil Survey Division, World Soil Resources, Washington, DC, 2010.)

(~10 million ha) and in Washington, Oregon, Idaho, northern California, and western Montana (Pewe, 1975; Rieger et al., 1979; Ping et al., 1989; Southard and Southard, 1991; Ugolini and Dahlgren, 1991; Goldin et al., 1992; Takahashi et al., 1993; McDaniel and Hipple, 2010). In the Pacific Northwest region of Washington, Idaho, and Oregon, most Andisols are forested and occur at mid-to-high elevations in cooler temperature regimes (McDaniel et al., 2005). Few Andisols are found in warmer temperature regimes because the summers are normally too hot and dry to allow sufficient weathering or leaching to produce the required andic properties.

Iceland contains ~7 million ha of Andisols. These represent the largest area of Andisols in Europe (Arnalds, 2004). Andisols also occur in France, Germany, Spain, Italy, and Romania (Buol et al., 2003; Kleber et al., 2004; Quantin, 2004; IUSS Working Group WRB, 2006; Arnalds et al., 2007). In New Zealand, ~3.2 million ha of Andisols occur on the North Island, the majority now supporting agriculture or forestry (Parfitt, 1990; Lowe and Palmer, 2005). Japan has ~6.9 million ha of Andisols (Wada, 1986; Takahashi and Shoji, 2002).

Some soils classified as Andisols are also found in humid areas not associated with volcanic activity such as in the southern Appalachian Mountains, parts of Kyushu (Japan), Scotland, Spain, and the Alps. These soils have large quantities of Al or Fe associated with humus (see Section 33.3.3.2) and similar management constraints as those of soils formed from volcanic ejecta and also key out as Andisols. These attributes further highlight the importance of realizing that Andisols are not classified on parent material but on the properties acquired during weathering and leaching. By the same token, soils other than Andisols, such as Entisols, Inceptisols, Spodosols, Mollisols, Oxisols, Vertisols, Alfisols, or Ultisols, may form in association with volcanic or pyroclastic materials (e.g., Shoji et al., 2006).

33.3.3 Andisol Properties

33.3.3.1 Morphological Features

Most Andisols have distinct morphological features. They usually have multiple sequences of horizons (Figure 33.14) resulting from the intermittent deposition of tephras and ongoing top-down soil formation referred to as upbuilding pedogenesis (see Section 33.3.5.1). A horizons are typically dark, often overlying reddish brown or dark yellowish brown Bw cambic horizons. Buried A–Bw sequences are common (Figure 33.14). Layers representing distinct tephra-fall events are common, often manifested as separate Bw horizons or as BC or C horizons if the tephra shows limited weathering or is relatively thick. Horizon boundaries are typically distinct or abrupt where these thicker layers occur.

Andisols are usually light and easily excavated because of their low bulk density and weakly cohesive clay minerals. The high porosity allows roots to penetrate to great depths. Andisols generally have granular structures in A horizons, but the structure in Bw horizons is generally weak subangular blocky, often crushing readily to crumb structure. Some Andisols (Udands) formed in areas of high rainfall have higher clay contents while soils that are subjected to wet and dry cycles form prismatic structures. At higher water contents, soils containing as little as 2% allophane have a characteristic greasy feel (Parfitt, 2009), an indication of sensitivity.

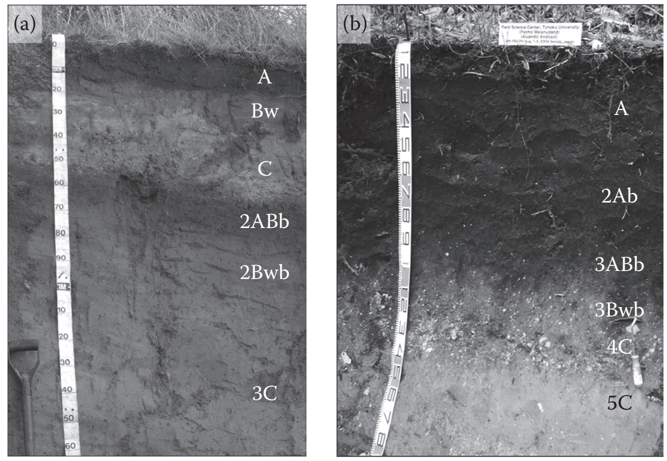

FIGURE 33.14 Andisol profiles (a) allophanic Taupo soil (Udivitrand) from New Zealand (scale divisions = 10 cm) and (b) nonallophanic Tohoku Farm soil (Melanudand) from Japan with thick (~1 m), dark, strongly humified horizons (melanic epipedon).

33.3.3.2 Mineralogical Properties

Tephra parent materials weather rapidly to form nanominerals that are responsible for many of the unique physical and chemical properties associated with Andisols. Although a wide range of clay minerals can be found in Andisols (such as gibbsite, kaolinite, vermiculite, smectite, crystalline Fe oxides such as hematite and goethite, and cristobalite), those of greatest interest are allophane, imogolite, ferrihydrite, and the Al– and Fe–humus complexes because they confer the characteristic andic properties (Dahlgren et al., 2004; Parfitt, 2009).

Allophane is nearly X-ray amorphous, but under an electron microscope it is structured over short distances, appearing as nanoparticles of hollow spheres 3.5–5 nm in diameter that have the chemical composition $(1–2)SiO_2 \cdot Al_2O_3 \cdot (2–3)H_2O$ (Figure 33.15a) (Wada, 1989; Churchman, 2000; Brigatti et al., 2006; Theng and Yuan, 2008). The most common type of allophane is the so-called Al-rich allophane with an Al:Si molar ratio of ~2 (it is sometimes called proto-imogolite allophane). There is also Si-rich allophane with an Al:Si ratio ~1 (also referred to as halloysite-like allophane).

Imogolite has the composition $(OH)SiO_3 \cdot Al_2(OH)_3$ and has both long- and short-range order. Under an electron microscope, it appears as long smooth and curved hollow threads or tubules with inner and outer diameters of ~0.7 and 2 nm, respectively

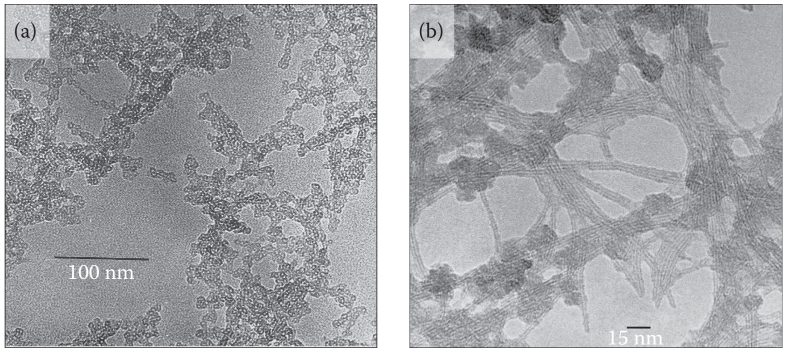

FIGURE 33.15 Micrographs of (a) allophane and (b) imogolite (external diameter of nanotubes is ~2 nm). (Reproduced with the kind permission of the Mineralogical Society of Great Britain and Ireland from a paper by Parfitt, R.L. 2009. Allophane and imogolite: Role in soil biogeochemical processes. Clay Miner. 44:135–155. With permission.)

(Figure 33.15b). These nanotubes typically appear as bundles of two or more threads 10–30 nm thick and several micrometers long (Theng and Yuan, 2008). Imogolite in Japan can be seen with the naked eye as a whitish gel film infilling pores in coarse pumice particles (Wada, 1989).

Allophane and imogolite both have high surface areas, ranging from 700 to 1500 m^2 g^{-1} (Parfitt, 2009), and this feature, coupled with their variable surface charge characteristics and exposure of $(OH)Al(OH_2)$ groups at wall perforations (defects), explains their strong affinity for water, metal cations, organic molecules, and other soil minerals (Harsh et al., 2002; Theng and Yuan, 2008). Even small amounts contribute huge reactive surface areas in soils (Lowe, 1995). Allophane and imogolite are soluble in ammonium (acid) oxalate solution, and the Si dissolved is used to estimate their contents in soils (Parfitt and Henmi, 1982; Parfitt, 2009). *Soil Taxonomy* uses oxalate-extractable Al (and Fe) to help define andic soil properties (see Figure 33.16 and Section 33.3.4). Allophane content of B horizons is quite variable, ranging from about 2% in slightly weathered or metal–humus-dominated systems to >40% in well-developed Andisols. It typically increases with depth in upper subsoils, usually being highest in the Bw and buried horizons. But in many Andisol profiles in New Zealand, allophane decreases and halloysite concomitantly increases with depth in lower subsoils either because of the downward migration of Si into lower profiles or because of changes in climate during upbuilding, or both. Imogolite is more commonly found in B horizons under carbonic acid weathering regimes than in A horizons where organic acid weathering dominates (Dahlgren et al., 2004). Allophane may occur dispersed as groundmass, as coatings, bridges, or infillings (in vesicles or in root channels), or it may be disseminated through pseudomorphs of glass or feldspar grains (Jongmans et al., 1994, 1995; Bakker et al., 1996; Gérard et al., 2007).

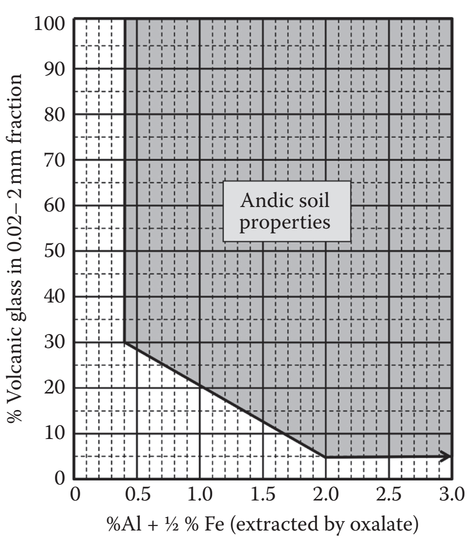

FIGURE 33.16 Andic soil properties as defined by *Soil Taxonomy*. (From Soil Survey Staff. 2010. Keys to soil taxonomy. 11th edn. USDA-NRCS, Washington, DC.)

Ferrihydrite is common in many Andisols, especially those associated with more basic parent materials, has a composition of $Fe_5HO_8 \cdot 4H_2O$ and imparts a reddish brown color (hues of 5YR–7.5YR; Bigham et al., 2002). Made up of spherical nanoparticles 2–5 nm in diameter (Schwertmann, 2008), ferrihydrite has large, reactive surface areas ranging from ~200 to 500 m^2 g^{-1} (Childs, 1992; Jambor and Dutrizac, 1998). Its abundance is commonly estimated from the amount of Fe extracted by ammonium oxalate solution multiplied by 1.7 (Parfitt and Childs, 1988). It is a widespread and characteristic component of young Fe-oxide accumulations precipitated from Fe-rich solutions in the presence of organic matter, such as in Iceland (Arnalds, 2004), and elsewhere, including New Zealand, Japan, and Australia where its precipitation may be inorganic or bacteria-driven (Childs et al., 1991; Lowe and Palmer, 2005). Ferrihydrite can transform to hematite via solid-state transformation or goethite through dissolution and reprecipitation (Schwertmann, 2008).

Metal–humus complexes are significant components of some Andisol colloidal fractions. These Al- and Fe-organic complexes are immobile and accumulate in dark or black surface horizons where organic materials are abundant, and dark (melanic) horizons may extend to depths as much as 2 m (see Figure 33.14b). Metal–humus complexes represent the active forms of Al and Fe in nonallophanic Andisols as described below (Dahlgren et al., 2004).

Halloysite is a relatively fast-forming 1:1 layer silicate that often exhibits tubular or spheroidal morphology (White and Dixon, 2002; Joussein et al., 2005). Its formation is favored in seasonally dry environments where higher Si concentrations are maintained (Shoji et al., 1993). These include areas of lower rainfall, restricted drainage, and Si-rich parent materials (Lowe, 1986; Churchman, 2000; Churchman and Lowe, 2011, Section 20.1). Halloysite surfaces are characterized by some permanent negative charge, allowing retention of cations across a wide range of pH values.

The soil solution in Andisols in a range of locations may contain large amounts of dissolved Si, which leads to the formation by nucleation of secondary silica minerals from the saturated solution (Ping et al., 1988; Shoji et al., 1993; Ping, 2000; Nanzyo, 2002, 2007; Waychunas and Zhang, 2008). Termed laminar opaline silica, this material is circular or elliptical in shape (0.2–0.5 μm diameter) and extremely thin. Precipitation of the silica may be aided by evaporation or freezing of soil water, or via plant-related processes related to Si uptake and recycling (Lowe, 1986; Drees et al., 1989; Churchman, 2000; Henriet et al., 2008). Such silica polymorphs can be distinguished from biogenic forms of silica (phytoliths) because the latter have more complex shapes inherited from biological cells (Kondo et al., 1994; Nanzyo, 2007).

Andisols dominated by allophane with subordinate imogolite and ferrihydrite in upper horizons are referred to as *allophanic Andisols*. These contrast with a second, strongly acid group known as *nonallophanic Andisols* in which metal–humus complexes dominate the colloidal mineralogy. Nonallophanic Andisols are common in Japan especially where they account for about 30% of soils formed on tephras (Takahashi and Shoji, 2002) and are known in around 20 other countries (Saigusa and Matsuyama, 2004). Examples of soils from each group are shown in Figure 33.14. In Table 33.15, the Thingvallasveit and Tirau soils are examples of allophanic Andisols; the Tohuku Farm soil is an example of a nonallophanic Andisol. The mineralogical differences between these two groups of Andisols lead to several important different physical and chemical properties (especially the strong acidity of nonallophanic Andisols) and significant management implications (Dahlgren et al., 2004).

In the silt and sand fractions of Andisols, the dominant components are volcanic glass (a mineraloid) and various primary minerals. The glass particles (shards) which, like shattered glass, have sharp angles and edges, are very abrasive. However, these glass particles are usually coated with colloidal materials including allophane, ferrihydrite, and other Fe oxides and their humus complexes, which all contribute to aggregate formation. It is noteworthy that volcanic glass is often quite vesicular and porous in nature (as is pumice), and thus can retain water and has more chemical activity than other common sandy materials (Ping, 2000; Lowe and Palmer, 2005).

33.3.3.3 Chemical Properties

One of the common characteristics of Andisols is accumulation of relatively large quantities of organic matter, both in the allophanic (moderate pH) and nonallophanic Andisols (low pH; Table 33.15). Allophanic Andisols typically contain up to ~8%–12% C, whereas nonallophanic soils may contain up to ~25%–30% C (Mizota and van Reeuwijk, 1989). The residence time of C in Andisols, as measured by ^{14}C, is much greater than that of other soil orders (Parfitt, 2009). In addition, upbuilding pedogenesis leads to the storage of C in lower parts of profiles, and especially in buried A horizons that are sealed off and isolated from most surface processes.

Andisols are almost always acid, with most pH (H_2O) values ranging from 4.8 to 6.0 (Shoji et al., 1993; Dahlgren et al., 2004). Uncultivated, nonallophanic Andisols with high organic matter contents typically have a pH (H_2O) < 4.5 (IUSS Working Group WRB, 2006).

One of the key factors affecting the chemistry of Andisols is the variable surface charge associated with the colloidal fraction (Qafoku et al., 2004). Electrical charges on colloid or nanoparticle surfaces can be either positive or negative, and change as a function of pH. Surfaces have positive charge at lower pH and retain anions, while cations are retained by the negative charge at higher pH. This variable charge greatly affects the behavior of ions that are retained in soils. Cation retention capacity of Andisols makes the soils susceptible to metal and radioactive fallout ^{137}Cs pollution (Adamo et al., 2003; Sigurgeirsson et al., 2005). Andisols also exhibit anion exchange properties, which can be important for nutrient retention (e.g., NO_3^- and SO_4^{2-}) (Shoji et al., 1993).

Measured cation exchange capacity of Andisols is relatively high, often 20 to >50 $cmol_c\ kg^{-1}$. However, because the dominant colloids have variable charge, much of this CEC is pH dependent. This is especially true in allophanic Andisols (Dahlgren et al., 2004) and means that CEC decreases with decreasing pH. And because most Andisols are acid, CEC measurements made at pH 7 or 8.2 will be artificially high. In andic soils of the Pacific Northwest of United States, average CEC values determined using unbuffered extractants (effective CEC or ECEC) are approximately one-fourth of those determined at pH 8.2, 6.5 $cmol_c\ kg^{-1}$ vs. 26.4 $cmol_c\ kg^{-1}$ (McDaniel et al., 2005). This phenomenon needs to be considered when measuring CEC and base saturation or interpreting these data. The relatively low ECEC of Andisols can limit their ability to retain and exchange Ca, Mg, and K. Some representative cation exchange characteristics of Andisols are presented in Table 33.15.

The active Al and Fe compounds in Andisols (allophane/imogolite, metal–humus complexes, and ferrihydrite) also have the ability to sorb and strongly bind anions such as phosphate and fluoride (Shoji et al., 1993; Dahlgren et al., 2004; Parfitt, 2009). Much of this sorption is not reversible, leading to large quantities of phosphate being rendered unavailable for plant uptake. As described in Section 33.3.4, the amount of P retention in soils is used to define andic soil properties (Soil Survey Staff, 2010). Similarly, quantities of active Al and Fe compounds can also be estimated by reacting soil with NaF solution. Sorption of F^- releases OH^- into solution, thereby raising the pH. A resultant pH greater than ~9.5 indicates the presence of allophane/imogolite

TABLE 33.15 Selected Chemical and Mineralogical Properties of Representative Andisols

Horizon	Depth (cm)	pH H_2O	pH NaF	Clay (%)	Organic C %	ECEC[a]	CEC pH 7 ($cmol_c$ kg^{-1})	Sum of Exch. Bases	[b]Exch. Al^{3+}	P Retention	Al in Al–Humus Complexes[c] (%)	Allophane[d]	Ferrihydrite[e]
Thingvallasveit Series[f] (Haplocryand)—Iceland													
A1	0–12	5.4	11.0	—	7.9	—	31.9	9.9	—	99	—	13	8.5
A2	12–28	5.7	10.7	1	7.6	—	44.6	13.1	—	99	0.1	16	11.1
Bw1	28–61	5.6	10.8	—	7.4	—	43.8	10.1	—	99	—	15	12.7
Bw2	61–68	5.7	11.1	3	4.2	—	32.3	6.5	—	99	—	16	6.0
2Bw3	68–87	5.6	10.5	—	2.1	—	24.8	4.4	—	98	—	17	4.4
2C	87–142	5.5	10.3	1	0.8	—	11.8	2.8	—	99	—	13	3.6
Tirau Series[g] (Hapludand)—New Zealand													
Ap	0–18	5.6	9.6	19	7.9	14.6	29.7	14.4	0.2	88	0.7	9	1.0
Bw1	18–32	6.2	9.9	11	2.0	5.6	11.3	5.6	0	98	0.2	14	1.2
Bw2	32–48	6.2	9.8	13	1.0	5.2	10.5	5.2	0	98	0.2	16	1.2
Bw3	48–65	6.4	9.5	20	0.5	6.1	12.5	6.1	0	91	0.1	10	0.7
2BCb	65–90	6.4	9.4	18	0.5	7.5	13.2	7.5	0	86	0.1	8	0.8
Tohoku Farm[h] (Melanudand)—Japan													
A1	0–14	4.2	10.5	18	11.5	13.3	52.2	2.3	11.0	83	1.5	0.3	—
2A2	14–30	4.6	11.6	5	8.0	8.4	49.1	0.4	8.0	93	2.5	1.1	—
3A3	30–57	4.8	11.6	12	13.2	8.4	56.1	0.4	8.0	93	2.7	1.4	—
4Bw1	57–80	5.2	11.3	8	2.8	2.0	18.4	0.1	1.9	93	0.9	6.7	—
4Bw2	80–126	5.2	11.0	16	0.9	0.9	13.7	0.2	0.7	94	0.6	7.4	—
4Bw3	126–160	5.3	10.9	20	0.6	0.8	10.9	0.2	0.6	89	0.5	7.0	—
4C	160–200	5.3	10.9	22	0.5	1.8	11.5	0.4	1.4	84	0.5	4.9	—

[a] Effective cation exchange capacity.
[b] Exchangeable Al^{3+} extracted with KCl.
[c] Estimated using pyrophosphate-extractable Al.
[d] Estimated using oxalate-extractable Si (Parfitt, 1990).
[e] Estimated using oxalate-extractable Fe (Parfitt and Childs, 1988).
[f] Data are from Arnalds et al. (1995).
[g] Data are from Bakker et al. (1996).
[h] Data (Pedon #86P0091) are from Soil Survey Staff 2011.

and/or Al–humus complexes, and because of this, NaF field test kits can be used for field identification of Andisols (Fieldes and Perrott, 1966; IUSS Working Group WRB, 2006).

33.3.3.4 Physical Properties

Unique physical attributes of Andisols are related to structural assemblages of hollow spheres and tubular threads as mineral entities into resilient, progressively larger (silt-sized) aggregated domains. This aggregation results in low density, high porosity, high surface area, and high soil water retention even at low water potentials. The structural arrangement accounts for the low thermal conductivity of andic materials, which is three to four times less than that of the phyllosilicates in other mineral soils. It also accounts for the thixotropic and sensitivity character of these soils and several irreversible changes in physical properties that occur upon drying (Ping, 2000; Neall, 2006).

Andisols have low bulk density, usually <0.9 g cm^{-3}, because of the high organic matter and nanomineral contents (Table 33.16) and well-developed aggregation. This results in good tilth and makes them excellent rooting media. On the other hand, low bulk densities result in low weight-bearing capacity and make Andisols highly susceptible to wind and water erosion when surface cover is removed or degraded (Ping, 2000; Nanzyo, 2002; Dahlgren et al., 2004; Neall, 2006). Because of the nature of volcanic ejecta and its distribution, many Andisols in proximal locations contain appreciable amounts of gravel and stones, and coarse-grained tephra layers may have adverse effects on hydrological properties by interrupting capillary movement of water. Some additional adverse physical properties include a high glass content that can reduce the quantity and biodiversity of soil organisms such as earthworms, and the presence occasionally of impenetrable horizons such as thin Fe pans (placic horizons) in higher rainfall areas (Neall, 2006).

Andisols typically exhibit high water retention because of the presence of allophane, ferrihydrite, and metal–humus complexes, which have high surface areas as noted previously. As a result, moisture contents of many Andisols can exceed 100% on a weight basis, even at soil moisture tensions of 1500 kPa—this feature is illustrated by the data for the Hilo soil (Table 33.16). Water retention is greatest in Andisols that have undergone significant weathering and hence have high clay contents.

Most field-moist Andisols have a greasy feel when rubbed between the fingers and exhibit smeariness—these can be both indicators of the properties known generally as sensitivity and thixotropy (Soil Survey Staff, 1975; Torrance, 1992). Thixotropy is a reversible gel–sol transformation that occurs when shear forces are applied to a moist soil. The applied shear force causes the soil to abruptly lose strength, sometimes to the point of behaving as a fluid. When the shear force is removed, the soil will recover some or all of its original strength. Sensitivity (the term used more commonly by engineers) is the ratio of undisturbed to disturbed (remolded) shear strength, that is, the maximum strength of

TABLE 33.16 Selected Physical Properties of Representative Andisols

		Texture					Bulk Density (g cm^{-3})		1500 kPa H_2O (%)	
Depth (cm)	Horizon	Field	Lab	Sand (%)	Silt (%)	Clay (%)	Oven Dry	Moist	Air-Dry	Moist
Bonner Series[a] (Vitrixerand)—Idaho										
0–4	A	Sil	Sil	30.6	63.8	5.6	0.75	0.68	20.3	—
4–20	Bw1	Sil	Sil	38.1	58.0	3.9	—	—	10.0	—
20–48	Bw2	Sil	Sil	39.3	59.0	1.7	0.96	0.94	9.9	—
48–69	Bw3	—	Cosl	52.6	44.1	3.3	—	—	6.8	—
69–89	Bw4	—	Cosl	52.7	44.8	2.5	0.80	0.80	6.8	—
89–152	2C	—	Cos	88.6	9.8	1.6	—	—	2.2	—
Tirau Series[b] (Hapludand)—New Zealand										
0–18	Ap	Sil	L	35	46	19	0.75	—	23.3	31.1
18–32	Bw1	Sl	Sil	33	56	11	0.71	—	15.7	33.7
32–48	Bw2	Sl	Sil	33	54	13	0.69	—	16.5	33.6
48–65	Bw3	Sl	L/Sil	30	50	20	0.79	—	22.0	34.7
65–90	2BCb	Sil	Sil	20	62	18	0.87	—	20.4	33.9
Hilo Series[c] (Hydrudand)—Hawaii										
0–18	Ap1	Sicl	Sl	66.2	29.7	4.1	—	—	28.4	54.1
18–36	Ap2	Sicl	Cos	86.8	12.1	1.1	0.82	0.51	27.7	107.2
36–60	Bw1	Sicl	Cos	91.8	7.9	0.3	1.66	0.41	25.0	112.6
60–92	Bw2	Sicl	Cos	93.1	6.9	—	1.61	0.25	25.1	132.5
92–108	Bw3	Sicl	Cos	94.9	4.7	0.4	1.41	0.30	26.0	122.4

Sil, Silt loam; Sicl, Silty clay loam; Cosl, Coarse Sandy loam; Cos, Coarse sand; Sl, Sandy loam; L, Loam.

[a] Data (Pedon #78P0553) are from Soil Survey Staff 2011.

[b] Data are from Bakker et al. (1996).

[c] Data (Pedon #89P0658) are from Soil Survey Staff 2011.

an undisturbed specimen compared with the residual strength remaining after force or strain is applied (Mitchell and Soga, 2005; Neall, 2006). In sensitive materials, the original strength is generally not recovered—unlike thixotropic materials—after removal of the force. Both sensitivity and thixotropic behavior are best expressed in Andisols having high water retention and lack of layer silicates to provide cohesion and can pose significant engineering problems.

Many Andisols exhibit irreversible changes upon drying. Allophane nanospheres collapse upon dehydration and form larger aggregates that do not break down upon rewetting. This phenomenon can cause crust formation at the soil surface during hot dry periods. It also results in well-known unreliable particle size analyses of air-dried Andisols. Normally, clay content is underestimated and sand and silt contents are overestimated (Ping et al., 1989; Dahlgren et al., 2004). However, reliable sand-, silt-, and crystalline clay-size fraction data were obtainable for andesitic Andisols in New Zealand by Alloway et al. (1992) who analyzed grain-size distributions of residual material following selective dissolution of nanominerals and organic constituents via ammonium oxalate. Other irreversible changes that occur upon desiccation include increases in bulk density, decreases in water retention, and increases in cohesive strength. Note the difference in water retention and bulk density values between dried and moist samples in Table 33.16.

Plasticity in Andisols is different from that of soils containing layer silicate clay minerals. Generally, field-moist Andisols have high liquid (60%–350%) and plastic limits (70%–180%) (Warkentin and Maeda, 1980; Neall, 2006). The low plasticity index (0–10) also clearly separates Andisols from other soils. Plasticity measurements can be used as an index of physical behavior in Andisols and as a substitute for particle size analysis, which usually is not reliable. Air-dry samples, on the other hand, often show low plasticity because of the irreversible changes on drying and behave like sandy soils.

33.3.4 Classification of Andisols

Andisols are classified on the basis of having andic soil properties, which are quantitatively defined in *Soil Taxonomy* (Soil Survey Staff, 2010). In general, andic soil properties consist of combinations of properties that develop as tephras and other volcanic materials weather. These include relatively low bulk density values (≤0.90 g cm^{-3}), relatively high phosphate retention (>25%–85%), the presence of volcanic glass, and the presence of nanoscale weathering products containing Al, Fe, and Si (Soil Survey Staff, 2010). The criterion of percentage Al extracted by ammonium oxalate (Al_o) plus half the percentage Fe extracted by ammonium oxalate (Fe_o) from short-range order nanominerals is used to quantitatively define andic properties. The quantity 0.5 Fe_o is used to normalize the criterion because Fe has an atomic weight (56), approximately twice that of Al (27). The majority of soils with a glass content and percentage of Al_o + $0.5Fe_o$ that falls within the shaded area of Figure 33.16 have andic soil properties. Soils possessing at least 36 cm of material with andic soil properties are classified as Andisols (Soil Survey Staff, 2010). It is emphasized that freshly deposited volcanic ash does not have andic soil properties and would not be classified as an Andisol. It is not until some weathering has occurred—sufficient to generate Al_o + $0.5Fe_o$ totaling at least 0.4%—that andic properties are developed.

Andisols are separated into suborders primarily on the basis of soil moisture and/or temperature regimes (Table 33.17). Aquands are poorly drained and have a water table at or near the soil surface for much of the year. These soils are usually restricted to low-lying landscape positions and have dark surface horizons. Because of excessive wetness, Aquands typically require drainage in order to be used for crop or pasture production.

Gelands are very cold Andisols that have a mean annual soil temperature (MAST) ≤0°C (Soil Survey Staff, 2010). Relatively little is known about the distribution of these soils, but they are found at higher latitudes in areas of either current or recent volcanic activity.

Cryands are cold Andisols having a cryic temperature regime (0°C < MAST ≤ 8°C and summers are cool; Soil Survey Staff, 2010). They are found at higher latitudes and higher elevations. In the United States, Cryands are found mainly in Alaska and mountainous regions of the Pacific Northwest (Ping, 2000; McDaniel and Hipple, 2010). They are extensive in Iceland (Arnalds and Kimble, 2001) and also on the Kamchatka Peninsula and occur in mountainous regions elsewhere including in eastern Africa, the Andes, and (uncommonly) New Zealand (Ping, 2000; Lowe and Palmer, 2005). Globally, there are ~26 million ha of Cryands, and they are the third most common suborder, representing ~28% of Andisols (Soil Survey Staff, 1999; Wilding, 2000).

Torrands are Andisols of very dry environments where moisture for plant growth is very limited. This lack of soil moisture slows down weathering processes and leaching, thereby inhibiting development of andic soil properties. Torrands are found in Oregon and Hawaii in the United States. They are the least extensive of any of the Andisol suborders, with only ~100,000 ha occupying the global ice-free land area (<1% of Andisols).

Xerands occur in temperate regions with xeric soil moisture regimes, which are characterized by cool, moist winters and very warm, dry summers (Mediterranean climates). In the United States, they occur primarily in northern California, Oregon, Washington, and Idaho where they have formed under coniferous forest. Elsewhere, Xerands occur in scattered localities including Italy, Canary Islands, Argentina, and South Australia (Broquen et al., 2005; Lowe and Palmer, 2005; Inoue et al., 2010). Xerands are uncommon globally (~4% of Andisols).

Vitrands are the only suborder that is not defined by a climatic regime. These Andisols are relatively young and only slightly weathered. They tend to be coarse-textured and have a high content of volcanic glass that may be strongly vesicular or pumiceous. In the United States, Vitrands are found in Washington, Oregon, and Idaho. They also occur in Argentina (Broquen et al., 2005) and are common in the North Island of New Zealand (Lowe and Palmer, 2005). Globally, there are ~28 million ha of Vitrands (~31% of Andisols) making them and

TABLE 33.17 Listing of Andisol Suborders and Great Groups

Suborders	Great Groups
Aquands—poorly drained Andisols with a water table at or near the surface for much of the year	
	Gelaquands—Aquands of very cold climates (mean annual soil temperature <0°C)
	Cryaquands—Aquands of cold climates (0°C < mean annual soil temperature ≤8°C)
	Placaquands—Aquands with a thin pan cemented by Fe, Mn, and organic matter
	Duraquands—Aquands with a cemented horizon
	Vitraquands—Aquands with coarse textures dominated by glassy materials
	Melanaquands—Aquands with a thick, dark, organic matter-rich surface layer
	Epiaquands—Aquands with a perched water table
	Endoaquands—Aquands with a groundwater table
Gelands—Andisols of very cold climates (mean annual soil temperature ≤0°C)	
	Vitrigelands—Gelands with coarse textures dominated by glassy materials
Cryands—Andisols of cold climates (0°C < mean annual soil temperature ≤8°C)	
	Duricryands—Cryands with a cemented horizon
	Hydrocryands—Cryands with very high water-holding capacity
	Melanocryands—Cryands with a thick, very dark, organic matter-rich surface layer
	Fulvicryands—Cryands with a thick, organic matter-rich surface layer
	Vitricryands—Cryands with coarse textures dominated by glassy materials
	Haplocryands—other Cryands
Torrands—Andisols of very dry climates	
	Duritorrands—Torrands with a cemented horizon
	Vitritorrands—Torrands with coarse textures dominated by glassy materials
	Haplotorrands—other Torrands
Xerands—Andisols of temperature regions with warm, dry summers and cool, moist winters	
	Vitrixerands—Xerands with coarse textures dominated by glassy materials
	Melanoxerands—Xerands with a thick, very dark, organic matter-rich surface layer
	Haploxerands—other Xerands
Vitrands—slightly weathered Andisols that are coarse textured and dominated by glassy material	
	Ustivitrands—Vitrands of semiarid and subhumid climates
	Udivitrands—Vitrands of humid climates
Ustands—Andisols of semiarid and subhumid climates	
	Durustands—Ustands with a cemented horizon
	Haplustands—other Ustands
Udands—Andisols of humid climates	
	Placudands—Udands with a thin pan cemented by Fe, Mn, and organic matter
	Durudands—Udands with a cemented horizon
	Melanudands—Udands with a thick, very dark, organic matter-rich surface layer
	Hydrudands—Udands with very high water-holding capacity
	Fulvudands—Udands with a thick, organic matter-rich surface layer
	Hapludands—other Udands

Source: Soil Survey Staff. 2010. Keys to soil taxonomy. 11th edn. USDA-NRCS, Washington, DC.

Udands (similar in extent) the two most extensive suborders (Soil Survey Staff, 1999; Wilding, 2000).

Ustands occur in tropical or temperature regions that have ustic soil moisture regimes. The ustic soil moisture regime is characterized by an extended dry period, but moisture is normally present at a time when conditions are suitable for plant growth (Soil Survey Staff, 2010). Ustands are of fairly limited extent in the United States and are found mainly in Hawaii. They also occur in Mexico, on Pacific islands, and in eastern Africa (Dubroeucq et al., 1998; Takahashi and Shoji, 2002). About 6.3 million ha of Ustands are known globally (~7% of Andisols).

Udands are Andisols with a udic moisture regime, which is common to humid climates. Udands are thus characterized by well-distributed precipitation throughout the year and limited periods of soil moisture stress. In the United States, Udands are found in western parts of Washington and Oregon and in Hawaii. Elsewhere, they occur commonly on other parts of the Pacific Rim including in Patagonia (Argentina), Mexico, Japan, the Philippines, Indonesia, and New Zealand (Takahashi and Shoji, 2002; Van Ranst et al., 2002; Broquen et al., 2005; Prado et al., 2007). In total, Udands occupy nearly ~28 million ha of ice-free land globally (~30% of Andisols; Soil Survey Staff, 1999; Wilding, 2000).

Each of the suborders is further separated into great groups. These are listed and described in Table 33.17. A variety of characteristics are used to define great groups, including the presence or absence of certain types of soil horizons, moisture and temperature regimes, glass content and texture, and water retention.

There are also soils of other orders that have been influenced to a lesser degree by andic materials (Parfitt, 2009). Andic soil properties have developed, but their distribution within the soil is not of sufficient thickness for the soils to be classified as Andisols. These soils are therefore classified as andic subgroups of other orders. In cases where andic materials have been extensively mixed with other parent materials, or andic properties are only weakly expressed, soils are classified as vitrandic subgroups of other soil orders (Soil Survey Staff, 2010). In the United States, such soils are extensive in the Pacific Northwest where they are transitional to higher-elevation, forested Andisols (McDaniel and Hipple, 2010).

33.3.5 Formation of Andisols

33.3.5.1 Parent Material and Stratigraphy

In most cases, the parent materials from which Andisols have formed are of explosive volcanic origin (rather than effusive) and thus fragmental and unconsolidated. Such materials range in size from ash (<2 mm) and lapilli (2–64 mm) through to large angular blocks or part-rounded bombs (>64 mm)—that is, from fine dust to boulders. Collectively, these materials are termed pyroclastic deposits or tephra (Gk *ashes*) (Alloway et al., 2007). Tephra deposits typically are loose and very coarse and thick close to source vents but become markedly finer and thinner with increasing distance away from source so that at more than ~100 km from vent most comprise mainly ash-sized material (equivalent to sand or finer particles). The accumulation at a particular site of numerous tephra deposits from sequential eruptions from one or more volcanoes leads usually to the formation of Andisols with distinctive layered profiles and buried soil horizons, forming multisequal profiles (Figure 33.17). Such layered profiles, together with their andic soil properties, are special features of Andisols. Study of the layers and attaining ages for them (tephrostratigraphy) is an important aspect of understanding Andisol formation (Lowe and Palmer, 2005; Lowe and Tonkin, 2010).

During periods of quiescence between major eruptions, soil formation takes place, transforming the characteristics of the unmodified tephra via normal top-down pedogenesis whereby the materials are altered in a downward-moving front to form subsoil horizons. However, when new tephras are added to the land surface, upbuilding pedogenesis takes place. The frequency and thickness of tephra accumulation (and other factors) determine how much impact the top-down processes have on the ensuing profile character, and if "developmental" or "retardant" upbuilding, or both, will take place. Two contrasting scenarios can be considered.

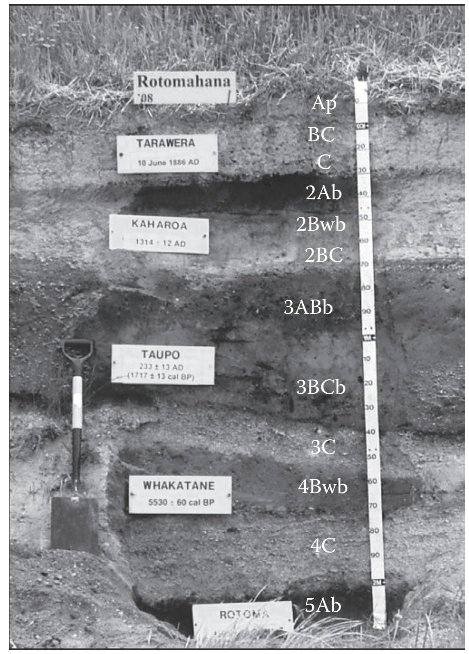

FIGURE 33.17 Udivitrand (Rotomahana soil) from New Zealand composed of multiple tephra deposits and associated buried soils and horizons; scale divisions = 10 cm; the oldest tephra layer (Rotoma) at bottom of exposure is 9505 ± 25 calibrated years BP.

In the first scenario, successive thin tephra deposits (ranging from millimeters to centimeters in thickness) accumulate incrementally and relatively infrequently, so that developmental upbuilding ensues. Such a situation occurs typically at distal sites. The thin materials deposited from each eruption become incorporated into the existing profile over time. Top-down pedogenesis continues as the tephras accumulate but its effects are lessened because any one position in the sequence is not exposed to pedogenesis for long before it becomes buried too deeply for these processes to be effective as the land surface gently rises (Figure 33.18). This history thus leaves the tephra materials with a soil fabric inherited from when the tephra was part of the surface A horizon or subsurface Bw horizon (Lowe and Palmer, 2005; Lowe and Tonkin, 2010). Each part of the profile has been an A horizon at one point, as illustrated in Figure 33.18.

In the second scenario, tephra accumulation is more rapid, as occurs in locations close to volcanoes, or when a much thicker layer (more than a few tens of centimeters) is deposited from a powerful eruption. In the latter case, the antecedent soil is suddenly buried and isolated beyond the range of most soil-forming processes (i.e., it becomes a buried soil horizon or paleosol). A new soil will thus begin forming at the land surface in the freshly deposited material. This scenario typifies retardant upbuilding, which recognizes that the development of the now-buried soil has been retarded or stopped, and the pedogenic "clock" reset to time zero for weathering and soil formation to start afresh. An example of a multisequal Andisol profile formed via retardant upbuilding pedogenesis since ~9500 calendar

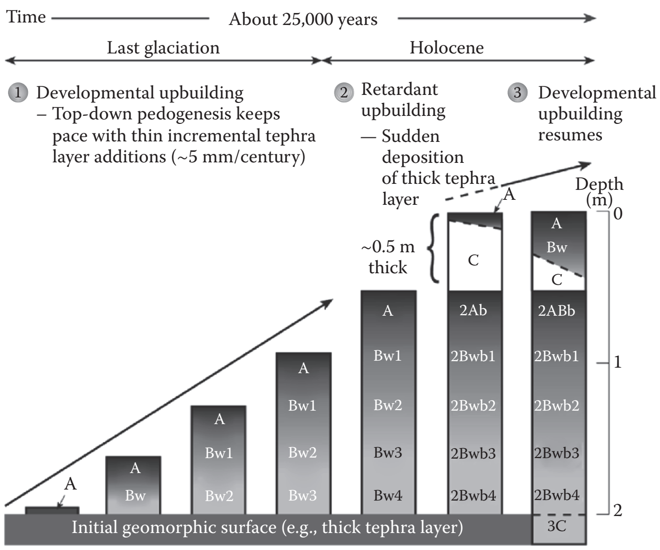

FIGURE 33.18 Model of upbuilding pedogenesis profiles over ~25,000 years in the Waikato region, New Zealand. (From Lowe, D.J., and P.J. Tonkin. 2010. Unravelling upbuilding pedogenesis in tephra and loess sequences in New Zealand using tephrochronology, p. 34–37. *In* Proc. IUSS 19th World Soil Congress, Symposium 1.3.2 Geochronological techniques and soil formation, Brisbane, Australia August 1–6, 2010. Published on DVD and online at: http://www.iuss.org. With permission.)

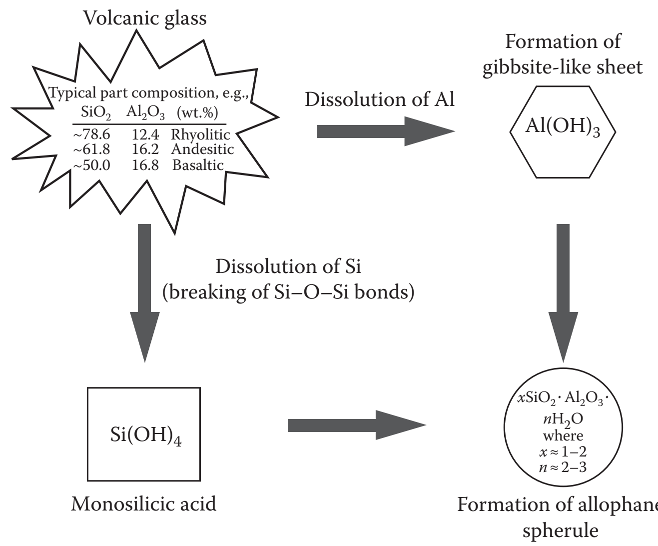

FIGURE 33.19 Volcanic glass compositions and dissolution of Al and Si to form allophane. (After Hiradate, S., and S.-I. Wada. 2005. Weathering processes of volcanic glass to allophane determined by ^{27}Al and ^{29}Si solid-state NMR. Clays Clay Miner. 53:401–408.)

(cal) years ago is shown in Figure 33.17. Each of five successive tephra deposits shows the imprint of top–down pedogenesis, as depicted by their horizonation. But the sudden arrival of a new deposit every few thousand years or so on average has buried and effectively isolated each of the weakly developed "mini" soil profiles as the land surface rises.

In addition to stratigraphic factors, Andisols are markedly affected by the mineralogical and physicochemical compositions of the parent tephras, or associated deposits derived from remobilization of volcanic and other material (collectively termed volcaniclastic deposits). For example, the marked influence of windblown dust (mainly basaltic glass) on soil properties in Iceland was described by Arnalds (2010). Tephras differ widely according to the chemical makeup of magmas of the volcanoes that generated them. The chemistry of magmas, especially Si content, governs the way a volcano erupts. Three main magma types and resulting eruptives can be identified according to their chemical composition—rhyolitic (≥~70% SiO_2), andesitic/dacitic (~50%–70% SiO_2), and basaltic (≤~50% SiO_2). All magmas generate volcanic glass, a noncrystalline, easily weatherable mineraloid, and various other primary silicate minerals (Shoji et al., 1993; Nanzyo, 2002; Smith et al., 2006; Alloway et al., 2007; De Paepe and Stoops, 2007). Glass especially provides much of the Si and Al required to dissolve and re-form as allophane or other aluminosilicate clay minerals, and the amounts differ according to the magma composition as shown in Figure 33.19. Feldspars also release Si and Al via weathering for clay formation.

Generally, the basaltic and intermediate tephras tend to weather more readily than rhyolitic tephras, and in all cases glasses weather very quickly (Neall, 1977; Kirkman and McHardy, 1980; Colman and Dethier, 1986; Hodder et al., 1996; Shikazono et al., 2005). Compared with hard rock, the fragmental tephra components, especially vesicular glass and pumice fragments, have a much greater surface area and higher porosity and permeability, and so break down to constituent compounds very readily. As this weathering occurs, various elements including Si and Al are released into chemical solution either for subsequent leaching, complexing with humic materials, plant uptake, or synthesis (neoformation) into relatively stable nanominerals and other clay minerals (Lowe, 1986; Vacca et al., 2003; Dahlgren et al., 2004). Andisols developed on (base-rich) basaltic eruptives are inherently more fertile than those on more siliceous eruptives (Wolff-Boenisch et al., 2004). Weakly weathered tephras give rise to the glass-rich Andisols that belong to the Vitrand suborder.

33.3.5.2 Climate

Climate plays an important role in Andisol formation, and soil moisture and temperature regimes are used to differentiate all the suborders except Vitrands (Ping, 2000). The majority of Andisols, as reported previously, are found under udic soil moisture regimes, and around half are found in the tropics, with the rest occurring in temperate or boreal regions. Climatic conditions help govern the combinations of processes, collectively referred to as andisolization, that occur in soils developing on tephras. Andisolization is the in situ formation of andic soil materials comprising nanominerals composed of "active" Al, Si, Fe, and humus (Dahlgren et al., 2004). The process is discussed in more detail in Section 33.3.6.

The essential conditions for the formation of allophane are the activity of silicic acid in the soil solution, the availability of Al species, and the opportunity for coprecipitation (Figure 33.19). These conditions are controlled largely by the leaching regime, the organic cycle, and pH, which, in turn, are potentially influenced by numerous environmental factors including rainfall, drainage, depth of burial, parent tephra composition and accumulation rate, dust accession, type of vegetation and supply of humic substances, and human activities (such as burning vegetative cover), together with thermodynamic and kinetic factors (Parfitt and Saigusa, 1985; Lowe, 1986, 1995; Hodder et al., 1990; Chadwick et al., 2003; Dahlgren et al., 2004; Parfitt, 2009). Availability of Al, derived mainly from the dissolution of glass or feldspars, is assumed to be unlimited in this model, though potentially more is available from andesitic and especially basaltic tephras than rhyolitic tephras (Lowe, 1986; Shoji et al., 1993; Cronin et al., 1996). In contrast, in pedogenic environments rich in organic matter and with pH ≤ 5, humus effectively competes for dissolved Al, leaving little Al available for coprecipitation with Si to form allophane or halloysite (Shoji et al., 1993; Dahlgren et al., 2004; see Section 33.3.5.4).

In New Zealand, both mineralogical and soil solution studies on soils derived from tephras extending across a rainfall gradient showed that rainfall, coupled with through-profile drainage, helps govern Si concentration [Si] in soil solution and thus the likelihood of allophane being formed or not (Parfitt et al., 1983; Singleton et al., 1989; Parfitt, 2009). The Si leaching model shown in Figure 33.20 is summarized as follows: where [Si] is less than ~10 ppm (mg L^{-1}), allophane is formed; where [Si] is greater than ~10 ppm, halloysite is formed. If [Si] is close to ~10 ppm, then either allophane or halloysite may predominate. A profile throughflow threshold of approximately 250 mm $year^{-1}$ of drainage water likely controls [Si]—less than ~250 mm $year^{-1}$ means that the loss of Si is insufficient for Al-rich allophane to form and halloysite (or Si-rich allophane) forms instead (Parfitt et al., 1984; Lowe, 1995).

It is emphasized, however, that allophane and other nanominerals can form under lower rainfalls (~700 mm $year^{-1}$) such as in Mexico, Iceland, and South Australia (Arnalds et al., 1995;

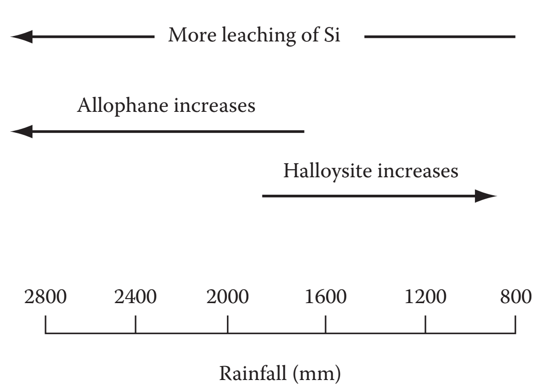

FIGURE 33.20 Simplified allophane–halloysite rainfall leaching model.

Dubroeucq et al., 1998; Ugolini and Dahlgren, 2002; Lowe and Palmer, 2005). As well, other factors such as parent tephra composition and vegetation are also important.

33.3.5.3 Topography

Andisols are found on all types of topography and at a wide range of elevations (from sea-level to >3000 m). Because of their strong association with the products of volcanism, Andisols occur most commonly in volcanic landscapes, which can range from mountainous to hilly terrains on the flanks of stratovolcanoes, dome complexes, shield volcanoes, and scoria or tuff cones through to rolling or essentially flat-lying landscapes, some associated with plateaus of tephra-draped welded ignimbrite sheets derived from large caldera eruptions, such as occur for example in the western United States, Japan, and northern New Zealand (Lowe and Palmer, 2005). Because tephras are carried long distances by the wind, and can easily be reworked by wind and water, they can be deposited in a variety of sedimentary and other landscapes including dune plains, fluvial terraces or plains, and flat intermontane valleys (Ping, 2000; Lowe and Palmer, 2005). The thickest tephra-fall deposits tend to be downwind from the volcanic source areas on rolling or flattish land surfaces (Lowe and Palmer, 2005). On hilly landscapes, tephras are variable in thickness because of erosion (especially during the last glaciation), and profiles typically are thinner on slopes than on more stable geomorphic surfaces (Neall, 2006). The effects of specific topographic positions on Andisol formation have been evaluated, for example, by Navarette et al. (2008) in the Philippines.

33.3.5.4 Vegetation

Because Andisols occur in all moisture and temperature regimes on different landscapes at different elevations, vegetation is accordingly highly variable. In general, Andisols are usually moderately acid to very acid and so vegetation associated with acid soils will predominate. Because of their relatively high water-holding capacities, Andisols, even in xerophytic areas, are likely to support more luxuriant vegetation than other nonandic soils in the same environment. When the vegetation cover is grass, a more humus-rich profile usually develops. A thick, dark surface known as a melanic epipedon is common in many Andisols that have formed under grasses. Melanic epipedons contain ≥6% organic C as a weighted average, have black soil colors (melanic index ≤1.70), and attain a thickness of ≥30 cm (Soil Survey Staff, 2010). Andisols with melanic epipedons are usually classified as Melanudands, an example of which is shown in Figure 33.14b.

Although the vegetation on Andisols is quite variable, it clearly has a major influence on the type of Andisol formed (Ping, 2000). In the Pacific Northwest and parts of Alaska, and in New Zealand, Udands generally formed under forest. However, Udands are known to form under other vegetation types such as grasses in south central Alaska (Ping, 2000) and in Japan, where pampas grass (*Miscanthus sinensis*), known as "susuki," has formed Melanudands, Sitka spruce (*Picea sitchensis*) Hydrudands, and blue joint grass (*Calamagrostis canadensis*) Fulvudands. The organic matter is dominated by humic acids in the first cases and by fulvic acids in the latter (Ping et al., 1989; Shoji et al., 1993; Nanzyo, 2002; Dahlgren et al., 2004).

In Japan and elsewhere, the humic acids in many Andisols and associated soils are characterized by their stability and aromatic (humified) structure (Shoji et al., 1993; Dahlgren et al., 2004; Hiradate et al., 2004; Parfitt, 2009). These features arise from the presence of labile and active metals, chiefly Al and Fe, which are able to bind with humic substances to form nanoparticles of Al- and Fe-humic acid complexes that are very resistant to degradation or leaching (Hiradate et al., 2004; Basile-Doelsch et al., 2005). As well as strong bonding, the metal-humic acid complexes are also evidently protected by Al toxicity to microorganisms or enzymes, physical encasement within abundant micropeds, and a P deficiency of microorganisms arising from high P retention (Ugolini and Dahlgren, 2002; Parfitt, 2009). The resultant very dark or black melanic epipedons can contain as much as 25%–30% organic carbon (Drijber and Lowe, 1990; Nanzyo, 2002; Hiradate et al., 2004).

Most Xerands, such as those in central Washington, Oregon, and northern California, have formed under mixed vegetation of grass and forest whereas those in South Australia formed under dense to semi-open woodland and grassland/shrubland or fernland (Ping, 2000; Lowe and Palmer, 2005; Takesako et al., 2010). Torrands formed under shrubby or grass vegetation, and Vitrands under grass or forest in North America and Mexico and under broadleaf–podocarp forest in New Zealand. Cryands generally formed under forest, but in Iceland and other high-latitude polar regions including the Aleutians, these soils formed under tundra or moss, heath, grass, and "desert" covers (Arnalds et al., 1995; Ping, 2000; Arnalds and Kimble, 2001). In the tropics, the vegetation is often rain forest, particularly on the slopes of volcanoes where grassland and savanna types also occur.

Some black A horizons on Andisols in New Zealand, typically ≤20 cm thick, have been attributed largely to the effects of bracken fern (*Pteridium* spp.), which replaced much of the original forest after human-induced deforestation by Polynesian burning from about AD 1300 (Newnham et al., 1999; McGlone et al., 2005; Lowe, 2008). In northern Idaho, establishment of bracken fern following forest canopy removal is associated with decadal-scale changes in soil properties. These include increased soil C, darker soil colors, lower pH, and increased organic forms of active Al (Johnson-Maynard et al., 1997).

33.3.5.5 Time

Andisols generally are found on relatively young parent materials in humid climates because, on older and more stable landscapes, more weathering would normally have taken place resulting in the transformation to other soil orders such as Ultisols or Oxisols. The drier and cooler the environment, the longer it usually takes for an Andisol to form (Shoji et al., 2006). Tephras or volcanic materials, either very young or in extremely dry areas, and not sufficiently weathered to form the minimum requirements for andic soil materials, are placed in the Entisol order. An A horizon can form in newly deposited tephra in a few hundred years or so whereas the development of BC and then Bw horizons beneath, it may take several hundreds to several thousands of years according to environmental conditions and the nature of the tephra (Shoji et al., 1993; Ping, 2000; Lowe and Palmer, 2005; Lowe, 2008). However, in northeast Kodiak Island, Alaska, a Cryand formed on tephra erupted in the 1912 Mount Katmai event already has thin E and Bw horizons (Ping, 2000), and soils on young basaltic tephra surfaces in Iceland (<100 years) are classified as Andisols (Arnalds and Kimble, 2001).

In situations where just a single eruption event has taken place to produce a parent deposit, then Andisols developed in that material have the same age as the eruption. However, where Andisols comprise upbuilding sequences of tephras and soil horizons (as commonly occurs), the maximum age of the soil-profile constituents depends on the depth at which the profile "base" is drawn. In considering the uppermost 1–2 m of Andisols, some generalizations about their *composite* ages can be made. For example, in Japan, many Andisols are ~5,000–10,000 cal years old although some are considerably older (~25,000 cal years old; Takesako and Muranaka, 2006). In southeast Alaska, Cryands may be up to ~15,000 cal years old (Ping, 2000). In Iceland, most of the Andisols range in age from a few decades to ~10,000 cal years old (Arnalds et al., 1995). In New Zealand, there is a much wider range of ages on Andisols. Most of the Vitrands are ~700 years old (on rhyolitic Kaharoa Tephra) and ~1,800 years old (on rhyolitic Taupo Tephra); extensive Udands date back ~20,000–25,000 cal years and some are as old as ~60,000 years. In all these countries and others, tephrochronology—the identification and correlation of tephras and their application as a linking and dating tool (Alloway et al., 2007; Lowe, 2011)—has been used to provide age models for the Andisols. These age models rely mainly on radiocarbon and other techniques including dendrochronology and depositional age-depth modeling as well as wiggle-match dating (Lowe, 2011). Historical accounts provide ages for the youngest tephras (Alloway et al., 2007).

33.3.6 Pedogenic Processes in Andisols

The importance of upbuilding and polygenesis in the formation of Andisols has already been described. In this section, the collective process of andisolization in tephra materials with a significant content of volcanic glass is outlined. Andisolization leads to the formation of a fine colloidal fraction dominated by nanoscale minerals, notably allophane, ferrihydrite, and metal–humus complexes, with typically subordinate amounts of imogolite. Nanoscale minerals are defined as having at least one dimension in the nanorange, which is 1–100 nm (Hochella, 2008). These nanoscale minerals are composed of active forms of Al, Si, Fe, and humus and impart andic soil properties (Theng and Yuan, 2008). Although volcanic glass is a common component in many Andisols, it is not a requirement of the Andisol order and some soils develop andic properties without the influence of glass (Soil Survey Staff, 1999). Soil solution and micromorphological studies show that, in contrast to podzolization, translocation of Al, Fe, organic matter, and clays is normally minimal in Andisols (Jongmans et al., 1994; Soil Survey Staff, 1999; Ugolini and Dahlgren, 2002; Stoops, 2007).

Allophane and imogolite (both aluminosilicates) and ferrihydrite (an iron hydroxide) are formed by the synthesis of soluble Al, Si, or Fe that are released from glass or other minerals (mainly feldspars or various mafic minerals) via rapid dissolution and hydrolysis by carbonic acid (H_2CO_3) normally under moderately acid and humid conditions. As soil solutions become rapidly oversaturated with respect to these nanominerals, they are precipitated preferentially because their nucleation is kinetically favored over that of less-soluble, long-range-order (crystalline) phases (Ugolini and Dahlgren, 2002; Theng and Yuan, 2008). Thermodynamic stability diagrams show that imogolite and very likely Al-rich allophane are more stable than halloysite over a wide range of Si activity, with the latter more stable than both imogolite and Al-rich allophane only at high Si activity (Farmer et al., 1991; Lowe and Percival, 1993; Hodder et al., 1996; Harsh, 2000). The central role of the Si leaching model in helping govern Si activity (i.e., concentration of $Si(OH)_4$ or monosilicic acid) was discussed earlier. Al-rich allophane forms at pH values between ~5 and 7, a value of at least ~4.8 being required for it to precipitate (Lowe, 1995; Dahlgren et al., 2004). Al undergoes hydrolysis and polymerizes in gibbsite-like octahedral sheets as $(OH)_3Al_2$ and then combines with Si in tetrahedral coordination as O_3SiOH to form imogolite or allophane (Figure 33.19; Hiradate and Wada, 2005; Theng and Yuan, 2008). Allophane can form rapidly and has been observed to precipitate in a few months on the face of open soil pits containing rhyolitic tephra (Parfitt, 2009).

Note that a long-standing previous model in which allophane was suggested to weather or "crystallize" by solid-state transformation to form halloysite with time (because the latter often occurs in older, that is, deeper, tephras and associated soil horizons) has been largely discounted and replaced by this Si leaching model and its variants. For allophane to alter to halloysite requires a complete rearrangement of the atomic structures, and this could only occur by dissolution and reprecipitation processes because the allophane would need to "turn inside out" so that Si-tetrahedra are on the outside, not inside, the curved Al-octahedral sheets (Figure 33.21; Parfitt et al., 1983; Ildefonse et al., 1994; Lowe, 1995; Churchman, 2000; Hiradate and Wada, 2005; Parfitt, 2009). The effect of time is clearly subordinate because glass can weather directly via dissolution either to allophane or halloysite depending on glass composition and both macro- and microenvironmental conditions, not time (Lowe and Percival, 1993; Lowe, 1995; Parfitt, 2009). As well, allophane can persist for hundreds of thousands of years under favorable conditions (Stevens and Vucetich, 1985; Churchman and Lowe, 2011, Section 20.1).

The formation of ferrihydrite in Andisols and other soils has been described in a number of papers including Schwertmann and Taylor (1989), Childs (1992), Lowe and Percival (1993),

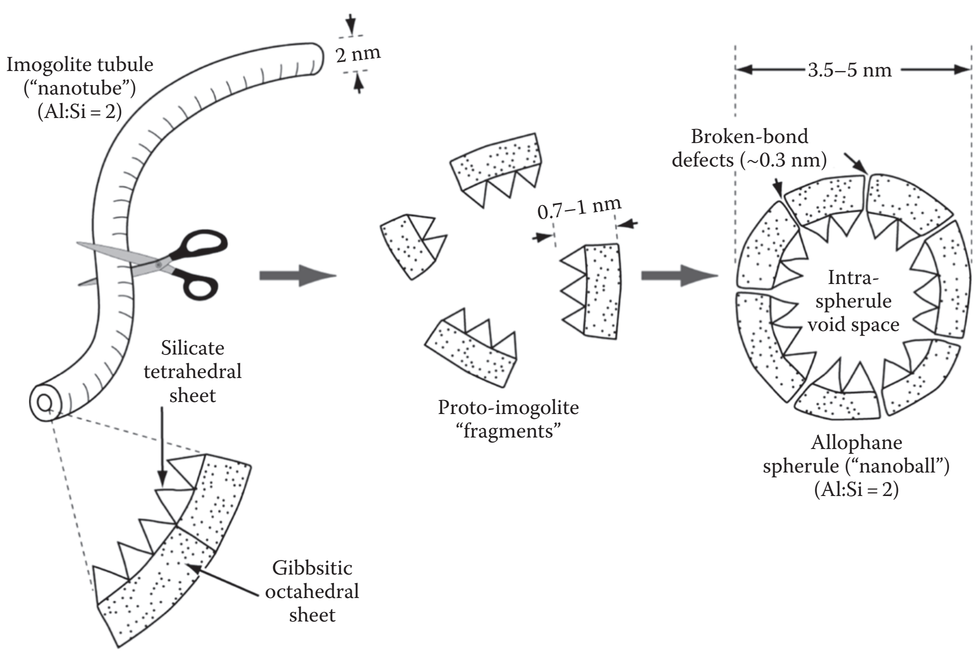

FIGURE 33.21 Diagram of imogolite tubes and allophane spherules. "Proto-imogolite" fragments have an imogolite atomic structure over a short range and link to form porous, hollow nanospheres with water molecules occupying the intra-spherule interior as well as being adsorbed to the outer AlOH surface (Parfitt, 2009). (After Lowe, D.J. 1995. Teaching clays: From ashes to allophane, p. 19–23. *In* G.J. Churchman, R.W. Fitzpatrick, and R.A. Eggleton (eds.) Clays: Controlling the environment. CSIRO Publishing, Melbourne, Australia.)

Churchman (2000), Bigham et al. (2002), Nanzyo (2002), and Dahlgren et al. (2004). After release of structural Fe from mafic primary minerals such as pyroxenes, amphiboles, biotite, and olivine (and presumably also from glass), Fe^{2+}-containing soil water oxidizes relatively quickly to Fe^{3+}, which immediately hydrolyzes to ferrihydrite. This occurs in the presence of organic matter, silicate, or phosphate, all of which inhibit the formation of more thermodynamically stable crystalline iron oxides. Ferrihydrite can redissolve readily by reduction (Schwertmann, 2008).

The availability of Al^{3+} is the critical factor regulating the formation of nonallophanic Andisols, which form preferentially in environments rich in organic matter and with pHs of ~5 or less, and which typically contain 2:1 layer silicate clays (Shoji et al., 1985, 1993; Nanzyo, 2002; Ugolini and Dahlgren, 2002; Dahlgren et al., 2004). At these pHs, organic acids are the dominant proton donors lowering pH and aqueous Al^{3+} activities through the formation of Al–humus complexes (also Fe–humus complexes, but less commonly). Under these conditions, carboxyl groups of humus and the 2:1 layer silicates (derived mainly from windblown dust) effectively compete for dissolved Al, leaving little Al available for coprecipitation with Si to form allophane or imogolite (Mizota and van Reeuwijk, 1989; Dahlgren et al., 2004). The preferential incorporation of Al into Al–humus complexes (and hydroxyl-Al interlayers of 2:1 layer silicates) has been termed the *antiallophanic* effect (Shoji et al., 1993). In Japan, allophanic Andisols tend to be formed where parent tephras are more basic (basaltic to andesitic) and rainfall is <1000 mm per year, so that higher pH values (>5) are favored (Dahlgren et al., 2004). Their distribution pattern is also attributed in part to proximity to volcanoes and the frequency therefore of tephra accretion vs. exotic dust accession (Saigusa and Matsuyama, 2004): sites closer to volcanoes effectively received a "top up" of Al through more regular deposition of weatherable tephras (resulting in allophanic Andisols) than at distal sites where the Al became quickly depleted (resulting in nonallophanic Andisols) through both humus complexing and "dilution" from the ongoing fallout of 2:1 mineral-bearing dust blown from China.

Changes in elemental composition during andisolization in Japan were studied by Nanzyo and Takahashi (2005).

33.3.7 Management of Andisols

Andisols support a much greater population than their limited extent would suggest. One reason for this is that crop productivity on Andisols is very high due to good physical properties and relatively high native fertility in some locations. Periodic additions of fresh tephra, especially when intermediate to basic in composition, can resupply potential nutrients and maintain favorable fertility levels in areas where other soils tend to be depleted in nutrients. For example, a "dusting" of andesitic ash fallout from the 1995–1996 Mt. Ruapehu eruptions in New Zealand added between 30 and 1500 kg ha^{-1} of sulfur as well as small quantities of Se, Mg, and K to substantial land areas (Lowe and Palmer, 2005). Consequently, in many areas, few inputs with the exception of P have been required for successful agricultural production on many Andisols. Andisols are also productive forest soils. In the Pacific Northwest region of United States, the presence of tephra (ash) mantles and their water-holding capacity is closely linked to forest productivity (Kimsey et al., 2008). Vitrands in New Zealand also support very productive plantation forestry (Ross et al., 2009).

Nevertheless, in other locations such as New Zealand where rainfall and leaching are generally high, and where siliceous tephras low in metals are common, many Andisols have low fertility because of a range of chemical limitations both acquired and inherited. As well as P fixation, exchangeable bases (especially K) tend to be low and other elemental deficiencies may include Mg, S, Co, and others (Lowe and Palmer, 2005). The high productivity of these soils is thus maintained by regular fertilizer input in combination with their generally excellent physical properties.

Most Andisols are acid and where acidity is particularly high, high levels of exchangeable Al may inhibit crop growth (Al toxicity). This is especially a problem in nonallophanic Andisols (Shoji et al., 1993; Shoji and Takahashi, 2002; Dahlgren et al., 2004). In such instances, addition of lime can reduce the amounts of organically complexed Al and exchangeable Al (Takahashi et al., 2006), or acid-tolerant crops must be selected (Michaelson and Ping, 1987; Nanzyo, 2002; Shoji and Takahashi, 2002). Where Andisols have been acidified by long-term N inputs, serious consequences for crop production can arise because as pH decreases, CEC also decreases. As a result, such soils are able to hold few base cations (Ca, Mg, and K) for plant uptake (Sumner and Hylton, 1993).

Requirements for N vary according to the ease with which organic matter is mineralized in these soils. Nitrogen mineralization is slower in allophanic Andisols than in nonallophanic Andisols because organic matter interacts with and is stabilized by allophane in the former (Parfitt, 2009). Liming often promotes the release of substantial amounts of N from nonavailable reserves, thereby improving N fertility.

As noted above, the high P-fixing capacities of most Andisols necessitate inputs of P for successful agricultural production (Michaelson and Ping, 1990; Dahlgren et al., 2004; Lowe and Palmer, 2005). Such inputs in the form of rock phosphate are effective on very acid Andisols but in any event, soluble P sources should be band-placed or otherwise protected from intimate mixing with the soil (Sumner and Hylton, 1993).

Other nutrients may sometimes be deficient in Andisols depending of the chemical composition of the parent material. One notable example occurs in the central North Island of New Zealand where Vitrands, formed in the silica-rich Taupo and Kaharoa tephras, are deficient in micronutrients including Co, Se, Cu, B, I, and Mo. Inherently low Co in these pumiceous parent tephras led to low Co levels in soils and herbage, and ultimately to a deficiency in ruminant animals (sheep and cows) that developed a serious and commonly fatal wasting disease known as "bush sickness" (Lowe and Palmer, 2005; Neall, 2006). This term referred to a variety of symptoms exhibited by livestock stemming from their inability to produce vitamin B12, of which Co is an essential component (Cornforth, 1998). The association of this condition with the Vitrands was recognized in the early 1930s (Grange and Taylor, 1932), and the cause identified a few years later. Acquired Co deficiency also occurred in some Udands and non-Andisol orders where high rainfall and strong leaching were the cause. Subsequent use of cobaltized superphosphate, and other methods, corrected the problem (Neall, 2006).

Andisols, in general, are noted for their good physical properties including high water-holding capacity, free drainage, good tilth and friability, resistance to water erosion, stable aggregation, low bulk density and high porosity, and hence good aeration. Consequently, they are usually excellent media for seedling emergence and root growth and proliferation. It should be emphasized that these properties can and often are altered through management (e.g., Candan and Broquen, 2009). Where uncultivated, undisturbed Andisols may be resistant to erosion, but loss of vegetative cover, compaction, and subsequent reduction in infiltration can lead to severe runoff and erosion including deep gullying in Vitrands (Lowe and Palmer, 2005). Agricultural production on Andisols can result in erosion and compaction, and no-till or reduced tillage farming practices are recommended to minimize these problems in agricultural systems (Dahlgren et al., 2004). Heavy livestock grazing can also compact Andisols (Schlichting, 1988). In the Pacific Northwest of United States, compaction resulting from mechanized timber harvesting can reduce infiltration and water-holding capacity of Andisols (Cullen et al., 1991). Severe erosion from harvested sites can also significantly reduce water-holding capacity to the point where site productivity is drastically diminished (McDaniel et al., 2005). In Iceland, Cryands have been overgrazed, denuded, and subject to severe wind erosion (Arnalds et al., 1995; Arnalds and Kimble, 2001).

As noted earlier, properties of sensitivity and thixotropy can cause Andisols to behave in a fluid-like manner when loading pressures are applied. Thus, Andisols are susceptible to failure when disturbed on slopes, which can cause them to reach the liquid limit, generating landslides (Basile et al., 2003; Neall, 2006). Mantling of harder bedrock by andic soils and the layering associated with tephra deposits and buried soils can form planes of failures (Arnalds, 2008).

Because of their sorptive capacity for anions and metals, Andisols offer potential environmental benefits as active filters. High nitrate adsorption by allophanic Andisols was correlated with retarded nitrate transport rates under coffee plantations in Costa Rica (Reynolds-Vargas et al., 1994; Ryan et al., 2001). Reduced phosphorus leaching losses from surface-applied domestic effluent in New Zealand have been attributed the relatively high P sorption capacity of Andisols (Barton et al., 2005). Similarly, the sorption of arsenate by Andisols offers potential for purifying drinking water or for remediating As-contaminated soils (Arai et al., 2005; Theng and Yuan, 2008). McLeod et al. (2008) have also demonstrated that New Zealand Andisols leached very small amounts of surface-applied microbial tracers via bypass flow, attributable to their sorptive capacity and the fine, porous nature of the soils. These and other studies suggest that Andisols may play an important role in reducing groundwater contamination.

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33.4 Entisols

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33.4.1 Introduction

Entisols are mineral soils that commonly exhibit minimal pedogenic alteration (Soil Survey Staff, 1999, 2010). This taxonomic order was established to preserve the early Russian "azonal" concept for soils with undeveloped profiles crossing bioclimatic zones (Smith, 1986). Typically described as A-C, A-Cr, or A-R profiles, Entisols retain many physical and chemical properties of the associated parent materials. Most A horizons classify as ochric epipedons, but with Ap horizons classifying as anthropic in some human-disturbed lands and as agric in some agricultural lands. Occasionally, Entisols have histic epipedons if affiliated with sulfidic materials in coastal regions with persistently high water tables. Highly disturbed soils from human activity may have fragments of former diagnostic horizons classified as Entisols.

Subsurface B horizons may be described to indicate a pathway of pedogenesis (e.g., Bw), but the properties must not meet the minimum requirements of any subsurface diagnostic horizon (e.g., cambic). Exceptions are the recognition of a weathered albic horizon as long as genetically related illuvial horizons (e.g., argillic and spodic) do not occur within 2 m of the surface, and cryoturbation in Gelisols that prevents diagnostic horizons from forming. Despite minimal horizonation, however, Entisols must develop physical and chemical properties capable of supporting plants. This characteristic distinguishes Entisols from nonsoil areas such as rock outcrop, water, ice, or badlands.

33.4.2 Distribution

Entisols most often occur where the influence of the soil-forming factors retards pedogenesis. Landforms where Entisols form are marshes and estuaries, modified urban areas, sand dunes, floodplains, and arid regions with steep or rocky slopes (Figure 33.22). These landforms are associated with wetlands, inert sandy parent materials, rapid rates of aggradation or erosion, and intensive human activity (Buol et al., 1997). Because of the geographic extent of these environments, Entisols are the most abundant of all soil orders covering approximately 16% of the world's ice-free land mass (Soil Survey Staff, 1998). Up to 60% of Entisols occur in temperate and tropical regions, and many mapped in circumarctic regions have been replaced by the Gelisol order.

Entisols are replete in buried contexts throughout the Phanerozoic (see Birkeland, 1999). Recognition of Entisols is important for understanding Quaternary landscape evolution and preservation potentials of the buried archaeological record (Holliday, 2004). Paleo-Entisols in the pre-Quaternary rock record serve as sequence stratigraphic markers (McCarthy and Plint, 1998; Atchley et al., 2004), enhance interpretations of pedofacies relations (Kraus, 2002), provide inferences regarding colonizing ecosystems (Retallack, 1994), and serve as repositories of paleobotanical and paleontological information (see Retallack, 2001).

33.4.3 Formation

Ripening and melanization are the principle soil-forming processes that initiate the development of surface diagnostic horizons and Entisols (Buol et al., 1997). Grossman (1983) reports data showing that bulk densities in alluvial settings initially increase because of frequent wetting and drying cycles, which leads to compaction. This process is similar to ripening as freshly deposited sediments from running water become dehydrated, but is followed by ped formation and increased porosity (Pons and van der Molen, 1973). Melanization contributes organic materials to the parent material by way of plant and animal activity promoting structural development that is accompanied by increased porosity and minor release

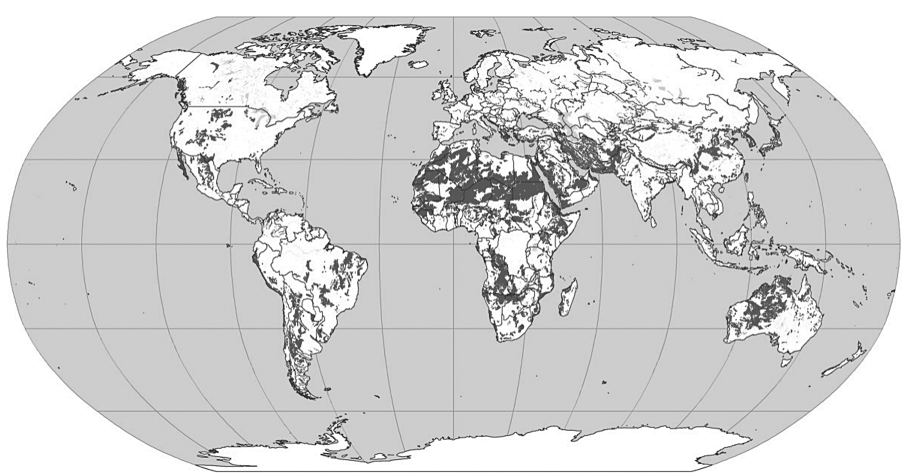

FIGURE 33.22 Global distribution of Entisols. (Courtesy of USDA-NRCS, Soil Survey Division, World Soil Resources, Washington, DC, 2010.)

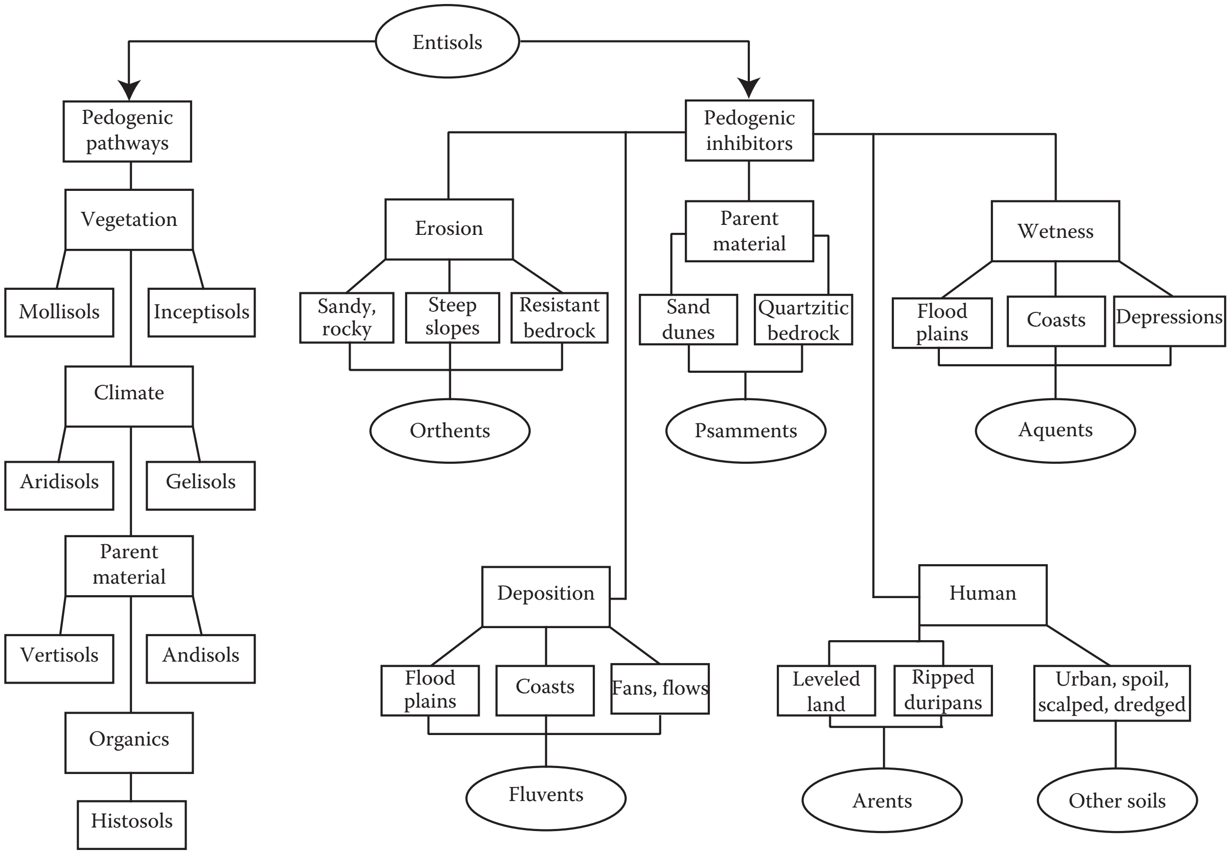

FIGURE 33.23 Flow diagram illustrating pedogenic pathways of Entisols to other soil orders and pedogenic inhibitors conditioning the formation of Entisol suborders.

and leaching of soluble constituents. In resistant bedrock, this process increases porosity as rock material is weathered and replaced by organic material. The processes of ripening and melanization in the formation of Entisols can proceed as long as mollic or umbric epipedons or subsurface diagnostic horizons do not form.

Entisols serve as the precursor to all other soil orders during their evolutionary history (Figure 33.23) (Soil Survey Staff, 1999, 2010). In grasslands, Mollisols with mollic epipedons form from Entisols when steady state has been attained in the accumulation of organic matter. In forest settings, ochric epipedons and Inceptisols may form because of relatively low, subsurface biomass production. Entisols develop into Aridisols outside of floodplains in areas with aridic soil moisture regimes. Vertisols form from Entisols in shrink-swell parent materials where wet–dry cycles are prevalent, and Andisols form from Entisols in parent materials with large quantities of volcanic ash in cool and humid climates. Histosols may form from Entisols with the accumulation of significant amounts of organic matter in association with high water tables and anaerobic conditions.

33.4.4 Suborders

In contrast to most other soil orders, suborders of Entisols are not defined by precipitation or temperature regime. Rather, Entisol suborders are designed primarily to differentiate conditions that impede subsurface horizon development (Smith, 1986; Soil Survey Staff, 1999, 2010). The five Entisol suborders are Aquents, Arents, Psamments, Fluvents, and Orthents (Figures 33.23 and 33.24). The great group and subgroup lower taxonomic levels recognize extragrades and intergrades to other soil moisture regimes and diagnostic conditions (Table 33.18).

33.4.4.1 Aquents

Aquents are wet Entisols. They exist widely in the United States, except in Arizona and New Mexico, with the largest concentrations in southern Florida (Soil Survey Staff, 1990). Worldwide, Aquents are not extensive, covering approximately 0.1% of the Earth's land. Areas where Aquents occur include salt marshes of coastal tidal areas, freshwater marshes and swamps, alluvial flats and backswamps of floodplains, and outwash plains that receive new deposits of sediment at frequent intervals. Depending on where Aquents are located, some can be used for agricultural production when drained. Many undrained areas of Aquents are used for pasture or wildlife habitat.

Aquents are defined as having aquic conditions and sulfidic materials within 50 cm of the mineral soil surface, or permanent saturation and a reduced matrix below a depth of 25 cm from the mineral soil surface, or texture and color requirements if a densic, lithic, or a paralithic contact, or the presence of aquic conditions for some time in some years in a layer between 40 and 50 cm from the

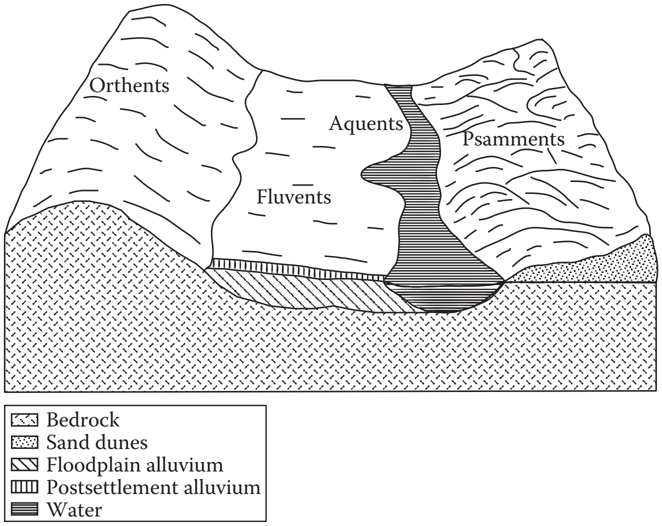

FIGURE 33.24 Block diagram illustrating typical landscape positions where Entisol suborders form: Orthents on steep slopes associated with bedrock, Fluvents in floodplains that include postsettlement alluvium, Aquents in poorly drained riparian conditions, and Psamments in sand dunes. Arents not shown.

mineral soil surface (Soil Survey Staff, 1998). Aquents can range in texture from sandy to clayey. One common feature of all Aquents is their saturation with water at or near the surface for much of the year, which presents many engineering and land use challenges.

Wetness in Aquents is normally inferred by the color distribution generated by the redox state of Fe. In most Aquents, this process produces redoximorphic features (Fe concentrations or depletions) within the upper 50 cm of the surface. Redoximorphic features are most common in nonsandy Aquents. The sandy Aquents (Psammaquents) often lack redoximorphic features because of the very low Fe content of the quartz-rich parent materials. Many Aquents have hydric soil indicators (thin layer of muck on the soil surface, strong sulfidic odor, or a gray and reduced soil matrix in the upper 15 cm of the soil; USDA-NRCS, 1996). Florida law and county listings elsewhere show that some Aquents (Sulfaquents, Hydraquents, and Psammaquents) are considered as hydric soils (Florida Statutes, 1995).

33.4.4.1.1 Lower Taxonomic Levels

Aquents are associated with landscapes that accumulate soil water. They exist in cryic to isohyperthermic soil temperature regimes. Sulfaquents and Hydraquents are located along coastal areas in

TABLE 33.18 Listing of Suborder, Great Groups, and Subgroups in the Entisol Order

Suborder	Great Group	Subgroup
Aquents	Sulfaquents	Haplic, Histic, Thapto-Histic, Typic
	Hydraquents	Sulfic, Sodic, Thaptic-Histic, Typic
	Cryaquents	Aquandic, Typic
	Psammaquents	Lithic, Sodic, Spodic, Humaqueptic, Mollic, Typic
	Fluvaquents	Sulfic, Vertic, Thapto-Histic, Aquandic, Aerie, Humaqueptic, Rhodic, Mollic, Typic
	Epiaquents	Aerie, Humaqueptic, Mollic, Typic
	Endoaquents	Sulfic, Lithic, Sodic, Aerie, Humaqueptic, Mollic, Typic
Arents	Ustarents	Haplic
	Xerarents	Sodic, Duric, Alfic, Haplic
	Torriarents	Sodic, Duric, Haplic
	Udarents	Alfic, Ultic, Mollic, Haplic
Psamments	Cryopsamments	Lithic, Aquic, Oxyaquic, Vitrandic, Spodic, Lamellic, Typic
	Torripsamments	Lithic, Vitrandic, Haploduridic, Ustic, Xeric, Rhodic, Typic
	Quartzipsamments	Lithic, Aquodic, Aquic, Oxyaquic, Ustoxic, Udoxic, Plinthic, Lamellic Ustic, Lamellic, Ustic, Xeric, Spodic, Typic
	Ustipsamments	Lithic, Aquic, Oxyaquic, Aridic, Lamellic, Rhodic, Typic
	Xeropsamments	Lithic, Aquic Durinodic, Aquic, Oxyaquic, Vitrandic, Durinodic, Lamellic, Dystric, Typic
	Udipsamments	Lithic, Aquic, Oxyaquic, Spodic, Lamellic, Plaggeptic, Typic
Fluvents	Cryofluvents	Andic, Vitrandic, Aquic, Oxyaquic, Mollic, Typic
	Xerofluvents	Vertic, Aquandic, Andic, Vitrandic, Aquic, Oxyaquic, Durinodic, Mollic, Typic
	Ustifluvents	Aquertic, Torrertic, Vertic, Anthraquic, Aquic, Oxyaquic, Aridic, Udic, Mollic, Typic
	Torrifluvents	Ustertic, Vertic, Vitrixerandic, Vitrandic, Aquic, Oxyaquic, Duric Xeric, Duric, Ustic, X, Anthropic, Typic
	Udifluvents	Aquertic, Vertic, Andic, Vitrandic, Aquic, Oxyaquic, Mollic, Typic
Orthents	Cryorthents	Lithic, Vitrandic, Aquic, Oxyaquic, Lamellic, Typic
	Torriorthents	Lithic Ustic, Lithic Xeric, Lithic, Xerertic, Ustertic, Vertic, Vitrandic, Aquic, Oxyaquic, Duric, Ustic, Xeric, Typic
	Xerorthents	Lithic, Vitrandic, Aquic, Oxyaquic, Durinodic, Dystric, Typic
	Ustorthents	Aridic Lithic, Lithic, Torrertic, Vertic, Anthraquic, Aquic, Oxyaquic, Durinodic, Vitritorrandic, Vitrandic, Aridic, Udic, Vermic, Typic
	Udorthents	Lithic, Vitrandic, Aquic, Oxyaquic, Vermic, Typic

Source: Soil Survey Staff. 1998. Global soil regions. USDA-NRCS, Lincoln, NE.

tidal marshes. Sulfaquents, as the name implies, have sulfidic materials. Hydraquents are thixotropic soils, meaning that they cannot support appreciable overburden pressures (*n* value >0.7). Psammaquents have a texture of loamy fine sand or coarser in all layers within the control section for the family particle size class. Fluvaquents occur on floodplains and are defined by the amount of organic C and its distribution in the soil. Cryaquents form in the colder climatic areas of Alaska and other higher latitudes that receive significant snowfalls, and soil development is slow.

33.4.4.1.2 Genesis

In association with wet conditions, Aquents form where pedogenetic development does not occur or is very slow. The soil must be saturated for long periods, and in some areas, organic matter can accumulate to create a histic epipedon. Saturation in Aquents can be episaturated (perched water table) or endosaturated (apparent water table). The time needed for Aquents to develop can be short. On floodplains and other areas experiencing rapid deposition, Aquents commonly have stratified layers of sand, silt, clay, and organic matter. Aquents forming on floodplains may also have buried histic epipedons. Aquents with high clay content may have vertic properties.

Aquents from Florida (Pellicer series: fine, smectitic, nonacid, hyperthermic Typic Sulfaquents), Egypt (Edko series: fine-loamy, mixed, nonacid, hyperthermic Mollic Fluvaquents), and Alaska (Beluga series: coarse-loamy, mixed, superactive, nonacid, Typic Cryaquents) are selected to represent Aquents from different geographic areas and to reflect different genetic pathways. The Pellicer soils form in clayey marine sediments in tidal marshes along the Atlantic coastal areas of Florida (Carlisle et al., 1981). These soils are very poorly drained and subject to daily tidal flooding. The representative Aquent from Egypt has anthric saturation and redoximorphic features in the form of brownish Fe accumulations. The soil was sampled in a depressional area south of Lake Burullus near Resheed City in northern Egypt (Rasheed et al., 1992). The Edko soil developed in recent fluviolacustrine deposits in an aridic climate. The Beluga soils are extensive on floodplains and alluvial fans in southern Alaska and form in stratified alluvium and colluvium derived from soft clayey shale and sandstone. These soils are poorly drained and saturated with water at the surface for portions of the year.

33.4.4.1.3 Physical and Chemical Properties

Physical properties such as particle size distribution and chemical properties such as organic C content, pH, extractable bases, and CEC_7 are presented for the Pellicer and Edko soils (Table 33.19). The sand, silt, and clay content of the Pellicer varies considerably within the profile, reflecting the different conditions under which the materials were deposited. Organic C content of this soil is high in the surface horizon as a result of the accumulation of organic materials under wet conditions. The Pellicer soils are high in S content (1%–4%), and this is reflected in the extremely low pH values (<3.5).

The sand, silt, and clay content of the Edko soil is fairly evenly distributed through the profile. In contrast to the Pellicer soils, the pH of the Edko soil is high (≥8.4) due to the high Na content.

33.4.4.1.4 Subaqueous Soils

An amendment to *Soil Taxonomy* to better classify subaqueous soils has been developed by Dr. Mark Stolt, Professor of Pedology at the University of Rhode Island (Soil Survey Staff, 2010). The formative element Wass is derived from the German (Swiss) word "wasser" for water and is used for subaqueous Histosols and Entisols. Six great groups within Wassents are proposed keying out in the order: Frasiwassents, Psammowassents, Sulfiwassents, Hydrowassents, Fluviwassents, and Haplowassents. The depth range of the water column where these soils occur is not known but a depth of 2.5 m below the surface has been set for soil survey inventory. Excluded are soils in areas where the water column is too thick (>2.5 m) for the growth of rooted plants. Subaqueous soils occur in protected estuarine coves, bays, inlets, and lagoons.

Only recently have pedologists considered these substrates to be soil material. Consequently, very few subaqueous soils have been characterized, and relationships between subaqueous soils and associated subtidal landforms have not been extensively studied. Recent subaqueous soil projects have been undertaken or are in progress along the Gulf coast in Florida and Texas and the Atlantic Coast in Connecticut, Delaware, Florida, Maine, Maryland, Massachusetts, New Hampshire, and Rhode Island. Work along the southeast Atlantic Coast of Florida (Fischler, 2006) has included mapping and classification of subaqueous soils near Cedar Key, Florida (Ellis, 2006), quantifying human activities that either destroy or create an environment in which it is difficult to reestablish the fragile seagrass vegetation, and studying the biogeochemical characteristics of subaqueous soils as related to aquatic vegetation in a river flowing into the Gulf of Mexico (Saunders, 2007).

33.4.4.2 Arents

Arents are former soils physically disturbed by the actions of humans (Soil Survey Staff, 1999, 2010). Arents are restricted to those soils "which have, in one or more layers between 25 and 100 cm from the mineral soil surface, 3% or more by volume fragments of diagnostic horizons that are not arranged in any discernible order (Soil Survey Staff, 1999, 2010)." Two examples of the transformation of preexisting soils into Arents follow.

Duripans were ripped by heavy equipment to improve the soils for the growth of almond nut trees in California (R.W. Simonson, personal communication, based on observations in the 1960s). According to Dr. Simonson, the ripping was possible only with moderately developed duripans, whereas thick, strongly developed duripans were impossible to rip. The value of the land was greatly improved by the ripping, such that successful commercial almond production became possible. As a result of the duripan ripping, the great group classification of the soils changed from Durixeralfs to Xerarents.

Another example of Arent soil formation occurred in western Tennessee where formerly gullied land was leveled with heavy equipment for agricultural purposes. Fragments of argillic horizons from the original Alfisols permitted the new soils to be identified as Arents after the land disturbance. Thus, Hapludalfs (i.e., Memphis soil series) were reclassified to Udarents following the leveling.

TABLE 33.19 Soil Characterization Data for the Pellicer Clay (Carlisle et al., 1981) and Edko Soil (Rasheed et al., 1992) of the Aquent Suborder

Horizon	Depth (cm)	Very Coarse Sand (%)	Coarse Sand (%)	Medium Sand (%)	Fine Sand (%)	Very Fine Sand (%)	Silt (%)	Clay (%)	pH (H_2O)	OC[a] (%)	Ca[b] ($cmol_c\ kg^{-1}$)	Mg ($cmol_c\ kg^{-1}$)	Na ($cmol_c\ kg^{-1}$)	K ($cmol_c\ kg^{-1}$)	CEC_7 ($cmol_c\ kg^{-1}$)	Base Saturation (%)
Pellicer Clay (Typic Sulfaquents)																
A	0–51	0.1	0.5	2.1	3.5	9.3	23.6	60.9	2.7	14.3	11.7	24.3	27.7	1.0	129.6	50
C1	51–89	0.1	1.1	15.9	21.6	11.3	16.6	33.4	3.2	7.1	6.8	15.7	20.7	1.1	78.6	56
C2	89–140	0.1	2.6	38.7	34.1	10.9	7.5	6.1	3.1	0.8	1.3	3.3	5.0	0.2	15.4	63
Edko Soil (Mollic Fluvaquents)																
A1	0–6	—	4.6	—	37.4	—	26.4	31.6	8.6	3.3	12.0	16.0	64.6	1.1	—	—
A2	6–23	—	4.0	—	29.1	—	36.0	30.9	8.4	1.6	14.0	24.0	67.2	0.9	—	—
C	23–56	—	0.4	—	29.9	—	34.8	29.9	8.6	0.9	10.0	14.0	65.2	4.5	—	—
2C	56–76	—	0.4	—	29.9	—	34.8	29.9	8.6	0.9	10.0	14.0	65.2	4.5	—	—

Source: Soil Survey Staff. 2011. National Cooperative Soil Characterization Database. Available online at http://ssldata.nrcs.usda.gov (Accessed July 8, 2011.)

[a] OC is organic carbon.

[b] Ammonium acetate extractable Ca, Mg, Na, K, and CEC (cation exchange capacity), buffered at pH 7.

33.4.4.2.1 Soils in Human-Deposited or Human-Exposed Parent Materials

Humans affect soils in many ways. These soils form in mine spoil, dredge spoil, and other recently deposited urban and industrially related materials, as well as in bomb craters in war zones (Hupy and Schaetzl, 2006). Some Entisols have been brought into existence where humans have deposited or exposed soil parent material by earth-moving operations. Some of these operations include highway construction, surface mining, landfill construction, or dredging.

In many soils with human-deposited or human-exposed parent materials, fragments of diagnostic horizons are not identifiable; thus, a classification of Arents would not be appropriate. Such soils commonly have been classified in the suborder of Orthents and subdivided into great groups according to the soil moisture regime (i.e., into Udorthents in the eastern United States; Soil Survey Staff, 1999, 2010). This classification is insufficient for many soil scientists because it does not reflect the role of humans in the genesis of the soils (Short et al., 1986a; Strain and Evans, 1994). Although these soils are commonly thought to be highly variable, studies have shown degrees of variability similar to that of some soils in naturally deposited parent materials (Short et al., 1986b).

Many soils forming in materials deposited by bulldozers or other earth-moving equipment have an irregular distribution of organic matter with depth. In theory, these soils would classify as Fluvents at the suborder taxonomic level (Soil Survey Staff, 1999, 2010). In contrast to Fluvents, however, human-disturbed soils are not stratified by natural processes, are not flooded in most years, and have a higher density. Despite the irregular organic matter depth distribution, these soils behave more like Orthents than Fluvents and have been forced, in some soil surveys (Smith, 1976), into Orthents by ignoring the irregular organic matter depth distribution.

Because of the inadequacies of the present classification system, new suborders and/or subgroups of Entisols have been proposed for soils in human-deposited or human-exposed parent materials (Fanning and Fanning, 1989; Chapters 24, 28, and Appendix 3). An international committee for anthropogenic soils (ICOMANTH) has shown considerable interest in adding new taxonomic classes. The committee, however, has not determined at which taxonomic level human-influenced soils should be recognized. For example, Kosse (1988) and Strain and Evans (1994) have proposed classifying these soils at the order level as Anthosols. Regardless, the Arent suborder discussed above established a precedence for recognition at the suborder level. Fanning and Fanning (1989) have supported this viewpoint.

West Virginia workers (Sencindiver, 1977; Smith and Sobek, 1978) proposed the suborder Spolents (for spoil) for soils formed in materials deposited from surface mining operations. During development of the soil survey of Washington, DC, Smith (1976) proposed a suborder of Urbents for soils formed in miscellaneous urban fill that had artifacts such as brick or concrete in the particle size control section (generally between a depth of 25 and 100 cm). The Garbents suborder was proposed for soils formed in garbage of an organic nature within a depth of 2 m from the soil surface, such that the soils would readily subside and generate methane under anaerobic conditions (Fanning and Fanning, 1989; Fanning, 1991). Fanning and Fanning (1989, Chapter 24) proposed special diagnostic soil characteristics (i.e., urbic, garbic, spolic, and dredged material; and the scalped land surface) for taxonomic classification. These characteristics were employed to propose special subgroup classes, such as Urbic Garbic Udorthents for "sanitary landfills where the cover material qualified as urbic materials" (Fanning and Fanning, 1989, Appendix 3). It was proposed that soils on scalped land surfaces (cuts from an engineering perspective) could be placed in Scalpic subgroups.

Soils created by dredging operations commonly qualify as Sulfaquents immediately after sulfidic dredged materials have been deposited (Fanning and Fanning, 1989). Under aerobic conditions at dredged material sites, sulfuric horizons commonly form from sulfidic materials in a few weeks or months. This causes the soils to become Sulfaquepts, a great group of Inceptisols. Thus, the period that such soils remain as Entisols, before becoming a soil of a different order, may be very brief.

With the present strong interest in improving knowledge about the classification of human-influenced soils, much progress may be expected in the near future. This is needed as soil surveys of lands with human-influenced soils are becoming more widespread. In ongoing urban surveys, many of the problems associated with the soils discussed in this section will be encountered. For example, sulfide-bearing clays have been used as landfill cover materials, which can lead to the formation of acid sulfate soils on landfills (Kargbo et al., 1993; Fanning and Burch, 1997). A more resolved classification system can enhance future interpretations and mapping techniques of soil surveys in these areas. Further, it has been suggested (Fanning, 1990) that a discipline of pedotechnology is needed to improve the design of soils brought into existence through the action of humans.

33.4.4.3 Psamments

Psamments are Entisols with sandy parent materials. They typically show little or no evidence of soil development with horizon sequences of AC or A/AC-C. An exception, however, are areas (e.g., central Florida ridge) where Psamments represent intensely weathered albic horizons with argillic and spodic horizons forming at depths greater than 2 m below the surface. Consequently, Psamments generally have an ochric epipedon, but lack any diagnostic subsurface horizons within the upper 2 m depth.

Psamments cover 3.4% of the Earth's land. Major desert regions, such as in northern Africa (Sahara), Saudi Arabia, western Australia, and southern Mongolia (Gobi), have Psamments. In the United States, these soils exist in all states. They are extensive in the Central Ridge of Florida, in the Sandhills in western Nebraska, in the sandsheet in south Texas, in central Wisconsin, and throughout the arid and semiarid areas of the western United States on the leeward sides of pluvial lakes.

Generally, Psamments have low amounts of plant nutrients, low water-holding capacity, and rapid permeability. Irrigation is normally necessary to maintain economical crop production. With management, however, they can be used for citrus production, for growing sand pines, and even for real estate development. In arid and semiarid regions, Psamments are ecologically fragile, and disturbance to rangeland vegetation growing on them can initiate feedback processes leading to severe desertification (Monger and Bestelmeyer, 2006).

Natural vegetation associated with Psamments in Florida consists of bluejack oak (*Quercus incana*), blackjack oak (*Quercus marilandica*), turkey oak (*Quercus laevis*), longleaf threeawn (*Aristida stricta*), and bluestem (*Andropogon* sp.). In the major cattle-growing states of Texas and Nebraska, Psamments are used for spring and summer grazing of native grasses. In central Wisconsin, truck crops are an important use.

33.4.4.3.1 Lower Taxonomic Levels

Psamments are restricted to soils that have a texture of loamy fine sand or coarser in the control section (family particle size class) and <35% by volume of particles >2 mm diameter (Soil Survey Staff, 1999, 2010). The latter distinguishes Psamments from sandy Orthents. Sandy loam lamellae are permitted in Psamments, as long as cumulatively they do not meet the requirements of an argillic horizon.

Psamments occur in a wide range of climates. Five of the six great groups reflect a climatic/geographic bias. Only Quartzipsamments do not indicate a geographic location, but rather the dominant mineralogy, quartz. Psamments exist in cryic or pergelic (Cryopsamments) soil temperature regimes but also are located in aridic (Torripsamments), ustic (Ustipsamments), xeric (Xeropsamments), and udic (Udipsamments) soil moisture regimes.

Subsurface features present in some Psamments include a lithic contact (Lithic), aquic or oxyaquic conditions, lamellae (Argic), plinthite, or durinodes. Thus, subsoil features vary greatly, and the variation is noted in the subgroup taxonomic name. Examples include Lithic Cryopsamments, Plinthic Quartzipsamments, and Aquic Durinodic Xeropsamments.

At the subgroup taxonomic level, climatic/geographic conditions are also described. As an example, Aridic Ustipsamments are located in several soil temperature regimes, but must have a period of time in which the soil is dry in all parts. The length of time that the soil must be dry is defined by the soil temperature regime.

33.4.4.3.2 Genesis

Genetically, Psamments develop in areas high in sand content (>85%). The parent material can be geologically young, as in eolian or alluvial materials or old, as in sandy residual bedrock. Sand dunes of either marine (Florida) or eolian (Saudia Arabia or Nebraska) origin are common parent materials. Some Psamments form in materials weathered from sandstone bedrock (e.g., South Dakota or Montana). Depending on the mineralogy of the sand, Psamments (Quartzipsamments) can be very resistant to weathering.

33.4.4.3.3 Representative Psamments

Psamments from Florida (Astatula series: hyperthermic, uncoated, Typic Quartzipsamments), Nebraska (Valentine series: mixed, mesic Typic Ustipsamments), and Saudi Arabia (Torripsamments) were selected to represent different geographic areas and, therefore, different genetic pathways. But morphologically, these soils are very similar and they are excessively drained.

Astatula, Valentine, and the Torripsamments from Saudi Arabia have thick C horizons with thin A horizons. If an A horizon is present, it is light colored and <15 cm thick. The A horizon is subject to continual mixing by wind action. Astatula soils form in eolian and marine sands and are extensive in peninsular Florida on upland slopes that range from 0% to 30%. Valentine soils form in sand dunes in the Sandhills of Nebraska on slopes that range from 0% to 60%. The sand dunes generally are oriented in a northwest to southeast direction (prevailing wind direction). In addition to Nebraska, Valentine soils occur extensively in Kansas, Montana, New Mexico, South Dakota, Texas, and Wyoming (>2 million ha). The Torripsamments from Saudi Arabia form in sand dunes and also in alluvial sandy deposits on plains and stream terraces. The dunes range in height from <2 to >10 m.

33.4.4.3.4 Physical and Chemical Properties of Representative Psamments

Physical and chemical properties are presented for the Astatula in Florida and for the Valentine in Nebraska (Table 33.20). Astatula soils (Quartzipsamments) have low organic C content and are strongly acid (pH < 5.5) and strongly leached as a consequence of high rainfall (1270 mm year^{-1}). Torripsamments from Saudi Arabia with <100 mm year^{-1} precipitation (data not shown have an accumulation of bases and are alkaline in reaction (pH > 8.5)). The sand size fraction of the Astatula and Valentine series are dominated by fine sand. Clay content of these Psamments is <5%.

33.4.4.4 Fluvents

Fluvents are Entisols that form in alluvium of rapidly aggrading floodplains, fans, deltas, and in some cases mudflows. World coverage of Fluvents is 2.2%, principally within major river valleys (see Lindbo, 1997).

When rates of deposition in alluvial settings approximate rates of pedogenesis, A-C soil profiles develop. In many instances, the A horizon is an ochric epipedon, but in others the surface horizon may be bedded (Soil Survey Staff, 1999, 2010). Alluvial processes prevent the formation of subsurface diagnostic horizons because of variations in flood magnitude and frequency leading to fluctuations in organic carbon content with depth. This occurs because greater quantities of allogenic organic matter tends to be associated with clayey rather than sandy flood deposits (Grossman, 1983). To taxonomically capture these conditions, organic carbon content in Fluvents must decrease irregularly with depth, or be above 0.2% at a depth of 125 cm.

Fluvents are differentiated from other Entisol suborders by (1) having textures finer than loamy very fine sand and less than 35% rock fragments (Psamments); (2) occurring on slopes less than 25% and without a densic, lithic, or paralithic contact within 25 cm (Orthents); (3) not having aquic conditions (Aquents); (4) having less than 3% fragments of diagnostic horizons in no discernable order (Arents); and (5) having any soil temperature regime that excludes the presence of gelic materials (Soil Survey Staff, 1999, 2010).

TABLE 33.20 Soil Characterization Data for the Astatula Fine Sand (Sodek et al., 1990) and Valentine Fine Sand (Soil Survey Staff, 1966) of the Psamment Suborder

Horizon	Depth (cm)	Very Coarse Sand	Coarse Sand	Medium Sand	Fine Sand	Very Fine Sand	Silt	Clay	pH[a] (H_2O)	OC[b] (%)	Ca[c]	Mg	Na	K	CEC_7	Base Saturation (%)
		(%)									($cmol_c\ kg^{-1}$)					
Astatula Fine Sand (Typic Quartzipsamments)																
A	0–13	—	0.1	18.8	76.7	2.1	0.9	1.4	5.5	0.5	0.5	0.1	0.0	0.0	2.7	23
C1	13–64	—	0.1	18.4	76.5	2.4	1.6	1.5	5.2	0.2	0.0	0.0	0.0	—	1.5	5
C2	64–175	—	0.1	19.7	75.8	2.2	0.6	1.6	5.0	0.1	0.0	0.0	0.0	—	0.7	5
C3	175–203	—	0.1	19.4	76.2	2.0	0.6	1.7	5.0	0.0	0.0	0.0	0.0	—	0.9	6
Valentine Fine Sand (Typic Ustipsamments)																
A	0–10	0.2	4.2	12.7	50.2	24.7	4.1	3.9	6.0	0.1	3.6	0.6	0.1	0.3	5.9	78
AC	10–25	0.1	4.6	13.3	55.1	21.6	2.2	3.1	6.2	0.5	2.1	0.4	0.1	0.3	3.7	78
C1	25–45	—	3.6	11.2	56.2	24.0	1.8	3.2	6.5	0.2	2.0	0.6	0.1	0.1	0.1	88
C2	45–81	—	4.4	18.6	58.9	14.7	1.0	2.4	6.7	0.1	1.5	0.2	0.1	0.2	2.4	79
C3	81–137	0.1	3.8	10.9	59.1	22.1	1.3	2.9	6.8	0.0	1.8	0.8	0.1	0.1	2.8	100

Source: Modified from Eswaran, H., P.F. Reich, J.M. Kimble, F.H. Beinroth, E. Padmanabhan, and P. Moncharoen. 2000. Global carbon stocks, p. 15–25. *In* R. Lal, J. Kimble, A. Mtimet, H. Eswaran, and H. Scharpenseel (eds.) Global climate change and pedogenic carbonates. Lewis Publishers, Boca Raton, FL.

[a] Determined in a 1:1 H_2O solution for the astatula fine sand and in a 1:5 Cl solution for the valentine fine sand.

[b] OC is organic carbon.

[c] Ammonium acetate extractable Ca, Mg, Na, K, and CEC (cation exchange capacity), buffered at pH 7.

33.4.4.4.1 Other Taxonomic Levels

Great groups of the Fluvents are differentiated by either soil temperature or soil moisture regime (Soil Survey Staff, 1999, 2010). Fluvents occur in udic (Udifluvents), ustic (Ustifluvents), xeric (Xerofluvents), and aridic (Torrifluvents) moisture regimes (see Sidhu et al., 1994). They also occur in cryic temperature regimes (Cryofluvents) and in colder areas without permafrost (Gelifluvents). Fluvents form in all other soil temperature regimes (frigid, mesic, thermic, and hyperthermic), including those with minimal annual temperature change (iso).

At the subgroup taxonomic level, typical Entisols intergrade to the Andisol (andic), Mollisol (mollic), Aridisol (aridic), and Vertisol (vertic) orders, and to other soil moisture and temperature regimes not used at the great group categorical level. For additional information on the family taxonomic classification of Entisols, see the Soil Survey Staff (1999, 2010).

33.4.4.4.2 Climate, Vegetation, and Topography

Fluvents are soils in disequilibrium with the regional influence of climate, vegetation, and in some cases topography. Thus, Fluvents can occur in most climates associated with most vegetation communities and on a variety of alluvial slopes. Climate can, however, indirectly influence the chemical and physical properties of Fluvents. Alluvium transported into floodplains, fans, deltas, or mudflows is derived from erosion of soils within the surrounding drainage basin where properties are more likely to reflect regional climate conditions. These preconditioned sediments can expedite the transformation of Entisols to other soil orders.

Flood frequency and magnitude in many areas appear to be increasing because of urbanization and other changes in land use practices. Many workers have observed this hydrological phenomenon where Inceptisols, Mollisols, and Vertisols of inactive or slowly aggrading floodplains are being buried by Historic sediments coined "postsettlement alluvium" (Daniels and Jordan, 1966; Scully and Arnold, 1981; Grossman, 1983; Knox, 1987; Phillips, 1997; Gomez et al., 2004; Carson, 2006; Lecce, 2007; Rustomji, 2007). If the thickness of postsettlement alluvium, or alluvium deposited by any process, buries a soil by more than 50 cm, these soils are classified as Fluvents because of minimal development. In most cases, if a recent flood deposit is less than 50 cm thick, it is treated as a mapping phase (Soil Survey Staff, 1999, 2010).

33.4.4.4.3 Time

Ripening processes lead to the formation of Fluvents within a few hundred years after drainage from the construction of polders in the Netherlands (Pons and van der Molen, 1973). This is consistent with a compilation of data from quasi-stable floodplains globally (Buol et al., 1997) showing that ochric epipedons commonly form within several hundred years.

Under conditions of landscape stability, Entisols may quickly evolve into other soil orders. In subhumid climates of central Texas and Oregon, for example, Entisols transform into Mollisols and Inceptisols, respectively, within 600 years (Parsons et al., 1970; Nordt and Hallmark, 1990). Further, Entisols transform into Inceptisols and Vertisols within 400 years in slowly aggrading floodplain sediments of the Brazos River in central Texas (Waters and Nordt, 1995). In some humid regions, however, mollic epipedons and Mollisols can develop in as little as 100 years (Ruhe et al., 1975) whereas in other humid regions it takes Entisols 500–600 years to develop into Inceptisols (Bilzi and Ciolkosz, 1977; Scully and Arnold, 1981). Further, Fluvents transform into Inceptisols on levee and point bar facies of the lower Mississippi River valley in less than 600–800 years (Aslan and Autin, 1998). In a subhumid climate of Australia, Walker and Coventry (1976) reported that floodplains that flooded once each year maintained soil development in the Entisol stage, whereas when flood frequencies were on the order of 1–10 years, Mollisols began forming. Fluvents in arid regions can persist for a thousand years in low carbonate parent material and for up to several thousand years in carbonatic parent material (Gile, 1975). Beyond these temporal thresholds, subsurface diagnostic horizons and Aridisols begin to form.

33.4.4.4.4 Parent Material

The distribution of alluvial facies in a floodplain, fan, or deltaic system is dependent on stream type (Chorely et al., 1985), which strongly influences the distribution of pedofacies (see Autin and Aslan, 2001, for example). Sedimentological packages of mixed load, meandering streams in humid regions typically display a fining upward sequence consisting of lower channel, intermediate point bar, and upper floodbasin or levee facies. In some cases, Fluvents form entirely within one of these facies and do not exhibit a noticeable textural change within the solum. If a Fluvent develops in a gradational point bar to floodbasin sequence, for example, a fining upward textural profile will be inherited. In other situations, reverse grading may be encountered. For example, a levee may be deposited over a floodbasin facies, producing a Fluvent with a coarsening upward textural distribution. In addition, suborders of Entisols other than Fluvents may form in floodplains. Psamments sometimes form in sandy channel and point bar deposits and Aquents in poorly drained topographic positions. In contrast to meandering streams, bedload streams or streams near mountain fronts will have a preponderance of Psamments and coastal streams a preponderance of Aquents.

Figure 33.25 is a pedostratigraphic column illustrating some of these fluvial geomorphic principles. The stratigraphic column displays two fining upward point bar to overbank sequences. The first terminates in a buried Fluvent with relict rooting patterns. The Fluvent formed in the upper sedimentological package is a mappable Entisol in the modern floodplain. However, it is buried by postsettlement alluvium. As depicted, the modern buried Fluvent is taxonomically classified, with the postsettlement alluvium characterized as a mapping phase because its thickness is <50 cm (see also Figure 33.24).

33.4.4.4.5 Physical and Chemical Properties

Many alluvial soils are inherently fertile and support some of the most important agricultural centers of the world (Edelman and Van Der Voorde, 1963; Radwanski, 1968; Sidhu et al., 1977). In fact, one-half of Entisols in alluvial settings support up to 1/3 of

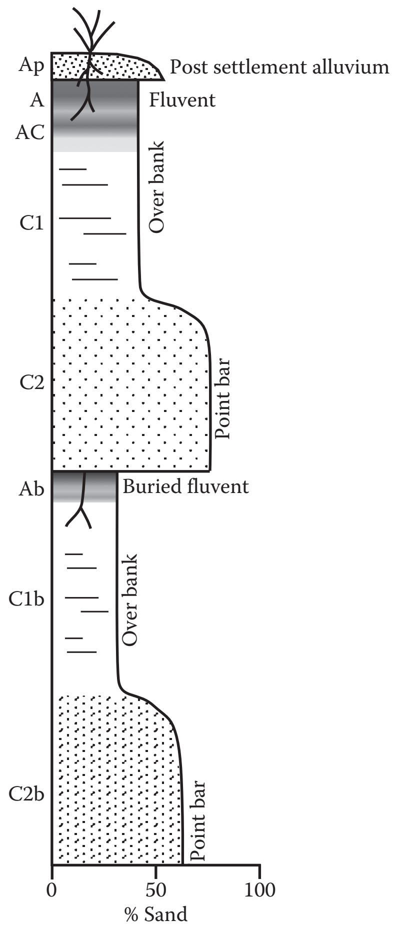

FIGURE 33.25 Pedostratigraphic diagram demonstrating fluvial geomorphic principles in relation to a buried and modern Fluvents in a floodplain setting covered by postsettlement alluvium.

the human population for food production (Buol et al., 1997). In the tropics, however, upland soils tend to be depleted in nutrients such that their alluvial counterparts produce low-fertility floodplain soils (Edelman and Van Der Voorde, 1963). This presents even greater food challenges to agricultural production in tropical regions.

As further examples, a floodplain Udifluvent and Torrifluvent display A-C profiles with textural and organic carbon stratification with depth (Table 33.21). There are similarities in clay, silt, and sand content between the two profiles, but the organic carbon content, largely detrital, is greater in the Udifluvent. Consequently, CEC_7 is slightly greater in the Udifluvent than the Torrifluvent. Perhaps the greatest difference in the two profiles is that the Udifluvent is carbonate free with the Torrifluvent having 11%–12% $CaCO_3$ equivalence. This results in a neutral pH and base saturation of ~60%–80% in the udic floodplain soil and a pH of between 8.1 and 8.5 and base saturation of 100% in the torric floodplain soil. The neutral pH of the Udifluvent would lead to greater plant availability of a number of nutrients such as iron and phosphorous (Brady and Weil, 2002). These two examples exemplify the variety of properties possible in Fluvents, largely inherited from the parent material. Soils surrounding the Udifluvent from Missouri are Alfisols and Inceptisols (Larsen, 2002) with Aridisols and other carbonate-rich soils dominating the drainage basin of the Torrifluvent in southern New Mexico (Gile et al., 1981).

33.4.4.5 Orthents

Orthents are Entisols that do not fit the other suborder categories (e.g., they are the “other Entisols”). They are not water saturated (Aquents), the result of deep mixing or other disturbances by humans (Arents), dominated by sandy textures (Psamments), or stratified recent water-deposited sediments (Fluvents). Instead, most soils that key into this category are on recent erosional surfaces where all former diagnostic horizons have been truncated (Figure 33.26). Other Orthents have formed in recent landslides, mudflows, glacial deposits, piedmont slope deposits, and young terraces, or are in areas of recent solifluction, volcanic deposits, thin regolith over hard rocks, or weakly cemented rock, such as shale (Soil Survey Staff, 1999, 2010).

Orthents occupy approximately 10.5% of the Earth’s land surface. They are more extensive than the combined area of the other Entisol suborders, which together occupy an area of about 5.7%. Orthents also have the most taxonomic subdivisions, reflecting lithologic properties inherited from exposed regolith, as well as properties related to climate, depth to bedrock, wetness, shrink-swell, and bioturbation (Soil Survey Staff, 1999, 2010). They occur in any climate and under any vegetation. Orthents in arid and semiarid climates are commonly used for rangeland, wildlife habitat, and irrigated agriculture. In humid regions, Orthents are used for pasture, hay production, cropland, and as forests.

33.4.4.5.1 Other Taxonomic Levels

Orthents are subdivided at the great group level based on the occurrence of udic, ustic, xeric, or aridic moisture regimes, or the cryic temperature regime (Soil Survey Staff, 1999, 2010). At the subgroup level, Orthents are divided according to many features that convey extragrade properties of the soil or to denote transitional intergrades to other taxonomic categories. Extragrades include lithic, lamellic, and vermic properties. Intergrades include durinodic (intergrade to Durids), oxyaquic (intergrade to Aquic subgroups), and aquic (intergrade to Aquents).

33.4.4.5.2 Occurrence and Genesis

Orthents are common in mountainous regions with arid and semiarid climates. An example of these Orthents is illustrated in Figure 33.27 as a block diagram of stepped geomorphic surfaces bordering a major floodplain mapped as Fluvents. Mountains and hills contain Orthents and areas of rock outcrop as the result of rapid erosion and truncation of former diagnostic horizons that were subsequently deposited downslope as colluvium. Widespread erosion of late Pleistocene mountain soils began in the middle Holocene around 7000 years ago in the southwestern United States (Gile et al., 1981). Such sediment makes up the mid-to-late Holocene stepped surface. The elevation of this surface was controlled by the former and slightly higher base level of the river floodplain. A recent lowering of the base level has caused the modern arroyo network to downcut into the Orthents of the Holocene deposits.

TABLE 33.21 Characterization Data of a Udifluvent (Phelps County, Missouri) and Torrifluvent (Dona Ana County, New Mexico) (Soil Survey Staff, 2008) of the Fluvent Suborder

Horizon	Depth (cm)	Sand	Silt	Clay	OC[a]	CEC_7[b] ($cmol_c\ kg^{-1}$)	$CaCO_3$ (%)	pH H_2O
		(%)						
Udifluvent								
A	0–13	89	8	3	0.39	2.9	0	6.9
C1	13–23	62	28	10	1.15	8.5	0	7.1
C2	23–43	51	36	13	1.05	9.1	0	7.3
C3	43–58	57	33	10	0.57	6.2	0	7.3
C4	58–81	69	24	7	0.48	4.5	0	7.4
C5	81–89	95	3	2	0.06	1.1	0	7.2
C6	89–117	97	1	2	0.02	0.7	0	7.0
C7	117–132	94	3	3	0.08	1.1	0	7.0
C8	132–152	98	0	2	0.01	0.8	0	7.1
Torrifluvent								
Ap1	0–10	69	20	11	0.67	8.2	11	8.1
Ap2	10–31	73	18	9	0.44	6.6	11	8.3
C1	31–51	81	12	7	0.28	5.2	11	8.4
C2	51–71	75	19	6	0.30	5.6	11	8.4
C3	71–107	83	13	4	0.17	4.2	11	8.4
C4	107–152	84	12	4	0.15	3.9	12	8.5

[a] OC is organic carbon.
[b] Ammonium acetate extractable Ca, Mg, Na, K, and CEC (cation exchange capacity) buffered at pH 7.

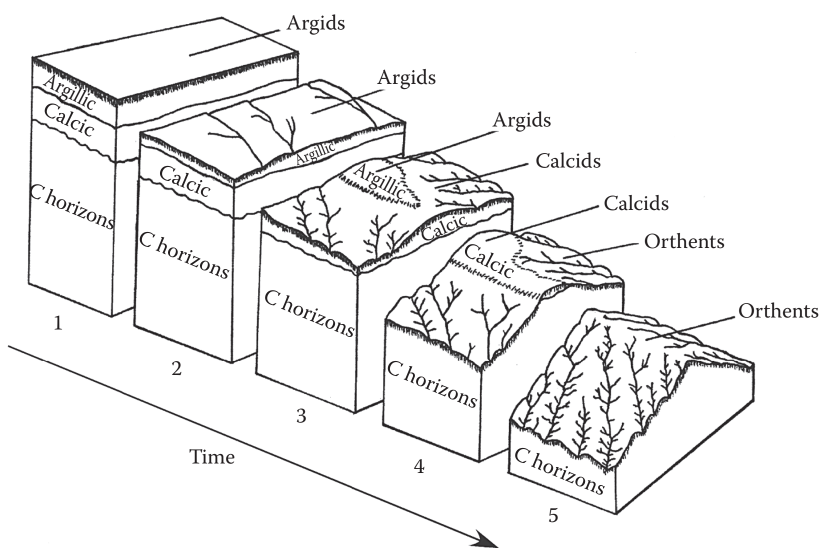

FIGURE 33.26 Formation of Orthents by the truncation of diagnostic horizons. Block diagram illustrates an Aridisol with argillic and calcic horizons that are progressively eroded, thus converting the suborders from Argids to Calcids to Orthents. (Modified from Gile, L.H., J.W. Hawley, and R.B. Grossman. 1981. Soil and geomorphology in the basin and range area of southern New Mexico—Guidebook to the desert project. New Mexico Bureau of Mines and Mineral Resources Memoir 39. New Mexico Bureau of Mines and Mineral Resources, Socorro, NM.)

Another location for Orthents in this landscape is the backslope of scarps (risers) associated with the stepped geomorphic surfaces (Figure 33.27). These erosional surfaces are truncating diagnostic horizons of an Aridisol and exhuming underlying sediments beneath the late Pleistocene stepped surface (tread). The rate of erosion on this geomorphic surface is greater than the rate of pedogenesis, and, thus, these soils have no diagnostic horizons other than the ochric epipedon. Sediments generated by erosion on the backslope are deposited on the footslope as unstratified colluvium, which is an example of a geomorphic surface having linked erosional and constructional components. Orthents associated with the erosional component are the same age as Orthents associated with the constructional component because pedogenesis would have started at the same time for both.

Once landscapes have stabilized the type of parent material can affect the rate at which diagnostic horizons form, and,

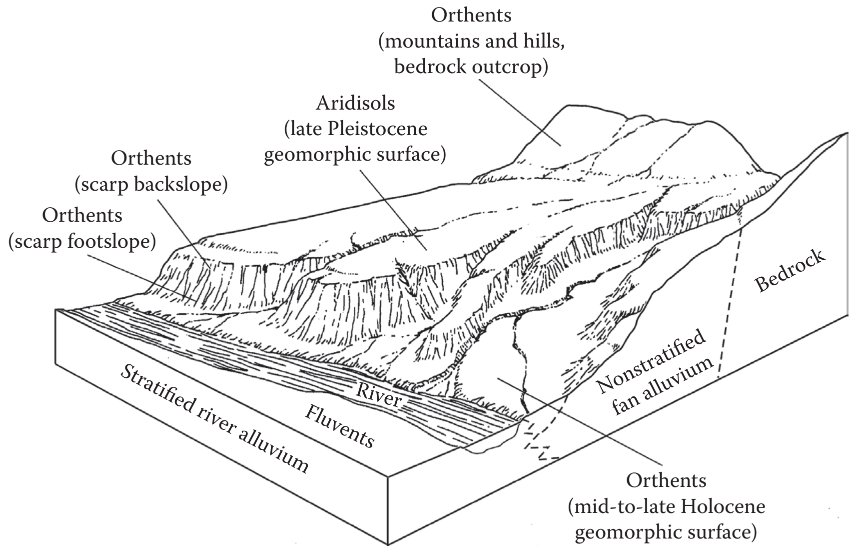

FIGURE 33.27 Block diagram of stepped geomorphic surfaces bordering a major floodplain illustrating the erosional and depositional occurrence of Orthents. (Modified from Gile, L.H., and R.B. Grossman. 1979. The desert project soil monograph. Doc. PB80-135304. National Technical Information Service, Springfield, VA.)

thus, the length of time a soil remains an Orthent. First, parent materials that contain large amounts of detrital carbonate can curtail the formation of argillic horizons by keeping clay flocculated and less mobile (Gile et al., 1981). Second, parent materials containing plutonic rocks with weatherable minerals and cleavage, such as granites, disintegrate along mineral contacts and produce argillic horizons at a rate faster than parent materials containing volcanic rock, such as rhyolite. This rock type, for example, resists weathering because they lack planes of weakness along which water can enter.

33.4.4.5.3 Physical and Chemical Properties

In contrast to soils classified according to their pedogenically produced diagnostic horizons, the physical and chemical properties of Orthents are mainly inherited from their parent materials. As a result, the range of Orthent properties is wide and reflects the properties of the unconsolidated material in which they occur, combined with bioclimatic factors. Coarse-silty Udorthents in Iowa, for example, have A horizons with organic carbon contents as high as 1.2% and cation exchange capacities of 13.6 $cmol_c\ kg^{-1}$ (Soil Survey Staff, 1975). In contrast, sandy-skeletal Torriorthents in New Mexico have A horizons with organic carbon contents as low as 0.4% and cation exchange capacities of 7.5 $cmol_c\ kg^{-1}$ (Gile and Grossman, 1979).

33.4.5 Carbon Stocks

Entisols store approximately 353 Pg (10^{15} g) of total soil carbon (TSC), which is 14.2% of all soil carbon globally (Table 33.22). Of this, 90 Pg (5.8%) of the global soil carbon pool is contained in the organic fraction and 263 Pg (27.7%) in the inorganic fraction. Of the TSC within the Entisol order, 75% is held in the soil inorganic carbon (SIC) pool and 25% in the soil organic carbon (SOC) pool (see also Wilding et al., 2002). This distribution reflects the presence of calcium carbonate in both pedogenic forms and as detritus in the fine-earth fraction inherited from the parent material (Rabenhorst et al., 1984; Nordt et al., 1998; Kraimer and Monger, 2009). Further, there is more SIC stored in Entisols than any other soil order except Aridisols (Eswaran et al., 2000). In contrast, Entisols store a smaller proportion of the SOC carbon pool globally than most other soil orders.

Orthents store the most TSC (10% globally) of the Entisol suborders (Table 33.22). The SIC pool is high because of the widespread distribution of Orthents in arid lands. After Orthents, Fluvents store more TSC than Psamments and Aquents. The low accumulation of SOC in Aquents is misleading in that their areal extent is limited, even though great groups and subgroups may contain histic and mollic intergrades.

Sequestration rates of atmospheric CO_2 in Entisols are low because of geomorphic processes that limit pedogenesis and the

TABLE 33.22 Total Soil Carbon, Soil Organic Carbon, and Soil Inorganic Carbon Stocks by Entisol Suborder in Petagrams (Pg) and Percent

	TSC		SOC		SIC	
Suborder	Pg	%	Pg	%	Pg	%
Aquents	2	0.1	2	0.2	0	0.0
Psamments	41	1.6	8	0.5	33	3.5
Fluvents	60	2.4	33	2.1	27	2.8
Orthents	250	10.0	47	3.0	203	21.4
Total	353	14.2	90	5.8	263	27.7

Source: Modified from Eswaran, H., P.F. Reich, J.M. Kimble, F.H. Beinroth, E. Padmanabhan, and P. Moncharoen. 2000. Global carbon stocks, p. 15–25. *In* R. Lal, J. Kimble, A. Mtimet, H. Eswaran, and H. Scharpenseel (eds.) Global climate change and pedogenic carbonates. Lewis Publishers, Boca Raton, FL.

Arents undetermined.

incorporation of carbon. However, if processes that limit pedogenesis are removed, more organic carbon could be stored in surface horizons during the process of melanization. As with any parent material rich in calcium carbonate, the formation of pedogenic carbonate-carbon would not result in a net accumulation of atmospheric CO_2 (see Nordt et al., 2000). However, Entisols forming in parent materials with weatherable calcium silicate minerals could in the process of forming pedogenic carbonate sequester atmospheric CO_2 at a greater rate than currently.

33.4.6 Conclusions

Entisols are the most widespread soil order in the world forming in a variety of regions reflecting parent material characteristics and geomorphic processes that limit development to A-C or A-R profiles. They typically have ochric epipedons and no subsurface diagnostic horizons. Entisols provide engineering and agricultural challenges in areas that are wet (Aquents or Wassents), sandy (Psamments), flooded (Fluvents), steep or rocky (Orthents), or disturbed by humans (Arents). In contrast, Entisols, especially Torrifluvents, have played an important role in the origins of agricultural centers of early civilizations, such as those along the Nile, Euphrates, Tigris, Indus, and Huang He rivers.

With quasi-landscape stability in parent materials containing weatherable minerals on relatively shallow slopes, Entisols quickly form into one of the other taxonomically defined soil orders accompanied by a variety of physical and chemical changes that includes increased carbon sequestration and storage. Further, Entisols hold more TSC than any other soil order except Aridisols. Fossil Entisols are common in the geologic record as Paleosols and provide information about paleoenvironmental conditions. Entisols capture the essence of challenging pedological problems for the future in response to an ever increasing population and other global change issues related to land use.

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33.5 Inceptisols

Wayne H. Hudnall

33.5.1 Introduction

Inceptisols serve two important functions in *Soil Taxonomy*. First, they group soils with incipient soil development (L. *Inceptum*, beginning), and second, they ensure that the classification system is fully bifurcated, that is, all soils fall within some taxa of the classification system. Because of the latter, Smith (1986) referred to Inceptisols as the wastebasket order. They are the repository for all soils that do not meet class differentiae of other orders. While the central concept of Inceptisols is that of soils in cool, very warm humid and subhumid regions that have weekly developed subsoil horizon (cambic) and a light colored surface horizon (Ochric), the order contains a wide variety of soils (Soil Survey Staff, 1998). In some areas, they contain soils with minimal development, while in other areas they represent soils with well-expressed diagnostic horizons that fail the criteria of other orders. All soils that have a plaggen epipedon are Inceptisols (Soil Survey Staff, 1999).

Inceptisols, because of their broad and inclusive nature, served as the precursor for new soil orders (Andisols and Gelisols) and new taxa in other orders. For example, shrink/swell soils with aquic conditions and those with cryic soil temperature regimes previously placed within Inceptisols are now Vertisols (Aquerts and Cryerts). Inceptisols with isomesic or warmer *iso* temperature regimes, previously called Tropepts, are now distributed among other great groups and families of Inceptisols with *iso* temperature regimes. Inceptisols as markers of past cultural habitats and anthropogenic activities (Plaggepts) are now termed Anthrepts, and Ochrepts, a previous suborder of Inceptisols, have now been incorporated into Inceptisols suborders reflecting soil moisture bias. Subaqueous soils formed in freshwater will be classified as Wassepts. Table 33.23 illustrates the dynamics of these changes and how suborders of Inceptisols have evolved with increased knowledge over the past 40 years.

The central concept of Inceptisols includes soils that have undergone modifications of the parent material by structural development and alteration sufficient to differentiate them from Entisols. Pedogenesis may include eluviations (losses of constituents); the translocation of clay, iron, silica, aluminum, carbonates, bases, and organic matter; and the formation of redoximorphic features by aquic (hydromorphic) conditions. Inceptisols occur in all known climates, except under aridic and pergelic conditions, and have many kinds of diagnostic horizons and epipedons. Hence, while the central concept expresses soils with incipient development, Inceptisols are an order with inordinate diversity. No attempt will be made herein to consider their full diversity. Rather, examples of the great range in physical, chemical, biological, and mineralogical attributes of this order pertinent to land use and behavior will be illustrated. For a more extensive coverage of these attributes, the reader is referred to online resources (http://ssldata.nrcs.usda.gov/).

TABLE 33.23 Inceptisols Suborders Reflecting the Evolution of Soil Taxonomy

7th Approximation (1960)[a]	Soil Taxonomy (1975)[b]	Soil Taxonomy (1998)[c]	Keys to Soil Taxonomy (2010)[d]
Andepts	Andepts	Aquepts	Anthrepts
Aquepts	Aquepts	Ochrepts	Aquepts
Ochrepts	Ochrepts	Plaggepts	Cryepts Gelepts
Umbrepts	Plaggepts	Tropepts	Udepts
	Tropepts	Umbrepts	Ustepts
	Umbrepts		Xerepts

Source: Soil Survey Staff. 2010. Keys to Soil Taxonomy. 11th edn. U.S. Government Printing Office, Washington, DC.
[a] Soil Survey Staff (1960).
[b] Soil Survey Staff (1975).
[c] Soil Survey Staff (1998). Soil
[d] Soil Survey Staff (2010).

33.5.2 Distribution

The worldwide distribution of Inceptisols is illustrated in Figure 33.28. Because these soils lack a sharply focused central concept, they occur indiscriminately globally. Where Inceptisols reflect youthfulness, they occur in high-gradient mountainous regions, along major river systems as deltaic/fluvial plains, and as soils developed from carbonate-rich bedrocks or sediments in positions of geomorphic instability. Where Inceptisols serve to bifurcate *Soil Taxonomy* there in no particular pattern to their occurrence, but they are commonly dispersed among more strongly developed Alfisols, Ultisols, and Oxisols in tropical and subtropical regions (Figure 33.28).

33.5.3 Formation of Inceptisols

Because soils placed in Inceptisols are varied and represent many different pedogenic processes and combinations of soil-forming factors, examples will be given that typify central concepts or exhibit unique characteristics.

33.5.3.1 Parent Material

Foss et al. (1983) state that Inceptisols develop on geologically young sediment or landscapes and/or under environmental conditions that inhibit soil development (e.g., coarse-textured siliceous deposits that are resistant to weathering or parent materials within cool climates that inhibit pedogenesis). Given the above constraints, parent materials include almost all types of igneous, metamorphic, and sedimentary materials (residuum, alluvium, colluvium, loess/eolian, glacial drift, etc.). Inceptisols are excluded from soils having a sandy texture throughout because it is believed that pedological features indicative of eluviations, weathering, translocation, and transformation (e.g., development of color Bw horizons) take a very short time to form in subsoils with such low specific surface area coarse-textured materials. The kinds and arrangement of horizons, and their chemical, physical, and mineralogical properties, are controlled to a major degree by the kinds of parent material given from which Inceptisols have developed. For example, fine-textured parent materials give rise to clayey Inceptisols, acidic parent materials result in base-poor (Dystr)

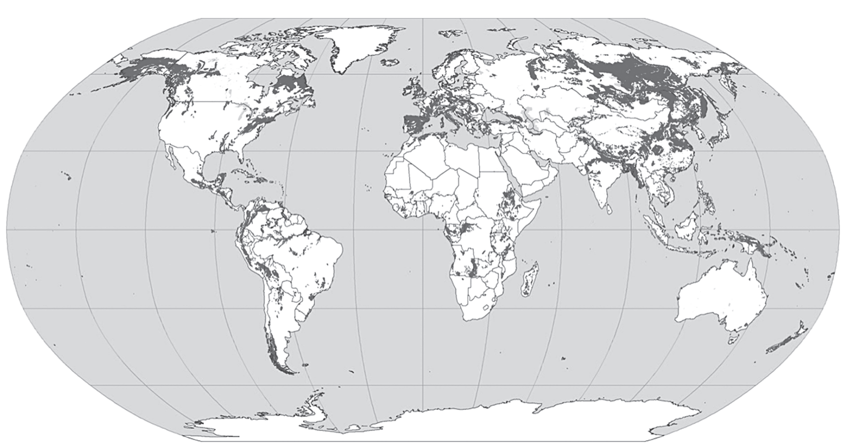

FIGURE 33.28 Global distribution of Inceptisols. (Courtesy of USDA-NRCS, Soil Survey Division, World Soil Resources, Washington, DC, 2010.)

Inceptisols, and calcareous parent materials form neutral or alkaline, base-rich (Eutr) Inceptisols. Depth, color, organic carbon content, and drainage frequently reflect the combination of geomorphic position, landscape instability, and nature of the parent materials.

Table 33.24 illustrates the diversity in selected physical, chemical, and mineralogical properties of soils developed from six different parent materials [e.g., Mississippi River alluvium (*Fluvaquentic Epiaquepts*); noncalcareous glacial outwash (*Aeric Fragiaquepts*); Permian red bed shales (*Vertic Calciustepts*); basic igneous rock (Oxic Haplustepts); colluviums weathered from sandstone, siltstone, and shale (Typic Dystrudepts); and Coastal Plains alluvium (Typic Sulfaquepts)]. Abridged attributes of these soils follow with detailed morphological, physical, chemical, and mineralogical data found online at Soil Survey Data Mart.

33.5.3.1.1 Fluvaquentic Epiaquepts (Conciene Series)

These soils developed from loamy alluvial sediments that have weathered into soils with moderately expressed cambic horizons. Their nearly level to depressional landforms favors restricted drainage, especially under humid climates. Under slightly better drained conditions, soils developed in this alluvium may form argillic horizons in a relatively short time (hundreds to several thousand years). This is because the eluviation–illuviation process proceeds relatively rapidly in such permeable, base-rich, medium textured parent sediments.

TABLE 33.24 Physical and Chemical Properties of Selected Pedons in the Inceptisols Order

Horizon	Depth (cm)	Clay	Sand	OC	Extr. Fe	CEC/Clay	WRD (cm cm^{-1})	CEC (cmol$_c$ kg^{-1})	BS	CaCO$_3$	pH$_{H_2O}$	COLE (cm cm^{-1})
		(%)							(%)			
Conciene Series (S88-LA-047—001) (Fine, Silty, Mixed, Superactive, Hyperthermic, Fluvaquentic Epiaquepts), LA												
Ap1	0–14	13.8	20.4	1.94	0.6	1.25	0.25	17.3	100	—	6.8	0.024
Bg2	51–78	15.9	17.6	0.30	0.4	0.91	0.22	14.5	100	1	7.7	0.009
Bssgb	180–215	50.4	4.3	0.71	1.0	0.69	0.19	34.9	100	1	7.5	0.116
Vernon Soil Series (S80-TX-253—001) (Fine, Mixed, Semiactive, Thermic, Vertic Calciustepts), TX												
A1	0–6	31.5	23.7	1.46	0.9	0.65	—	20.6	100	7	8.3	—
Bca2	45–65	40.5	13.8	0.21	1.7	0.34	0.08	14.8	100	23	8.5	0.042
Cr2	160–190	52.9	5.2	0.04	1.9	0.48	—	25.5	100	11	8.7	—
Kahana Series (S89-HI009—004) (Very Fine, Kaolinitic, Semiactive, Isohyperthermic Oxic Haplustepts), HI												
Ap	0–38	71.9	3.3	1.20	5.7	0.21	—	14.8	26	—	4.7	—
Bo2	56–87	59.8	7.1	0.45	5.9	0.23	—	14.0	71	—	6.2	—
Bo4	120–155	48.3	12.1	0.33	5.9	0.28	—	13.7	80	—	6.3	—
Volusia Series (S86-NY-025—001) (Fine-Loamy, Mixed, Mesic, Aeric Fragiaquepts), NY												
Ap	0–20	—	—	4.50	2.0	—	—	22.5	44	—	5.4	—
Bw	20–29	—	—	0.95	1.8	—	—	11.3	33	—	5.5	—
Eg	29–42	—	—	0.27	1.2	—	—	6.3	30	—	5.2	—
2Bx3	90–145	—	—	0.16	1.5	—	—	10.9	68	—	6.2	—
Fedscreek Series (S83-KY-195—018) (Coarse-Loamy, Mixed, Semiactive, Mesic, Typic Dystrudepts), KY												
A	0–9	13.5	38.0	2.17	1.2	0.75	0.28	10.1	37	—	5.0	1.3
Bw2	42–76	14.7	43.6	0.20	1.2	0.34	0.17	5.0	20	—	4.8	0.4
BC	122–152	17.4	36.8	0.43	1.5	0.41	0.13	7.2	31	—	4.9	0.5
Gimhae Series (S85-FN-515—012) (Fine-Silty, Mixed, Active, Mesic, Typic Sulfaquepts), Korea												
Ap1	0–9	33.2	4.6	1.97	1.5	0.34	0.19	11.3	80	—	5.1	0.2
Bg1	15–32	34.4	5.2	1.59	1.9	0.38	—	13.2	42	—	4.0	—
Bg3	66–125	19.4	29.2	1.64	0.9	0.58	0.28	11.3	100	—	3.6	1.4
Vidhas Soil (S94-FN-120—009) (Fine-Silty, Mixed, Superactive, Thermic, Fluventic Haploxerepts), Albania												
Ap	0–25	21.4	12.4	0.98	0.9	0.75	0.17	16.1	100	20	8.1	0.6
Bw1	49–83	31.1	0.9	0.59	0.9	0.66	0.18	20.5	100	20	8.3	1.6
Bw2	83–116	59.8	1.9	0.69	1.3	0.52	0.09	31.2	100	15	8.2	1.4
Moran Series (S85-CO-051—005) (Loamy-Skeletal, Mixed, Superactive, Humic Dystrocryepts), CO												
Oi	8–0	—	—	16.6	—	—	—	—	—	—	—	—
A1	0–19	17.5	46.1	5.07	2.3	1.33	—	23.3	33	—	4.8	—
A2	19–64	13.1	54.0	3.36	2.3	1.40	—	18.4	24	—	5.2	—
Bw	64–98	14.5	50.7	3.05	2.3	1.32	—	19.2	22	—	5.2	—

Source: Soil Survey Staff. 2011. National Cooperative Soil Characterization Database. Available online at http://ssldata.nrcs.usda.gov Accessed 7/8/2011.
OC, organic C; WRD, water retention difference; BS, base saturation; COLE, coefficient of linear expandability.

33.5.3.1.2 Vertic Calciustepts (Vernon Series)

The calcareous clays and shales of the Permian red beds restrict water percolation and profile development. Well-drained soils that develop in these sedimentary materials have a calcic or salic horizon whereby lithogenic carbonates and soluble salts have been translocated to subsoils and precipitated as pedogenic products at the terminus of the wetting front. The depth of carbonate and salt translocation, degree of calcic or salic horizon expression, and thickness of the cambic horizons are conditioned by the landform position, texture, and permeability of the parent materials.

33.5.3.1.3 Oxic Haplustepts (Kahana Series)

These well-drained, clayey soils have developed in basic igneous rocks in Hawaii. Under a warm, tropical environment, the weathering of nonresistant mafic-rich minerals and fine grained to amorphous materials has been rapid. An oxic-like (sesquioxide rich, subactive clay) cambic subsoil has developed. On a stable landform position, under an ustic moisture regime, subsequent weathering over time will result in an Oxisol with low activity clay.

33.5.3.1.4 Aeric Fragiaquepts (Volusia Series)

These soils have developed from weakly or noncalcareous glacial outwash of stratified sand, silt, and gravel. Upon weathering, these poorly sorted materials form loamy pans that have a brittle or fragic rupture upon deformation (fragipans). Weathering discontinuities in these materials enhanced by textural stratification are believed to result in a weak cementation between closely packed skeleton grains (Smeck et al., 1989). The bonding agents between the grains are believed to be hydrous oxides of Al and Si and/or translocated clay coatings or bridges. Restricted drainage would complement transport discontinuities in these soils to form weathering discontinuities.

33.5.3.1.5 Typic Dystrudepts (Fedscreek Series)

These well-drained soils form under udic moisture regimes in loamy colluviums weathered from sandstone, siltstone, and shale. The parent materials consist of acid channery-sized skeletal materials. Because the chemical and physical buffering capacity of the colluvium is low, and the parent materials are siliceous clasts, very acid conditions develop upon weathering. Profile development within solum, however, is minimal because the preweathered clasts are resistant to further decomposition and disintegration. Hence, release of clay and other products of weathering for subsequent translocation and transformation are limited. Again, the nature of the parent material has been pivotal in governing the intensity, modes, and mechanisms of pedogenesis in these coarse-textured, highly permeable, acid upland soils.

33.5.3.1.6 Typic Sulfaquepts (Gimhae Series)

These soils are common along coastlines with tidal influence. Here, the parent materials are medium textured alluvial deposits, usually associated with deltas or depositional regions of rivers that empty into the ocean. These wet environments are favorable for the accumulation of metal sulfides (e.g., pyrite), which upon drainage (natural or artificial) oxidize to form sulfuric acid and other acid sulfate weathering products (Van Breeman, 1982). Sulfaquepts are important soils because of their land use, distribution, and behavior are conditioned by their acidic character.

33.5.3.2 Climate

Inceptisols occur in subhumid to humid climates from equatorial to tundra regions where, if they have permafrost within 2 m, they are replaced by Gelisiols (Soil Survey Staff, 1998, 2010). Inceptisols cannot have an aridic (torric) moisture regime unless they have an Anthropic epipedon, which causes them to become Anthrepts.

Soil temperature and moisture are used at the suborder level in *Soil Taxonomy* to reflect major controls on current soil-forming processes. Such controls influence not only rates, modes, and mechanisms of physical and chemical weathering, but also biological and geomorphic processes that influence relief, landscape stability, topography, and soil drainage. Because Inceptisols form over a wide range of climatic conditions, excluding aridic and pergelic, the climatic impact on Inceptisols is similar to that of most other orders.

Soil moisture regime is used as a differentiation for four Inceptisols suborders (Aquepts, Udepts, Ustepts, and Xerepts), soil temperature regime for one suborder (Cryepts), and past human cultural impacts for one suborder (Anthrepts). For Cryepts, it was judged that temperature constraints on pedogenesis, use and management were more important than moisture constraints, even though both are important; the soil temperature regime was used as a differentiation at the family level for all orders except Gelisols and those soils that have a cryic soil temperature regime. Anthrepts have no particular climatic bias and are of limited geographical extent.

Aquepts form under aquic or excess wetness conditions (saturation with water, periodic reduction, and formation of contemporaneous redoximorphic features, redox concentrations and/or depletions). Udepts form under better drained, strongly leaching environments, where precipitation exceeds evapotranspiration. These soils generally lack salts and carbonates in the solum and have undergone moderate to strong weathering. Generally, summer periods of soil moisture deficit are insufficient to negatively impact mesophytic agricultural crops. Ustepts are formed under drier climates with significant periods of summer soil moisture deficit. Precipitation is less than evapotranspiration, at least during significant periods of the summer. They commonly exhibit incomplete leaching, weak to moderate weathering intensities, and pedogenic salts and/or carbonates in the lower solum. Moisture deficit is sufficiently severe to limit crop growth during summer months. Xerepts are similar to Ustepts but are formed in climates where winter precipitation is dominant. These soils are common in Mediterranean regions of Europe and along the Pacific coastal zones of western United States. The Cryepts serve as an interface between the cold soils of high latitudes with pergelic conditions (permafrost within 2 m) and cold soils of temperate region or orographic mountainous zones without permafrost within 2 m.

In Table 33.24, the climate effects on soil classification and corresponding properties are illustrated for a few pedons. Vertic Calciustepts (*Vernon series*) and Oxic Haplustepts (*Kahana series*) are examples of soils with an ustic moisture regime and contrasting cambic horizon composition, development, and chemical characteristics. The Calciustepts are incompletely leached, have carbonates throughout the pedon, and have a clay mineral composition that favors shrink/swell activity with changing moisture content. Comparison between Fluventic Haploxerepts (*Vidhas series*) and Humic Dystrocryepts (*Moran series*) reflecting contrasting temperature and precipitation regimes due to elevation differences. Although parent materials are different, climate is the major determinant governing increased organic matter contents in the Humic Dystrocryepts, which formed under udic moisture and cryic temperature regimes about 1050 m elevation. This is compared to the Fluventic Haploxerepts formed under xeric moisture and thermic temperature regimes at an elevation of about 90 m. The Dystrocrypets have accumulated much greater organic carbon contents in the Histic layer (Oi) overlaying and umbric epipedon and a low base subsoil. In contrast, the Haploxerepts have accumulates much less organic matter in the ochric epipedon, and the presence of carbonates distributed throughout the profile implies weak leaching potentials. Climate, in this example, is considered an intensity factor whereby the rate of soil development increases as temperature and rainfall increase, but decreases as either decreases. Limit soil moisture during periods of optimal soil weathering limits profile development in the Haploxerepts. In contrast, the Dystrocryepts have optimal moisture for pedogenesis, but the soil temperature is limiting.

33.5.3.3 Topography

The topography associated with many Inceptisols is a sloping to strongly sloping landscape subject to erosion rates that are equal to or greater than soil formation rates. Many Inceptisols occur on hillsides of major mountains throughout the world. Profile developments here are constrained by unstable landforms and cool climates. Youthful Inceptisols are common also on nearly level fluvial, lacustrine, and deltaic plains (or associated depressions) where profile development is limited by age of the sediments and/or shallow groundwater tables. Coastal estuaries are also probable locations for Inceptisols, especially those where metallic sulfides and sulfates are likely to accumulate. Subaqueous condition favor the formation of the cambic horizon of the soils formed before subsidence and sea level rise. Processes that are active for subaqueous conditions that form cambic horizons have not been extensively studied.

33.5.3.4 Vegetation

It is difficult to isolate vegetation as an independent soil-forming factor governing the synthesis of Inceptisols because vegetation is dependent on climate and topography. Interdependency of these factors is easily demonstrated by comparing the Humic Dystrocryepts (*Moran series*) and Vertic Calciustepts (*Vernon series*) of Table 33.24. Dystrocryepts support mountain meadow and alpine vegetative communities such as Kobres (*Koberesia* sp.), Asters (*Aster* sp.), and Yarrow (*Achillea* sp.) under cold, moist climates. Under hotter, drier conditions of the Calciustepts, native vegetation is mixed, midgrass prairie consisting of buffalograss (*Buchloe dactyloides*), Grama grasses [blue (*Bouteloua gracilis*), hairy (*B. hirsuta*), and sideoats (*B. curtipendula*)], and tobosa grass (*Hilaria mutica*). Primary productivity, accumulation of soil organic matter, and leaching potentials are governed by effective soil moisture and temperature regimes. Effects on soil properties are remarkably evident in these two soils (Table 33.24). Primary productivity for the Dystrocryepts is limited by soil temperature regime while a seasonal soil moisture deficit limits primary productivity for Calciustepts. Higher organic matter accumulation, low turnover rates, and moderate to strong leaching potentials (acid subsoils) are consequences in Dystrocryepts. In contrast, lower organic matter contents, high turnover rates, and less effective leaching potentials (calcareous solum) occur in Calciustepts.

33.5.3.5 Time

Many examples are available in the literature to demonstrate that time is a soil-forming factor (Jenny, 1941, 1980). Some might argue that factors other than time have more influence on soil formation. While this may be true in some cases, time in combination with the intensity factor (climate) has been responsible for many attributes of Inceptisols. Inceptisols are usually considered immature soils, but some have advanced stages of development and are just short of completing their evolutionary cycle from Entisols to Oxisols. For example, Oxic Haplustepts are similar to Oxisols but either have oxic conditions too deep in the soil (greater than 150 cm) or have CEC charge characteristics just a little too high for oxic horizon requirements—a prerequisite for all Oxisols. Upon erosion and truncation, some Inceptisols have soil properties at depths associated with old stable landforms prior to dissection. Further, several taxonomic classes of Inceptisols have morphological features that could be interpreted either as youthful or an advanced stage of pedogenesis. Arenic Eutrudepts is one such example in which the soils may have developed from recent sandy sediments or may be residuum from weakly consolidated sandstone on old stable landforms.

33.5.4 Morphology

Inceptisols have many kinds of diagnostic horizons and epipedons, but the most common sequence of horizons is an ochric or umbric epipedon overlying a cambic or fragipan horizon. However, a cambic horizon is not required if an umbric, histic, or plaggen epipedon is present or if there is a fragipan, duripan, placic, calcic, petrocalcic, gypsic, petrogypsic, salic, or sulfuric horizon (Soil Survey Staff, 1998, 2010). They can also have anthropic and mollic epipedons but not argillic, kandic, or natric horizons unless buried below recent sediments. Oxic horizons are permitted only if the upper boundary is deeper than

150 cm. Spodic horizons are permitted only if they are less than 10 cm thick or if the upper boundary is deeper than 50 cm.

It is clear from the above discussion that the number and kinds of processes active in Inceptisols and the definition of Inceptisols are unavoidably complex because the order serves dual functions in grouping soils of minimal pedogenic development and bifurcating *Soil Taxonomy*. In summary, they are youthful soils in rejuvenated landscapes but pedogenetically advanced on more stable landforms. Their attributes are highly variable and seldom uniquely definitive.

33.5.5 Pedogenic Processes

Pedogenic processes common to Inceptisols have been briefly considered under Section 33.5.3. Such processes are converged in greater detail in Chapters 30 and 31. The processes germane to Inceptisols are combinations and complexes of subprocesses and reactions found in most other soil orders. The nature of these processes, including their terminology, is reviewed by Marbut (1921), Kellogg (1936), Byers et al. (1938), Simonson (1959), Buol et al. (1980), Arnold (1983), Fitzpatrick (1986), and Fanning and Fanning (1989).

Biological processes including establishment of macro and micro fauna and flora, microbial and fungal mineralization, decomposition (mummification) and accumulation of organic matter, biotic transformations (e.g., ammonification, nitrification, denitrification, and nitrogen fixation), and pedoturbation are important in epipedon and cambic horizon genesis. Chemical processes including hydration, hydrolysis, solution, mineral synthesis, and oxidation/reduction reactions are responsible for both epipedon and cambic horizon synthesis. Physical processes contribute to the formation of pedogenic structure (obliteration of at least 50% of the "rock" structure) and transport of materials within the solum. They include aggregation, expansion/contraction (shrink/swell), freeze–thaw, and mass transport by solution or suspension (translocation).

Formation of a cambic horizon requires conversion of rock structure to soil structure (peds) and evidence of at least one of the following: oxidation/reduction (aquic conditions); neoformations of clay; translocation of salts, carbonates, clay, and/or organic–metal complexes; or liberation of free iron oxides. The cambic horizon is a prerequisite for most Inceptisols, but exceptions occur for certain epipedons or subsurface diagnostic features (Section 33.5.3).

Illuviation and/or neosynthesis of clay and organic–metal complexes cannot be of such magnitude that argillic, natric, kandic, or spodic horizons are formed, unless buried. Processes responsible for formation of oxic horizons include desilication, neoformation of low activity clays, and/or residual concentration of secondary oxides and oxyhydroxides of iron and aluminum (Kellogg, 1936; Sivarajasingham et al., 1962; Cline, 1975; Jenny, 1980). Acid sulfate soils are formed by processes termed sulfidization and sulfurization. These processes are responsible for the accumulation of metal sulfides in soils and sediments, oxidation of these sulfides, formation of sulfuric acid, ferrolysis, soil acidification, and mineral dissolution (Kittrick et al., 1982; Van Breeman, 1982; Fanning and Fanning, 1989).

33.5.6 Classification

In the 1938 classification scheme (Baldwin et al., 1938) and in most subsequent revisions in the United States prior to *Soil Taxonomy*, Inceptisols were included in a number of great soil groups including Ando, Hydro Humi Latosols, Humic Gley, Tundra, Half Bog, Sols Bruns Acids, Brown forest, Regosols, Lithosols, Aluvial, and numerous other minor groups (Foss et al., 1983). Even at that time, this was a very extensive and diverse taxa of soils. In the Canadian soil classification system, the wet Inceptisols (Aquepts) were classified as Gleysols but most of the better drained members were Brunisols. In the legend of the FAO/UNESCO soil map of the world (FAO-UNESCO, 2003), most well-drained Inceptisols would be classed as Cambisols or Phaeozems if the base saturation is low. The wet Inceptisols would be classed mostly as Gleysols except those with acid sulfate conditions, which would be grouped with Fluvisols.

Inceptisols suborders, great groups, and subgroups are represented in Table 33.25. Rationale for these taxa has been considered previously in this chapter and in Chapter 32. The following discussion of suborders has been taken directly from *Soil Taxonomy* (1998) to illustrate the general nature and characteristics of Inceptisols suborders.

33.5.6.1 Anthrepts

These are the Inceptisols that are the more less freely drained that have either an anthropic or a plaggen epipedon. Most of these soils have been used as cropland or places of human occupation for many years. They can have almost any temperature regime, and almost any vegetation. Most of them have a cambic horizon. Only two subgroups are recognized at this time, those with plaggen epipedons indicative of ancient cultural sites (*Plaggenanthrepts*), and those with anthropic epipedons reflecting other long-term human impacts on soils such as irrigated farmlands or housing areas. Most of these soils are in Eurasia or Northern Africa.

33.5.6.2 Aquepts

These are the wet Inceptisols. Their natural drainage is poor or very poor and, if they have not been artificially drained, groundwater is at or near the soil surface at some time during normal years but typically not during at all seasons. They mostly have a gray to black surface horizon and gray subsurface horizon with redox concentrations that begin at a depth of less than 50 cm. A few have a brownish surface horizon that is less than 50 cm thick.

Most Aquepts have developed in late Pleistocene or younger deposits in depressions, nearly level plains, or floodplains. They occur from the equator to latitudes with discontinuous permafrost. The common features of most of these soils are the gray and rusty colors of redoximorphic features at a depth of 50 cm or less and the shallow groundwater or artificial drainage. They may have almost any particle size class except fragmental, and any reaction class, any temperature regime except pergelic, and almost any

TABLE 33.25 Listing of Suborders, Great Groups, and Subgroups for Inceptisols Order

Suborder	Great Groups	Subgroups
Aquepts	Sulfaquepts	Salidic, Hydraquentic, Typic
	Petraquepts	Histic Placic, Placic, Plinthic, Typic
	Halaquepts	Vertic, Aquandic, Duric, Aeric, Typic
	Fragiaquepts	Aeric, Humic, Typic
	Cryaquepts	Sulfic, Histic Lithic, Lithic, Vertic, Histic, Aquandic, Fluvaquentic, Aeric Humic, Aeric, Humic, Typic
	Vermaquepts	Sodic, Typic
	Humaquepts	Hydraquentic, Histic, Aquandic, Cumulic, Fluvaquentic, Aeric, Typic
	Epiaquepts	Vertic, Aquandic, Fluvaquentic, Fragic, Aeric, Humic, Mollic, Typic
	Endoaquepts	Sulfic, Lithic, Vertic, Aquandic, Fluvaquentic, Fragic, Aeric, Humic, Mollic, Typic
Anthrepts	Plagganthrepts	Typic
	Haplanthrepts	Typic
Cryepts	Eutocryepts	Humic Lithic, Lithic, Andic, Vitrandic, Aquic, Oxyaquic, Lamellic, Xeric, Ustic, Humic, Typic
	Dystrocryepts	Humic Lithic, Lithic, Andic, Vitrandic, Aquic, Oxyaquic, Lamellic, Spodic, Xeric, Ustic, Humic, Typic
Gelepts	Humigelepts	Lithic, Andic, Aquic, Oxyaquic, Fluventic, Turbic, Eutric, Typic
	Dystrogelepts	Lithic, Andic, Aquic, Fluventic, Turbic, Eutric, Typic
	Haplogelepts	Lithic, Andic, Aquic, Fluventic, Turbic, Eutric, Typic
Ustepts	Durustepts	Typic
	Calciustepts	Lithic Petrocalcic, Lithic, Torrertic, Vertic, Petrocalcic, Gypsic, Aquic, Aridic, Udic, Typic
	Dystrustepts	Lithic, Andic, Vitrandic, Aquic, Fluventic, Oxic, Humic, Typic
	Haplustepts	Aridic Lithic, Lithic, Udertic, Torrertic, Vertic, Andic, Vitrandic, Anthraquic, Aquic, Oxyaquic, Oxic, Lamellic, Torrifluventic, Udifluventic, Fluventic, Gypsic, Haplocalcidic, Calcic, Udic, Calcic, Aridic, Dystric, Udic, Typic
Xerepts	Durixerepts	Aquandic, Andic, Vitrandic, Aquic, Entic, Typic
	Calcixerepts	Lithic, Vertic, Petrocalcic, Sodic, Vitrandic, Aquic, Typic
	Fragixerepts	Andic, Vitrandic, Aquic, Humic, Typic
	Dystroxerepts	Humic Lithic, Lithic, Aquandic, Andic, Vitrandic, Anthraquic, Fragiaquic, Fluvaquentic, Aquic, Oxyaquic, Fragic, Fluventic Humic, Fluventic, Humic, Typic
	Haploxerepts	Humic Lithic, Lithic, Vertic, Aquandic, Andic, Vitrandic, Gypsic, Aquic, Lamellic, Fragic, Fluventic, Calcic, Typic
Udepts	Sulfudepts	Typic
	Durudepts	Aquandic, Andic, Vitrandic, Aquic, Typic
	Fragiudepts	Andic, Vitrandic, Aquic, Humic, Typic
	Calciustepts	Lithic Petrocalcic, Lithic, Torrertic, Vertic, Petrocalcic, Gypsic, Aquic, Aridic, Udic, Typic
	Dystrudepts	Lithic, Andic, Vitrandic, Aquic, Fluventic, Oxic, Humic, Typic

Source: Soil Survey Staff. 2010. Keys to Soil Taxonomy. 11th edn. U.S. Government Printing Office, Washington, DC.

vegetation. Most of them have a cambic horizon, and some have a fragipan. It is possible that some have a plaggen epipedon.

Table 33.25 lists the nine great groups of Aquepts. These great groups are based upon limiting features or horizons that impact use, management, and behavior. Specifically, these include soils with acid sulfuric horizons within 50 cm of the surface associated with oxidation of metal sulfides (*Sulfaquepts*); soils with a restrictive cemented subsurface horizon that forms a continuous phase within 100 cm of the soil surface (*Petraquepts*); salty (salic) or alkali (natric) subsoil horizons (*Halaquepts*); soils with fragipan (*Fragiaquepts*); soils with a cryic temperature regime (*Cryaquepts*); soils with strongly bioturbated layers by macrofauna such as crayfish, worms, and mammals (*Vermaquepts*); soils with darkened, organic-enriched surface horizons (*Humaquepts*); soils with a perched water table (*Epiaquepts*); soils that are saturated from a groundwater source (*Endoaquepts*).

33.5.6.3 Cryepts

Cryepts are the cold Inceptisols of high mountains or high latitudes. They cannot have permafrost. The vegetation is mostly conifers or mixed conifers and hardwood trees. Few of them are cultivated. These soils may be formed in loess, drift, alluvium, or solifluction deposits, mostly late Pleistocene or Holocene in age. They commonly have a thin, dark brownish ochric epipedon and a brownish cambic horizon. Some have bedrock within 100 cm of the surface. In the United States, these soils are moderately extensive in the high mountains of the West as well as other mountainous areas of southern Alaska and the world.

Two great groups of Cryepts are recognized (Table 33.25). Those that are calcareous or have high base saturation are *Eutocryepts*, and Cryepts with low base saturation are *Dystrocryepts*.

33.5.6.4 Udepts

Udepts are mainly the more or less freely drained Inceptisols that have a udic or perudic moisture regime. They formed on nearly level to steep surfaces mostly of late Pleistocene or Holocene age. Some, where the soil moisture regime is perudic, formed in older deposits. Most them had or now have a forest vegetation, but some have shrub or grass. A few have been formed from

Mollisols by truncation of the mollic epipedon (*Eutrudepts*), mostly under cultivation. Most of them have an ochric or umbric epipedon and a cambic horizon with low base saturation (*Dystrudepts*). These were the Sols Bruns Acides of earlier classification schemes (*Fragiudepts*), or a duripan (*Durudepts*). In the United States, Udepts are most extensive on the Appalachian Mountains, the Allegheny Plateau, and the West Coast; they also occur extensively in Eurasia.

33.5.6.5 Ustepts

Ustepts are the more or less freely drained Inceptisols that have a ustic moisture regime. They have dominantly summer precipitation or an isomesic, hyperthermic, or warmer temperature regime. Many of these soils are calcareous at a shallow depth and have a Bk or calcic horizon (*Calciustepts*). A few have formed from Mollisols by truncation of the mollic epipedon, mostly under cultivation. Most of them have an ochric or umbric epipedon and a cambic horizon (*Haplustepts*). Some have a duripan (*Durustepts*), especially in areas where a labile silica source is associated with pyroclastic deposits or volcanic ash. Ustepts with low base saturation (*Dystrustepts*) occur in areas with polycycled preweathered acidic sediments or outcrops of acid bedrocks common in West Africa and isolated areas of the United States. The native vegetation commonly was grass, but some supported tress and savannas. Ustepts are of moderate extent in the United States. They are most common on the Great Plains mostly in Montana, Texas, and Oklahoma. They occur extensively in ustic sectors of the Americas, Eurasia, West Africa, and several island countries.

33.5.6.6 Xerepts

Xerepts are the more or less freely drained Inceptisols that have a xeric moisture regime, dominantly winter precipitation. They have a frigid, mesic, or thermic temperature regime. They formed mostly in Pleistocene or Holocene deposits or on steep slopes. Many of these soils are calcareous at a shallow depth and have a Bk, a calcic, or a petrocalcic horizon (*Calcixerepts*). Others have low base saturation (*Dystrixerepts*). Most of them have an ochric or umbric epipedon and a cambic horizon (*Haploxerepts*). Some have a duripan (*Durxerepts*) and a few have a fragipan (*Fragixerepts*). The native vegetation was commonly coniferous forest on those with a thermic temperature regime. These soils are of moderate extent in the United States. They are most common near the West Coast in the States of California, Oregon, Washington, Idaho, and Utah. They are the major soils of the Mediterranean regions of Eurasia and Northern Africa.

33.5.7 Physical, Chemical, and Mineralogical Properties

Because of the diverse nature of Inceptisols, few generalized statements can be made about their physical, chemical, biological, and mineralogical properties. Their properties are nearly as inclusive as all the other soil orders collectively. Inceptisols span the global regions from intensively weathered to minimally developed soils. For example, they are acidic to alkaline in reaction, weak to strongly physic-chemically buffered, have low to high organic matter contents, low to high hydraulic transport functions, low to high water retention values, fertile to infertile in nutrient status, etc. The *Soil Survey Manual* (Soil Survey Staff, 1993) identifies class ranges for the above classes and for many other soil attributes described and measured. Table 33.24 illustrates the kind of magnitude of diversity vested in many of these Inceptisols attributes. The database online for Inceptisols (http://soildatamart.nrcs.usda.gov/) further documents this aspect.

The reader is referred to discussions of other orders in Sections 33.2 through 33.13 for physical, chemical, biological, and mineralogical attributes likely to be associated with Inceptisols. For example, those Inceptisols with andic surface materials are similar to Andisols; those with vertic shrink/swell features to Vertisols; those with calcic, gypsic, salic, petrocalcic, and petrogypsic horizons to Aridisols; those with translocation and/or neosynthesis of clay or organic–metal colloids to Alfisols, Ultisols, and Spodosols; those with high sesquioxide contents and/or low activity clay mineral suites to Oxisols; and those with cryic temperature regimes to Gelisols.

The sand and silt mineralogy commonly consists of quartz, mica, and feldspars with minor components of weatherable heavy minerals (opaque). The opaque minerals in soils derived from basic bedrocks are feldspars, hematite, magnetite, ilmenite, and rutile. Inceptisols rich in sesquioxides (Oxic subgroups) comprise oxyhydroxides of Fe, Al, and Ti including gibbsite, hematite, goethite, boehmite, rutile, and anatine. Those soils associated with drier climates (Ustepts and Xerepts) contain soluble salts (e.g., $NaCO_3$, $NaSO_4$, and NaCl), gypsum and carbonates (calcite and dolomite). Inceptisols with andic and vitrandic materials (pyroclastics) contain amorphous or short-range order minerals such as allophone, ferrihydrite, and glasses. Phyllosilicate clay minerals range from smectite, mica, kaolinite, chlorite, vermiculate, and mixed layer assemblages to rather unique suites dominated by serpentine and glauconite. In acid sulfate soils (e.g., Sulfaquepts and Sulfudepts), jarosite is a common constituent that marks very acid conditions associated with metal sulfide oxidation. Pedogenic gypsum is also common in these soils.

33.5.8 Management

For management purposes, Inceptisols can be subdivided into three land uses, namely forestry, pasture production, and agronomic cropping.

33.5.8.1 Forestry

Most Inceptisols under forested land use occur in mountainous regions on slopes ranging from 3% to 90%. On steep terrains, management systems other than natural regrowth are environmentally unacceptable and practically impossible. Most forested Inceptisols have carbon contents sufficiently high as long as the surface is not eroded, that indigenous nutrient recycling is sufficient to sustain growth without external fertilizer amendments. On less sloping terrain, many large, commercial timber companies find fertilizer amendments, especially phosphorous applications, to be economically beneficial. Management techniques to control competition from unwanted species and disease suppression commonly involve controlled burns or physical removal of dead or unwanted species. Harvesting methods depend on the slope gradient; they range from surface skidding to aerial removal by helicopter or cable lifts. Harvest schemes may involve clear cutting or selective harvesting. Damages from erosion and compaction of harvest operations on less sloping Inceptisols are commonly ameliorated with surface tillage operations on less sloping Inceptisols are commonly ameliorated with surface tillage operations to break up surface crusts, compacted zones and to establish soil and water conservation buffers.

33.5.8.2 Pasture Production

Most of the Inceptisols under pasture management are planted to improved pastures that respond to N, P, and K fertilization, especially if the pasture is intensely managed for animal or forage production. Areas, which occur under native range, or under shifting traditional agriculture as in developing countries, should be managed using best management practices for stocking rate, pasture rotation, and fallow period. These vary from one region to another, not only in terms of soil conditions but also in terms of the ability of the operator to provide investment inputs. Commonly under traditional agriculture, the inputs are minimum and the soil dictates the best management and utilization practices to be followed. Inceptisols with an ustic or xeric moisture regime are limited by insufficient rainfall. Irrigation may be used to supplement natural precipitation on some Inceptisols, but the cost of this investment for improved pastures may make it economically unattractive. However, many examples of irrigated improved pastures occur in Europe and the United States where the economics are favorable and adequate high quality aquifers are available. Surface or subsurface drainage may be required to remove surface water or lower the water table for optimal forage production, especially for Aquepts.

33.5.8.3 Agronomic Cropping

Inceptisols are the major soils on which agronomic crops are produced in some parts of the world. Aquepts, Cryepts, Udepts, Ustepts, and Xerepts are very productive and valuable agricultural land resources, if properly managed. Crops produced depend on climate (length of growing season, dependability of precipitation, seasonal periods of soil moisture excess or deficit, photoperiodism, degree days, etc.), but because of Inceptisols diversity, most major food, feed, and fiber crops are included. Irrigation is used on vegetable, citrus, and other important cash or specialty crops when summer periods of moisture deficit become extreme. The highly weathered Inceptisols within the tropics are used for sugarcane, pineapple, cotton, and some upland rice production with many of these crops being produced under irrigation. Most of these soils respond to N, P, and K fertilizers and some soils require lime in order to neutralize acidity and sustain production.

Cryepts are used mostly for cereal crops because of the short growing season, photoperiodism, and few number of degree days. Sometimes these soils are used for cool season vegetables and forage production. Where these soils interface with Gelisols, they are constrained by cold soil temperatures, freeze–thaw heaving, and erosion induced by wind, water, and mass movement (solifluction). They may be under alpine meadow.

Aquepts used for agronomic production usually require surface or internal drainage but depending on soil temperature, a wide range of crops are grown including cotton, sugarcane, rice, and corn in the subtropical and tropical climates and corn, soybeans, sorghum, and some cereal crops in temperate climates. Sulfaquepts (and Sulfudepts) require specialized management because when they are drained mainly for lowland rice production, the sulfuric horizon produces large amounts of sulfuric acid. Because the reactions involved were not well understood when most of these soils were drained, vast wastelands were created. Management schemes, which reduce acid production, have been developed by Ponnamperuma et al. (1973), Cisse et al. (1993), Van Breeman (1982), and Coly (1996). Reclamation and management practices for acid sulfate soils have been proposed by Van Breeman (1982), Rimwanich and Suebsuri (1983), and Coly (1996). The formation of a sulfuric horizon should be prevented or minimized by controlling the depth of the water table upon drainage (keeping the soils saturated to prevent pyrite oxidation). Sulfaquepts can be reclaimed by inducing reducing conditions within the sulfuric horizon and/or flushing the soil to remove the sulfuric acid. The flushing can be accomplished using a combination of saline and freshwater to remove the soluble salts. The addition of $CaCO_3$ is required to neutralize any acidity produced, and $CaSO_4$ is required when saline water has been used for reclamation to remove excess sodium (Cisse et al., 1933; Coly, 1996).

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33.6 Gelisols

J.G. Bockheim

C. Tarnocai

33.6.1 Introduction

Gelisols, which are the permafrost-affected soils, constitute the 12th and newest soil order. They comprise 18 million km^2 or about 13% of the Earth's land surface and occur in the Arctic, Subarctic, Boreal, Antarctic, Subantarctic, and some alpine regions under cold continental, subhumid or semiarid, and arid conditions (Bockheim et al., 1994). They are either unvegetated or support continuously vegetated tundra, subarctic and boreal forest, and some alpine tundra. Gelisols are of global concern because they contain many protected areas, support numerous indigenous populations who depend on the land and surrounding oceans for sustenance, and may be subject to considerable impacts from human development (oil, coal and gas exploration, and mining) and global warming.

Gelisols are defined as soils having permafrost within 100 cm of the soil surface, or gelic materials within 100 cm of the soil surface and permafrost within 200 cm of the soil surface (Soil Survey Staff, 2010). Gelic materials, in turn, are seasonally or perennially frozen mineral or organic soil materials that have evidence of cryoturbation (frost churning), ice segregation, or cracking from cryodesiccation. Gelic materials contrast with other kinds of materials, such as andic or spodic materials, in being defined entirely on the basis of physical and thermal characteristics, rather than chemical properties.

A representative Gelisol is shown in Figure 33.29. Gelic materials occur in both the active layer and the upper part of the permafrost as evidenced by cryoturbation, denoted in soil descriptions by the subscript jj (y in the Canadian system).

33.6.2 Permafrost and the Occurrence of Gelisols

Gelisols only occur in areas containing permafrost within 100 cm of the soil surface if the soil is not cryoturbated or 200 cm of the surface if the soil is cryoturbated. Permafrost is defined

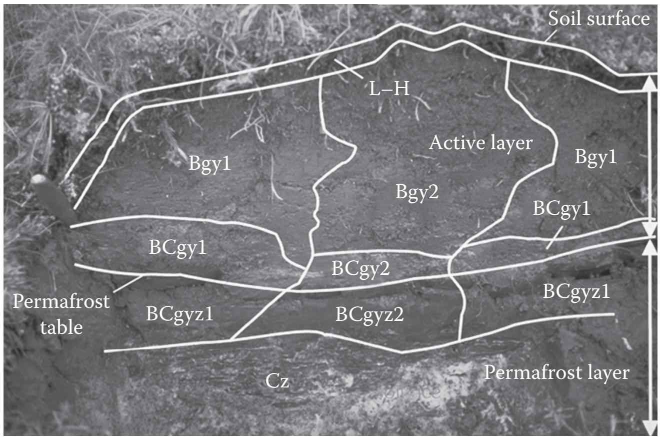

FIGURE 33.29 An aquaturbel developed on an earth hummock in northern Canada. The Canadian soil horizon nomenclature is used, where L–H = Oi–Oa, y = jj, and z = j in *Soil Taxonomy*.

here as a thermal condition in which a material (including soil materials) remains below 0°C for 2 or more years in succession. Permafrost may be ice-cemented (designated in soil descriptions with the subscript fm) or in the case of insufficient interstitial water, dry (designated as f). In the frozen layer, a variety of ice lenses, vein ice, segregated ice crystals, and ice wedges may be evident. An important consideration is that the permafrost is in dynamic equilibrium with the environment.

Permafrost, which comprises about 13% of the Earth's surface and 24% of the northern hemisphere, is differentiated into four zones on land in the circumpolar regions: continuous (91%–100% cover), discontinuous (51%–90% cover), sporadic (10%–50% cover), and isolated (<10% cover) patches (Figure 33.30).

The distribution and thickness of permafrost are influenced by natural surface features, such as snow and vegetation cover, topography, and bodies of water, but as Figure 33.30 shows, are most affected by regional climate.

33.6.3 Description and Distribution of Gelisols

There are three suborders within the Gelisol order (Soil Survey Staff, 2010): Histels, Turbels, and Orthels that are differentiated on the basis of organic matter content and for mineral soils whether or not there is cryoturbation. Distribution of Gelisols is depicted in Figure 33.31. Suborders, great groups, and subgroups are shown in Table 33.26.

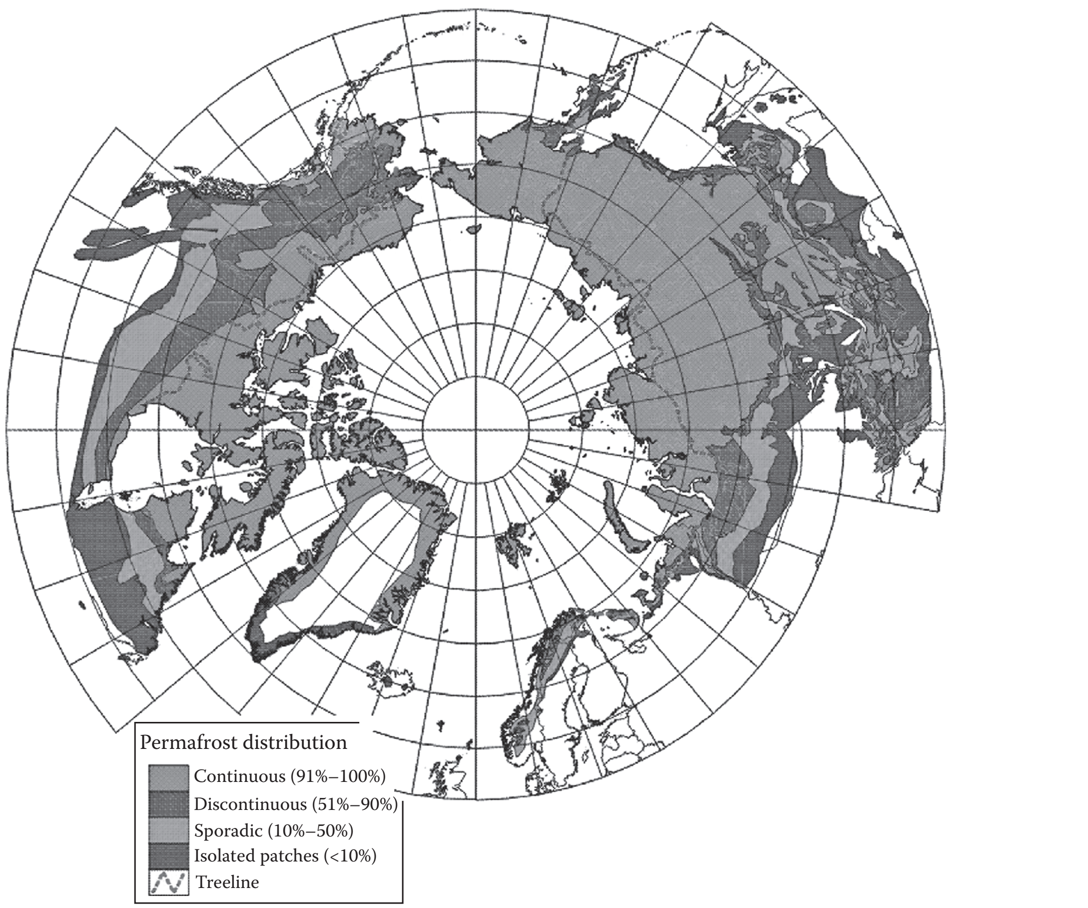

FIGURE 33.30 Distribution of permafrost in the northern hemisphere. (From Brown, J., O.J. Ferrians, Jr., J.A. Heginbottom, and E.S. Melnikov. 1997. Circumarctic map of permafrost and ground ice conditions, 1:10 000 000 scale map. U.S. Geological Survey International Permafrost Association. U.S. Geological Survey, Washington, DC.)

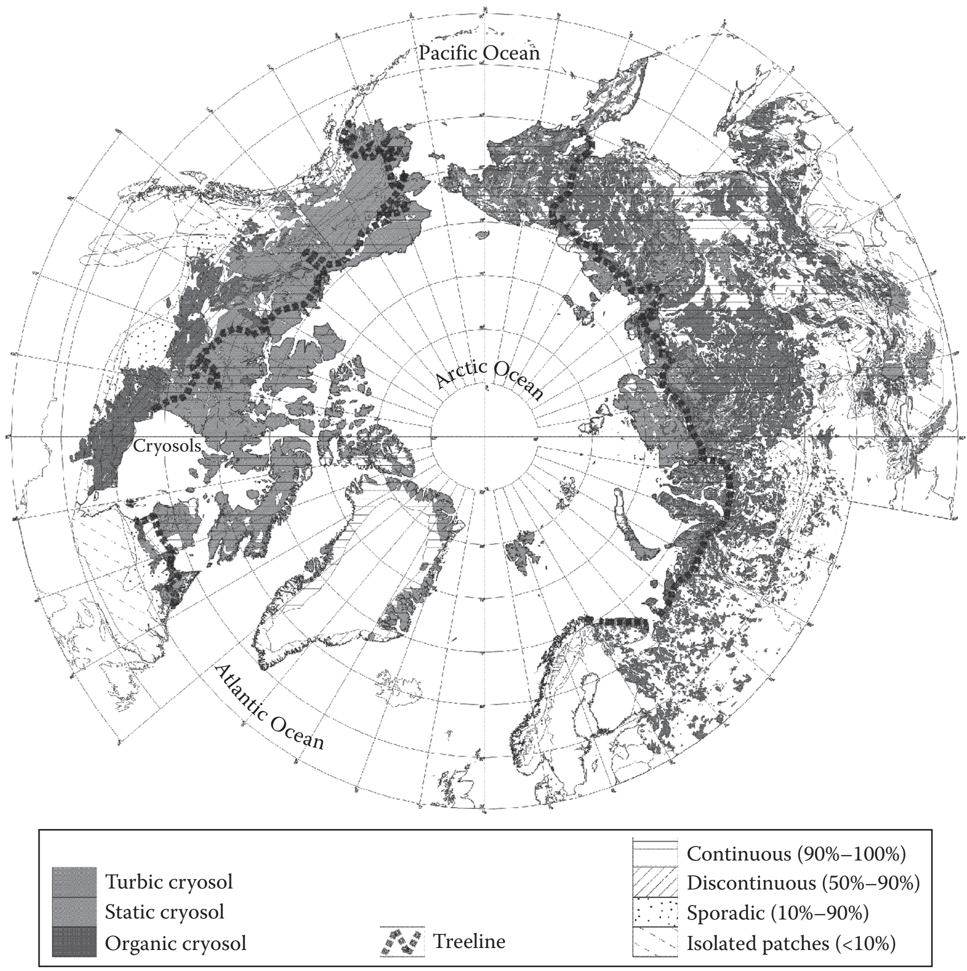

FIGURE 33.31 Distribution of Gelisols by dominant suborder in the circumarctic region. (From Tarnocai, C., D. Swanson, J. Kimble, and G. Broll. 2007. Northern circumpolar carbon database. Digital Database, Research Branch, Agriculture and Agri-Food Canada, Ottawa, Canada. With permission.)

33.6.3.1 Histels

Histels are Gelisols that have 80% or more (by volume) of organic materials from the soil surface to a depth of 50 cm, or to a restricting layer (Figure 33.32a). Histels otherwise meet the requirements of a Histosol except for the presence of permafrost within the upper 100 cm. Histels comprise 15% of the Gelisols in the circumarctic (Tarnocai et al., 2007; Figure 33.31). Most of these soils are located in the Mackenzie River Valley and the Hudson Bay lowlands of Canada. In general, Histels mostly occur in the Boreal, Subarctic, and Low Arctic regions. Histels are commonly associated with palsas (a peaty permafrost mound containing a core of alternating layers of segregated ice and peat or mineral soil), peat plateau, peat hummocks, and low- and high-centered lowland polygons. These soils have been discussed fully by Zoltai and Tarnocai (1971), Tarnocai (1972, 1973), Zoltai and Tarnocai (1974, 1975), Everett (1979), and Bockheim et al. (2004).

Histels are divided into five great groups, including the Folistels, Glacistels, Fibristels, Hemistels, and Sapristels. With the exception of the Glacistels, these great groups follow the suborders in the Histosol order (Section 34.2). The Glacistels contain a glacic layer within 50 cm of the surface that is 30 cm or more thick and contains 75% or more ice (by volume).

33.6.3.2 Turbels

Turbels are mineral soils that occur in areas with patterned ground. Patterned ground is a general term for any ground surface with a discernibly ordered, more or less symmetrical, morphological pattern of ground and, where present, vegetation (Washburn, 1980). These soils show marked influence of cryoturbation (Figure 33.32b) and occur throughout the circumpolar regions. They are the dominant suborder and account for more than 57% of the Gelisols mapped in the circumarctic (Figure 33.31). Field and laboratory data for these soils are contained in publications by Douglas and Tedrow (1960), Tedrow et al. (1968), Tedrow (1970), Zoltai and Tarnocai (1974), Pettapiece (1975), Tarnocai and Smith (1992), Bockheim and Hinkel (2007), and Simas et al. (2008).

Most horizons and layers within the active layer of Turbels are strongly affected by cryoturbation. Turbels commonly contain irregular or broken horizons, involutions, organic matter that

TABLE 33.26 Listing of Suborders, Great Groups, and Subgroups of the Gelisol Order

Suborder	Great Group	Subgroups
Histels	Folistels	Lithic, Glacic, Typic
	Glacistels	Hemic, Sapric, Typic
	Fibristels	Lithic, Terric, Fluvaquentic, Sphagnic, Typic
	Hemistels	Lithic, Terric, Fluvaquentic, Typic
	Sapristels	Lithic, Terric, Fluvaquentic, Typic
Turbels	Histoturbels	Lithic, Glacic, Ruptic, Typic
	Aquiturbels	Lithic, Glacic, Sulfuric, Ruptic–Histic, Psammentic, Typic
	Anhyturbels	Lithic, Glacic, Petrogypsic, Gypsic, Nitric, Salic, Calcic, Typic
	Molliturbels	Lithic, Glacic, Vertic, Andic, Vitrandic, Cumulic, Aquic, Typic
	Umbriturbels	Lithic, Glacic, Vertic, Andic, Vitrandic, Cumulic, Aquic, Typic
	Psammoturbels	Lithic, Glacic, Spodic, Typic
	Haploturbels	Lithic, Glacic, Aquic, Typic
Orthels	Historthels	Lithic, Glacic, Fluvaquentic, Fluventic, Ruptic, Typic
	Aquorthels	Lithic, Glacic, Sulfuric, Ruptic–Histic, Andic, Vitrandic, Salic, Psammentic, Fluvaquentic, Typic
	Anhyorthels	Lithic, Glacic, Petrogypsic, Gypsic, Nitric, Salic, Calcic, Typic
	Mollorthels	Lithic, Glacic, Vertic, Andic, Vitrandic, Cumulic, Aquic, Typic
	Umbrorthels	Lithic, Glacic, Vertic, Andic, Vitrandic, Cumulic, Aquic, Typic
	Argiorthels	Lithic, Glacic, Natric, Typic
	Psammorthels	Lithic, Glacic, Spodic, Typic
	Haplorthels	Lithic, Glacic, Fluvaquentic, Aquic, Fluventic, Typic

Source: Soil Survey Staff. 2010. Keys to soil taxonomy. 11th edn. USDA-NRCS, Washington, DC.

usually accumulates on the surface of the permafrost, oriented rock fragments, and silt caps and silt-enriched subsoil horizons. Permafrost occurs within 200 cm of the soil surface. These soils are differentiated into seven great groups that link them with other orders not containing permafrost, including the Histoturbels (40%–80% organic materials by volume in the upper 50 cm), Aquiturbels (aquic conditions), Anhyturbels (anhydrous conditions) (anhydrous conditions refer to soils of cold deserts and other areas with permafrost [often dry permafrost] and low precipitation [usually <50 mm $year^{-1}$ water equivalent]. Anhydrous soil conditions are similar to the aridic and torric soil moisture regimes except that the soil temperature is <0°C [Soil Survey Staff, 2010]), Molliturbels (mollic epipedon), Umbriturbels (umbric epipedon), Psammoturbels (sandy texture), and Haploturbels (other Turbels).

33.6.3.3 Orthels

Orthels are mineral soils containing permafrost within the upper 100 cm, but they lack cryoturbation (Figure 33.32c). These soils comprise <3.6% of the Gelisols mapped in the circumarctic and occur primarily in areas with dry permafrost such as floodplains and the dry valleys of Antarctica. They are divided

(a) (b) (c)

FIGURE 33.32 Examples of the three suborders of Gelisols: (a) Histel, (b) Turbel, and (c) Orthel. (Photo courtesy by J. Bockheim.)

TABLE 33.27 Comparison of Soil Taxa among Soil Taxonomy, the Canadian System, and the Revised *World Reference Base for Soil Resources*

Soil Taxonomy (Soil Survey Staff, 2010)	World Reference Base for Soil Resources (2006)	Canadian System (Agric. Canada Expert Committee on Soil Survey, 1987)
Gelisol (order)	Cryosol (soil group)	Cryosol (order)
Histels (suborder)		Organic Cryosols (great group)
Folistels	Cryic Folic Histosols	No equivalent
Glacistels	Glacic Histosols	Glacic Organic Cryosols
Fibristels	Cryic Fibric Histosols	Fibric Organic Cryosols
Hemistels	Cryic Hemic Histosols	Mesic Organic Cryosols
Sapristels	Cryic Sapric Histosols	Humic Organic Cryosols
Terric Fibristels	Cryic Fibric Histosols (Terric)	Terric Fibric Organic Cryosols
Terric Hemistels	Cryic Hemic Histosols (Terric)	Terric Mesic Organic Cryosols
Terric Sapristels	Cryic Sapric Histosols (Terric)	Terric Humic Organic Cryosols
Turbels (suborder)		Turbic Cryosols (great group)
Histoturbels	Turbic Cryosols	Gleysolic Turbic Cryosols
Aquiturbels	Turbic Cryosols (Reductaquic)	Gleysolic Turbic Cryosols
Anhyturbels	Turbic Cryosols (Natric, Salic, Calcic)	No equivalent
Molliturbels	Turbic Crysols (Eutric)	Brunisoilc Turbic Cryosols
Umbriturbels	Turbic Cryosols (Dystric)	Brunisolic Turbic Cryosols
Psammoturbels	Turbic Cryosols	Regosolic Turbic Cryosols
Haploturbels	Turbic Cryosols	Orthic Turbic Cryosols
Glacic subgroups	Glacic Cryosols	No equivalent
Orthels (suborder)		Static Cryosols
Historthels	Histic Cryosols	Gleysolic Static Cryosols
Aquorthels	No equivalent	Gleysolic Static Cryosols
Anhyorthels	Salic, Calcic, Anhyorthels	No equivalent
Mollorthels	Mollic Cryosols	Brunisolic Static Cryosols
Umbrorthels	Umbric Cryosols	Brunisolic Static Cryosols
Argiorthels	No equivalent	No equivalent
Psammorthels	Arenic Cryosolos	Regosolic Static Cryosols
Haplorthels	Haplic, Cambic, Cryosols	Regosolic Static Cryosols
Glacic subgroups	Glacic Cryosols	No equivalent
No equivalent	Technic Cryosols	No equivalent
No equivalent	Hyperskeletic Cryosols	No equivalent
Lithic subgroups	Leptic Cryosols	No equivalent
Vitrandic subgroups	Vitric Cryosols	No equivalent
Spodic Psammorthels	Spodic Cryosols	No equivalent

into great groups that parallel the Turbels, that is, Historthels, Aquorthels, etc. Subgroups within each great group of Gelisols are listed in Table 33.27.

33.6.4 Cryopedogenic Processes

Cryopedogenic processes that lead to gelic materials are driven by the physical volume change of water to ice, moisture migration along (1) thermal, (2) hydrostatic pressure, (3) solute concentration, and (4) electrical potential gradients in the frozen (or unfrozen) system (Marion, 1995), or thermal contraction of the frozen material by continued rapid cooling. These processes include freezing and thawing, cryoturbation, frost heaving, cryogenic sorting, thermal cracking, and ice segregation (Tedrow, 1977; Washburn, 1980; Figure 33.33).

It should be emphasized that cryopedogenic processes are soil-forming processes characteristic of soils with permafrost and should not be viewed as operating against the other soil-forming processes in lower latitude soils; rather, they are distinctive processes producing horizons and properties that are uncommon to other soil orders (Bockheim et al., 2010). Processes common to the other soil orders operate in Gelisols but at a lesser magnitude because of the dominance of cryopedogenic processes.

33.6.5 The Pedon as a Basic Soil Unit

The pedon is the basic soil unit for sampling in *Soil Taxonomy* (Soil Survey Staff, 1999); it is an especially important concept for describing, classifying, and sampling Gelisols. The pedon is defined so as to encompass the full cycle of patterned ground

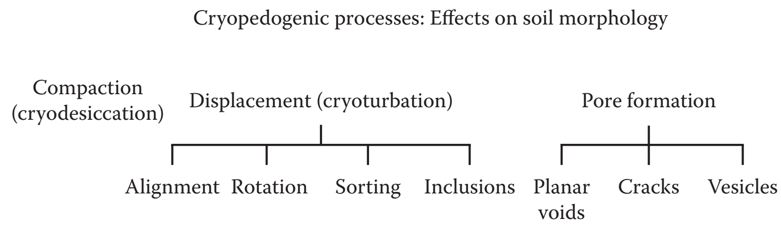

FIGURE 33.33 Conceptual framework showing the interrelationships of the effects of cryopedogenic processes on development of specific fabric types in Gelisols. (Revised from Fox, C.A. 1994. Micromorphology of permafrost-affected soils, p. 51–62. *In* J.M. Kimble and R.J. Ahrens (eds.) Proc. Meet. Classif. Correl. Manage. Permafrost-Affected Soils. USDA-SCS, National Soil Survey Center, Lincoln, NE. With permission.)

with a 1 or 2 m linear interval or a half cycle with a 2–7 m cycle (Figure 33.34a). This interval is suitable for most patterned ground features such as earth hummocks, circles, nets, and non-sorted polygons. In the case of large-scale (>7 m) ice-rich polygons, such as those that occur along the Alaskan Coastal Plain, two pedons are delineated: one within the center of the polygon and the other within the ice wedge (Figure 33.34b).

If no patterned ground exists, the pedon is arbitrarily selected but approximates about 1 m^2 in area.

Scaled sketches of a pedon showing soil horizons, including patches of cryoturbated material, should be drawn on graph paper in the field, or digital images should be taken and the horizons annotated directly on the image (Figure 33.29). Samples should be collected from each horizon across the pedon and composited for subsequent laboratory characterization. In the case of highly cryoturbated soils, the areal percentage of each horizon is reported in soil descriptions rather than depth intervals (Kimble et al., 1993).

33.6.6 Historical Approaches to Classification of Permafrost-Affected Soils

Tedrow (1977) provided a comprehensive review of early approaches to classification of soils in the cold regions, emphasizing the zonal systems. Of the soil taxonomic systems used today, only the approaches of Canada (Agriculture Canada Expert Committee on Soil Survey, 1987), the United States (Soil Survey Staff, 2010), and *World Reference Base (WRB) for Soil Resources* (2006) recognize soils with permafrost in a separate order (Table 33.27).

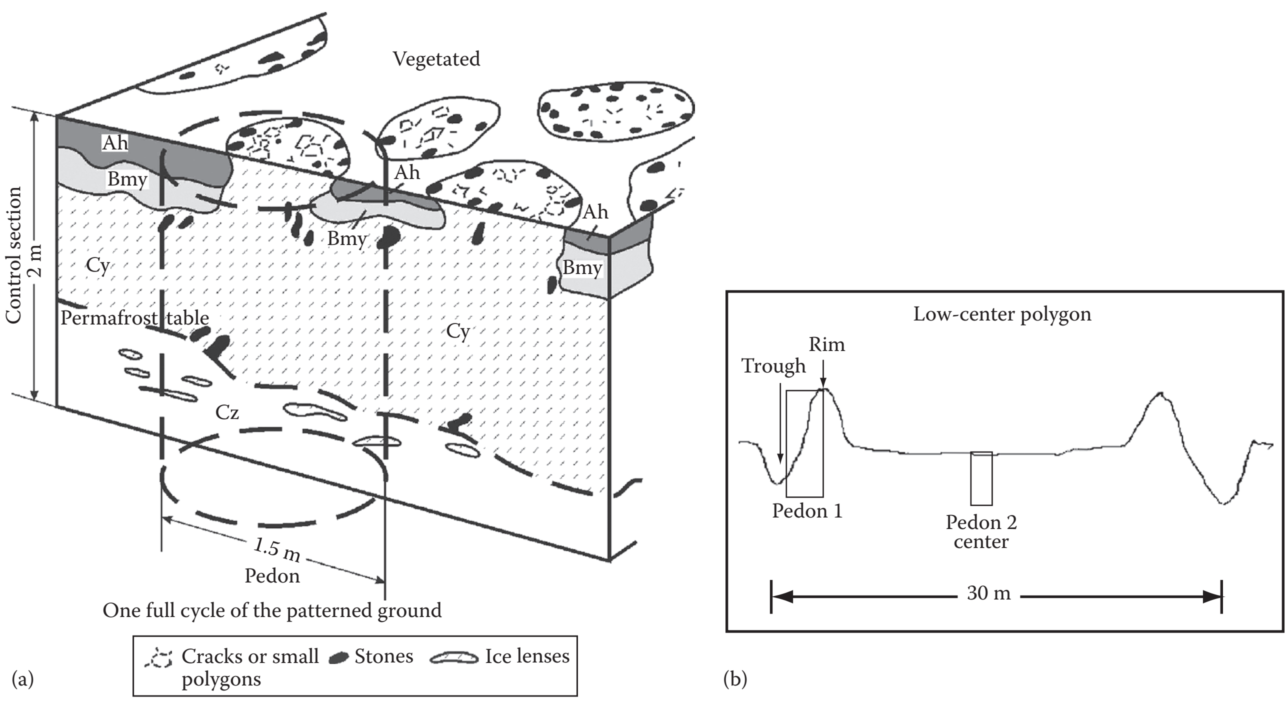

FIGURE 33.34 The pedon concept as it applies to Gelisols and patterned ground form, including (a) a nonsorted circle (From Tarnocai, C. and Smith, C.A.S., The formation and properties of soils in the permafrost regions of Canada, in: D.A. Gilichinsky (ed.), *Proceedings of the First International Conference on Cryopedology*, Russian Academy of Science, Pushchino, Russia, 1992. With permission.) and (b) a large-scale, low-centered polygon. (From Bockheim, J.G., L.R. Everett, K.M. Hinkel, F.E. Nelson, and J. Brown. 1999. Soil organic carbon storage and distribution in arctic tundra, Barrow, Alaska. Soil Sci. Soc. Am. J. 63:934–940. With permission.) The Canadian soil horizon nomenclature is used, where Ah = A, Bmy = Bwjj, Cy = Cjj, and Cz = Cf in *Soil Taxonomy*.

Whereas permafrost-affected soils are identified as Gelisols (from the Greek word, gelid, meaning very cold) in *Soil Taxonomy*, they are called Cryosols (from the Greek word, kraios, meaning cold or ice) in the other two systems. The Canadian and U.S. systems divide permafrost-affected soils into three categories: organic soils, cryoturbated mineral soils, and other mineral soils. The soils are further delineated on the basis of key properties that link them with soils of lower latitudes. In the WRB, the Cryosol soil group is divided into 16 soil units. Fifteen suffix qualifiers are used to link Cryosols with soils of lower latitudes.

33.6.7 Properties of Gelisols

Gelisols/Cryosols encompass a vast array of soils in terms of chemical and mineralogical properties. However, with the exception of the Histels, Gelisols are uniform in terms of physical and thermal properties.

33.6.7.1 Macromorphology

The most common macroscopic soil features are due to cryoturbation and include irregular or broken horizons and incorporation of organic matter in lower horizons, especially along the top of the permafrost. Oriented stones and displacement of soil materials are common in Gelisols. Freezing and thawing produce granular and platy structures in surface horizons and blocky, prismatic or massive structures in subsurface horizons. The massive structure is due to cryostatic pressure and desiccation that develop when the two freezing fronts, one from the surface and the other from the permafrost, merge during freeze back in the autumn. The perennially frozen layer commonly contains ground ice in the form of segregated ice crystals, vein ice, ice lenses and wedges, and thick ground ice.

The granular, platy, or blocky structures of the surface mineral horizons are also the result of cryopedogenic processes such as the freeze–thaw process and vein-ice formation (ice segregation process). The subsurface horizons often have massive structures and are associated with higher bulk densities, especially in fine-textured soils.

Almost all Gelisols contain ice in the form of crystals, lenses, layers (vein ice), wedges, or massive ground ice, often to a thickness of several meters. Soil texture is one of the factors controlling ice content in mineral soils with fine-textured Gelisols generally having higher ice contents than coarse-textured soils. Coarse-textured soils often have a relatively low ice content; however, they may contain ice wedges in the form of polygons. Histels have an ice content of 60%–90% on a volume basis (Zoltai and Tarnocai, 1975).

The active layer that is subject to annual thawing and refreezing lies above the permafrost and not only supports biological life, but also protects the underlying permafrost. The thickness of the active layer is controlled by soil texture and moisture, thickness of the surface organic layer, vegetation cover, aspect, and latitude.

Dilitancy, often confused with thixotropy, is common in soils with high silt content, greatly affects the trafficability of the soil, and is frequently present during the thaw period. Dilitancy is a property of granular masses expanding due to the increase of space between rigid particles upon displacement of the particles. When dilitant soils dry out, a characteristic vesicular porosity develops. Salt crusts, patches, and pans are common in polar desert soils of the high arctic and cold desert soils of Antarctica (Tedrow, 1977; Bockheim, 1997).

33.6.7.2 Micromorphology

When viewed in thin sections, Gelisols contain a variety of fabrics resulting from desiccation and displacement due to alignment, rotation, sorting, and inclusions (Figure 33.33). These features are accompanied by planar voids, cracks, and vesicles during pore formation. More specifically, the fabric of Gelisols varies from granular (granic and granoidic) in the surface horizons to mainly porphyroskelic with fragmic and fragmoidic components in subsurface mineral horizons (Pawluk and Brewer, 1975; Fox, 1985; Smith et al., 1991; Schaefer et al., 2008). The micromorphology of Gelisols can show evidence of matrix displacement and movement, with resultant reorganization of skeleton grains into circular or elliptical patterns, producing an orbicular fabric (Fox and Protz, 1981). Ice lensing and vein ice development lead to the formation of lenticular fabrics, while cryoturbation and cryodesiccation can lead to granic or granoidic fabrics (Smith et al., 1991). In addition, suscitic and conglomeric fabrics also are common in Gelisols (Fox and Protz, 1981).

33.6.7.3 Thermal Characteristics

Soil temperatures for a Gelisol located north of the arctic treeline in Canada are shown in Figure 33.35. The unique thermal signature that separates Gelisols from all other soils is the presence of a perennially frozen layer, usually below 50 cm. Because of this frozen layer, Gelisols have a steep vertical temperature gradient. If these soils are associated with certain types of patterned ground, the horizontal temperature gradient can also be large. For example, in the case of Gelisols associated with earth hummocks, the soil temperature at the center of the hummock can decrease from 12°C at the surface to 0°C at 50 cm during the summer months. Soil temperatures at comparable depths under the interhummock depression, <1 m away and at equivalent depths, can be 5°C–7°C lower (Tarnocai and Zoltai, 1978).

33.6.7.4 Physical and Chemical Characteristics

Histels contain weakly to well-decomposed moss, sedge, woody, and amorphous organic material. The pH of these soils ranges from 2.5 to 7.0 (Table 33.28). Histels derived from moss peats (Fibristels) tend to have lower pH values than other organic soils with permafrost. Turbels and Orthels developed from calcareous parent materials have a pH_{CaCl_2} of 7 and a high base saturation. On the other hand, where these soils are developed from acidic parent materials, pH_{CaCl_2} < 5.5.

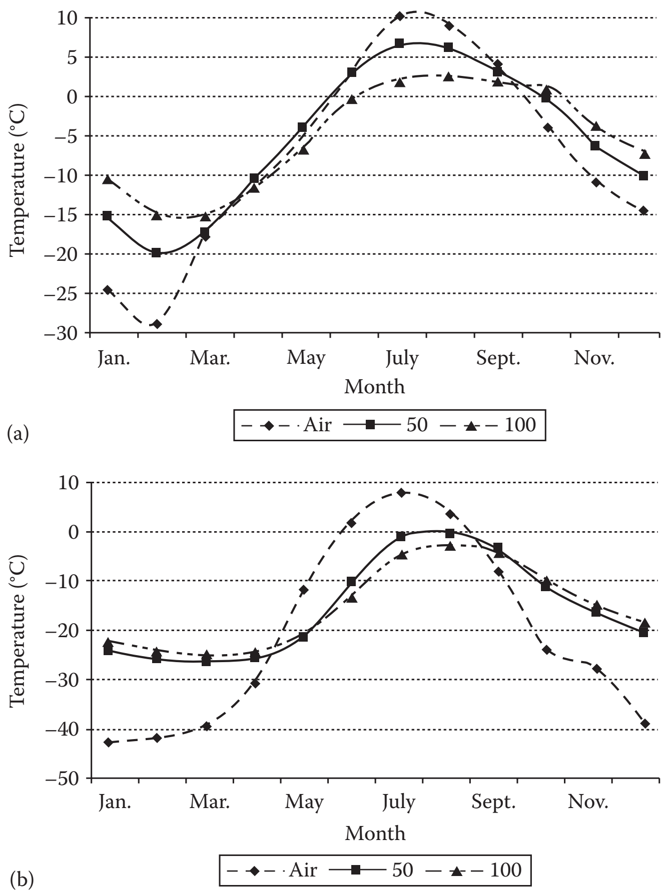

FIGURE 33.35 Soil temperature variation in Gelisols from (a) Auyuittuq National Park, Baffin Island, Canada (66°23′30″N, 65°29′20″W) and (b) Lake Hazen, Quttinirpaaq National Park, Ellesmere Island, Canada (81°49′15″N, 71°33′17″W) during 1999. (From Tarnocai, C. 2008. Arctic permafrost soils, p. 3–16. *In* R. Margesis (ed.) Permafrost soils. Soil Biology Series. Vol. 16. Springer-Verlag, Berlin, Heidelberg, Germany. With permission.)

Anhyturbels and Anhyorthels in polar and cold deserts often have high salt contents and electrical conductivities. Although these soils have many features common to Aridisols (Section 33.10), they have patterned ground, dry permafrost in the upper meter over ice-cemented permafrost, and show evidence of thermal cracking (Bockheim, 1997).

Gelisols, especially the Histels and Turbels, often contain large amounts of organic C in the upper m, with values commonly ranging from 3 kg C m^{-3} in Haplorthels to over 100 kg C m^{-3} in Sapristels and averaging about 50 kg C m^{-3} (Bockheim et al., 1999, 2004; Tarnocai et al., 2007). This is similar to Histosols of other regions (Section 33.2). Less than half of the organic C in Turbels is in the active layer, with the remainder in the upper 30–50 cm of permafrost (Bockheim et al., 1997). Much of this organic carbon exists below 1 m, the depth to which carbon stores are normally reported.

The particle size distribution of Gelisols varies from clayey to coarse gravelly sand. The composition of the fine-earth fraction is commonly dependent on the composition and age of the parent materials.

33.6.8 Special Problems in Managing Gelisols

Gelisols present special problems in terms of management, not only because of frost churning, heaving, sorting, and cracking, but also because melting of segregated ice following a disturbance to the thermal regime leads to subsidence, or thermokarst. To preserve the integrity of structures (buildings, roads, and pipelines) in permafrost soil, it is important to maintain the negative thermal balance of the soil. This is achieved by using special construction methods. For agricultural development, it is important to determine the ice content of the soil; otherwise, after clearing of the land or within a few years after cultivation begins, severe subsidence and thermokarst can develop (Péwé, 1982).

33.6.9 Global Warming and Trace Gas Emissions

General circulation models (GCMs) predict that with a projected twofold increase in atmospheric CO_2 by the year 2050, the mean air temperature of the Earth's surface could increase by 1.5°C–4.5°C (Maxwell and Barrie, 1989); however, warming at the high northern latitudes could be on the order of 4°C–5°C. Indeed, sea ice variations over the past several decades are compatible with a distinct warming of air temperatures in the arctic, especially during the winter and spring (Arctic Climate Impact Assessment, 2004). In addition, permafrost temperatures in northernmost Alaska have increased by 2°C–4°C during the last few decades (Lachenbruch and Marshall, 1986).

As mentioned previously, Gelisols are large C sinks. There is concern that warming in the circumpolar regions could increase the thickness of the active layer and enhance heterotrophic respiration, releasing additional CO_2 to the atmosphere (Oechel and Billings, 1992; Shaver et al., 1992; Waelbroeck et al., 1997), and Gelisols would become a C source. However, Bockheim (2007) has pointed out that continued warming in the arctic could accelerate cryoturbation and enable some soils to store more organic C at depth than at present, thereby mitigating some of the loss of CO_2 to the atmosphere from increased soil respiration.

33.6.10 Human-Caused Disturbances

Arctic regions contain vast energy reserves, including fossil fuels (coal, oil, and gas), biomass, and hydropower. The extraction of fossil fuels and minerals and deforestation may have dramatic long-term effects on arctic ecosystems as they have a low resistance and resilience to disturbance (Reynolds and Tenhunen, 1996). As the world's population continues to expand, there will be increased pressure for development at the high latitudes. For example, Siberia already contains several cities that were developed in permafrost and have in excess of 500,000 inhabitants.

The circumarctic contains numerous indigenous people who are dependent on the terrestrial and marine ecosystems for

TABLE 33.28 Analytical Characteristics of Selected Gelisols

Horizon	Depth (cm)	pH	Bulk Density (mg m^{-3})	C (%)	N (%)	Clay (%)	Silt (%)	Sand (%)	Extr. P (mg L^{-1})	Ca (cmol$_c$ kg^{-1})	Mg (cmol$_c$ kg^{-1})	Na (cmol$_c$ kg^{-1})	K (cmol$_c$ kg^{-1})	Exch. Acid (cmol$_c$ kg^{-1})	Exch. Al (cmol$_c$ kg^{-1})	CEC NH_4OAc (cmol$_c$ kg^{-1})	Base Sat. (%)
Typic Haploturbel; Dry Acidic Tundra (Empetrum–Betula–Dryas–Arctostaphylos); Till/Congelifractate; 68°37′N, 149°18′W																	
Ajj	2–46	4.91	0.56	8.68	0.53	9.8	53.2	37.0	0.7	2.4	1.2	TR	0.1	39.4	11.5	36.1	10
Bwjj	4–40	4.83	0.87	2.3	0.13	18.4	46.2	35.4	1.2	0.7	0.8	TR	TR	25.2	9.1	20.5	8
BCjj	0–34	5.14	0.68	9.31	0.26	11.0	55.3	33.7	3.5	0.7	0.4	0.1	0.1	34.5	9.7	2.6	5
C	8–90	6.94	ND	0.58	0.05	5.2	44.4	50.4	5.4	7	1.2	TR	0.1	3.1	0	10.8	77
Typic Molliturbel; Dry Nonacidic Tundra (Carex–Dryas–Tomentypnum); Loess/Outwash; 70°11′N, 149°17′W																	
A	1–21	6.93	ND	5.66	0.37	14.0	26.8	59.1	0.1	61.3	2.6	TR	0.1	11.2	0	69.3	92
Oa/Ckjj	21–42	7.55	ND	15	0.96	16.0	16.1	67.9	0.5	47.8	1.2	0.1	0.1	3.6	0	30.3	100
2 Ck	42–61	7.88	ND	2.37	0.04	5.7	6.1	88.2	0.1	35.4	0.8	0	TR	0	0	3.3	100
Ruptic–Histic Aquiturbel; Moist Nonacidic Tundra (Dryas–Salix–Eriophorum); Loess/Outwash; 69°27′N, 148°40′W																	
Oi	0–8	8.37	0.12	35.7	1.25				15.2	84.1	15.4	0.2	1.5	16	0.1	113	89
Oejj1	0–40	7.81	0.20	24.2	1.25				2.1	84.3	6.8	0.3	0.2	11.6	0.1	102	89
Bw	0–12	8.01	1.07	3.55	0.24	19.8	70.0	10.2	0.0	47.1	3.1	TR	0.1	0.8	0	22.1	100
Bg	0–88	7.92	1.19	3.8	0.29	23.5	66.9	9.6	0.5	36.3	2	TR	0.2	3	0	24.3	100
Oejj2	0–50	7.55	0.43	20.12	1.29				1.1	81.1	5.1	0.1	0.3	13.7	0.1	97.8	89
2Ajj	0–85	7.73	1.09	7.27	0.39	16.7	69.3	14.0	1.8	55.1	2.3	TR	0.2	3.7	0	29.5	100
2Cd/Oejjfm	65–85	7.40	ND	10.93	0.74	18.1	70.1	11.8	3.0	69.5	3.8	TR	0.2	7.4	0	50.8	100
Typic Umbrorthel; Moist Acidic Prostrate Shrub–Grass Tundra (Salix–Carex–Polytrichum); Residuum; 68°46′N, 149°35′W																	
A	0–8	4.21	0.57	7.13	0.48	3.8	15.6	80.6	10.3	1.9	1.2	0	0.3	21.4	3.3	18.1	19
Bw	8–22	4.05	1.30	2.05	0.12	5.2	15.9	78.9	1.8	0.5	0	TR	TR	13.9	3.9	9.2	7
BC	22–31	4.64	1.38	0.44	0.05	1.2	11.9	86.9	7.5	0.2	0	0	TR	4.6	1.5	3	7
Typic Histoturbel; Moist Acidic Tussock Tundra (Eriophorum–Sphagnum–Betula); Colluvium; 68°37′N, 149°19′W																	
Oi	0–9	4.46	0.04	44.34	0.85				34.0	11.5	6.3	0.8	3.1	110	2.5	136	16
Oe1	9–17	5.06	0.10	39.8	1.40				2.4	9.3	2.4	0.3	0.4	93.5		104	12
Oe2	17–23	5.37	0.37	24.3	1.34				1.2	3.9	1.6	0.1	0.3	63.2		59.9	10
Bg	23–40	5.17	1.46	3.04	0.11	20.0	41.9	38.1	0.4	0.5	0.4	TR	TR	14.7	4.2	15.8	6
BCg	40–48	5.70	1.09	3.46	0.14	20.5	39.8	39.7	0.6	0.5	0.4		TR	15	3.9	14.7	6
Cg/Oefm	48–80	6.39	ND	4.24	0.19	19.0	38.9	42.1	1.5	0.7	0.4	TR	TR	16	3.3	14	9
Typic Fibristel; Wet Acidic Tundra (Eriophorum–Sphagnum); Organic Basin Deposits; 68°37′N, 149°19′W																	
Oi1	0–13	4.48	0.03	43.56	0.65				10.4	1.9	0.8	0.1	0.3	57.2	4.1	109	3
Oi2	13–33	4.52	0.19	44.82	2.58				4.0	8	2.6	0.1	0.2	107	4.8	77.1	14
Oe	33–44	4.66	0.35	25.2	1.17				0.6	3.4	0.7	TR	TR	75.2	6.9	51.3	8
Oa	44–62	4.99	0.52	18.5	1.02				0.5	3.6	0.7	TR	TR	58.2	9	46.6	9

Source: Soil Survey Staff. 2011. National Cooperative Soil Characterization Database. Available Online at http://ssldata.nrcs.usda.gov.Accessed July 8, 2011.

food, fuel, and shelter. Special efforts must be made to ensure that these ecosystems remain sustainable. The polar regions are less diverse biologically than other life zones. However, ancient (~2.5 million years old) permafrost contains viable microorganisms that may give clues to the evolution of microbes (Gilichinsky, 1993).

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33.7 Vertisols

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33.7.1 Introduction

Vertisols are clayey soils that exhibit significant volume change due to shrink and swell processes as the soil dries and wets. Most Vertisols have smectitic mineralogy, but Vertisols with mixed or kaolinitic mineralogy also occur. The intrinsic shrink/swell behavior is the dominant soil-forming process of Vertisols and results in significant close-range spatial and temporal variability of soil properties, possibly more so than in any other soil order. As a result, Vertisols are often difficult to manage for agricultural, natural resource, and engineering applications. A substantial body of information about Vertisols exists in the literature, including summaries (e.g., Coulombe et al., 2000; Blokhuis, 2002), and more in-depth discussions about specific aspects of Vertisols (e.g., Ahmad, 1983; Probert et al., 1987; Wilding and Puentes, 1988; Kimble, 1991; Ahmad and Mermut, 1996; Coulombe et al., 1996b).

33.7.2 Distribution and Formation

Geographically, Vertisols occur in more than 100 countries and occupy 309 Mha (10^6 ha), which corresponds to about 2% of the ice-free land area (Table 33.1; Figure 33.36; USDA-NRCS, 2009). Globally, Vertisols are most extensive in India (80 Mha), Australia (70 Mha), and Sudan (50 Mha), followed by the United States (18 Mha), Ethiopia (13 Mha), and China (12 Mha) (Dudal and Eswaran, 1988; Isbell, 1991; Blokhuis, 2002).

In the United States, Vertisols are reported in 25 states and territories (Coulombe et al., 1996b, 2000), primarily west of the Mississippi River and in the Black Belt region of the southern states. Vertisols are most extensive in Texas (6.5 Mha), South Dakota (1.5 Mha), California (1 Mha), and Montana (0.6 Mha). The remaining states each have less than 0.25 Mha of Vertisols and vertic intergrades.

Vertisols are formed from a wide variety of parent materials of varying age and under a wide range of climatic conditions (Coulombe et al., 1996a, 1996b). A common attribute is a soil-forming environment that favors preservation of silica and basic cations and that produces episodic or periodic wetting and drying to allow cracks to form. Most Vertisols occur

FIGURE 33.36 Global distribution of Vertisols. (Courtesy of USDA-NRCS, Soil Survey Division, World Soil Resources, Washington, DC, 1998.)

on shallow slopes in uplands and on a variety of alluvial landforms. Natural vegetation of Vertisols is typically grassland and savanna, probably because most woody plant roots cannot survive shearing during wetting and drying, but mixed/deciduous forested and scrub/shrub can occur, depending on the ecological succession of the vegetation and human impact.

33.7.3 Morphology

33.7.3.1 Color

In general, Vertisols are dark in color, particularly in the upper horizons. Typically, they exhibit a moist Munsell value of 3 or less and chroma of 2 or less in the matrix, and have hues of 7.5YR or yellower. The dark colors are generally attributed to the close association of organic matter with clay minerals and to various compounds of Mn, Fe, and Ti (Ahmad and Mermut, 1996). Vertisols with black, gray, brown, or red pigmentation have been reported in various regions of the world. In the most recent *Keys to Soil Taxonomy* (Soil Survey Staff, 2010), dark Vertisols are considered the norm; soil color differentiates Chromic subgroups, which require a value of 4 or more (moist), or 6 or more (dry), or a chroma of 3 or more. The chroma requirement does not apply to the subgroup of Aquerts. Lighter soil colors are usually associated with the accumulation of calcium carbonate, gypsum, or more soluble salts, especially in subsurface horizons. Reddish redox concentrations can occur in poorly drained Vertisols; reddish matrix colors and high chroma are often associated with Vertisols on better drained landscapes, and in less humid climates, and possibly in Vertisols formed from parent material with low Mn content.

33.7.3.2 Texture

Vertisols are fine-textured soils. In general, Vertisols must have at least 30% clay in the fine-earth fraction in the upper 50 cm or to a densic, lithic, or paralithic contact, a duripan, or a petrocalcic horizon if any of those features is shallower (Soil Survey Staff, 2010). Clay content of Vertisols can be as high as 90%, particularly for those derived from pyroclastic deposits. Soils that do not meet the clay content requirements, or that do not exhibit other required properties (slickensides or wedge-shaped peds, and cracks, described below), but that exhibit considerable shrink/swell activity as measured by the coefficient of linear extensibility (COLE), often are classified as vertic intergrades to other soil orders. Particle size distributions for selected Vertisols are shown in Table 33.29.

TABLE 33.29 Particle Size Distribution of Selected Vertisols Formed from a Wide Variety of Parent Materials

Regions	% Sand 2.00–0.50 mm	% Silt 0.50–0.002 mm	% Clay <0.002 mm	References
Australia	16.0 (2.5)	33.6 (3.7)	50.3 (5.5)	Boettinger (1992)
India	13.8 (3.7)	21.9 (2.6)	61.0 (5.2)	Hirekurubar et al. (1991)
Sudan	16.8 (8.4)	19.9 (10.1)	63.1 (9.7)	Blokhuis (1993)
Texas	9.6 (7.3)	36.3 (6.9)	53.5 (3.7)	Kunze et al. (1963), Yule and Ritchie (1980a), and Hallmark et al. (1986)
West Africa	18.4 (11.6)	27.7 (8.0)	54.0 (15.4)	Beavington (1978) and Yerima et al. (1988)
California	3.4 (0.4)	34.6 (1.2)	62.0 (1.2)	NSSL Pedon No. 94P0056
El Salvador	10.8 (4.0)	31.2 (7.3)	58.1 (8.5)	Yerima et al. (1985)
West Indies	17.5 (9.7)	10.6 (3.0)	74.9 (11.4)	Ahmad and Jones (1969)
Uruguay	26.3 (5.6)	27.0 (8.1)	46.7 (13.3)	Lugo-Lopez et al. (1985)

Data are reported as mean (sd).

33.7.3.3 Differentiating Physical Properties and Features

In addition to the clay content requirements, Vertisols must exhibit significant shrink/swell characteristics to develop diagnostic soil structure and features in the soil profile. Evidence of shrink/swell behavior includes slickensides or wedge-shaped aggregates with the long axis inclined at 10°–60° from the horizontal (Soil Survey Staff, 2010). Slickensides are more or less planar features that exhibit shiny and grooved surfaces at the interface of the peds (Figure 33.37) that indicate soil movement and shearing. The presence of slickensides in a soil horizon is typically indicated by the morphological horizon designation, Bss. Wedge-shaped aggregates are soil structural units that generally result from limited formation of slickensides bounding ped surfaces (Figure 33.38). The presence of cracks (Figure 33.39) that open and close in response to drying and wetting is also a requirement for Vertisols (Soil Survey Staff, 2010).

Gilgai and diapir are other features that may be present in Vertisols. Gilgai (Figure 33.40) is an Australian aboriginal term used to describe microtopography (Figures 33.41 and 33.42) that

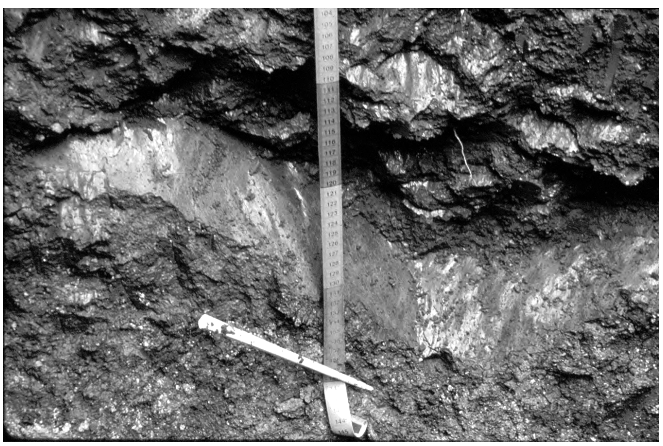

FIGURE 33.37 The glossy slickensides show evidence of subsoil shrink-swell and shearing. (Photograph from Eswaran et al., 1999).

FIGURE 33.38 Wedge-shaped peds are a result of subsoil shearing in many Vertisols. The bar represents 5 cm. (Photograph courtesy of Susan B. Southard, USDA-NRCS).

FIGURE 33.39 Cracks that open and close periodically are a required morphological feature of Vertisols. These cracks range from about 1 to 6 cm wide. (Photograph from Eswaran et al., 1999).

FIGURE 33.40 Vertisol microtopography (gilgai) is especially evident during the wet season. (Photograph from Eswaran et al. 1999).

FIGURE 33.41 Cross-section of a Vertisol exposed in a trench. Darker-colored bowls, or microlows, are evident on the left and right sides of the exposure. A lighter-colored diapir, or microhigh, is visible between the bowls to the right of the tape measure. Strings have been used to delineate regions of distinctive morphology. The tape measure is 180 cm long (Photograph from Eswaran et al., 1999).

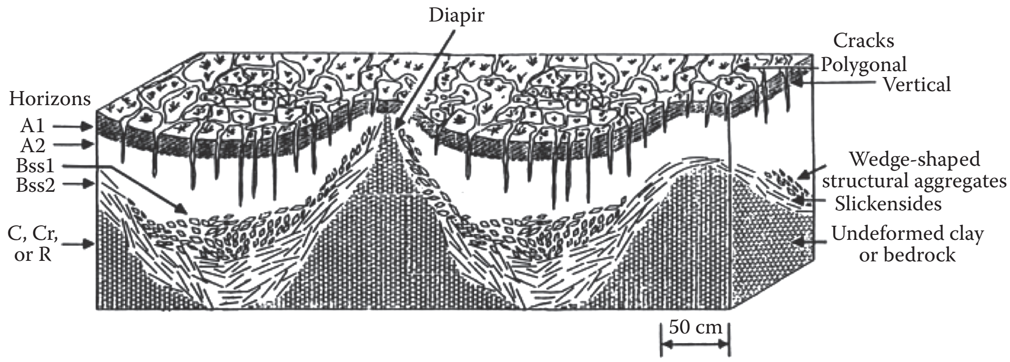

FIGURE 33.42 Schematic representation of Vertisol morphology and horizon arrangement. (From Coulombe, C.E., L.P. Wilding, and J.B. Dixon. 1996b. Overview of Vertisols: Characteristics and impacts on society. Adv. Agron. 57:289–375. With permission.)

consists of mounds (microhigh), shelves (intermediate), and depressions (microlow), either in circular, linear, or complex patterns. Prior to 1992, *Soil Taxonomy* used gilgai as a differentiating characteristic of Vertisols, but, that criterion was eliminated because not all Vertisol landscapes have gilgai. Diapir, also called mukkara in Australia, in this context was adopted from geology terminology, where the term is used to describe large-scale intrusions of ductile material (often mud or salt) into more rigid, surrounding rock. In Vertisols, a diapir is a smaller-scale intrusion of soil material that penetrates from the subsoil through the upper layers and can be identified by a contrast in color and/or texture. Diapirs may occur independent of gilgai; however, when gilgai are present, the tip of the diapir coincides with the mound or microhigh.

33.7.4 Genesis

The dominant processes that lead to the characteristic properties of Vertisols is shrinking and swelling induced by changes in soil water conditions. The shrink/swell action can be observed from macroscopic to submicroscopic scales of resolution (Figure 33.43). The shrink/swell phenomena are clearly related to the changes in the interlayer spacing of expansible phyllosilicate minerals, principally smectites, by sorption and desorption of interlayer water. However, changes in the interlayer spacing of smectites contribute only 10%–30% of the change in volume when saturated with Ca (Greene-Kelly, 1974; Tessier, 1984). Tessier (1984) reported that the type of microstructure, for example, crystallite, domains, or quasicrystals (Figure 33.43) and the surface area of the clay mineral phases govern the shrink/swell potential. The microstructure and the porosity accommodate the changes in water content and potential of the clay-water systems. Other factors affecting shrink/swell phenomenon include particle rigidity, stress history, predominance of cations present on the exchange sites (e.g., Na vs. Ca), the electrolyte concentration of the soil solution, and the amount and location of the charge in the clay mineral crystallographic structure, which may cause variation in the structural organization and the interlayer distance among smectites.

Tessier (1984), Quirk (1994), and Coulombe et al. (1996a, 1996b) extensively reviewed the models and factors involved in shrink/swell processes. According to Coulombe et al. (1996b), the application of the Coulomb–Mohr theory of shear failure constitutes the best model to account for the formation and occurrence of features and structure in Vertisols. Shear failure is more likely to occur under moist or wet than dry conditions, since a lower normal stress is required to promote shear failure. Even though shrink/swell processes are a common phenomenon among Vertisols, other pedogenic processes also occur, such as accumulation of salts (e.g., salic, natric, calcic, and gypsic great groups), silica (duric great group), and depth of saturation by water (epi and endo great groups).

33.7.5 Classification

Vertisol is the widely accepted term to designate fine-textured soils with extensive shrink/swell behavior. The term originates from *vertere* (L.), meaning to churn or turn over and *sol* for soils (Dudal and Eswaran, 1988). Coulombe et al. (1996b) discussed the history and evolution of the Vertisol order in the U.S. soil classification systems. Vertisols were first introduced as a soil order in the *7th Approximation* (Soil Survey Staff, 1960), and differentiating criteria were subsequently modified in *Soil Taxonomy* (Soil Survey Staff, 1975, 1999). Prior to the *7th Approximation*, most Vertisols were identified as Grumosols, Regosols, Rendzina, or Alluvial soils (Soil Survey Staff, 1975). In 1960, Vertisols were the first order in the key and were differentiated on the basis of clay content, cation exchange capacity, cracks, and evidence of shrink–swell in the form of gilgai, slickensides, or wedge-shaped aggregates. Two suborders, Aquerts and Usterts, were identified, and an important property for identifying great groups was the presence of a self-mulching surface horizon (Figure 33.44; "Grumic" great groups, e.g., Grumaquerts) or of a platy or massive surface crust ("Mazic" great groups, e.g., Mazusterts). In the first edition of *Soil Taxonomy* (Soil Survey Staff, 1975), Vertisols were placed as the fourth order in the key, subgroups keyed in the sequence: Xererts, Torrerts, Uderts, and Usterts (no Aquerts), and great groups were identified on the basis of soil color, not grumic or mazic properties (e.g., Chromoxererts, with moist soil chroma ≥1.5 in the upper 30 cm, otherwise Pelloxererts). Revisions based on the efforts of ICOMERT (the International Committee on the Classification of Vertisols) added the Aquerts

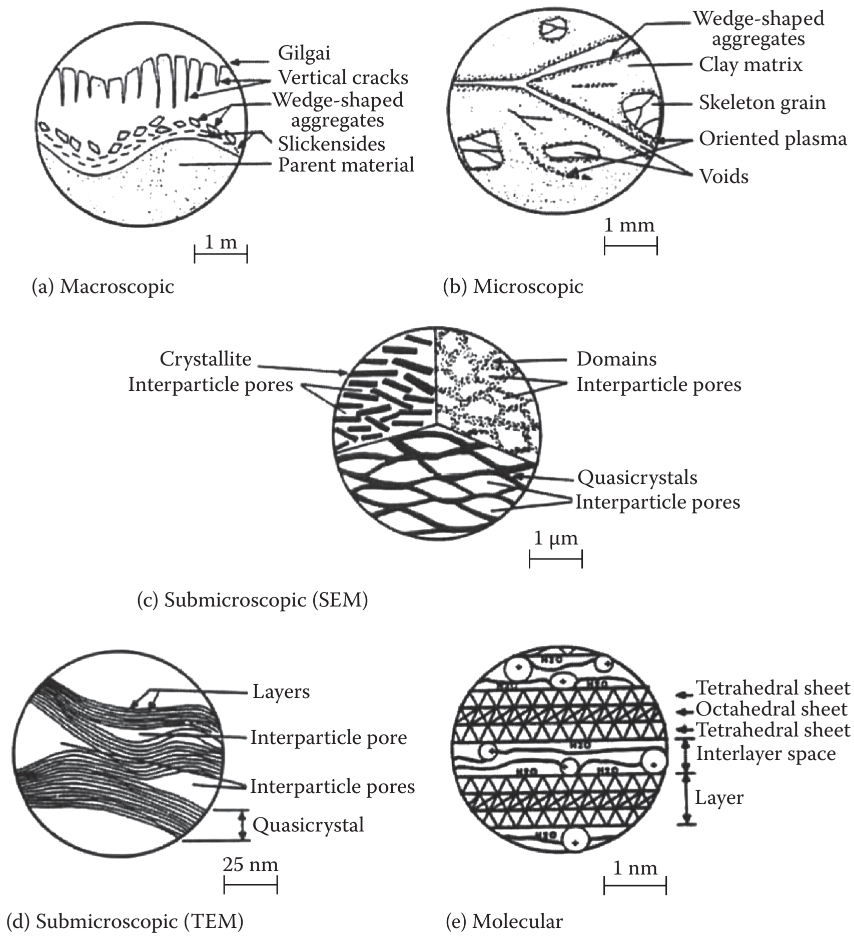

FIGURE 33.43 Shrink-swell features depicted at various scales of resolution. (From Coulombe, C.E., L.P. Wilding, and J.B. Dixon. 1996b. Overview of Vertisols: Characteristics and impacts on society. Adv. Agron. 57:289–375. With permission.)

FIGURE 33.44 Vertisol surface mulch of very fine and fine angular blocks and granules.

and Cryerts suborders and included some of the subsurface diagnostic horizons as criteria for differentiation of great groups (Soil Survey Staff, 1999).

The Vertisol order currently keys out as the sixth order, based on the following (Soil Survey Staff, 2010):

Other soils that have:

1) A layer 25 cm or more thick, within 100 cm of the mineral soil surface, that has *either* slickensides *or* wedge-shaped peds that have their long axes tilted 10 to 60 degrees from the horizontal; *and* 2) A weighted average of 30 percent or more clay in the fine-earth fraction either between the mineral soil surface and a depth of 18 cm or in an Ap horizon, whichever is thicker, *and* 30 percent or more clay in the fine-earth fraction of all horizons between a depth of 18 cm and either a depth of 50 cm or a densic, lithic, or paralithic contact, a duripan, or a petrocalcic horizon if shallower; *and* 3) Cracks that open and close periodically.

The order is subdivided into 6 suborders, 24 great groups, and 158 subgroups (Soil Survey Staff, 2010; Table 33.30). The suborders are identified on the basis of aquic conditions, cryic soil temperature regime, and the frequency and duration of soil cracking. The cracking criterion allow inferences about the soil moisture regime (as implied by the suborder names), but soil moisture regimes in Vertisols are not well characterized due to cracking and slow permeability when wet, leading to heterogeneous wetting and drying of the soils and preferential bypass flow. Because

TABLE 33.30 Listing of Suborders, Great Groups, and Subgroups for the Vertisol Order

Suborders	Great Groups	Subgroups
Aquerts	Sulfaquerts	Salic, Sulfic, Typic
	Salaquerts	Aridic, Ustic, Leptic, Entic, Chromic, Typic
	Duraquerts	Aridic, Xeric, Ustic, Aeric, Chromic, Typic
	Natraquerts	Typic
	Calciaquerts	Aeric, Typic
	Dystraquerts	Sulfaqueptic, Aridic, Ustic, Aeric, Leptic, Entic, Chromic, Typic
	Epiaquerts	Halic, Sodic, Aridic, Xeric, Ustic, Aeric, Leptic, Entic, Chromic, Typic
	Endoaquerts	Halic, Sodic, Aridic, Xeric, Ustic, Aeric, Leptic, Entic, Chromic, Typic
Cryerts	Humicryerts	Sodic, Typic
	Haplocryerts	Sodic, Chromic, Typic
Xererts	Durixererts	Halic, Sodic, Aquic, Aridic, Udic, Haplic, Chromic, Typic
	Calcixererts	Lithic, Petrocalcic, Aridic, Leptic, Entic, Chromic, Typic
	Haploxererts	Lithic, Halic, Sodic, Aridic, Aquic, Udic, Leptic, Entic, Chromic, Typic
Torrerts	Salitorrerts	Aquic, Leptic, Entic, Chromic, Typic
	Gypsitorrerts	Chromic, Typic
	Calcitorrerts	Petrocalcic, Leptic, Entic, Chromic, Typic
	Haplotorrerts	Halic, Sodic, Leptic, Entic, Chromic, Typic
Usterts	Dystrusterts	Lithic, Aquic, Aridic, Udic, Leptic, Entic, Chromic, Typic
	Salusterts	Lithic, Sodic, Aquic, Aridic, Leptic, Entic, Chromic, Typic
	Gypsiusterts	Lithic, Halic, Sodic, Aridic, Udic, Leptic, Entic, Chromic, Typic
	Calciusterts	Lithic, Halic, Sodic, Petrocalcic, Aridic, Udic, Leptic, Entic, Chromic, Typic
	Haplusterts	Lithic, Halic, Sodic, Petrocalcic, Gypsic, Calcic, Aridic Leptic, Aridic, Leptic Udic, Entic Udic, Chromic Udic, Udic, Leptic, Entic, Chromic, Typic
Uderts	Dystruderts	Aquic, Oxyaquic, Leptic, Entic, Chromic, Typic
	Hapluderts	Lithic, Aquic, Oxyaquic, Leptic, Entic, Chromic, Typic

Source: Soil Survey Staff. 2010. Keys to soil taxonomy. 11th edn. USDA-NRCS, National Soil Survey Center, Lincoln, NE.

of low permeability and water conductance via bypass flow along macrovoids, Vertisols rarely saturate in their entirety (Jacob et al., 1997; Nordt and Driese, 2009). The great group categories reflect additional dominant soil processes and resulting diagnostic horizons, and subgroups represent intragrade or intergrade concepts toward other soil orders or categories. Vertisols are not required to have a specific diagnostic horizon, although many have mollic epipedons and cambic horizons.

Vertisols are also recognized in other international soil classification systems such as FAO–UNESCO (1988, 1998) and the World Reference Base (WRB) for Soil Resources (IUSS Working Group WRB, 2007) and the French systems (CPCS, 1987; Baize and Girard, 1996). Some variations in the criteria used to differentiate Vertisols may occur from one classification system to another, but the *Soil Taxonomy* criteria for Vertisols (no vertic horizon) and the following WRB criteria (IUSS Working Group WRB, 2007) for a vertic horizon and for identification of Vertisols are essentially the same.

33.7.5.1 Vertic Horizon

The vertic horizon (from Latin vertere, to turn) is a clayey subsurface horizon that, as a result of shrinking and swelling, has slickensides and wedge-shaped structural aggregates. Diagnostic criteria (1) contains 30% or more clay throughout; (2) has wedge-shaped structural aggregates with a longitudinal axis tilted between 10° and 60° from the horizontal; (3) has slickensides; and (4) has a thickness of 25 cm or more.

Vertisols key out as the sixth Reference Soil Group as follows: Other soils having (1) a vertic horizon starting within 100 cm of the soil surface; and (2) after the upper 20 cm have been mixed, 30% or more clay between the soil surface and the vertic horizon throughout; and (3) cracks that open and close periodically.

33.7.6 Mineralogical Properties

Most Vertisols are dominated by smectitic mineralogy. However, Vertisols containing abundant kaolinite, illite, halloysite, hydroxy-interlayered smectite, or a combination of interstratified clay minerals are also found in various regions of the world (Coulombe et al., 1996a). Recognition of Vertisols with these mineralogical characteristics has caused geoscientists to reconsider the Vertisol paradigm and the mechanisms involved in the shrink/swell processes of soils that are not dominated by smectite. As discussed earlier, shrink/swell behavior includes expansion and collapse of interlayer space, but also involves changes in microstructure and porosity associated with changes in water content and potential (Tessier, 1984; Wilding and Tessier, 1988; Coulombe et al., 1996a).

The smectites that dominate most Vertisols tend to concentrate in the fine clay fraction and contribute to high particle surface area and shrink–swell potential. Characteristics of smectites in Vertisols, which may also apply to other clay minerals and other soil orders, include (1) relatively high thermodynamic stability, (2) high layer charge (>0.45 per half unit cell), which may be as high as vermiculite or illite and may impact the availability of some ions such as K and NH_4, and (3) a high Fe content in some crystallographic structures such as nontronite or beidellite, particularly when derived from basaltic parent materials containing abundant ferromagnesian minerals (Coulombe et al., 1996a). A CEC/clay ratio ≥0.60 is typical of soils in the smectitic mineralogy class at the family taxonomic level (a CEC/clay ratio ≥0.60 identifies the "superactive" CEC activity family class [Soil Survey Staff 2010], which suggests that the clay fraction is dominated by smectites).

33.7.7 Chemical Properties

33.7.7.1 pH

Vertisols are commonly neutral to alkaline in reaction (Table 33.31), given that the majority are derived from calcareous or base-rich parent materials. However, acidic Vertisols (Ahmad, 1985) do exist and are recognized as Dystric great groups of the Aquerts, Uderts, and Usterts (pH values of 4.5 or less in 0.01M $CaCl_2$ [5.0 or less in 1:1 water] and electrical conductivity less than 4.0 dS m^{-1}). Some Vertisols have a sulfuric horizon with a pH or 3.5 or less (Soil Survey Staff, 2010) produced by the oxidation of sulfides and resulting formation of sulfuric acid, or they contain sulfidic materials that can oxidize and acidify the soil (e.g., Sulfaquerts).

33.7.7.2 Cation Exchange Capacity and Exchangeable Cations

Cation exchange capacity (CEC) of Vertisols generally ranges from about 30 to about 60 $cmol_c$ kg^{-1} (soil) and depends on the amount and type of mineral phases and soil organic matter (SOM) present (Table 33.31). Calcium and Mg commonly are the main exchangeable cations. In some cases, Na is an abundant exchangeable cation, and when exchangeable Na percentage (ESP) exceeds 15, particle dispersion and soil structure deterioration can occur. Particle dispersion can occur at ESP values less than 15 if soluble salts are absent and the resulting soil solution electrolyte concentration is low. The presence of soluble salts and gypsum can maintain solution electrolyte concentrations at high enough levels to prevent soil dispersion even at ESP values greater than 15. Under acidic conditions, Al and H may become more prevalent in the soil solution and replace exchangeable Ca and the other base cations. An increase in the concentration of exchangeable Al and H and a decrease in CEC indicate that chemical weathering has occurred, via dissolution of primary and secondary (clay) minerals (Ahmad, 1985).

33.7.8 Soil Biochemical Properties

Biochemical properties are most strongly influenced by the quantity and type of organic matter, and land use (Table 33.32). Generally, organic C content ranges from 3 to 60 g kg^{-1}, but most commonly from 3 to 35 g kg^{-1} (Coulombe et al., 1996a), and varies with (1) climatic conditions (higher under more humid and cooler regimes); (2) microtopography (higher in microlows than in microhighs; Kunze and Templin, 1956; Wilding and Tessier, 1988; Nordt et al., 2004); (3) depth (generally decreases with depth in the soil profile, which refutes the argument that complete self-mixing occurs in Vertisols, or at least indicates that the rate of mixing is slower than the rate of soil organic C accumulation and redistribution; Wilding and Tessier, 1988; Southard and Graham, 1992; Coulombe, 1997); and (4) tillage (higher in noncultivated than in cultivated areas; Kunze and Templin, 1956; Skjemstad and Dalal, 1987; Coulombe, 1997; Chevallier et al., 2000; Nordt and Wilding, 2009).

Soil organic matter has some influence on soil color, but black Vertisols may contain as little as 3 g kg^{-1} organic C. The contribution of soil organic matter to the dark color is the result of strong clay–organic complexes in which smectites are coated

TABLE 33.31 Chemical Characteristics[a] of Selected Vertisols Formed from a Wide Variety of Parent Materials

Regions	pH (1:1 Water)	CEC ($cmol_c$ kg^{-1} Soil)	Extractable Bases ($cmol_c$ kg^{-1} Soil)				References
			Ca	Mg	K	Na	
Australia	7.0 (0.6)	33.7 (2.5)	22.4 (3.4)	15.7(3.4)	0.1 (0.1)	0.9 (0.4)	Boettinger (1992)
India	8.2 (0.2)	55.8 (4.7)	53.4 (5.6)	[b]	0.9 (0.1)	1.4 (0.5)	Hirekurubar et al. (1991)
Sudan	7.3 (0.4)	40.6 (11.0)	n/a	n/a	n/a	2.95 (3.9)	Blokhuis (1993)
Texas	7.4 (0.9)	41.0 (5.4)	52.9 (23.0)	6.8 (3.5)	0.7 (0.3)	2.7 (3.4)	Kunze et al. (1963) and Hallmark et al. (1986)
West Africa	7.4 (0.8)	30.3 (8.0)	17.9 (7.9)	7.5 (3.1)	0.7 (0.4)	3.2 (3.3)	Beavington (1978) and Yerima (1986)
California	8.0 (0.6)	45.7 (0.9)	26.8 (2.5)	17.1 (1.3)	0.8 (0.3)	1.9 (1.2)	NSSL Pedon No. 94P0056
El Salvador	6.7 (0.6)	45.5 (3.4)	33.9 (10.3)	11.6 (2.5)	0.5 (0.2)	1.9 (1.5)	Yerima et al. (1985)
West Indies	7.3 (0.6)	48.0 (11.7)	51.8 (23.6)	8.1 (5.8)	0.4 (0.3)	2.4 (2.1)	Ahmad and Jones (1969)
Uruguay	6.5 (0.5)	33.9 (7.9)	23.1 (7.5)	8.0 (2.3)	0.3 (0.1)	0.3 (0.1)	Lugo-Lopez et al. (1985)

n/a, Not available.
[a] Data are reported as mean (sd).
[b] Extractable Mg was compiled with data for extractable calcium.

TABLE 33.32 Biochemical Characteristics of Surface Horizons of Selected Vertisols

Regions	Soil Series (and/or Parent Material)	Land Use	OC (g kg^{-1})	Total N (g kg^{-1})	C/N Ratio	Total P (mg kg^{-1})	References
Australia	Langlands-Logie clays	Virgin	23.0	2.41	9.5	n/a	Skjemstad et al. (1986)
	(Alluvial clayey sediments)	20 year cultivation	11.4	1.12	10.2	n/a	Skjemstad et al. (1986)
		35 year cultivation	8.9	0.89	10.0	n/a	Skjemstad et al. (1986)
		45 year cultivation	7.8	0.77	10.1	n/a	Skjemstad et al. (1986)
	Black Earth	10 year pasture	14.2	0.9–2.6	5.8–12.2	n/a	Jocteur Monrozier et al. (1991)
Texas	Houston Black	Virgin	20.9–48.2	1.37–3.53	13.7–21.1	0.35–0.55	Coulombe (1997)
		20 year restoration	13.7–24.0	0.87–1.45	18.4–19.8	0.23–0.29	Coulombe (1997)
		Pasture	15.5–33.3	0.97–2.02	18.2–25.8	0.29–0.44	Coulombe (1997)
		Cereal rotation	12.2–15.6	0.83–1.03	14.7–20.3	0.34–0.43	Coulombe (1997)
		Row crop rotation	15.6–19.3	0.96–1.45	13.4–19.4	0.29–0.48	Coulombe (1997)
Nigeria	Lacustrine plain	Cereal	0.26–0.58	0.06–0.08	5.2–7.2	5–10	Beavington (1978)
Trinidad	Princes Town clay	>100 year sugar cane	31.0–38.0	2.7–3.5	11	29–58	Ahmad and Jones (1969)
Jamaica	Carron Hall clay	Unimproved pasture	42.0–60.0	4.5–5.7	9–11	13–15	Ahmad and Jones (1969)
Barbados	Black soil	>300 year sugar cane	12.0–15.0	0.8–1.0	15	250–500	Ahmad and Jones (1969)
Antigua	Fitches clay	Mixed shrub/grasses	32.0–38.0	2.2–2.3	15–17	13–17	Ahmad and Jones (1969)
Dominica	Black soil	Shrub	11.0–28.0	1.0–2.1	11–13	2–4	Ahmad and Jones (1969)
California	Capay	Grass rangeland	12.8	1.56	8.2	n/a	NSSL Pedon No. 94P0056

n/a, Not available.

with humic materials in the presence of Ca (Singh, 1954, 1956). The resistance to degradation is due to this smectite–Ca–organic complex, rather than due to the recalcitrant chemical composition of the organic material (Skjemstad et al., 1986; Skjemstad and Dalal, 1987).

The C/N ratio generally ranges from 10 to 25 (Table 33.32). Microbial biomass may represent up to 5% of total organic C (Jocteur Monrozier et al., 1991). Polysaccharides and fungal hyphae tend to be concentrated at interparticle positions in the domains and quasicrystals of the clay minerals (Chenu, 1989; Chenu and Jaunet, 1990). Total P ranges widely (Table 33.32), mostly as a function of parent material composition, but may also reflect organic forms of P that vary as a function of SOM content and soil management.

Loss of organic C can occur rapidly after a change in land use from native conditions to cultivation. After a few decades or more of continuous cultivation, loss of organic C can be significant (Skjemstad et al., 1986; Skjemstad and Dalal, 1987; Puentes, 1990; Coulombe, 1997; Nordt and Wilding, 2009), but the trend can be reversed, at least to some degree, when cultivated fields are allowed to return to native conditions (Chevallier et al., 2000).

33.7.9 Physical Properties

33.7.9.1 Moisture Retention and Shrink/Swell Behavior

Soil moisture conditions are especially important in Vertisols because they dominate the physical properties and shrink/swell behavior. Due to the clay content, Vertisols exhibit high water retention characteristics at all water potentials (Table 33.33).

TABLE 33.33 Physical and Mechanical Characteristics[a] of Selected Vertisols from a Wide Array of Parent Materials

Regions	Water 0.033 MPa (Mass %)	Water 1.5 MPa (Mass %)	Bulk Density 0.033 MPa (Mg m^{-3})	Bulk Density Oven Dry (Mg m^{-3})	COLE (cm cm^{-1})	Liquid Limit (%)	Plastic Limit (%)	Plasticity Index (%)	References
India	38.3 (2.9)	17.8 (0.9)	n/a	1.3 (0.1)	n/a	61.0 (2.4)	34.0 (1.2)	26.9 (1.3)	Hirekurubar et al. (1991)
Texas	42.2 (3.2)	23.4 (2.3)	n/a	1.8 (0.2)	n/a	67.1 (3.8)	22.7 (4.3)	44.3 (3.6)	Kunze et al. (1963) and Yule and Ritchie (1980a)
California	33.5 (1.0)	21.9 (0.6)	1.31 (0.02)	1.85 (0.03)	0.12 (0.01)	n/a	n/a	n/a	NSSL Pedon No. 94P0056
West Africa	31.4 (6.5)	n/a	1.4 (0.1)	1.4 (0.1)	0.10 (0.03)	n/a	n/a	n/a	Yerima (1986)
El Salvador	n/a	n/a	1.1 (0.1)	1.1 (0.1)	0.16 (0.03)	n/a	n/a	n/a	Yerima et al. (1985)

n/a, Data not available.
[a] Data are reported as mean (sd).

Soil water content varies considerably depending on the amount and kind of clay, climatic conditions, and land use.

Shrink/swell behavior places major constraints on the management of Vertisols. Under normal field conditions, most of the volume change occurs at water potentials >−1.5 MPa (Yule and Ritchie, 1980a; Wilding and Tessier, 1988), but it is not clear whether volume change is isotropic (equidimensional) or anisotropic. Current knowledge suggests that individual undisturbed peds shrink and swell isotropically (Yule and Ritchie, 1980a; Cabidoche and Voltz, 1995). Yule and Ritchie (1980b) reported that shrinkage in large (73 cm diameter, 140 cm long) and small (10 cm diameter) cores of a Texas Vertisol was similar. However, under field conditions, Cabidoche and Voltz (1995) reported anisotropic volume change in a Vertisol from Guadeloupe (French West Indies) with slightly larger movements vertically than horizontally (anisotropy ratio ~0.8). They attributed the anisotropy to movement of the peds along oblique slickensides, which highlights the importance of spatial arrangement or geometry of the peds in a soil profile on shrink/swell behavior.

33.7.9.2 Bulk Density and Coefficient of Linear Extensibility

Bulk density values for selected Vertisols are presented in Table 33.33. Factors that influence bulk density include (1) water content, (2) clay content, (3) particle density of the mineral fraction, and (4) the method used to determine bulk density (e.g., core vs. clods). Bulk density is routinely used to estimate total porosity and other volumetric percentages and to assess structural conditions of soils. Bulk density should be coupled with morphological characteristics and other properties for a thorough assessment of soil structural conditions.

The coefficient of linear extensibility (COLE) is an index of soil shrinkage (Table 33.33). The COLE value is typically calculated from soil bulk densities measured at −0.033 and −1.5 MPa, assuming that soil shrinkage is isotropic [COLE = (bulk density dry/bulk density moist)$^{1/3}$ – 1]. COLE is correlated with total and fine clay contents, surface area, water retention at field capacity, and ESP (Anderson et al., 1973; Yule and Ritchie, 1980a; Yerima et al., 1989).

33.7.9.3 Consistence and Atterberg Limits

Consistence refers to "the degree and kind of cohesion and adhesion that soil exhibits, and/or the resistance of soil to deformation or rupture under an applied stress" (Schoeneberger et al., 2002). The consistence of Vertisols ranges from hard to extremely hard when dry, from friable to very firm when moist, and from moderately sticky and moderately plastic to very sticky and very plastic when wet. Vertisols exhibit very narrow ranges of consistence and soil water content favorable for workability and trafficability.

The Atterberg limits (liquid limit, plastic limit, and plasticity index; Chapter 3) are empirical tests used by geoscientists and engineers to determine the mechanical characteristics of a soil and are approximated by the soil consistence properties described above. Values for Vertisols are presented in Table 33.33. In general, values of Atterberg limits increase as the clay content increases and decrease in the order: smectite > illite > kaolinite.

33.7.9.4 Structure/Porosity and Related Properties

Soil processes such as gas exchange, heat transfer, water flow, and movement of solutes and organics are dependent on structure and porosity. Vertisols tend to form large pores (macro- and mesopores, i.e., cracks) during soil drying and shrinkage. These cracks have a profound impact on several soil processes. They allow oxygen at atmospheric partial pressure to penetrate to the depth of cracking, they allow greater nitrification rates than in the rest of the soil mass (Kissel et al., 1974), and they promote movement of water and solutes through preferential or bypass flow (Kissel et al., 1973, 1974; Bouma, 1983; Bouma and Loveday, 1988). Lin et al. (1996) reported, for a Texas Vertisol, that macropores (>0.5 mm effective diameter) and mesopores (0.06 to <0.5 mm effective diameter) contributed 89% and 10% of the water flow, respectively. Several soil properties influence infiltration rates (Table 33.34). In clayey soils that shrink and swell, hydraulic conductivity (K) does not meet the assumption of uniform flow in a homogenous soil (Darcy's law), so infiltration rates are often preferred over K to characterize water movement (Lin, 1995; Lin and McInnes, 1995). Nonetheless, infiltration rates can be highly variable in space and time and remain dependent on a variety of factors including land use and the method used to measure infiltration rate.

33.7.10 Land Use and Management Considerations

33.7.10.1 Natural Resources

33.7.10.1.1 Agriculture

For many regions of the world, Vertisols are highly productive and sustainable resources for various crop production systems, including a wide array of food and fiber crops. Nonetheless, Vertisols in developed and developing countries are subject to degradation due to intensive and/or inappropriate management practices. Soil properties may partially recover from the effects of cultivation, but it is unlikely that native conditions can be rejuvenated even after a restoration period of 20 years or more (Puentes, 1990; Coulombe, 1997; Chevallier et al., 2000; Nordt and Wilding, 2009). One of the difficulties in successfully managing Vertisols involves some limitations on technology transfer. Despite the common properties of high clay content, shrink/swell behavior, and cracking, many Vertisols require site-specific management for crop production (Ahmad and Mermut, 1996) due to variations in climate, soil chemistry, and cropping systems.

33.7.10.1.2 Wetlands

Vertisols, especially those on the wetter end of the climatic spectrum (i.e., Aquerts and Uderts), can potentially be used for the creation or restoration of wetland habitats (Chapter 20

TABLE 33.34 Infiltration Rates under Saturated Conditions for Surface Horizons of Selected Vertisols with Reference to Land Use and Method

Regions	Soil Series (and/or Parent Material)	Land Use	Method (References)	Infiltration ($\mu m\ s^{-1}$)	References
India	(Chlorite schist)	n/a	Double ring infiltrometer (Marshall and Stirk, 1950)	3.3	Hirekurubar et al. (1991)
	(Shale)	n/a	Double ring infiltrometer (Marshall and Stirk, 1950)	5.6	Hirekurubar et al. (1991)
	(Granite gneiss)	n/a	Double ring infiltrometer (Marshall and Stirk, 1950)	0.8	Hirekurubar et al. (1991)
	(Deccan trap)	n/a	Double ring infiltrometer (Marshall and Stirk, 1950)	1.2	Hirekurubar et al. (1991)
	(Limestone)	n/a	Double ring infiltrometer (Marshall and Stirk, 1950)	11.1	Hirekurubar et al. (1991)
Texas	Heiden	15–25 year pasture	Double ring infiltrometer (Bouwer, 1986)	0.70–0.94	Puentes (1990)
	Heiden	15–25 year cultivation	Double ring infiltrometer (Bouwer, 1986)	0.14	Puentes (1990)
	Houston Black	Virgin	Column method (Bouma, 1982)	1.86	Coulombe (1997)
			Ponded infiltrometer (Prieksat et al., 1992)	1.77	Potter (pers. comm.)
			Disc permeameter (Perroux and White, 1988)	2.13	Lin (pers. comm.)
		20 year restoration	Column method (Bouma, 1982)	1.04	Coulombe (1997)
			Ponded infiltrometer (Prieksat et al., 1992)	1.77	Potter (pers. comm.)
			Disc permeameter (Perroux and White, 1988)	2.29	Lin (pers. comm.)
		Pasture	Column method (Bouma, 1982)	0.61	Coulombe (1997)
			Ponded infiltrometer (Prieksat et al., 1992)	1.26	Potter (pers. comm.)
			Disc permeameter (Perroux and White, 1988)	1.86	Lin (pers. comm.)
		Cereal rotation	Column method (Bouma, 1982)	0.39	Coulombe (1997)
			Ponded infiltrometer (Prieksat et al., 1992)	1.75	Potter (pers. comm.)
			Disc permeameter (Perroux and White, 1988)	1.40	Lin (pers. comm.)
		Row crop rotation	Column method (Bouma, 1982)	0.88	Coulombe (1997)
			Ponded infiltrometer (Prieksat et al., 1992)	1.65	Potter (pers. comm.)
			Disc permeameter (Perroux and White, 1988)	1.37	Lin (pers. comm.)

Source: From Coulombe, C.E., L.P. Wilding, and J.B. Dixon. 2000. Vertisols, p. E-269–E-286. *In* M.E. Sumner (ed.) Handbook of soil science. CRC Press, Boca Raton, FL.; Based on 1997 estimates by the USDA-NRCS.

n/a, Not available.

of *Handbook of Soil Sciences: Resource Management and Environmental Impacts*). Besides the presence of hydric conditions, a requirement for successful establishment of desired wetland functions includes development of wetland hydrology and the presence of hydrophytic vegetation. The swelling of Vertisols when wet drastically reduces permeability and provides a mechanism to more easily manage surface waters to maintain hydric conditions, establish wetland hydrology, and provide an environment that favors hydrophytic vegetation.

33.7.10.2 Engineering

33.7.10.2.1 Environmental Engineering

Due to their extensive shrink/swell behavior, Vertisols are generally not suitable for solid waste and wastewater disposal. Water generally percolates too rapidly when the soils are dry and too slowly when the soils are wet. In either case, these conditions could contribute to contamination of surface and groundwater and result in potential environmental health hazards. For similar reasons, Vertisols that exhibit a permanent (endoaquic) or seasonally perched (epiaquic) water table would not be a suitable filter for wastewater disposal. Further, disposal facilities built on Vertisols are prone to damage by shrink/swell processes that can crack foundations or rupture pipes.

On the other hand, Vertisols can be used as constructed treatment wetlands for final disposal of tertiary effluent provided that wetland hydrology is maintained at saturation to enhance low soil permeability and to reduce groundwater recharge. Treatment wetlands are effective technologies for water quality improvement of various wastewaters as well as being beneficial for wildlife habitats (Kadlec and Knight, 1995).

33.7.10.2.2 Civil Engineering

The shrink/swell nature of Vertisols severely limits civil engineering applications. Vertisols must be stabilized for construction of buildings, roads, pipelines, and utility corridors. For example, several major cities in Texas (Dallas, Houston, San Antonio, and Corpus Christi) have been built on Vertisols. The swell/shrink properties of Vertisols make effective construction site stabilization very difficult. The most economic choice may be to select stable soil (not Vertisols) for construction sites. If construction on a Vertisol is necessary, on-site investigation is generally required. Morphological properties of these soils often indicate movement at depths between 2 and 4 m due to natural moisture variability (Coulombe et al., 2000). These depths of movement suggest that soil management is the best way to minimize damage to structures on Vertisols and other expansive soils. Homeowners can maintain moist soils around the foundations of their dwellings that are built on steel reinforced concrete slabs to minimize movement. Application of gypsum to reduce ESP minimizes excessive swelling of smectites.

Prior to construction of roads, streets, and many structures that are to be built on Vertisols, lime [$Ca(OH)_2$] stabilization of the soil is required to a depth of about 0.3 m or less (Eades and Grim, 1960; National Lime Association, 1991). Although this is beneficial, it obviously is only a partial treatment and complete stabilization would be extremely expensive. Larger multistory buildings sited on vertic soils and deep smectitic clays typically are supported by concrete piers with large bell-shaped feet at about 4 m depth to avoid contact with expansive soil.

Many other commercial methods to stabilize Vertisols are being promoted, some of which may be as successful as lime stabilization. One such treatment with a sulfonated naphthalene was tested under laboratory conditions on three Vertisols with some success (Marquart, 1995), but the results were inconsistent, depending on soil conditions and level of treatment.

It is imperative for construction workers to seriously consider the use of shoring materials and other safety precautions when a Vertisol is excavated. In Texas, from 1980 to 1985, 50 construction workers were killed due to destabilization of Vertisols that occurred during excavation (USDA-SCS, 1986).

33.7.11 Geological Studies of Lithified Paleosols Interpreted as Vertisols

33.7.11.1 Introduction

Vertisol-like paleosols here and elsewhere (hereafter referred to as "paleovertisols") are clay-rich paleosols that show abundant evidence for shrink–swell processes including slickensides, cracks, and strongly developed sepic–plasmic microfabrics. Vertisols are one of the few soil orders for which *Soil Taxonomy* can be easily applied to corresponding paleosols in the geological record (Mack et al., 1993). Paleovertisols have provided a wealth of information on paleoenvironment conditions, including Paleozoic (540 to 225 million years before present) levels of atmospheric CO_2 and changes in soil morphology accompanying the evolution and diversification of terrestrial land plants (Mora et al., 1996; Mora and Driese, 1999; Driese and Mora, 2001; Driese et al., 2005; Driese, 2009). Much of the interest in paleosols is driven by their potential to serve as climatic proxies in ancient environments.

Vertisols are characterized by macro- and micromorphologic characteristics formed in response to shrink–swell phenomena associated with either seasonal precipitation or seasonal soil moisture deficits. Many of these features preserve well in the ancient rock record in Paleovertisols and can be easily recognized in the field (in outcrop exposures and in cores) and in thin section; thus, Paleovertisols are increasingly reported in the geologic record, ranging in age from the Proterozoic (Rye and Holland, 2000; Driese, 2004) to the Cenozoic. Paleovertisols formed over a wide range of paleolatitudes, ranging from equatorial–low latitude (Appalachian basin; references cited below) to as high as 78° (Triassic of Antarctica; Retallack and Alonso-Zarza, 1998). Virtually every Paleozoic red bed formation in the Appalachian Basin stratigraphic succession has common occurrences of Paleovertisols or vertic intergrades of other soil orders (Mora and Driese, 1999; Driese et al., 2000a; Driese and Mora, 2001; Stiles et al., 2001). The wide spatial and temporal distribution of Paleovertisols reflects the potential for their formation under wide ranges of moisture regimes (from semiarid to humid); the only strict climatic requirement is that the moisture is seasonally distributed. The widespread distribution also reflects the predominantly physical controls on Vertisol formation; these soils are not limited to specific environments or ecosystems. Retallack (1994) compiled a nonlinear relationship that predicts annual rainfall based upon the depth from the soil surface to pedogenic carbonate horizons. This relationship, based on measurements in carbonate-bearing soils representing a variety of soil orders, was better constrained specific for Vertisols by Nordt et al. (2006). Stable carbon and oxygen isotope analysis of pedogenic carbonate in Vertisols helps to elucidate the soil ecosystem (dominance of C_3 vs. C_4 flora) and changes in soil water compositions that may indicate climate change (Kovda et al., 2003, 2006). Similar information may be preserved in the isotopic compositions of Paleosols, as well as constraints on paleoatmospheric pCO_2 using Cerling's (1991) carbon isotope paleobarometer (Ekart et al., 1999).

33.7.11.2 Distribution of Fossil Vertisols in Geologic Record

Modern Vertisols today only comprise about 2% of world soils, far less abundant than most other soil orders recognized in USDA *Soil Taxonomy*, yet they appear to comprise the majority of paleosols preserved in the geologic record. Possible hypotheses that might explain this apparent overrepresentation of Paleovertisols in the geological record include the following:

1. There a strong recognitional bias for Paleovertisols in outcrop exposures and in cores because of ease of identification (discussed previously).
2. There are significant preservational processes biased for Vertisols. Because Vertisols are known to form very rapidly (within tens to hundreds of years, in some cases) and mostly in low-lying floodplains and lacustrine floodbasins, not much time is necessary for subaerial exposure and pedogenesis, and the resultant formation of a paleosol with high preservation potential. In addition, Vertisols can form over wide ranges of latitudes (e.g., from North Dakota to south Texas in the United States), and under wide ranges of moisture regimes. As such, they may have been more widely distributed in the geologic past than they are now. Because the physical processes of clay shrink–swell dominate soil-forming processes, plants may be unnecessary in Vertisol genesis (Mora and Driese, 1999). A wider temporal range (i.e., pre-Silurian) is therefore possible. This contrasts for example, with the distribution of soils with mollic epipedons (mostly Mollisols), the first appearance of which in the Tertiary is related to the appearance of dryland grasses (Retallack, 1997).

TABLE 33.35 Diagenetic Alteration of Paleovertisols (High Clay-Content Paleosols with Extensive Shrink–Swell Features Such as Slickensides) Compared with a Surface Soil Analog (Houston Black Series, TX)

Name (Geologic Age)	Burial Depth (km)	Burial Temperature (°C)	Clay Mineral Assemblage	wt% K_2O
Houston Black (modern)	Surface	22° Mean annual temperature	Na–smectite	1.29
Pennington (325 Ma)	2–3	60–90	Illite (+ kaolinite + chlorite)	4.61
Hekpoort (2.25 Ga)	12–15	350	Sericite–muscovite–chlorite	9.50

Sources: Driese, S.G., C.I. Mora, and J.M. Elick. 2000a. The paleosol record of increasing plant diversity and depth of rooting and changes in atmospheric pCO_2 in the Siluro-Devonian, p. 47–61. *In* R.A. Gastaldo and W.A. DiMichele (eds.) Phanerozoic terrestrial ecosystems, a short course. The Paleontological Society Papers 6, New Haven, CT; Driese, S.G., C.I. Mora, C.A. Stiles, R.M. Joeckel, and L.C. Nordt. 2000b. Mass-balance reconstruction of a modern Vertisol: Implications for interpretations of geochemistry and burial alteration of paleovertisols. Geoderma 95:179–204; Rye, R., and H.D. Holland. 2000. Geology and geochemistry of paleosols developed on the Hekpoort Basalt, Pretoria Group, South Africa. Am. J. Sci. 300:85–141. Driese, S.G. 2009. Paleosols, pre-Quaternary, p. 748–751. *In* V. Gornitz (ed.) Encyclopedia of paleoclimatology and ancient environments. Kluwer Academic Publishers, New York.

Ma, millions of years; Ga, billions of years. Note potassium enrichment occurring over time and with increasing burial depths and temperatures.

33.7.11.3 Paleoclimate Reconstructions Based on Fossil Vertisols

Paleoclimate reconstructions based on fossil Vertisols primarily emphasize estimation of mean annual precipitation (MAP) using a variety of proxy measures, including (1) depth to pedogenic carbonate (Bk) horizon (Caudill et al., 1996; Nordt et al., 2006); (2) Fe content of pedogenic Fe–Mn nodules and concretions (Stiles et al., 2001); (3) chemical indices of weathering used to estimate paleoprecipitation based on bulk geochemistry (e.g., Chemical Index of Alteration minus Potash or CIA − K, Sheldon et al., 2002 [note: CIA − K = molar ratio of $Al_2O_3/(Al_2O_3 + Na_2O + CaO) \times 100$, and the linear regression has an $r^2 = 0.73$], but a new paleoprecipitation proxy specific for Paleovertisols termed CALMAG has been developed where CALMAG = molar ratio of $Al_2O_3/(Al_2O_3 + CaO + MgO) \times 100$, and the linear regression obtained has an $r^2 = 0.90$; Nordt and Driese, 2010a); (4) total element mass-flux and mass-balance calculations (Driese et al., 2000b; Stiles et al., 2003); and (5) transfer functions for reconstructing colloidal properties of Paleovertisols from bulk geochemistry Nordt and Driese (2010b). Paleolandscape reconstructions in which topography or hydrology is important variable (interpretation of paleocatenas) are commonly conducted at both local and more regional scales. Reconstructions of Phanerozoic pCO_2 employ the CO_2-carbonate paleobarometer of Cerling (1991) to interpret Paleozoic (Mora et al., 1996; Mora and Driese, 1999) and Mesozoic paleoatmospheres from fossil Vertisols (Nordt et al., 2002, 2003). This technique utilizes the $\delta^{13}C$ values measured from pedogenic carbonates, measurements of $\delta^{13}C$ values of paleosol organic matter (or an estimate thereof based on marine proxy records), and assumptions of soil productivity (soil-CO_2 production rates) to estimate paleoatmospheric pCO_2. These results are in good agreement with pCO_2 estimates based on long-term mass-balance carbon models (Berner, 2006). Other research directions include systematic studies of root diameter, depth, and density in Paleosols using root traces, and relating these changes to development of soil morphology (Driese and Mora, 2001), widespread deposition of black shales, and carbon sequestration.

Problems in interpreting paleoclimates from pre-Quaternary fossil Vertisols include extensive burial diagenetic alteration and some compaction (Sheldon and Retallack, 2001). Commonly cited diagenetic alteration processes include physical compaction and consequent modifications of original soil thickness and morphology, oxidation of soil organic matter, burial gleization (Fe loss, largely through microbial reduction), color modifications (intensification) related to dehydration and recrystallization of hydrous mineral phases (such as FeOOH), recrystallization of soil smectites to illites (or even to metamorphic mineral assemblages; Table 33.35), and exchange of oxygen isotopes between burial fluids and pedogenic carbonates and clays (Mora et al., 1998; Mora and Driese, 1999). Evolutionary changes in terrestrial plant and animal communities over time also create difficulties in paleosol interpretations because of attendant changes in morphological features such as root traces, which are (1) not present in Ordovician paleosols; (2) rhizomatous and fine in Silurian paleosols; and (3) larger, deeper, and more "modern" in Devonian and post-Devonian paleosols (Driese et al., 2000a, 2000b; Driese and Mora, 2001).

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33.8 Mollisols

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33.8.1 Introduction

Mollisols are generally characterized as soils with thick, dark surface horizons (mollic epipedons) resulting from organic C incorporation. The terms mollic and Mollisol are derived from the Latin *mollis*, that is, soft. While the soil taxon, Mollisols, is defined by an exact set of soil property criteria, these criteria arose from conceptual ideas and empirical observations of soil landscapes with thick, dark, and often fertile surface horizons. Mollisols can form under multiple environmental conditions that facilitate accumulation of organic C in the upper soil profile. Although Mollisols correspond to the Chernozem soils of the Russian and older U.S. classification systems, the Mollisol soil order includes soils that are beyond the central concept of Chernozems (Fanning and Fanning, 1989).

Mollisols are among the most important soils for food and fiber production due to relatively high levels of native fertility coupled with climatic conditions conducive to plant growth. For this reason, the characteristics of Mollisols have been studied extensively, with early efforts focusing on the Chernozems of the Russian Steppes. During the late nineteenth century, droughts and crop failures afflicted the Russian Chernozem region resulting in widespread famine. In an effort to determine if the droughts could be prevented or moderated, Dokuchaev (1883) initiated an extensive study of the climate, soils, vegetation, and topography of the Chernozem Steppes. Based on the physical evidence that he and his colleagues collected, Dokuchaev theorized that the Chernozems of the Russian Steppes formed under specific conditions of climate, vegetation, and topography. These ideas contradicted the in vogue theories that Chernozems originated under either aquatic or marsh conditions. Dokuchaev's classic studies not only provided insight into the potential processes responsible for the formation of Chernozems, and subsequently Mollisols, but also articulated a conceptual framework for soil genesis that is still one of the foundations of modern pedology.

33.8.2 Geography and General Characteristics

33.8.2.1 Geographic Distribution

Extensive areas of Mollisols are distributed throughout the midlatitudes of the world, predominantly on subhumid steppes and prairies (Figure 33.45). Mollisols occur under a wide range of climatic conditions including xeric, ustic, udic, and aquic soil moisture regimes and cryic, frigid, mesic, thermic, and hyperthermic soil temperature regimes. Spatially, Mollisols are the 8th (of 12) most common soil order and are estimated to cover

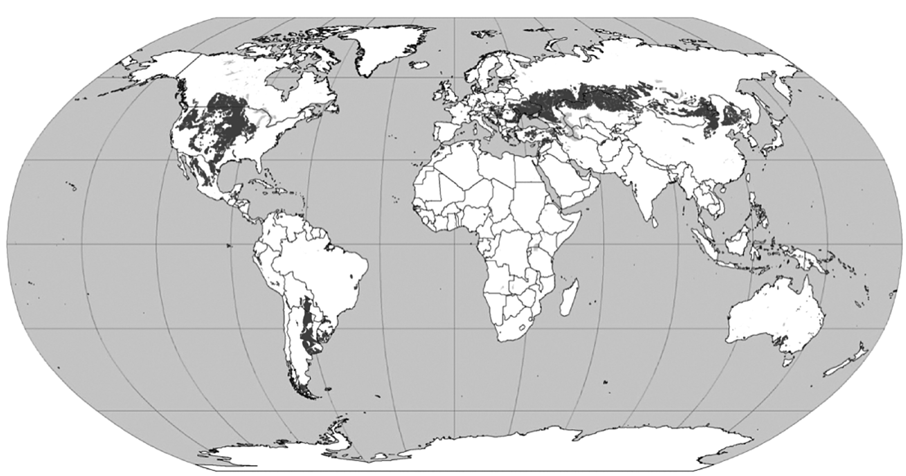

FIGURE 33.45 Global distribution of Mollisols. (Courtesy of USDA-NRCS, Soil Survey Division, World Soil Resources, Washington, DC, 1998.)

TABLE 33.36 Estimated Occurrence of Mollisol Suborders Worldwide by General Climatic Condition

Climatic Conditions	Major Geographic Locations	Mollisol Suborder	Approximate Land Area (km^2)
Semiarid	Great Plains of United States and Canada, Southern Russia, Pampas Region of Argentina, Yucatan Peninsula of Mexico	Ustolls	5,244,636
Humid	Black Sea Region of Russia, North-Central United States, Western Europe	Udolls and Rendolls	1,526,878
Cold	Mountainous regions of the Western United States and Canada, higher latitudes across North America, Kazakhstan, Ukraine, and Russia	Gelolls and Cryolls	1,163,797
Mediterranean	Turkey, Argentina, Palouse Region of United States, California	Xerolls	924,394
Seasonally saturated	Red River Valley of North, riparian zones, Gulf coastal plain, glaciated regions of the United States	Aquolls and Albolls	145,728

Source: Based on 1997 Estimates by the USDA-NRCS; NRCS Soil Survey Staff: http://soils.usda.gov/survey/

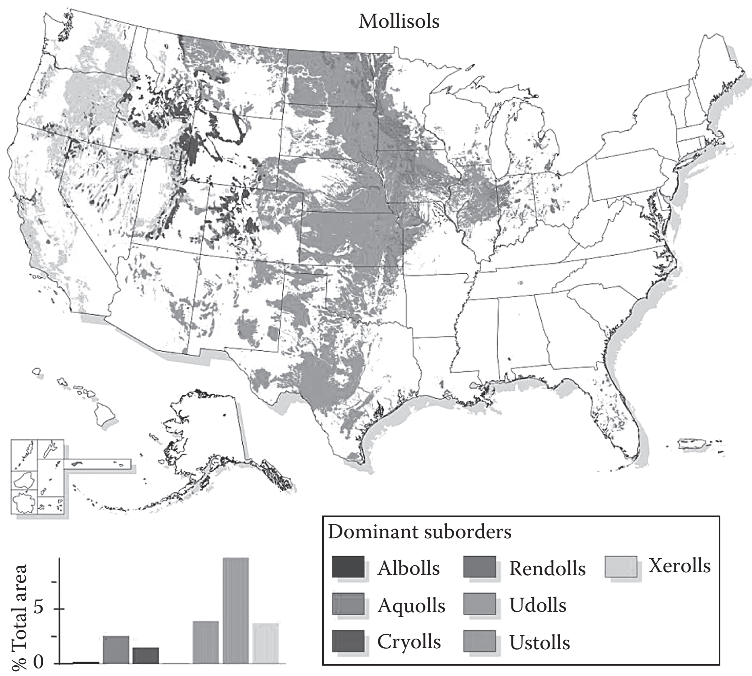

FIGURE 33.46 Distribution of Mollisols in the United States. (From http://soils.usda.gov/technical/classifcation/orders/mollisols_map.html).

9,005,433 km^2 or approximately 7% of the ice-free land area on Earth (Table 33.1). The distribution of Mollisol suborders can be distinguished by differences in climate, with the exception of wet Mollisols (Aquolls and Albolls), which typically occur in localized areas subject to saturated and reduced soil conditions (Table 33.36). The distribution of Mollisol suborders in the United States is shown in Figure 33.46. The close relationship of prairie ecosystems and Mollisols is apparent in the U.S. distribution where Mollisols are more common in the Western United States, west of the forested eastern half of the country. Noticeably, the "Prairie Peninsula" shows up in Figure 33.46 as the eastern expansion of Mollisols into Illinois and Indiana (Geis and Boggess, 1968).

33.8.2.2 General Characteristics of Mollisols

The general concept of Mollisols is that of dark colored soils of semiarid to subhumid grassland ecosystems. The dark color reflects soil organic matter (SOM) enrichment in the upper portion of the profile (Figure 33.47). The formation of dark surface horizons is termed melanization, which is actually a combination of several processes involving the addition of organic matter to the soil in the form of plant residues and its subsequent transformation into humus (Buol et al., 2003). Distinguishing features of Mollisols include the presence of a mollic epipedon and high base status (Soil Survey Staff, 2010). A mollic surface layer by itself is not diagnostic for Mollisols, as it can be a feature of soils belonging to other orders. Mollisols are differentiated in these cases by the presence of high base status (>50% base saturation) horizons that underlie the mollic epipedon, or by the lack of features associated with high shrink/swell clays, typical in Vertisols (Graham and Southard, 1983). A wide variety of subsurface horizons can occur beneath the mollic epipedon of Mollisols, including albic, cambic, argillic, calcic, petrocalcic, natric, duripans, and gleyed horizons. However, fragipans are unknown in Mollisols (Franzmeier et al., 1989). The nature of

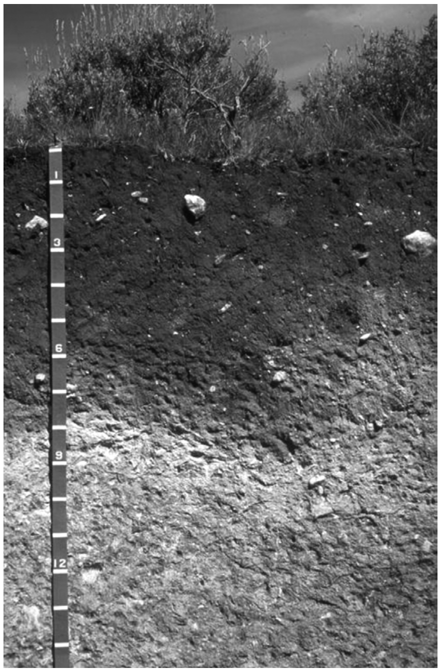

FIGURE 33.47 Soil profile of a Mollisol showing thick, dark mollic epipedon. The mollic epipedon extends to a depth of almost 60 cm; scale is in decimeters.

any diagnostic horizon is usually expressed at the great group level in classification with the exception of Albolls (albic) and Aquolls (gleyed).

A mollic epipedon is defined in *Soil Taxonomy* by several quantitative criteria including horizon thickness, organic C content, color, consistence, structure, and base status (Soil Survey Staff, 1999, 2010). Although a mollic epipedon must contain at least 1% SOM (0.6% organic C) on a weight basis, many Mollisols have higher SOM contents. A mollic epipedon must generally be a minimum of 18 cm thick for shallow and 25 cm thick for deep soils. Soils, which are very shallow, require only a 10 cm thick mollic epipedon (Soil Survey Staff, 2010). Munsell color requirements include a value (a measure of lightness/darkness) that is 3 or less for a moist sample and 5 or less for a dry sample. In both cases, the chroma (intensity or purity of color) is 3 or less. Thus, the colors of a mollic epipedon range from black or very dark brown when moist and are somewhat lighter (higher value) when dry. The color criteria were established as a field criterion to separate high organic matter soils formed under grasslands from their lower organic matter counterparts under forest. Mollic epipedons often have well-developed granular structure that is quite friable. By definition, a mollic epipedon cannot be both hard and massive when dry. A minimum of 50% base saturation is required for a mollic epipedon indicating a dominance of basic cations, primarily Ca and Mg, occupying cation exchange sites.

The high base status of Mollisols generally translates into a high level of native fertility. Both calcium and Mg, typically the dominant exchangeable cations, are required in fairly large quantities for plant growth. The pH values for Mollisols may be quite variable ranging from strongly acid (5.1–5.5) to strongly alkaline (8.5–9.0). However, many Mollisols have pH values somewhere in the middle of this range and are generally considered to be favorable for plant growth without widespread use of liming agents. Their high native fertility is one of the main reasons for their extensive use in agriculture throughout the world.

Mollisols generally have not undergone intensive weathering, and therefore, their mineralogy is often dominated by minerals inherited from their parent materials. Many Mollisols have formed in recently deposited parent materials such as glacial till and loess, which has not allowed sufficient time for significant mineral weathering to occur. Mineral weathering is also limited by the cooler temperatures and/or lack of moisture that are associated with many Mollisols. Clay mineralogy of Mollisols is typically dominated by 2:1 layer silicates, including clay-sized mica (illite), smectite, and vermiculite (Allen and Fanning, 1983). Generally, kaolinite, a 1:1 layer silicate, is commonly found in Mollisols although in small quantities (Allen and Hajek, 1989). However, soils dominated by kaolinite, such as the Ewa soil series from Hawaii (a fine, kaolinitic, isohyperthermic Aridic Haplustoll), are sometimes found (USDA-NRCS, 2009).

Many of the physical characteristics of Mollisols are influenced by SOM and the associated biological processes. The relatively high SOM content of Mollisols is conducive to the formation of water-stable aggregates, which typically have crumb or granular structure, and is formed and maintained by the interaction of plant roots, microbial structures, and biological molecules (Tate, 1992). This type of stable structure is important for good water infiltration and reduced susceptibility of soils to erosion. Organic matter plays a considerable role in enhancing the water-holding capacity of Mollisols, as it can hold up to 20 times its weight of water (Stevenson, 1994). The dark colors associated with Mollisols are attributable to the humic fractions of SOM, which are black and mask the lighter colors associated with fulvic fractions (Schultze et al., 1993).

33.8.3 Classification

Soils classified as Mollisols in *Soil Taxonomy* can be divided into 8 suborders and 37 great groups. While all these soils have a mollic epipedon, they are separated into the various suborders largely on the basis of their climatic regimes. Mollisols include soils that were formerly classified as Chernozems, Brunizems, Chestnut, and Rendzina soils and some soils formerly classified as Brown, Brown Forest, Humic Gley, and Planosols (Fenton, 1983). Table 33.37 lists the suborders, great groups, and subgroups in the Mollisol order. Tables 33.38 and 33.39 list the estimated areal extent of the Mollisol suborders and great groups in the NRCS database for the United States.

33.8.3.1 Albolls

Albolls are the sixth most commonly mapped Mollisol suborder (Table 33.38). They have a well-developed, light-colored horizon (albic) and a fluctuating water table. The albic horizon forms as pigmenting agents such as SOM, clays, and Fe oxides are

TABLE 33.37 Listing of Suborders, Great Groups, and Subgroups in the Mollisols Order

Suborder	Great Group	Subgroups
Albolls	Natralbolls	Leptic, Typic
	Argialbolls	Xerertic, Vertic, Argiaquic Xeric, Argiaquic, Xeric, Aquandic, Typic
Aquolls	Cryaquolls	Vertic, Histic, Thapto-Histic, Aquandic, Argic, Calcic, Cumulic, Typic
	Duraquolls	Natric, Vertic, Argic, Typic
	Natraquolls	Vertic, Glossic, Typic
	Calciaquolls	Petrocalcic, Aeric, Typic
	Argiaquolls	Arenic, Grossarenic, Vertic, Abruptic, Typic
	Epiaquolls	Cumulic Vertic, Fluvaquentic Vertic, Vertic, Histic, Thapto-Histic, Aquandic, Duric, Cumulic, Fluvaquentic, Typic
	Endoaquolls	Lithic, Cumulic Vertic, Fluvaquentic Vertic, Vertic, Histic, Thapto-Histic, Aquandic, Duric, Cumulic, Fluvaquentic, Typic
Rendolls	Cryrendolls	Lithic, Typic
	Haprendolls	Lithic, Vertic, Inceptic, Entic, Typic
Gelolls	Haplogelolls	Lithic, Andic, Aquic, Cumulic, Typic
Cryolls	Duricryolls	Argic, Calcic, Typic
	Natricryolls	Typic
	Palecryolls	Aquic, Oxyaquic, Abruptic, Pachic, Ustic, Xeric, Typic
	Argicryolls	Lithic, Vertic, Andic, Vitrandic, Abruptic, Aquic, Oxyaquic, Calcic Pachic, Pachic, Calcic, Alfic, Ustic, Xeric, Typic
	Calcicryolls	Lithic, Vitrandic, Petrocalcic, Pachic, Ustic, Xeric, Typic
	Haplocryolls	Lithic, Vertic, Andic, Vitrandic, Aquic Cumulic, Cumulic, Fluvaquentic, Aquic, Oxyaquic, Calcic Pachic, Pachic, Fluventic, Calcic, Ustic, Xeric, Typic
Xerolls	Durixerolls	Vertic, Vitritorrandic, Vitrandic, Aquic, Paleargidic, Abruptic Argiduridic, Cambidic, Haploduridic, Argidic, Argiduridic, Haplic Palexerollic, Palexerollic, Haplic Haploxerollic, Haploxerollic, Haplic, Typic
	Natrixerolls	Vertic, Aquic Duric, Aquic, Aridic, Duric, Typic
	Palexerolls	Vertic, Vitrandic, Aquic, Pachic, Petrocalcidic, Duric, Aridic, Petrocalcic, Ultic, Haplic, Typic
	Calcixerolls	Lithic, Vertic, Aquic, Oxyaquic, Pachic, Vitrandic, Aridic, Vermic, Typic
	Argixerolls	Lithic Ultic, Lithic, Torrertic, Vertic, Andic, Vitritorrandic, Vitrandic, Aquultic, Aquic, Oxyaquic, Alfic, Calcic Pachic, Pachic Ultic, Pachic, Argiduridic, Duric, Calciargidic, Aridic, Calcic, Ultic, Typic
	Haploxerolls	Lithic Ultic, Lithic, Torrertic, Vertic, Andic, Vitritorrandic, Vitrandic, Aquic Cumulic, Cumulic Ultic, Cumulic, Fluvaquentic, Aquic Duric, Aquultic, Aquic, Oxyaquic, Calcic Pachic, Pachic Ultic, Pachic, Torrifluventic, Duridic, Calcidic, Torripsammentic, Torriorthentic, Aridic, Duric, Psammentic, Fluventic, Vermic, Calcic, Entic Ultic, Ultic, Entic, Typic
Ustolls	Durustolls	Natric, Haploduridic, Argiduridic, Entic, Haplic, Typic
	Natrustolls	Leptic Torrertic, Torrertic, Leptic Vertic, Glossic Vertic, Vertic, Aridic Leptic, Leptic, Aquic, Aridic, Duric, Glossic, Typic
	Calciustolls	Salidic, Lithic Petrocalcic, Lithic, Torrertic, Udertic, Vertic, Petrocalcic, Gypsic, Aquic, Oxyaquic, Pachic, Aridic, Udic, Typic
	Paleustolls	Torrertic, Udertic, Vertic, Aquic, Pachic, Petrocalcic, Calcidic, Aridic, Udic, Calcic, Entic, Typic
	Argiustolls	Aridic Lithic, Alfic Lithic, Lithic, Aquertic, Torrertic, Pachic Udertic, Udertic, Pachic Vertic, Vertic, Andic, Vitritorrandic, Vitrandic, Aquic, Oxyaquic, Pachic, Alfic, Calcidic, Aridic, Udic, Duric, Typic
	Vermustolls	Lithic, Aquic, Pachic, Entic, Typic
	Haplustolls	Salidic, Ruptic-Lithic, Aridic Lithic, Lithic, Aquertic, Torrertic, Pachic Udertic, Udertic, Pachic Vertic, Vertic, Torroxic, Oxic, Andic, Vitritorrandic, Vitrandic, Aquic Cumulic, Cumulic, Anthraquic, Fluvaquentic, Aquic, Pachic, Oxyaquic, Torrifluventic, Torriorthentic, Aridic, Fluventic, Duric, Udorthentic, Udic, Entic, Typic
Udolls	Natrudolls	Petrocalcic, Leptic Vertic, Glossic Vertic, Vertic, Leptic, Glossic, Calcic, Typic
	Calciudolls	Lithic, Vertic, Aquic, Fluventic, Typic
	Paleudolls	Vertic, Petrocalcic, Aquic Pachic, Pachic, Aquic, Oxyaquic, Calcic, Typic
	Argiudolls	Lithic, Aquertic, Oxyaquic Vertic, Pachic Vertic, Alfic Vertic, Vertic, Andic, Vitrandic, Aquic Pachic, Pachic, Aquic, Oxyaquic, Lamellic, Psammentic, Arenic, Abruptic, Alfic, Oxic, Calcic, Typic
	Vermudolls	Lithic, Haplic, Typic
	Hapludolls	Lithic, Aquertic, Pachic Vertic, Vertic, Andic, Vitrandic, Aquic Cumulic, Cumulic, Fluvaquentic, Fluventic, Aquic Pachic, Pachic, Aquic, Oxyaquic, Vermic, Calcic, Entic, Typic

Source: Soil Survey Staff. 2006. Keys to soil taxonomy. 10th edn. USDA-NRCS., U.S. Government Printing Office, Washington, DC. NRCS Soil Survey Staff: http://soils.usda.gov/survey/

TABLE 33.38 Area of Mollisol Suborders as Mapped in the United States

Suborder	Acres	km²	%
Ustolls	397,898,253	1,610,239	45
Udolls	188,190,505	761,581	21
Xerolls	134,345,386	543,677	15
Aquolls	112,247,357	454,249	13
Cryolls	43,078,506	174,333	5
Albolls	7,750,437	31,365	1
Rendolls	1,400,059	5,666	0.2
Gelolls	30	0.1	0.0
Grand total	884,910,533	3,581,109	100

Source: NRCS Soil Survey Staff: http://soils.usda.gov/survey/

TABLE 33.39 Area of Mollisol Great Groups as Mapped in the United States

Great Group	Acres	km²	%
Argiustolls	169,731,080	686,878	19
Haplustolls	114,527,789	463,478	13
Hapludolls	94,569,463	382,709	11
Argiudolls	81,990,684	331,805	9
Calciustolls	64,759,491	262,073	7
Argixerolls	62,415,968	252,589	7
Endoaquolls	61,711,978	249,740	7
Haploxerolls	56,061,358	226,872	6
Paleustolls	38,498,924	155,800	4
Argiaquolls	24,051,906	97,335	3
Haplocryolls	22,774,652	92,166	3
Calciaquolls	20,982,190	84,912	2
Argicryolls	17,353,940	70,229	2
Natrustolls	10,145,698	41,058	1
Argialbolls	7,750,437	31,365	1
Durixerolls	6,806,457	27,545	1
Calciudolls	6,106,031	24,710	1
Palexerolls	4,687,101	18,968	1
Calcixerolls	4,000,851	16,191	0.4
Natrudolls	3,803,116	15,391	0.4
Epiaquolls	2,290,433	9,269	0.3
Natraquolls	1,890,157	7,649	0.2
Calcicryolls	1,632,530	6,607	0.2
Paleudolls	1,601,323	6,480	0.2
Palecryolls	1,281,435	5,186	0.1
Cryaquolls	1,219,124	4,934	0.1
Haprendolls	772,135	3,125	0.1
Cryrendolls	627,924	2,541	0.1
Natrixerolls	373,651	1,512	0.04
Durustolls	235,271	952	0.03
Vermudolls	119,888	485	0.01
Duraquolls	101,569	411	0.01
Duricryolls	35,949	145	0.00
Haplogelolls	30	0.1	0.00
Grand total	884,910,533	3,581,109	100

Source: NRCS Soil Survey Staff: http://soils.usda.gov/survey/

removed from mineral grains (primarily quartz) exposing their light color. In Albolls, the albic horizon typically occurs just below the mollic surface layer and directly above a less permeable horizon of clay accumulation, an argillic horizon. Albolls often have a water table at or near the surface for several months during the winter and spring. In the United States, Albolls are most extensive in loess deposits of the Midwest (Figure 33.46).

33.8.3.2 Aquolls

Aquolls are the sixth most commonly mapped Mollisol suborder (Table 33.38). They are the wettest Mollisols and are characterized by the presence of redoximorphic features below a thick and very dark brown or black epipedon. Depending on the duration of wet conditions and Fe reduction, expression of Fe-depleted zones can range from entire horizons with predominantly dull gray colors to horizons that have a strongly contrasting mottled gray color pattern. They typically occupy the lower lying areas of the landscape where water accumulates and a high water table exists. The most extensive areas of Aquolls in the United States occur in the Red River Valley of the North, along floodplains and terraces of major rivers in the central part, and along the Gulf coastal plain (Figure 33.46). Most uses of Aquolls, whether urban or agricultural, are limited by wetness. Many Aquolls are suitable for agriculture if artificially drained. Aquolls typically have black surface horizons underlain by gray colors indicating reduction and/or depletion of Fe (Table 33.40). Organic C accumulates in the surface horizon due to reduced rates of organic matter decomposition resulting from seasonal anaerobic conditions. The presence or absence of carbonates in Aquoll profiles is quite variable and can be used as an indicator of wetland hydrology. Carbonates tend to be removed from recharge wetlands by leaching, whereas discharge wetlands are enriched often to the soil surface. Depressional basins in subhumid climates often have recharge hydrology resulting in leaching of carbonates and formation of argillic horizons (Mausbach and Richardson, 1994).

33.8.3.3 Rendolls

Rendolls form in highly calcareous parent materials of humid or cold regions, often under forest vegetation. These soils typically consist of a mollic surface layer overlying mineral material that contains >40% $CaCO_3$ by weight (Soil Survey Staff, 2010). Rendolls are associated with parent materials such as chalk, limestone, highly calcareous glacial till, and shell deposits. These soils are not extensive in the United States where only a few soils, <1% of all Mollisols, have been recognized (Table 33.38). However, Rendolls are extensive in Western Europe (Smith, 1986).

33.8.3.4 Gelolls

Recent changes in *Soil Taxonomy* have made the presence of permafrost in the soil profile the first criterion for classification at the order level (Soil Survey Staff, 2010). Soils with permafrost within 100 cm of the surface, or within 200 cm with gelic materials, classify as Gelisols (see Section 33.6). Mollisols with a gelic soil temperature regime, that is, mean annual soil temperature (MAST) ≤0°C, are classified as Gelolls; these were formerly classified as

TABLE 33.40 Selected Morphological, Physical, and Chemical Properties of an Aquoll from Southern Minnesota

Horizon	Depth (cm)	Texture	Moist Color	Clay (%)	pH	Organic C (%)	Cation Exchange Capacity ($cmol_c\ kg^{-1}$)	$CaCO_3$ (%)
Ap	0–28	Clay loam	Black	37.5	7.5	4.5	45.5	1
A	28–41	Clay loam	Black	35.9	7.2	1.3	34.9	Trace
AB	41–55	Clay loam	Very dark gray	32.8	7.4	0.5	28.9	Trace
Bg	55–71	Clay loam	Olive gray	28.6	7.5	0.4	23.0	7
Cg1	71–105	Loam	Olive gray	26.3	7.5	0.2	20.5	11
Cg2	105–125	Loam	Gray	25.3	7.6	0.2	18.8	12

Location: Waseca County, Minnesota; mean annual air temperature: 6.2°C; mean annual precipitation: 836 mm; vegetation: Reed canary grass (*Phalaris arundinacea* L.), Cattails (*Typha* sp.); parent material: Wisconsinan-age glacial till; landscape position: depression, 0%–1% slope; classification: fine-loamy, mixed, mesic, Typic Endoaquoll.

Pergelic Cryoborolls (see the discussion on the fate of Borolls in Section 33.8.3.5). Gelolls are required to have a mean summer soil temperature (MSST) of ≤8°C, if no O horizon, or ≤5°C, if an O horizon is present; otherwise, they classify as Cryolls. Gelolls are rare and essentially not useful for agriculture. They are exclusively found at high elevations and latitudes. There is only one recognized soil series in the Geloll suborder; it is the Kanauguk series, a loamy-skeletal, mixed, mesic, superactive, subgelic Lithic Haplogeloll, found in Alaska. Consequently, only 30 ac (~0.1 km^2) of Gelolls are represented in the NRCS database (Tables 33.38 and 33.39).

33.8.3.5 Cryolls

Cryolls occur in cold climates having a mean annual soil temperature <8°C with cool summers, but do not have permafrost. Cryolls are a replacement for the Cryoboroll great group of Borolls that was discontinued when the Boroll suborder was removed from *Soil Taxonomy* in 1998 (Soil Survey Staff, 1998). The fate of other Great Groups of Borolls is not as clear because not all soils that were classified formerly as Borolls have been reclassified and entered into the NRCS database. However, in addition to the placement of Cryoborolls (except Pergelic Cryoborolls, which are now Gelolls) into Cryolls, the convention for placement of former Borolls into the revised taxonomy is somewhat complicated. Borolls were defined as having a frigid, cryic, or pergelic temperature regime. As a result Borolls that had a pergelic regime and permafrost are now Gelisols, and those with a pergelic temperature regime, but no permafrost, are now Gelolls as discussed in Section 33.8.3.4. Those with a cryic temperature regime would have gone to Cryolls as discussed above. Those with a frigid temperature regime would have gone to Udolls and Ustolls but not to Xerolls because Xerolls preceded Borolls in the seventh edition key to *Soil Taxonomy* (Soil Survey Staff, 1975).

Cryolls are relatively rare, ~5% of mapped Mollisols in the United States (Table 33.38) and generally are associated with a short growing season that limits their use for agricultural production. They can, however, be productive for some cold-tolerant crops such as potatoes and for small grains and hay production where practicable. Large quantities of organic C are typically associated with such soils giving a strong black color to the mollic epipedon. Cryolls are generally found at higher latitudes or elevations. In fact, the most northerly Mollisol described in the literature is a loamy-skeletal, micaceous, isofrigid, Typic Haplocryoll found under tundra vegetation in Arctic Sweden at 68°23.32′N, 18°20.40′E, and at 775 m elevation on a 78% slope (Darmody et al., 2000). Its mean annual soil temperature of 1.2°C barely keeps it out of the Gelolls.

33.8.3.6 Xerolls

Xerolls form in areas that receive the majority of precipitation during the winter and spring months when temperatures are cool (Mediterranean climates). Little precipitation is received during the growing season and, as a result, Xerolls frequently become dry at some point during the summer. This type of moisture regime is very effective for leaching $CaCO_3$ and other soil constituents in higher precipitation zones. Conversely, Xerolls of the drier areas are subject to long periods of soil moisture deficits during the growing season. In the United States, Xerolls comprise ~15% of the land area occupied by Mollisols and are restricted to the western states (Table 33.38, Figure 33.46). An example of a Xeroll profile from the Palouse region of northern Idaho is shown in Figure 33.48, with selected data in Table 33.41. Important features of this soil include the lack of $CaCO_3$ in the profile, acidic conditions, and dense subsoil horizons (bulk density >1.7 $g\ cm^{-3}$ at 94–152+ cm). Effective leaching by winter precipitation is responsible for the lack of $CaCO_3$ and the low pH. The dense subsoil horizons restrict downward water movement and result in the formation of perched water tables during the winter and spring months. Xerolls are used primarily for production of wheat, hay, and some timber.

33.8.3.7 Ustolls

Ustolls occur in subhumid or semiarid climates receiving significant amounts of spring and summer rain (350–900 mm annual precipitation), which is characteristic of the ustic soil moisture regime (Figure 33.46). As a result, the potential productivity of these soils for agriculture varies considerably. In the higher rainfall areas, drought-sensitive crops, such as corn and soybeans, can be produced, while in the drier regions, summer fallow, which allows recharge of soil moisture in a fallowed field (typically every other year), is often necessary for crop production (Soil Survey Staff, 1999). Ustolls are extensive throughout the central Great Plains and are the dominant suborder (~45% of the total Mollisol area) in the United States (Table 33.38). Data from

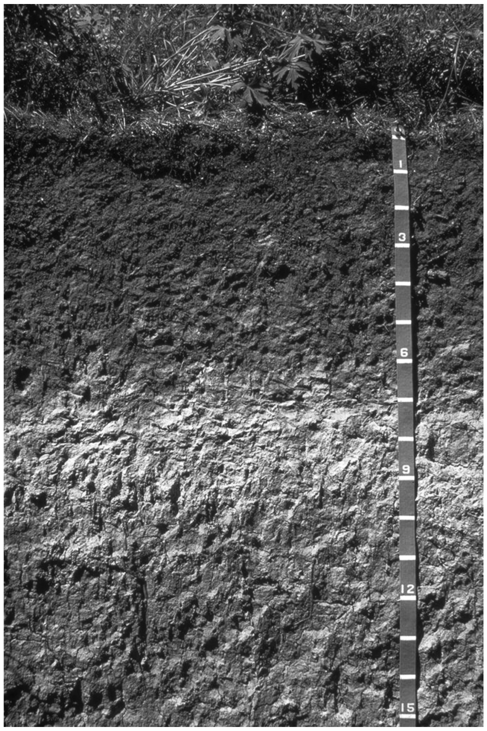

FIGURE 33.48 Soil profile of a Xeroll from northern Idaho. Accompanying data are in Table 33.41; scale is in decimeters.

a representative Argiustoll profile from Montana are presented in Table 33.42. Limited leaching in this profile has resulted in the removal of $CaCO_3$ from the upper 61 cm and the development of acid pH values near the surface. Base saturation increases with depth and is generally ~100% in lower horizons where $CaCO_3$ has accumulated. This particular pedon is from a frigid soil climate, and in this region, these soils are mainly used for winter wheat and hay production. However, often it is the dryness of Ustolls that limits agricultural productivity.

33.8.3.8 Udolls

Udolls occur in humid regions having typically formed in late Pleistocene or Holocene deposits under tall grass prairie (Figure 33.46). They are the second most commonly mapped suborder, accounting for about 21% of the Mollisols in the NRCS database (Table 33.38). Udolls are characterized by having moderate precipitation, well-distributed throughout the year (Soil Survey Staff, 1999). Because extended soil moisture deficits are not experienced, and their high natural soil fertility, most Udolls have been placed into agricultural production. Udolls are extensive in the Midwestern United States and comprise one of the most productive grain-producing regions of the world (Soil Survey Staff, 1999). The famous Morrow Plots on the University of Illinois campus, where corn has grown continuously for well over 100 years, are developed in Udolls (Darmody and Peck, 1996). Corn and soybeans, which are among the principal crops produced on Udolls, are generally not produced or produced with higher risks, on other Mollisols because of climatic limitations. Data for a Udoll from Iowa are presented in Table 33.43. Although the parent material is calcareous loess, the relatively high annual precipitation has completely removed $CaCO_3$ from the upper 175 cm. Calcium is still the dominant soil cation and, with Mg, occupies most of the CEC.

33.8.4 Pedogenic Processes

Soils classified as Mollisols develop under a variety of environmental conditions through several distinct genetic pathways. Soil genesis typically occurs over time frames where environmental conditions, and hence the theoretical equilibrium state, are variable. The morphological imprints left by grassland vegetation are ephemeral in pedogenic time frames, and hence, are probably indicative of the soil climate over the past few hundreds to thousands of years. The genesis of Mollisols will be discussed from two different perspectives (factors and processes of soil formation) in an attempt to elucidate the development of these unique soils.

33.8.4.1 Factors of Soil Formation

The factor approach applied by Jenny (1941, 1980) to the soil-forming factor model is predicated on the hypothesis that soil properties result from the interaction of at least five site and flux factors. This model proposes that soil bodies result from

TABLE 33.41 Selected Morphological, Physical, and Chemical Properties of a Xeroll (Southwick Series, 93-ID-29151) from Northern Idaho

Horizon	Depth (cm)	Moist Color	Sand (%)	Silt (%)	Clay (%)	Bulk Density (g cm^{-3})	pH	Organic C (%)
A	0–18	Very dark brown	6.9	72.2	21.0	—[a]	6.5	3.09
AB	18–38	Very dark grayish brown	7.9	72.7	19.3	1.25	5.9	1.69
Bw	36–71	Dark brown	7.7	70.0	22.3	1.30	5.8	1.20
BE	71–58	Brown	7.3	74.0	18.7	1.37	5.6	0.46
E	58–94	Grayish brown	8.7	78.3	13.0	1.48	5.5	0.46
Btb1	94–114	Brown	5.5	63.2	31.3	1.76	*5.2*	0.42
Btb2	114–152+	Brown	6.6	63.8	29.6	1.75	5.6	0.17

Location: 46° 44′N Lat., 116° 50′W Long., Latah County, Idaho; 13 km east of Moscow; elevation: 825 m; mean annual air temperature: 7°C; mean annual precipitation: 635 mm; vegetation: ponderosa pine, Idaho fescue, snowberry; parent material: late Pleistocene and Holocene loess; landscape position: near summit of gently sloping (4%) spur ridge; classification: fine-silty, mixed, superactive, mesic Oxyaquic argixeroll.

[a] Not sampled.

TABLE 33.42 Selected Morphological, Physical, and Chemical Properties of a Pedon Mapped as Shawmut, a loamy-skeletal, mixed, superactive, frigid Typic Argiustoll (S71 MT 31-1) from Montana

Horizon	Depth (cm)	Moist Color	Clay (%)	pH	Organic C (%)	Cation Exchange Capacity ($cmol_c$ kg^{-1})	$CaCO_3$ (%)	Base Saturation (%)
A1	0–6	Black	28.1	5.5	9.88	43.7	nd[a]	70
A2	6–18	Black	28.6	5.8	5.27	36.8	nd	80
Bt1	18–29	Dark brown	32.8	6.1	1.88	32.1	nd	84
Bt2	29–42	Brown	28.9	6.4	1.11	32.4	nd	100
Bt3	42–61	Brown	27.1	7.0	0.87	34.0	nd	100
Bk1	61–86	Brown	25.0	7.7	0.55	27.2	11	100
Bk2	86–130	Brown	26.2	8.0	0.28	26.8	8	100
2C	130–175	Brown	26.3	8.2	0.21	25.6	2	100

Sources: Data Courtesy of Dr. G.A. Nielsen, Montana State University, Emeritus; From Soil Survey Staff, *Soil Taxonomy: A Basic System of Soil Classification for Making and Interpreting Soil Surveys*, 2nd edn., USDA Agriculture Handbook No. 436, U.S. Government Printing Office, Washington, DC, 1999.

Location: Gallatin County, Montana; elevation: 1585 m; mean annual air temperature: 5°C; mean annual precipitation: 510 mm; vegetation: bluebunch wheatgrass, Ldaho fescue, prairie junegrass, sagebrush; parent material: mixed alluvium/colluvium; landscape position: alluvial/colluvial fan, 4% slope.

[a] Not detected.

TABLE 33.43 Selected Morphological, Physical, and Chemical Properties of a Udoll (S63 Iowa-83-2) from Iowa

Horizon	Depth (cm)	Texture	Moist Color	Clay (%)	pH	Organic C (%)	Cation Exchange Capacity ($cmol_c$ kg^{-1})	Exch. Ca ($cmol_c$ kg^{-1})
Ap	0–18	Silty clay loam	Very dark brown	30.4	5.6	2.20	22.0	13.9
A	18–33	Silty clay loam	Very dark brown	33.5	5.7	1.87	22.9	14.7
AB	33–46	Silty clay loam	Very dark grayish brown	32.8	5.8	1.11	21.6	14.8
Bw1	46–69	Silty clay loam	Dark brown/brown	30.4	5.8	0.58	20.0	14.8
Bw2	69–86	Silty clay loam	Dark brown/brown	28.2	5.9	0.33	20.7	14.7
BC1	86–110	Silt loam	Yellowish brown	26.9	5.9	0.21	20.4	14.6
BC2	110–125	Silt loam/silty clay loam	Yellowish brown/olive gray	28.0	6.0	0.17	20.7	15.2
C1	125–145	Silt loam	Brown/olive gray	26.9	6.0	0.11	20.8	14.6
C2	145–175	Silt loam	Yellowish brown/olive gray	25.7	6.2	0.10	19.7	13.8

Sources: Data from Soil Survey Staff, *Soil Taxonomy: A Basic System of Soil Classification for Making and Interpreting Soil Surveys*, USDA Agriculture Handbook No. 436, U.S. Government Printing Office, Washington, DC, 1975, Appendix 4; Data courtesy of USDA-Natural Resources Conservation Service, Soil Survey Division, World Soil Resources, 2009.

Location: Shelby County, Iowa; 135 km West of Des Moines; mean annual air temperature: 9°C; mean annual precipitation: 720 mm; vegetation: cropland; parent material: late Pleistocene loess; landscape position: axis of short interfluve, 3% slope; classification: fine-silty, mixed, mesic Typic Hapludoll.

the action of climate, organisms, and relief acting upon parent material over time and provides a useful conceptual framework by which soil differences can be attributed to state or site factors.

33.8.4.1.1 Organisms

To a large degree, Mollisols are distinguished from other soils on the basis of organisms, or more specifically, the overriding effects of grassland vegetation. The reason why grassland vegetation imparts distinctive soil morphologies is discussed later. While some Mollisols have developed under consistent vegetative conditions, abundant evidence suggests that many were forested during glacial and postglacial periods (Walker, 1966; Ruhe, 1970; Wright, 1970). Similar evidence also suggests that the climate has changed sufficiently during the Holocene to allow fluctuations between forest and prairie vegetation, especially along humid to subhumid ecotones (Walker, 1966; Geis and Boggess, 1968). Most soils classified as Mollisols (with the possible exception of some Rendolls, Aquolls, and Albolls) have probably had grassland vegetation at some time during their development. Because soil characteristics reflect the integration of the soil-forming factors over long time periods (centuries to millennia), the contemporary environment may or may not accurately reflect the conditions under which a soil has developed. Mollisols are common in transitional areas between humid and subhumid environments such as the north central United States. Changes in climate during the Holocene have probably resulted in the formation and subsequent degradation of Mollisols as forest encroached on areas that were previously prairies (Fenton, 1983).

The mixing and incorporation of organic matter within the mollic epipedon have also been attributed to soil faunal activity. Of particular importance are burrowing organisms, such as gophers, prairie dogs, worms, and ants (Thorp, 1948; Munn, 1993). The net effect of burrowing activities is to move organic matter lower in the soil profile and to mix soil materials, resulting in a rather homogeneous epipedon.

33.8.4.1.2 Climate

As previously suggested, climate is also a key factor in the formation of Mollisols, primarily through its effect on vegetation. Grassland ecosystems are associated with broad ranges in mean annual precipitation and temperature. Consequently, Mollisols occur under a wide range of soil temperature and moisture regimes. In the United Sates, these soils are found from as far south as Texas to the Canadian border and into Alaska. Mollisols are most common in the zone between the arid deserts of the Western states and the forests of the Midwestern states, a range of approximately 300 mm to more than 900 mm in annual precipitation (Figure 33.46).

The effects of climate on Mollisol formation can be demonstrated by comparing soil morphology along climatic gradients. Munn et al. (1978) examined a warm/dry-to-cool/moist sequence of grassland soils in Montana. They observed an increase in annual aboveground biomass production with increasing precipitation and decreasing temperature (Figure 33.49). Similarly, SOM content, which increased as well, is expressed by increasing thickness of the mollic epipedon. This suggests that the dark, organically enriched mollic epipedon provides a record of the balance between annual site production and microbial decomposition. As production increases with increasing precipitation, decreasing temperature favors the accumulation of SOM by decreasing microbial decomposition rates. Leaching of $CaCO_3$ in the soils was greater with increasing annual precipitation, indicating that depth to carbonate serves as an indicator of rainfall within a geographic region.

33.8.4.1.3 Relief

The local topography or hillslope position affects hydrologic processes, especially erosion and sedimentation and the timing, duration, and depth of seasonal soil saturation. Many glacial till landscapes of the eastern portion of the central United States contain both Mollisols and Alfisols, the "Prairie Peninsula" (Geis and Boggess, 1968). Aquolls and Albolls occur in the wetter, lower lying portions of the landscape (darker soil surface; Figure 33.50). Conversely, Alfisols occupy the adjacent higher positions having better drainage (Allen and Fanning, 1983).

Richardson et al. (1992) proposed a framework to view wetland hydrology based on differences in how water flows through depressional basins (recharge, flowthrough, or discharge). Differences in basin hydrology result in distinctive differences in soil morphology along hillslopes (Thompson and Bell, 1996; Bell and Richardson, 1997). Examination of a toposequence of Mollisols from south central Minnesota illustrates typical catena relationships (Figure 33.51). Along this hillslope continuum, the following changes occur from upper to lower landscape positions: (1) increased darkness (organic C) and thickness of the surface

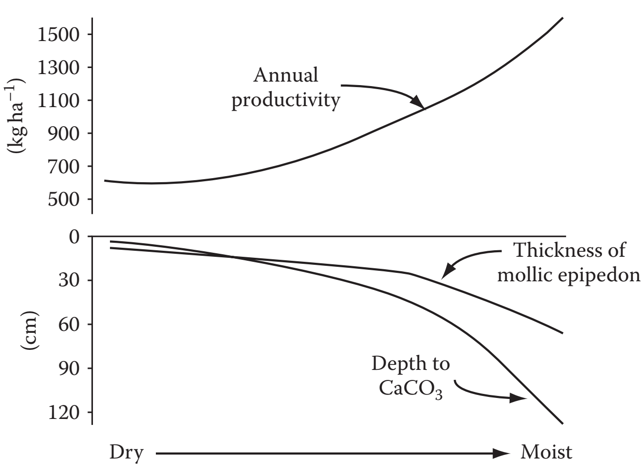

FIGURE 33.49 Generalized relationship between aboveground productivity in rangelands in Western Montana, depth to $CaCO_3$, and thickness of mollic epipedon. (Adapted from Munn, L.C., G.A. Nielsen, and W.F. Mueggler. 1978. Relationships of soils to mountain and foothill range habitat types and production in western Montana. Soil Sci. Soc. Am. J. 42:135–139. With permission of the Soil Science Society of America.)

FIGURE 33.50 Glacial till landscape in central Indiana (the easternmost portion of the "prairie peninsula") showing patterns of Alfisols (light tones) and Mollisols (dark tones). Wet Mollisols (Aquolls) occupy lower lying, wetter portions of the landscape. Alfisols are found in the higher, better drained positions. (Photo used with permission of the Soil Science Society of America.)

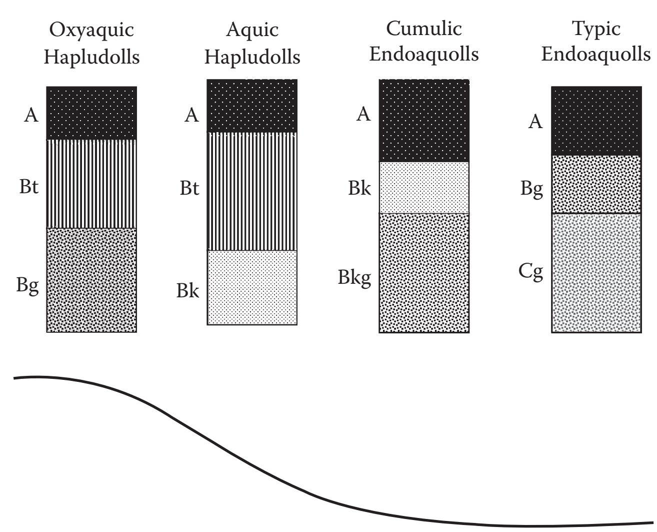

FIGURE 33.51 Toposequence of soil profiles along a hillslope in glacial tills near Waseca, Minnesota.

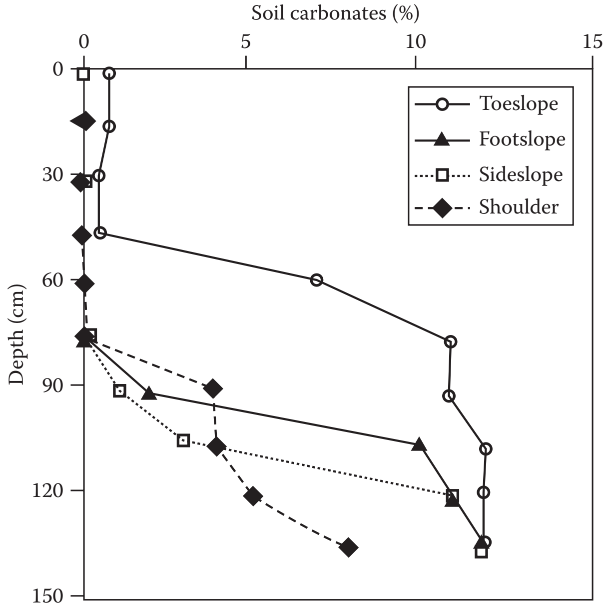

FIGURE 33.52 Depth distributions of soil carbonate concentrations at four hillslope positions in glacial till near Waseca, Minnesota.

horizon, (2) decreased depth to redoximorphic features, and (3) decreased chroma of the subsurface horizons. These morphological differences can often be traced back to differences in soil processes, such as erosion, deposition, and reduction that are affected by hydrology and the shape and position on the hillslope. Numerous studies have documented increased duration of saturation in lower landscape positions for similar landscapes (Khan and Fenton, 1994; Bell et al., 1995; Thompson and Bell, 1996). For the example from Waseca, Minnesota (Figure 33.52), higher carbonate concentrations between 45 and 90 cm at the toeslope are probably indicative of discharge hydrology and subsequent enrichment of the soil profile from carbonate laden groundwaters. The distribution of organic C with depth along the hillslope clearly indicates a trend of increasing organic C in the surface horizons from upper to lower hillslope positions (Figure 33.53). Organic C concentrations tend to converge at a depth of approximately 75 cm for all landscape positions where concentrations fall below 0.5%. These same spatial relationships cannot be assumed for all hillslopes as differences in soil stratigraphy, regional hydrology, and climate can cause different spatial patterns to develop.

33.8.4.1.4 Parent Material

Mollisols occur on a variety of substrate materials, but are most commonly associated with unconsolidated sediments from coastal, riverine, or glacial depositional environments, including loess (Soil Survey Staff, 1999). Loess is a common parent material of Mollisols in the Midwestern United States and in southern and central Russia. Duchaufour (1982) has suggested that the permeability and $CaCO_3$ content of loess is especially conducive to the formation of Mollisols. Many Mollisols have formed in calcareous parent materials resulting in appreciable quantities of $CaCO_3$ within the soil profile, particularly those Mollisols in

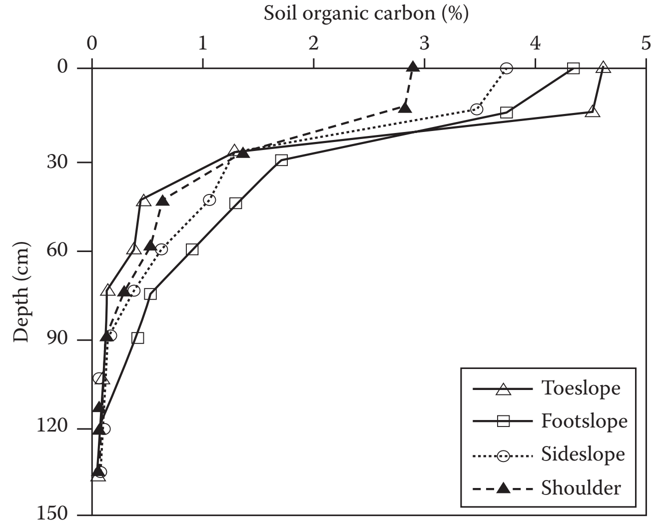

FIGURE 33.53 Depth distributions of soil organic carbon at four hillslope positions in glacial till near Waseca, Minnesota.

lower precipitation areas. Rendolls form exclusively in highly calcareous parent materials such as chalk, limestone, and shell deposits.

33.8.4.1.5 Time

Most Mollisols in the United States and Russia appear to be relatively young and have formed on late Pleistocene and Holocene deposits and surfaces (Soil Survey Staff, 1999). However, the recognition of Mollisols in intertropical areas may warrant reexamination of this concept. It has been shown that accumulation of SOM occurs relatively fast. Schafer et al. (1980) compared organic C levels in newly created mine soils with those of adjacent undisturbed grassland soils of eastern Montana and found that only 30 years was required for C to build up in the top 10 cm to the level in undisturbed soils. In Iowa, the rate of formation of a Udoll surface horizon was estimated at 0.08 cm year^{-1} (Buol et al., 2003). Foss et al. (1985) suggested that mollic epipedons developed in <900 years in the Red River Valley of the North in Minnesota and North Dakota. From these and other studies, it is clear that the formation of a mollic surface horizon can occur in relatively short time spans.

33.8.4.2 Genetic Pathways for Mollisol Formation (Soil Processes)

The accumulation of organic C in the upper soil profile is the salient morphological feature distinguishing Mollisols from other soil orders. The C content is high when the long-term rate of addition and retention exceeds that of decomposition. Rates of annual root production and decomposition in grasslands result in high rates of organic C turnover in the soil to depths often approaching 50–100 cm (Dahlman and Kucera, 1965; Dormaar and Sauerbeck, 1983). Thorp (1948) estimated that 113–409 kg ha^{-1} of raw OM (dry weight) were added annually for short grass and 136–500 kg ha^{-1} for tall grass prairies. Alternatively, soil organic C accumulates when rates of microbial decomposition are low due to either anaerobic (wet) conditions or cool

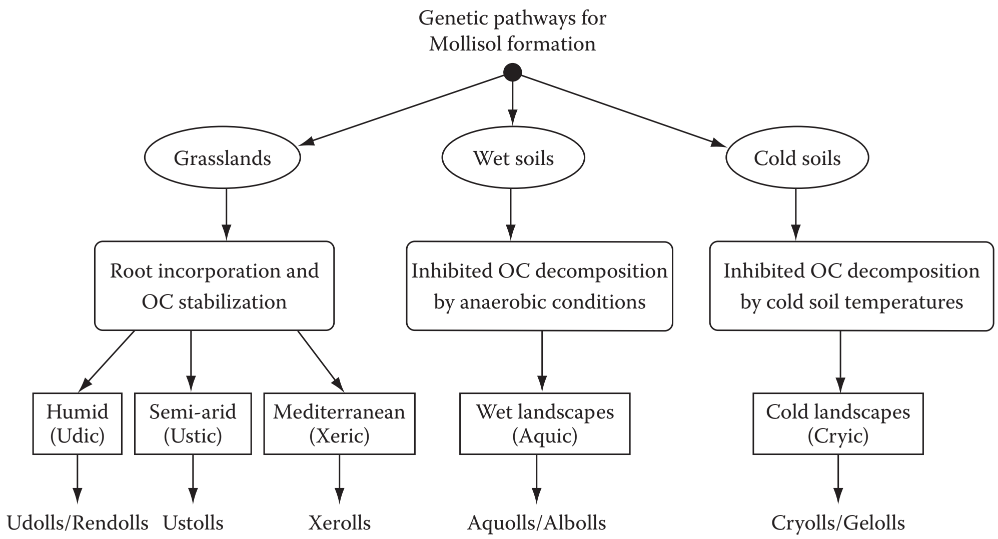

FIGURE 33.54 Theoretical pathways for Mollisol genesis.

soil temperatures (Jenny, 1930). Three primary pathways for Mollisol genesis based on the mode of organic C accumulation can be identified: (1) high rates of accumulation in grasslands, (2) low rates of decomposition under anaerobic (wet) conditions, and (3) low rates of decomposition in cold climates (Figure 33.54). These pathways are not necessarily mutually exclusive. For example, many wet soils may have also developed under grassland, but wet grassland soils frequently have higher organic C contents than drier soils in the same landscape (Figures 33.50 and 33.53).

33.8.4.2.1 Grassland Mollisols (Udolls, Ustolls, and Xerolls)

A comparison of the chemical composition and growth forms of forests and grasslands explains why soils under these covers have distinctive morphologies. In forests, organic C tends to accumulate on the soil surface from annual decomposition cycles of forest floor litter. Despite the fact that the forested soils (Alfisols) receive more annual precipitation, they contain less organic C throughout the mineral soil and lack the dark soil colors associated with Mollisols. Much of the organic matter associated with the forest soils is contained in litter layers lying on top of the mineral soil. Leachate from the decomposition of forest floor litter is often composed of carbonic, fulvic, and/or tannic acids facilitating translocation (eluviation) of soil materials deeper within the soil profile resulting in morphological features usually associated with Alfisols and Spodosols. These conditions and processes inhibit the accumulation and stabilization of organic C in the upper soil layers.

By contrast, photosynthesis and other metabolic processes in grassland vegetation quickly transport organic C to dense and fibrous root systems. Estimates indicate that grasslands have the highest annual additions of C to soil of any of the terrestrial ecosystems, including tropical forests (Bolin et al., 1979). Related research has demonstrated that the greatest biomass production in grassland ecosystems is below ground (Caldwell, 1975; Lauenroth and Whitman, 1977; Jenny, 1980). As such, Mollisols

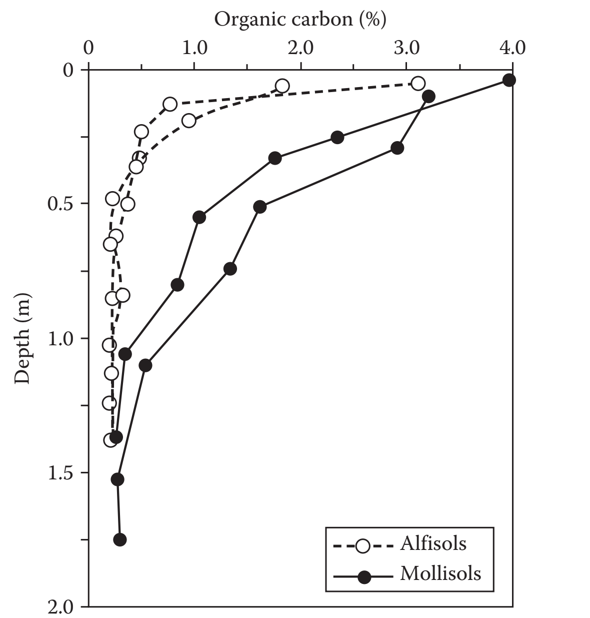

FIGURE 33.55 Depth distribution of soil organic carbon in Mollisols and Alfisols of the Palouse region of Northern Idaho. The Mollisols have formed under Idaho fescue grassland with approximately 580 mm of annual precipitation; the Alfisols have formed under grand fir forest with approximately 830 mm of annual precipitation. (Data courtesy of the Soil Characterization Laboratory, University of Idaho, Moscow, ID.)

have distinctive organic matter profiles with depth that differ markedly from Alfisols, where most of the biomass is produced aboveground (Figure 33.55). The annual proliferation of roots and their subsequent decomposition is responsible for the deep accumulation of SOM leading to the formation of mollic epipedons whose thickness is largely determined by the depth and amount of grass roots (Hole and Nielsen, 1968; Cannon and Nielsen, 1984). Incorporation of organic C is also facilitated by the mixing of near-surface soil horizons by ants (Formicidae), earthworms (*Lumbricus*), and other soil fauna (Curtis, 1959; Baxter and Hole, 1967). Schlesinger (1977, 1991) have estimated that temperate grassland soils have a mean organic matter content of 19.2 kg C m^{-2}, ranking them behind only soils of wetlands and tundra/alpine ecosystems on a global scale.

Grasses have another effect that differs from forest vegetation on soil genesis. Grasses are generally better base cyclers. That, in part, accounts for the higher fertility of Mollisols when compared with forest soils such as Alfisols. In addition, that difference also accounts for the absence of fragipans in Mollisols (Franzmeier et al., 1989).

Once added to the soil, organic materials undergo further decomposition by complex, microbially mediated processes that results in the formation of a relatively stable organic fraction. The large annual additions of C to Mollisols and subsequent cycling result in the formation of an active (bioavailable) as well as a very stable organic fraction. The humus contained in Mollisols appears to be more stable than that found in other soils (Stevenson, 1994), which is possibly related to the chemical composition of grasses and soil parent material. Grasses have a higher ratio of humic to fulvic acids (Glazovskaya, 1985). Novak and Smeck (1991) found higher concentrations of humic substances in the surface horizons of Mollisols compared with Alfisols in southwestern Ohio where these soils are adjacent in the same landscape. Minimal differences were found in fulvic acid content. The combination of humic substance in the presence of calcareous soil parent material results in the formation of Ca humates that are thought to bind silicate grains to organic C, which becomes stabilized (Evans and Russell, 1959; Stevenson, 1994). Formation of these complexes increases the resistance of SOM to physical disintegration, chemical extraction, and further biological change through microbial decomposition (Oades, 1989; Stevenson, 1994). Estimates as to the extent of organic matter–mineral associations in grassland soils vary. As much as 80% of the organic C in a mollic epipedon may be so closely associated with the mineral fraction that it cannot be separated by physical means (Greenland, 1965; McKeague et al., 1986). Researchers have been able to determine mean residence time (MRT) of organic fractions of Mollisols in several different environments. The MRT represents the average length of time that an organic fraction has been present in the soil. Oldest MRT values measured in soils are associated with Mollisols and Histosols (Oades, 1989). For example, MRT values between 1255 and 2973 years have been measured for Mollisols in Canada and the United States (Anderson and Paul, 1984; Hsieh, 1992).

The stabilization of organic C in mollic epipedons is also facilitated by soil clays. Organic C associated with fine clays is protected from further rapid decomposition lengthening the turnover time of otherwise labile humic substances from days to months, years, or even decades (Anderson and Paul, 1984). Anderson (1979) found that as much as 50% of the total humus in some grasslands is associated with clay. From these and other studies, it appears that texture is an important factor in determining the stable level of organic C in Mollisols as well as in other soils. Texture influences the water-holding capacity of a soil and the quantity of clays available to form

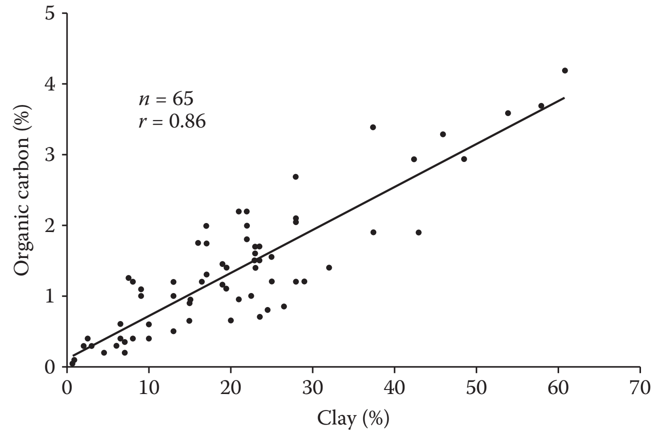

FIGURE 33.56 Relationship between clay and organic matter content in warm grassland soils of the southern United States. (Adapted from Nichols, J.D. 1984. Relation of organic carbon to soil properties and climate in the southern Great Plains. Soil Sci. Soc. Am. J. 48:1382–1384. With permission of the Soil Science Society of America.)

complexes with organic matter. Nichols (1984) observed a good correlation ($r = 0.86$) between soil organic C and clay content in 65 Mollisols and associated soils of Texas, Oklahoma, and New Mexico where mean annual temperatures exceed ~15°C (Figure 33.56). These data indicate that water-holding capacity and the protection afforded to humus through formation of mineral–organic complexes may control the equilibrium organic C contents in Mollisols of warmer regions. In contrast, clay content does not appear to exert such strong control of organic C contents in Mollisols of cooler regions. McDaniel and Munn (1985) found little correlation between SOM and clay in 137 Mollisols and associated soils in Montana and Wyoming having mean annual soil temperatures <8°C (Figure 33.57). Comparison of these data with those of Nichols (1984) suggests that it is temperature rather than moisture that becomes the determining factor in establishing an equilibrium SOM content. Furthermore, organic matter may be able to

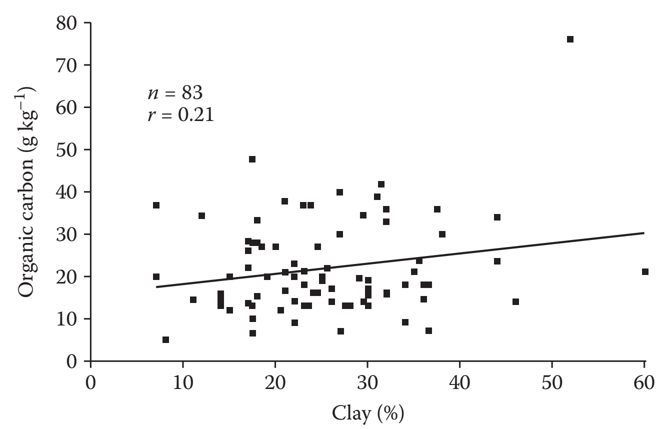

FIGURE 33.57 Relationship between clay and organic matter content in cool grassland soils of the northern United States. (Adapted from McDaniel, P.A., and L.C. Munn. 1985. Effect of temperature on organic carbon–texture relationships in Mollisols and Aridisols. Soil Sci. Soc. Am. J. 49:1486–1489. With permission of the Soil Science Society of America.)

persist in cold soils without the stabilizing influence afforded by complexation with clays (McDaniel and Munn, 1985).

33.8.4.2.2 Wet Mollisols (Albolls and Aquolls)

An alternate genetic pathway for the development of Mollisols involves organic C accumulation under anaerobic soil conditions due to reduced rates of organic matter decomposition (Ponnamperuma, 1972; Gambrell and Patrick, 1978; Tate, 1980). These Mollisols have an aquic soil moisture regime and are classified as Aquolls or Albolls. The decomposition of organic matter is less efficient under anaerobic conditions, and a net accumulation of SOM will occur if rates of biomass production are sufficiently high. If this rate of organic matter accumulation is high enough, organic soils (Histosols) will develop. In many landscapes, Mollisols are found at the transition between organic and mineral soils.

In general, Aquolls have black (N 2/0) surface horizons with predominant gleyed or depleted horizons directly beneath the surface horizon (Table 33.40). Increases in the duration of soil saturation and anaerobic conditions are usually associated with higher soil organic C contents (Khan and Fenton, 1994; Bell et al., 1995). Topographically, organic C increases downslope as evidenced by the darker soil surfaces in lower landscape positions (Figure 33.51) associated with higher surface concentrations of organic C in depressional areas (Figure 33.53). Thompson and Bell (1996) found good agreement between a profile darkness index based on the thickness and darkness of the surface horizons and the duration of saturation for soil hydrosequences at several locations in the United States. They also found that the profile darkness index was highly correlated with soil organic C for the soils that were derived from glacial deposits.

Soil saturation, the development of anaerobic conditions and subsequent reduction of Fe(III) to Fe(II), causes the formation of distinctive color patterns (redoximorphic features) in the horizons beneath the mollic epipedon. Biochemical reduction results in the translocation of Fe compounds, which are the primary determinants of soil color in low organic matter horizons. The processes involved in the development of redoximorphic features are discussed in Chapter 15. If the soil water is stagnant, Fe(II), which has a distinctive bluish or greenish color, may be present, thus creating a reduced (or gleyed) soil matrix. More commonly, the soil water is moving, and Fe(III) coatings on mineral grains will be reduced and removed as Fe^{2+} ions, leaving the mineral grains and revealing the dull, gray color of the uncoated minerals, thus forming redoximorphic depletions. Under conditions of fluctuating water tables, Fe(II) may be reoxidized creating discrete concentrations of orange or reddish Fe(III) (Vepraskas and Sprecher, 1997). Bell and Richardson (1997) discuss specific soil process and morphologies associated with Aquolls and Albolls.

33.8.4.2.3 Cold Mollisols (Gelolls and Cryolls)

The formation of Mollisols under cold conditions (i.e., cryic or gelic soil temperature regimes) is somewhat analogous to the formation of wet Mollisols. Low soil temperatures (similar to lack of soil O_2) reduce microbial activity and facilitate organic matter accumulation if the long-term rate of biomass production exceeds that of decomposition. Again, as previously covered, where permafrost is within 2 m of the soil surface, the soils are classified as Gelisols, and further classified as Histels if they are dominantly composed of organic soil materials (see Section 33.2). Where permafrost is lacking, or deeper than 2 m, and where rates of organic matter accumulation are high, the soils are dominated by organic matter and are in Cryo great groups of Histosols (Cryofibrists, Cryofolists, Cryohemists, and Cryosaprists). Cold Mollisols, which are classified as Cryolls or Gelolls, develop in areas where organic matter accumulation rates are somewhat lower than are necessary for organic soil materials to accumulate. Gelolls and Cryolls, along with their wet but warmer analogs, Aquolls and Albolls, often develop under grasslands, or tundra-like vegetation in the case of the cold Mollisols, but can develop under other types of vegetation where conditions are favorable.

33.8.4.3 Associated Pedogenic Processes

The following processes are also important in the formation of some Mollisols.

33.8.4.3.1 Carbonate Translocation

The dissolution of $CaCO_3$ and its subsequent precipitation as secondary carbonates lower in the soil profile is common in many calcareous parent materials where sufficient leaching takes place. Carbonates in layers of maximum accumulation are more finely divided than those associated with the parent material supporting secondary carbonate formation (Redmond and McClelland, 1959). Total precipitation is important in determining the depth at which $CaCO_3$ is deposited (Jenny, 1941). In many Xerolls, where precipitation is low but exceeds evapotranspiration especially in winter, $CaCO_3$ is readily mobilized and reprecipitated at depth. In the Palouse region of eastern Washington and northern Idaho, $CaCO_3$ has been removed from the upper 1.5 m of the profile in Xerolls receiving more than 530 mm of rainfall (Barker, 1981). In contrast, Ustolls of the Great Plains receiving comparable amounts of total rainfall typically have well-developed zones of $CaCO_3$ accumulation within 1 m of the soil surface. The Ustolls receive much of their precipitation during the growing season when evapotranspirational demand is high, resulting in less moisture being available for mobilization of $CaCO_3$. As previously discussed, spatial patterns of carbonate removal and accumulation along hillslopes can often be used to interpret soil hydrology.

33.8.4.3.2 Clay Translocation

The movement and accumulation of clay are common soil-forming processes that occur in soils occupying relatively stable landscape positions and depression-focused recharge depressional basins. Clay translocation causes differences in soil texture that can affect soil water movement and the subsequent genesis of certain soil morphological features. This process is thought to be important in the formation of Albolls. Appreciable clay movement does not occur until carbonates have been removed from the upper portions of the soil profile (Duchaufour, 1982; Fanning and Fanning, 1989). Carbonate removal releases clay previously cemented, which may be subject to further chemical alteration prior to movement (Fenton, 1983). In calcareous parent materials, subsoil clay accumulation commonly occurs immediately above the zone of $CaCO_3$ accumulation. In Mollisols, subsurface clay accumulation is highly variable and is not always observed.

In a study of Mollisols and associated soils in the forest steppe–dry steppe transition in Eastern Europe and the forest–prairie transition in the United States Great Plains, Bronger (1991) was unable to find micromorphological evidence of clay illuviation in Mollisols commonly described as having argillic horizons. While Nettleton et al. (1969) proposed that the lack of illuvial clay skins in some fine-textured soils was due to disruption by shrink/swell clays, Bronger (1991) suggested that increased clay contents in Mollisol subsoils are either lithologic discontinuities or are polygenetic in origin and, as such, are relict features that formed under past, moister climates. Sobecki and Wilding (1983) investigated Texas Coast Prairie Mollisols and found that argillic horizons were confined to microtopographic lows, suggesting depression-focused recharge. Weakly developed Bt horizons on microtopographic highs were determined to be relict features based on micromorphological analysis. They suggested that processes leading to both carbonate accumulation and clay illuviation in the same horizon were incompatible under the current humid climate.

33.8.4.3.3 Erosion and Sedimentation

Erosional processes result in the redistribution of surface soil material from upper to lower hillslope positions. This process often results in thickening of the mollic epipedons in concave slope positions. Soils with these overthickened surface horizons are often classified as cumulic at the subgroup level. Anthropogenically accelerated erosion causes significant reduction in the thickness of upland A horizons, which no longer meet the surface horizon thickness criteria for Mollisols. These eroded soils will typically be classified as Inceptisols or Alfisols (if an argillic horizon is present). Erosional losses as a genetic pathway for Alfisols or Inceptisols, which is contrary to the conceptual models for the genesis of these soils, has been a longstanding taxonomic dilemma in the classification of Mollisols (Fenton, 1983). The choice presented to soil mappers is between recognizing eroded phases of Mollisols, or classifying the soils based on the thinner epipedon resulting from anthropogenically accelerated soil erosion.

33.8.5 Land Use

Human impact on the use of Mollisols has increased dramatically over the past century. Early cultures had little influence on grassland soils other than the use of fire as a management tool. In the plains of North America, burning was used to stimulate the subsequent year's production and attract buffalo (Warkentin, 1969; Anderson, 1987). With the advent of the moldboard plow, Mollisols became widely used as arable soils. During the early 1900s, artificial drainage (tile drains and surface ditching) and flood protection measures lowered water tables sufficiently for the conversion of extensive areas of Aquolls and other marginally wet Mollisols to productive cropland. In the United States, where Mollisols make up 22% of the land area, much of the agriculturally based, westward settlement was a direct result of the ease with which productive Mollisols could be cleared by fire. The high level of native fertility and favorable physical properties meant that most areas of Mollisols were converted to row cropping. Virgin areas of Mollisols and associated prairies are now quite rare in the United States.

Mollisols are among the most agriculturally productive soils in the world. Although Mollisols make up only ~7% of the Earth's ice-free land area, a large part of the world's wheat and other small grains are grown on Ustolls and Xerolls (Troeh and Thompson, 1993). Because Mollisols are commonly found in subhumid climates, lack of sufficient soil moisture can limit the production of traditional agricultural crops. Soil moisture conservation strategies, such as fallowing and residue management, must be implemented for sustainable dryland farming. Unfortunately, the use of unsustainable farming practices not designed to protect the soil from erosion or to conserve soil organic C has led to a decline in soil quality and subsequent productivity in some regions of the world. Long-term cultivation has reduced the organic C content of Mollisols, and reductions of ~35% in C contents in 60–70 years of cultivation have been documented (Tiessen et al., 1982). These reductions are often associated with degradation of soil structure and increased susceptibility to erosion. More recent research into the impacts of agriculture on the nearly level Aquolls and Udolls in Illinois indicated that these soils are quite resilient. After an initial loss of SOM when the native prairie is cultivated, they retain much of their organic C even after a century of row crop production (David et al., 2009). However, in most cases, the conversion of prairie to cropland further increases the erosion potential and has resulted in devastating wind erosion during cyclic droughts in portions of Asia and North America, most famously in the serious wind erosion of Ustolls during the Dust Bowl in the 1930s. These processes of soil degradation result in loss of organic matter and organically bound nutrients and subsequent declines in soil productivity. The use of appropriate soil conservation practices can greatly reduce soil degradation and maintain soil productivity.

As population pressure has increased, conversion of Mollisols from arable to urban uses increased. Many of the same characteristics that make Mollisols preferable for cropland also favor many nonagricultural uses. The conversion of these prime agricultural soils, and the associated decline in the resource base for food and fiber production, has generated much debate in the United States and stimulated farmland preservation efforts.

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33.9 Spodosols

Randall Schaetzl

Willie Harris

33.9.1 Introduction

Spodosols are soils that typify the process of podzolization; they are roughly equivalent to Podzols in older U.S. classifications schemes (Baldwin et al., 1938; Muir, 1961; Petersen, 1976, 1984; McKeague et al., 1983). (The term Podzol is still widely used in many classification schemes outside of the United States.) In the process of podzolization, soluble compounds of Al and organic matter, often accompanied by Fe and Si, are translocated from an upper, eluvial zone to a lower, illuvial zone. As a result, a typical Spodosol profile has a bleached E horizon overlying a dark, reddish to brownish Bs, Bh, and/or Bhs horizon(s). The process is best expressed in humid climates, under vegetation assemblages that tend to produce acidic litter. Spodosols are most extensive in cool-cold climates, but also are common in subtropical and tropical regions, where eluvial and/or illuvial zones can be quite thick and well expressed (Daniels et al., 1975; Thompson, 1992; Figure 33.58). These thick illuvial zones constitute a substantial pool of C (Holzhey et al., 1975; Stone et al., 1993), one that has not yet been definitively tallied in many estimates of global C stocks.

The original Podzol concept derived from the ashy, light colored E horizon, having been derived from the Russian words “pod” (under) and “zola” (ash) (McKeague et al., 1983; Sauer et al., 2007). This name has often been mistakenly applied to all soils with a light colored, high-chroma E horizon, regardless of the nature of the underlying horizons. Thus, soils with a well-developed E horizon and undergoing various degrees of podzolization needed to be separated from those that embodied the core concept of the true podzolization process. Therefore, in an attempt to restrict the order Spodosols to soils formed mainly via podzolization, the Soil Survey Staff (1975, 1999) placed the taxonomic focus on B horizon character, chemistry, and morphology, for example, Stanley and Ciolkosz (1981), Mokma (1993). Important aspects of the current Spodosol classification scheme for the United States (Soil Survey Staff, 1999) are discussed in the following.

33.9.2 Processes Involved in the Formation of Spodosols—the Podzolization Suite

In podzolization, a combination of organic matter and Al, often in association with Fe^{3+}, are translocated from the upper profile to a lower, illuvial horizon (Petersen, 1976; DeConinck, 1980; Buurman and van Reeuwijk, 1984; Courchesne and Hendershot, 1997, Lundström et al., 2000a). Some literature suggests that podzolization includes not only translocation of metals (sesquioxides) and organic matter, but also the associated processes of (1) chemical weathering by organic and carbonic acids and (2) clay destruction and/or translocation (Fridland, 1958; Schaetzl and Anderson, 2005). Regardless, in both definitions, podzolization processes operate under acidic conditions, where weathering is intense in the upper profile, and where the weathering byproducts are translocated to the B horizon, or removed from the profile. As a result, an acidic E horizon is formed—impoverished in

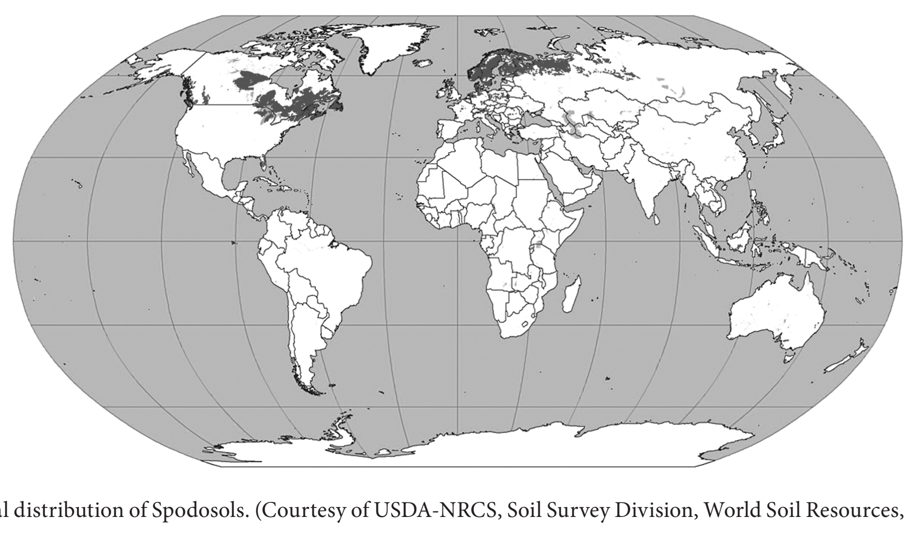


FIGURE 33.58 Global distribution of Spodosols. (Courtesy of USDA-NRCS, Soil Survey Division, World Soil Resources, Washington, DC. 2010.)

FIGURE 33.59 Relict treethrow pits and mounds in a landscape dominated by Spodosols. This landscape was originally forested but is now in pasture. It is likely that the microtopography has been muted somewhat by the decades of grazing at this site. Location: Northeastern Wisconsin. (Photo by R. Schaetzl.)

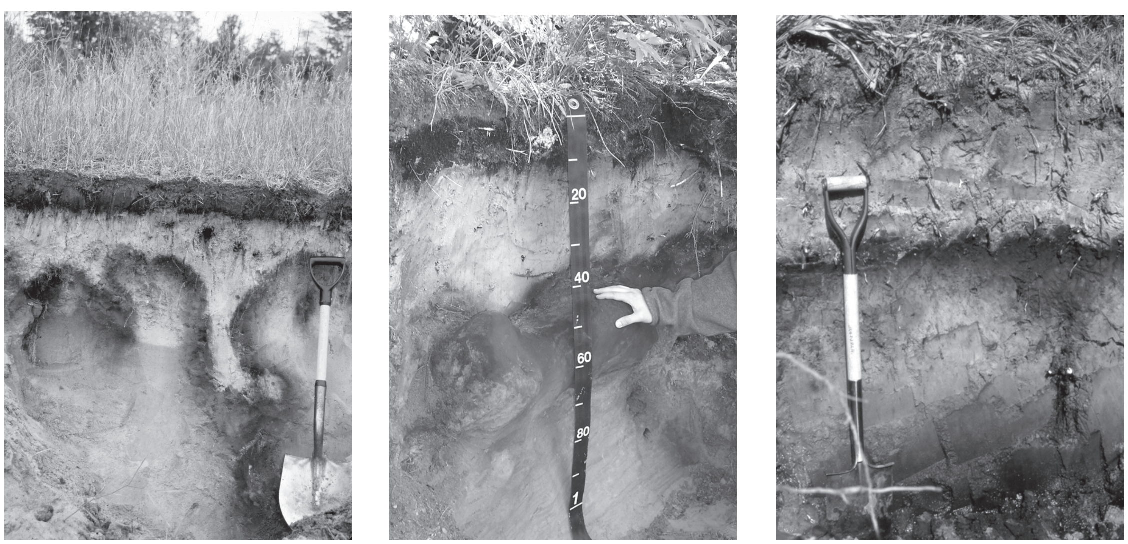
Al, organic matter and usually Fe. In many soils, the result of podzolization is a profile that classifies as a Spodosol (or Podzol). It is important to note, however, that podzolization is also a subsidiary or background process in many soils. For example, it may occur after, or in association with, acidification, lessivage, brunification, and various intensities of redox processes (DeConinck and Herbillon, 1969; Guillet et al., 1975; Ugolini et al., 1977; Li et al., 1998; Harris and Hollien, 1999, 2000; Hseu et al., 2004).

33.9.3 Characteristics of a Spodosol Profile

Spodosols often have distinct horizons with relatively abrupt or clear boundaries, usually due to low rates of pedoturbation by bioturbators, for example, worms, mammals, insects, which can blur boundaries (Langmaid, 1964), or break and distort them, as occurs locally during tree uprooting (Figure 33.59; Schaetzl, 1986; Schaetzl et al., 1990). Coarse soil textures, acidic conditions, and cool soil temperatures act in unison to retard faunalturbation. It is common for E–Bh horizon boundaries to be wavy or irregular, sometimes conforming to the shapes of previous tree roots, which might have influenced the percolation pathways of water in these sandy soils (Figure 33.60).

Due to the low rates of decomposition more than to high rates of litter production, Spodosols typically have thin to thick (2–20 cm), usually continuous, O horizons. Particularly thick O horizons are found under coniferous forests and heath vegetation. Decomposition rates of the litter decrease with temperature, favoring greater litter accumulation for cool region (frigid and cryic) Spodosols. Slow litter decomposition is also a consequence of the acidic litter character and its high contents of slowly degradable compounds, for example, lignins and waxes (Vance et al., 1986). Spodosol O horizons vary in their degree of decomposition, although Oi and Oe horizons, also referred to as *mor* litter, are common. The slow decomposition in the O horizon is performed by fungi, usually resulting in mycelial mats.

The A horizons in Spodosols are typically thin, especially in coarser-textured soils, where macroorganism activities are low. Bioturbation of organics from the O horizon into the mineral



(a) (b) (c)

FIGURE 33.60 Morphology of the E and Bs horizons some representative, well-developed, Spodosols. (a) A sandy, mixed, frigid, Entic Haplorthod, showing pronounced tonguing of the E and Bs horizons. Note that this soil has been plowed in the past and exhibits an Ap horizon. Subtropical Alaquods commonly exhibit similar, irregular boundaries. Location: NE Lower Michigan. (Photo by R. Schaetzl.) (b) A sandy, mixed, frigid, shallow, ortstein, Typic Durorthod, showing a thick, cemented Bsm horizon (ortstein). Location: Northern Michigan. (Photo by R. Schaetzl.) (c) A sandy, siliceous, hyperthermic, Aeric Alaquod, showing an upper and a lower Bh horizon, with contrasting upper boundaries (abrupt vs. gradual, respectively). Note that the E horizon above the lower Bh has not been as thoroughly eluviated as has the upper E horizon. Location: Northern Florida. (Photo by L. Daniels.)

soil, typically important to A horizon formation, is minimal in Spodosols. Thus, the primary pathways of A horizon formation are via translocation of organic materials into the mineral soil by percolating water and by in situ root decay. Neither of these processes is strong in Spodosols. However, small organic molecules, with varying degrees of solubility, can move through the A horizon and participate in chemical reactions in the middle and lower profile. Many Spodosols lack an A horizon entirely; the O horizon is directly overlying an E horizon.

The E horizon is a morphologic hallmark of the Spodosol order and of podzolization. The E horizon is lighter colored (higher Munsell values) than other horizons in the profile because (1) coatings bound by oxides and organic matter have been stripped from, or inhibited from coating, the quartz grains that dominate the mineralogy and (2) most of the dark, ferromagnesian minerals in the parent material have been chemically weathered. Many Spodosol E horizons are bleached to the point of being almost pure white. The E horizon is commonly the most acidic horizon in the Spodosol profile, having been leached of most of its base cations (Lundström et al., 2000b). It is also coarser textured than underlying horizons, partially due to clay eluviation, as well as intense chemical weathering that destroys some of the clays. Well-developed E horizons qualify as albic horizons (Soil Survey Staff, 1999). Although most Spodosols do have a well-developed, high-chroma E horizons, *Soil Taxonomy* (Soil Survey Staff, 1999) has placed taxonomic emphasis on the illuvial (B) horizon, for two reasons: (1) the E horizon is often disturbed by plowing, bioturbation, or erosion, and (2) in soils formed in loamy parent materials, the E horizon is often not well expressed, although the strong B horizon is clearly indicative of ongoing podzolization (Figure 33.61).

FIGURE 33.61 Profile of a loamy, mixed, frigid, Typic Haplorthod from Maine. Note the weak E horizon in this soil, despite evidence from the B horizon of intense podzolization and illuviation of humus and Fe. This type of Spodosol morphology is common in parent materials where silt and clay contents are high; in such soils the high surface area limits the rate at which podzolization can "strip" all the particles in the E horizon of their coatings. (Photo by R. Schaetzl.)

The reddish, reddish brown, or nearly black, B horizon in Spodosols is usually a Bs, Bh, or Bhs, indicative of illuvial sesquioxides, humus, or both (Mokma and Evans, 2000). In Spodosols with a high water table (Aquods), the B horizon may be low in ferrous iron (Fe^{++}) compounds, having been lost to the groundwater. The zone of maximum illuvial humus accumulation in most Spodosol B horizons is near the top, variously decreasing with depth. Thus, Bhs–Bs1–Bs2 horizon sequences are common, as illuvial carbon decreases with depth in the horizon. The upper boundary of B horizons is typically abrupt, often with wavy boundaries or tonguing, contrasting to the typically diffuse lower horizon boundaries (Figure 33.60). The wavy upper boundary reflects the preferential flow of percolating water, as it moves in tongues along preferred zones in coarse-textured soils, or as it follows root pathways (Figure 33.60). Schaetzl (1986, 1990) determined that at least some of these tongues are due to preferential infiltration beneath microtopographic depressions and pits (present or former) in the forest floor. Well-developed B horizons qualify as spodic horizons if they meet certain chemical and morphological criteria (Soil Survey Staff, 1999). Spodic horizons are characterized by illuvial Al and organic matter; more details are provided in Section 33.9.7.

Strongly-developed spodic horizons, for example, Bhsm, Bsm, can be cemented into ortstein (Figure 33.60; Wang et al., 1978; Dubois et al., 1990; Mokma, 1997). Ortstein is commonly so well cemented that it restricts root penetration, but when sandy it still maintains good permeability (Lambert and Hole, 1971). Ortstein is generally more planar in Aquods than in better drained soils and is usually associated with a water table (Mokma et al., 1994). The cementing agent in ortstein is generally assumed to be an amorphous, Al-dominated material (Lee et al., 1988), although Fe and Si may also play a role.

33.9.4 Spodosol Genesis: Details and Theories of the Podzolization Process

33.9.4.1 The General Suite of Processes

There are two primary theories of podzolization: (1) the protoimogolite theory and (2) the chelate-complex theory. Both endeavor to explain the mobilization, translocation and immobilization of the oxidized, metal ions (sesquioxides) and organic compounds in soils (Schaetzl and Anderson, 2005), as well as the variability in organic matter content in Spodosol B horizons. By association, podzolization theory must also explain the intense weathering typical of these soils. All podzolization theories assume that the soil parent material has been preconditioned or exists in a state conducive to the process: it has minimal clay and is acidic or has been acidified. It is also assumed that the vegetation produces litter rich in low-molecular-weight organic acids and fulvic acids, and that the climate is humid enough to initiate deep percolation, at least at some times of the year.

Intense weathering of minerals by organic acids is a central part of the podzolization process (Van Hees et al., 2000). As minerals weather, various cations are released to the soil solution. Base cations must be depleted before Al and Fe can be rendered mobile. If the soil pH is not low enough, that is, if too many base cations exist in the soil, Al and Fe cations released by weathering will precipitate as solid, less mobile forms (oxides, oxyhydroxides, and hydroxides); they will not readily translocate. Defining the exact mechanism by which these metal cations become mobile and illuviate is where the two main theories of podzolization diverge.

The protoimogolite theory focuses on the inorganic, colloidal transport of Al, Si, and Fe. The theory was formulated by the observation that Al and Fe can exist in some humus-poor Spodosols as amorphous, inorganic compounds, for example, imogolite and allophane (Farmer et al., 1980; Anderson et al., 1982; Farmer, 1982; Childs et al., 1983; Gustafsson et al., 1995). Imogolite and allophone are Al–Si hydroxyl gels also known as imogolite-type materials, or ITM (Lundström et al., 2000a). In this theory, ITM form from weathering products in O and E horizons and move with percolating water until immobilized, usually in the B horizon where higher (>5) pH values exist. Al is transported as a positively charged hydroxy aluminum silicate complex (Anderson et al., 1982). The Al and Fe in ITM have presumably been dissolved/weathered by noncomplexing organic and inorganic acids, as well as by readily biodegradable, small, complexing organic acids (Lundström et al., 2000b). Next, negatively charged, colloidal organic matter migrates from the upper profile into the B horizon and precipitates onto the positively charged ITM that are already there. This process is supported by thin section data that show dark organs surrounding allans (allophane-rich cutans) (Freeland and Evans, 1993), or Al- and Fe-rich cutans overprinted onto Si-rich cutans in B horizons (Jakobsen, 1989). Dissolution/weathering processes continue to act on ITM in the B horizon. Because ITM are more easily weathered than crystalline Fe oxides, ITM will continue to eluviate, eventually leaving behind an Fe-rich B horizon, as the lower B horizon becomes enriched in Al by the continued dissolution and remobilization of ITM rich in Al. In a hybrid model of sorts, Ugolini and Dahlgren (1987) and Dahlgren and Ugolini (1991) proposed that ITM are formed in B horizons in situ, as organometallic complexes illuviate and interact with an Al-rich residue or protoimogolite formed thereby CO_2 weathering (Lundström et al., 2000b). Low soil organic matter contents tend to favor the formation of ITM, whereas conditions in soils with abundant organic matter appear to inhibit its formation (Wang et al., 1986).

In the more traditional, widely accepted, chelate-complex (or fulvate complex) theory, organic acids form chelate complexes with Fe and Al cations, rendering these normally insoluble cations (and Si) soluble for translocation (Wright and Schnitzer, 1963; DeConinck, 1980; Buurman and van Reeuwijk, 1984, Dahlgren and Ugolini, 1989; Sauer et al., 2007). Normally, Fe^{3+} and Al^{3+} are not soluble in soils, but when chelated they become more soluble and are readily translocated in percolating water (Riise et al., 2000). In short, the process is driven by organic acids, as opposed to the dominantly inorganic pathway outlined by the protoimogolite theory.

Organic acids and phenolic compounds in soils generally can be categorized based on their molecular weights: low-molecular-weight (LMW) acids <1000 Da, fulvic acids (FAs) 1000–3000 Da, and humic acids >3000 Da (Krzyszowska et al., 1996; Lundström et al., 2000b). Many of these acids are readily produced during litter decay and carried in the soil solution. They also can be produced as root and fungal exudates. From experiments on aqueous extracts from plant litter, it has been long known that organic acids can dissolve ferric and aluminum oxides (Bloomfield, 1953; Schnitzer and Kodama, 1976; Kodama et al., 1983; Lundström et al., 1995). Many of these acids, however, also readily form metal chelates. Thus, central to the chelate-complex theory is the idea that organic acids produced in soils not only weather primary minerals but also chelate the cations released from them. As the chelate complexes move within the profile, they continue to chelate more metal cations, but remain soluble until a certain level of saturation (metal loading) is achieved (Petersen, 1976). At this point, the complex is rendered immobile and precipitates on ped faces, roots, and mineral surfaces. Immobilization (precipitation) of the chelate complexes occurs for any of a number of reasons: (1) increased pH values in the lower profile (Gustafsson et al., 1995), (2) microbial decomposition of the organometallic complex (Lundström et al., 1995), (3) stoppage of the percolating water at a water table, lithologic discontinuity or aquitard (DeConinck, 1980; Buurman and van Reeuwijk, 1984; Schaetzl, 1992; Schaetzl and Schwenner, 2006), (4) flocculation from continued metal loading and electrostatic charge reduction of complexes (DeConinck, 1980), and (5) sorption/precipitation of complexes within a matrix of finer illuvial materials also accumulated as part of the podzolization process (Holzhey et al., 1975; Dahlgren and Marrett, 1991; Harris et al., 1995; Li et al., 1998; Harris and Hollien, 1999, 2000). Once precipitated in the lower B horizon, aluminum can be released by additional microbial breakdown of the organometallic complexes, or it can combine with silica to form ITM. Iron so released will form less soluble ferric oxyhydroxides (Buurman and van Reeuwijk, 1984). With time, as more organometallic coatings accumulate in the B horizon, cutans on grain surfaces tend to thicken. Because they are not crystalline, the cutans tend to shrink and crack upon drying. Cracked grain coatings are diagnostic of illuvial organometallic compounds (DeConinck, 1980; Stanley and Ciolkosz, 1981).

In the initial stages of podzolization, Fe and Al cations are released from primary minerals (if present) so rapidly that any chelate complexes that form are quickly saturated and hence, immobilized. Eventually, however, a zone of depleted Fe and Al forms (the incipient E horizon). Newly formed organic chelates can then pass through this eluvial zone before becoming saturated. Therefore, the E horizon may be present in the early stages of podzolization, but not visible until it deepens past the depth of organic matter accumulation, that is, the A horizon. In sandy, upland soils, mixing is so minimal that A horizons are thin, and distinct E horizons can exist immediately below, perhaps within a few millimeter of the surface. Over time, the E horizon thickens, as (1) metal cations are continually stripped from the top of the B, remobilized, and redeposited lower and (2) the base of the E grows downward, below the depth of melanization. The B horizon continues to accumulate organic matter, Fe, and Al as cutans and as microaggregates, but the majority of the illuvial compounds reside near the top of the B horizon.

33.9.4.2 Groundwater Podzolization: Alaquods

Most Spodosols formed by processes described above are found in cool (mesic to cryic), midlatitude sites, on freely draining, upland landscape positions. However, large areas of well-expressed Spodosols also occur in subtropical and tropical regions where podzolization is associated with water tables (Brasfield et al., 1973; Dubroeucq and Volkoff, 1998; Harris, 2000; Watts and Collins, 2008). These Spodosols commonly classify as Alaquods, reflecting their wetter conditions and and prevalence of Al over Fe in the spodic horizon (usually a Bh horizon). In the southeastern United States, these wet Spodosols typically occur on nearly level, Coastal Plain landforms called "flatwoods" (Brasfield et al., 1973; Brooks, 1981). Flatwoods are poorly drained uplands where periodic chemical reduction and leaching favor Fe depletion. Flatwoods vegetation includes pine species (e.g., *Pinus palustrus*; *Pinus taeda*) with an understory dominated by saw palmetto (*Serenoa repens*) and gallberry (*Ilex glabra*) (Abrahamson and Hartnett, 1990), a plant community consistent with podzolization. However, Alaquods are not restricted to this plant community or to flatwoods landscapes. They can be found along some lake margins, seepage zones, floodplains, etc. with the common denominator being a seasonal water table.

A clue to the water table linkage is the morphological transition between Alaquods and better drained soils upslope (Garman et al., 1981; Tan et al., 1999). The common trend is for the thickness and morphological prominence of E and Bh horizons to diminish with diminishing water table influence, for example, shorter periods of saturation, greater depths to saturation, etc. (Figure 33.62). The Bh lightens in color and its upper boundary becomes shallower as the mean water table depth increases. This divergence between Bh horizon and the water table at transitions confirms that Bh horizons in these Alaquods do not form as a simple consequence of C precipitation with groundwater-entrained Al in quartz sand. Instead, true podzolic E horizons form, similar to those in Spodosols of cooler regions. Similarity of cooler and warmer Spodosols is also evidenced by morphological properties such as common irregular upper boundaries, for example, the root disturbance legacy effect on preferential flow as discussed above (Figures 33.59 and 33.60). Furthermore, near-surface water saturation

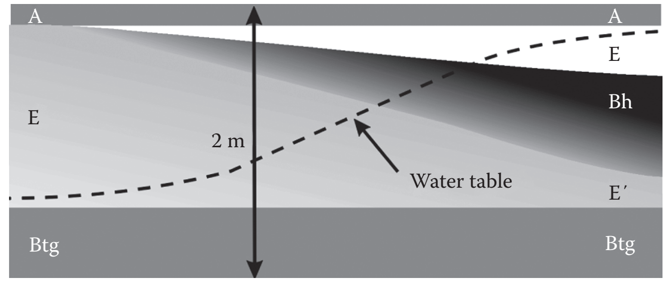


FIGURE 33.62 Schematic of a transition between Paleudult and Alaquod. Fading-upward of Bh and wedging out of E is typical of transitions between Aquods and better drained soils in northern Florida (Garman et al., 1981; Tan et al., 1999). The E of the Paleudult has sufficient coatings on sand grains to impart lower values and higher chromas than for the Alaquod E, in which grains have been essentially stripped of coatings.

is a dynamic seasonal condition rather than a static condition that would be required to explain abrupt white-to-black horizon boundaries.

The E–Bh boundary in Alaquods is also a boundary between zones of uncoated sand grains (A and E horizons) and coated grains (Bh and subjacent horizons; Figure 33.62). Sand-grain coatings are either stripped or prevented from forming in the upper profile under typical Alaquod hydrology, yet coatings occur in near-surface horizons of better drained sandy soils of the region (Harris et al., 1996). Hence, the origin, composition, and stability of grain coatings are pertinent to the mechanism of Alaquod formation. These coatings likely originate from mineral weathering in conjunction with the binding effects of secondary metal oxides. Most <50 μm-sized minerals in these soils are bound in coatings. The clay fraction of grain coatings in the Coastal Plain soils of the southeastern United States is commonly dominated by hydroxy-interlayered vermiculite, kaolinite, gibbsite, and quartz (Carlisle and Zelazny, 1974; Harris et al., 1987a, 1987b), with minor amounts of metal oxides that bind the coatings. These minerals, except for quartz, are nearly depleted in podzolic E horizons.

Experimental generation of Aquod-like E–Bh horizons has documented that colloidal transport, that is, eluviation–illuviation of sand-grain coatings, was a component of the process (Harris et al., 1995; Harris and Hollien, 2000). The coating destabilization occurred as a result of organo-complexation of the metal oxide-binding agents. Sand-grain stripping produced a light-gray podzolic E horizon and culminated in illuviation of the mobilized sand-grain coating materials to form a finer-textured zone subjacent to the E horizon. The illuvial zone (incipient Bh) provided conditions for sorption of organometal complexes and thus served as a template for Bh horizon development. The E–Bh horizon boundary contrast sharpened with time, from the darkening effect of continued C flux.

Colloidal transport in experimental formation of Aquod-like features does not prove that it occurs in the genesis of real Alaquods. However, it is consistent with the (1) crystalline clay mineralogy of Bh horizons, which typifies that of sand-grain coatings in nonpodzolized areas (Harris and Carlisle, 1985; Harris et al., 1987a; Harris and Hollien, 2000), (2) upward-trending Bh horizons in areas transitional to better drainage, as described above (Figure 33.62), (3) elevated silt and clay contents in Bh horizons, and (4) abrupt, irregular E–Bh boundaries that can be more easily accounted for by the physical accretion of fine particulates than by chemical precipitation alone. Indeed, other studies have reported evidence of particle migration in Spodosols (Calhoun and Carlisle, 1973; Guillet et al., 1975; Ugolini et al., 1977; Li et al., 1998; Harris and Hollien, 1999).

The role of the water table in Alaquod formation may be to bring about conditions unfavorable for the stability or formation of sand-grain coatings. Artificial generation of E- and Bh-like horizons was achieved only (1) in predisposed coated sands collected on wetter landscape positions, that is, bordering Alaquods, and (2) under fluctuating water table conditions (Harris et al., 1995). Lower Fe oxide contents and lower Al crystallinity are factors that could explain the predisposition of wetter landscape materials to form Aquod-like features, compared with materials from drier summit positions. Lower Fe contents would be a predisposing factor for sand-grain coating destabilization because Fe oxides serve as binding agents. Aluminum oxides were not depleted in predisposed materials and still served to bind coatings. However, the lower crystallinity of Al oxides, possibly the result of higher organic acid activity in the wetter positions, would render them more vulnerable to organic acid complexation (Huang and Violante, 1986). Thus, wetter landscapes predispose sand-grain coatings to undergo destabilization and to participate in podzolization processes theorized above, should the groundwater influence intensify.

A fluctuating water table is still required to experimentally form artificial E–Bh sequences, even in predisposed material. Hence, the water table appears to have a role in not only predisposing, but also actually triggering, podzolization. The triggering mechanism, however, is still uncertain. It may relate to the effects of the frequency and duration of water saturation in the affected zone. For example, anaerobic conditions could inhibit microbial degradation of complexing organic anions, such that these anions reach sufficient activity to mobilize metals.

33.9.4.3 Groundwater Podzolization: Deep, Thick Bh Horizons

The previous discussion on Alaquod formation conforms closely to the chelate-complex theory of podzolization. However, the formation of deep, thick Bh horizons, found in some subtropical coastal regions (Daniels et al., 1975; Thompson, 1992), may involve elements of the protoimogolite theory. These horizons are morphologically distinct from shallower, thinner Bh horizons (Figure 33.60) and their upper boundaries may be located within, or below, 2 m. There are sometimes multiple Bh horizons with varying degrees of coalescence, across zones up to several meters thick. Deep Bh horizons of some regions (1) darken gradually from the upper boundary downward, rather than being darkest at the top, (2) are not overlain by a well-expressed podzolic E horizons, (3) have consistently smooth upper boundaries, and (4) transgress soil drainage continua, extending below excessively drained sandy soils, as well as Alaquods. These distinctions suggest C immobilization by *in situ* inorganic Al (Farmer et al., 1983) at the approximate upper boundary of long-term saturation, adhering to the independent C-transport aspect of the latter theory. The absence of a well-expressed podzolic E horizon is evidence that immobilization of C by Al, as influenced by water table, could occur whether there has been C-induced transport of Al or not.

33.9.5 Factors That Influence the Formation of Spodosols

Podzolization occurs in humid climates, where precipitation exceeds evapotranspiration during some of the year, such that water frequently percolates through the profile, or stands in the profile at the water table (McKeague et al., 1983). Podzolization is best expressed under vegetation that produces acidic litter, such as coniferous forest and heather, and on coarse-textured parent materials that have minimal surface area. It is especially strong in the coniferous forests of the high midlatitudes. Cool temperatures in midlatitude and alpine areas keep evapotranspiration rates low and inhibit decomposition of the acidic litter. The discussion below focuses on how, as these factors change, the character of the profile also changes.

33.9.5.1 Vegetation

Podzolization is promoted under vegetation that produces acidic litter that decays slowly. The litter decay products, that is, organic acids of various kinds, not only help weather Fe- and Al-bearing primary minerals, but also can be efficient as metal chelates, because they release abundant amounts of water-soluble organic acids. Vegetation types that particularly promote podzolization include heath (heather), hemlock (*Tsuga canadensis*), kauri (*Agathis australis*), and pine (*Pinus* sp.), the litter from which releases acidic leachates (Bloomfield, 1953; Mackney, 1961; Nielsen et al., 1987a, 1987b; Madsen and Nørnberg, 1995; Mossin et al., 2001; Jongkind et al., 2007). Thick litter layers in various states of decay are typical in naturally vegetated Spodosol landscapes. Thick O horizons also help reduce variation in soil moisture contents and allow for more prolonged weathering and leaching of the mineral soil below. Barrett (1997) illustrated the importance of above-ground vegetation to podzolization by demonstrating that the process markedly slows after only a few decades of forest removal and litter disturbance (see also Hole, 1975).

Fire impacts Spodosol development by eliminating or reducing the O horizon volume, changing its chemistry and character, and altering the above-ground forest vegetation. Mokma and Vance (1989) observed that soils under vegetation that burns more frequently were much less developed than in areas of diminished fire frequency. Indeed, in parts of Michigan, repeated (ca. 5–50 years) fires have so retarded spodic development that most of the soils are Entisols, even after 15,000 years (Schaetzl et al., 2006). Schaetzl (2002) also noted that the areas in Michigan having maximal fire frequencies and minimal soil development also receive much less snow than other areas with better soil development, suggesting that snowpack and snowmelt impacts on soils development are often as important as fire frequency. Indeed, the two are often mutually reinforcing. Subtropical Alaquods are an exception to the fire-suppressing effect on podzolization; they predominate in fire-maintained flatwoods ecosystems.

Because most Spodosols form under forest vegetation, mixing by tree uprooting is a common, albeit localized and often pedon-specific, form of disturbance (Dunn et al., 1983; Schaetzl et al., 1990; Jonsson and Dynesius, 1993). Pit and mound topography is the common surface expression of this process, remaining on the landscape for centuries after the treethrow event (Figure 33.59; Schaetzl and Follmer, 1990; Kabrick et al., 1997; Small, 1997). This microtopography dramatically impacts infiltration pathways, directly and indirectly (Schaetzl, 1990). Soils are often much better developed beneath pits, where infiltration is focused, and litter and snowpacks are thicker. Tree uprooting distorts, breaks, mixes, and even inverts soil horizons, leading to high levels of soil spatial variability (Veneman et al., 1984; Schaetzl, 1986; Meyers and McSweeney, 1995). The spatial variation in soil development continues to evolve and even strengthen as a consequence of the subsequent microtopography and its effects on the redistribution of water and litter.

33.9.5.2 Climate

Spodosols on upland sites are best developed in cool, humid, continental climates, especially those with a pronounced, snowy winter. Cool temperatures favor vegetation types, for example, conifers and heather, that produce acidic litter, which promotes podzolization. Low (but not freezing) mean annual temperatures, coupled with an extended period of subfreezing temperatures and snow in winter, tend to slow litter decay processes and maximize the potential for organic acid production and its slow, steady translocation into and through the mineral soil. Wet local conditions in the tropics similarly affect litter decay, promoting Spodosol development. In warmer and/or drier climates, much of the litter is quickly oxidized and mineralized, resulting in minimal amounts of acid leachate for podzolization. Globally, podzolization and spodic development (on upland sites) is maximal in the frigid soil temperature regime, eventually weakening at the extreme cold end of the climate spectrum, in the gelic and pergelic soil temperature regimes. Here, vegetation transitions to tundra, and soils are frozen for longer periods of time. Similarly, podzolization on upland sites is weak in the slightly warmer, mesic soil temperature regime, where litter decay is rapid and where forest vegetation tends to include more broadleaf trees and base cyclers (Cann and Whiteside, 1955; Schaetzl and Isard, 1991). For example, in midlatitude locations near the "warm" end of their range, Spodosols located on slightly cooler sites tend to be better developed, especially with regard to B horizon character (Stanley and Ciolkosz, 1981). In the frigid soil temperature regime of northern Michigan, Hunckler and Schaetzl (1997) observed much better spodic development on cooler, northerly slope aspects. Thus, climate impacts podzolization mainly by affecting the vegetation that occupies the landscape and by impacting the nature of the pedohydrologic cycle.

In midlatitude areas, Schaetzl and Isard (1991) observed that podzolization is much more pronounced in areas that receive heavy, early snowfall, and accumulate thick snowpacks. This pattern suggests that snowpacks insulate the soils from freezing (Isard and Schaetzl, 1995), facilitating slow and steady infiltration during spring snowmelt. Snowmelt water is often rich in organic acids, having just percolated through fresh and slightly decomposed litter (Schaetzl and Isard, 1996). Additionally, areas with deep snowpacks have warmer soils, especially in winter, which foster weathering of primary minerals and release of metal cations to the soil solution (Isard et al., 2007). The melting snow also provides water for long, steady infiltration events, and thus, efficient translocation of metal chelates. In support of this assumption, Schaetzl (1990) observed that soil water during snowmelt contained more Fe and Al than water similarly sampled during warm season percolation events. Thicker snowpacks also reduce fire frequencies; fire has been shown to inhibit podzolization in some areas, as previously discussed.

33.9.5.3 Parent Material

Spodosols form best on acidic, coarse-textured parent materials, with low amounts of silicate clay and Fe-rich minerals. In these materials, base and metal cations are slowly released by weathering, and the metal cations are subject to chelation by organic acids in the profile and translocation to the B horizon. (The fate of base cations is varied; many are leached from the profile, some are biocycled.) In finer-textured, Fe-rich soils, cations (especially Fe) are released in abundance, although many are quickly adsorbed to clay minerals. As such, lessivage (clay translocation) and/or brunification (reddening), rather than podzolization, tends to dominate the pedogenic suite. Essentially, the great numbers of Fe and other cations released to the soil solution in these types of parent materials commonly overwhelm the ability of the soil to translocate them. As a result, distinct eluvial/illuvial zones do not readily form (Duchaufour and Souchier, 1978; Alexander et al., 1994; Figure 33.61). These soils commonly only exhibit a reddened, "color B" horizon with no, or only a thin, E horizon. Thus, as clay and Fe contents increase, translocation of metal cations and humus decreases and E horizons are less "bleached" (Cline, 1949; Gardner and Whiteside, 1952). Schaetzl and Mokma (1988) used this knowledge to develop a numerical, field- and morphology-based index of spodic development—the POD index.

Increased biological activity in finer-textured soils also makes it more difficult to retain a thick O horizon at the surface. Particularly, important is the increased amount of bioturbation in these finer-textured soils, which can blur or destroy incipient horizons. Basic (high pH) parent materials must generally be leached of free carbonates before podzolization can

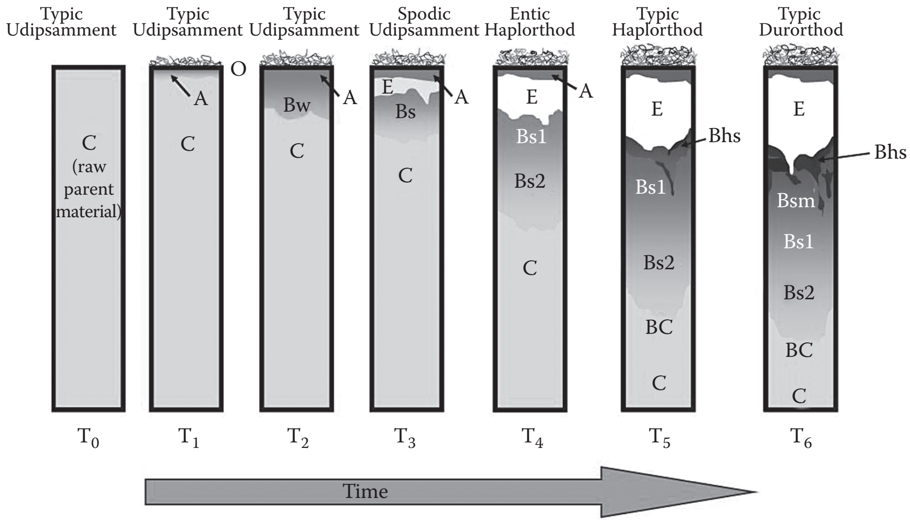


FIGURE 33.63 Typical horizonation/soil development sequence, from Entisols to well-developed Spodosols, on sandy, well-drained parent materials in the midlatitudes.

begin in earnest, as high pH values inhibit the formation of metal–chelate complexes and inhibit weathering and release of metal cations. Thus, podzolization will occur sooner in parent materials that are initially acidic. Coarse-textured parent materials also foster rapid infiltration and deep percolation, accelerating the development of an eluvial zone, and a deeper B horizon.

33.9.5.4 Time

Spodic morphologies develop along generally predictable pathways (Figure 33.63) and can form in as few as a few decades (Paton et al., 1976; Stützer, 1998), although it usually takes at least 3000–8000 years, under optimal conditions, for this development to reach the point where the soil classifies as a Spodosol (Franzmeier and Whiteside, 1963; Jauhiainen, 1973; Barrett and Schaetzl, 1992; Mokma et al., 2004; Sauer et al., 2008). E horizons typically form first, or are most apparent, often forming within a few centuries (VandenBygaart and Protz, 1995; Lichter, 1998; Barrett, 2001). Noticeable darkening and reddening of the B horizon take longer to occur, but is often visible within 1000–2000 years. After the Bs horizon has become well established, a Bhs horizon may form near the top of the illuvial zone. A chronosequence in beach deposits in Norway (Sauer et al., 2008) showed incipient podzolic features within 3800 years, but fully expressed spodoc horizons were observed only in the 6600-year-old soil. In some soils, cementation of parts of the B horizon (ortstein) may follow. Sauer et al. (2007) provide an excellent review of the rate of Spodosol formation.

33.9.5.5 Topography, Relief, and Drainage Class

In this chapter, we have differentiated between cool region Spodosols that commonly occur on freely drained upland landscape positions and subtropical Spodosols that are mainly restricted to wet upland landscapes. The commonality is "upland;" Spodosols, regardless of drainage, are located on landscapes that experience leaching and throughflow of soil water. Flatwoods of the southeastern United States are an example of poorly drained, leached uplands. However, some poorly drained leaching soil environments are not uplands, for example, toeslopes of sandy hills, and lake margins. There is a strong hydrologic–topographic control of Spodosol occurrence in the southeastern United States. Spodosols dominate on landscapes with fluctuating water tables but are absent, or only of isolated occurrence, on other landscapes, for example, sandhills, karst terrain. Podzolization in all settings involves gravity-driven translocation, which can also be affected by preferential flow. Materials move from eluvial to illuvial horizons and, barring disturbances such as tree uprooting or bioturbation, and excluding cation biocycling, do not readily move back toward the surface, even in wetter Spodosols. Indeed, Schaetzl and Mokma (1988) found that Spodosols in somewhat poorly drained positions were generally the best developed Spodosols, among drainage classes. Further, Alaquods can have strongly expressed E horizons (Watts and Collins, 2008).

33.9.6 Distribution of Spodosols

When considered geographically, the state factors discussed previously outline the global distribution of Spodosols. Thus, Spodosols occur mainly in cool, humid climates, generally under forest vegetation and on freely draining parent materials (Figure 33.58). The most widespread and dense concentrations of Spodosols occur in Canada and the northern Great Lakes region, Russia, Scandinavia and northern Europe. At appropriate (high) elevations, Spodosols are found on just about all of Earth's major mountain ranges. Aquods are also common on wet landscapes of the humid and subhumid tropics and subtropics. In areas where the climate or vegetation is marginal for the formation of well-developed Spodosols, weakly developed Spodosols or soils with spodic-like profiles can still be observed, provided the remaining state factors favor podzolization. For example, Spodosols exist in cold, polar deserts (Ugolini, 1986; Blume et al., 1997).

33.9.7 Characteristics and Classification of Spodosols

Most Spodosols meet the NRCS classification criteria by having a spodic horizon. The spodic horizon is characterized by illuvial accumulations of Al and organic matter, most of which is translocated from overlying eluvial horizons above, and accumulates in amorphous forms. Illuvial Fe is usually associated with the Al and organic matter, although Fe is not necessary to meet many of the various worldwide "spodic/podzol" classification criteria. These illuvial substances, rich in Al, Fe, and various forms of illuvial organic materials, are referred to as "spodic materials." By definition, spodic horizons must be composed of 85% or more spodic materials (Soil Survey Staff, 1999). The spodic material concept embodies the podzolization process by requiring the materials to have (1) relatively low pH values and high organic matter contents and (2) red colors (usually) if they underlie an albic (E) horizon. In soils that lack an albic horizon, the pH, organic matter content, and color criteria are still required, but the spodic materials must *also* exhibit one or more of the following properties: (1) cementation, (2) cracked coatings on sand grains, (3) significantly more Al and Fe than an overlying horizon, determined by ammonium oxalate extraction, or (4) high optical densities of the liquid ammonium oxalate extract (an indication of amorphous materials, especially organic matter; Stanley and Ciolkosz, 1981; Daly, 1982; Mokma, 1993).

Many soils with a spodic horizon also have andic properties and are placed in the Andisol order (Flach et al., 1980; Soil Survey Staff, 1999). Because Andisols and Spodosols are pedogenically similar, many Andic subgroups of Spodosols have been defined (Table 33.44), and vice versa. Provision is also made for soils in which the spodic horizon is shallow and thus has been plowed. If the Ap horizon of such soils contains 85% or more spodic materials, the soil may classify as a Spodosol instead of an Andisol.

Some Spodosols have only a placic horizon, or a placic and a spodic horizon. The placic horizon is a thin (1–25 cm thick), black to dark reddish, cemented pan with a sharp upper boundary but diffuse lower boundary; it is cemented by Fe (or Fe and Mn) and organic matter (Clayden et al., 1990; Hseu et al., 1999). The pan is often wavy and even may bifurcate. Placic horizons may form as reduced iron, mobilized in surface horizons, is translocated downward in the profile and

TABLE 33.44 Listing of All the Current (1999) Suborders, Great Groups, and Subgroups of Spodosols

Suborders	Great Groups	Subgroups
Aquods	Cryaquods	Aeric, Andic, Duric, Entic, Lithic, Pergelic, Placic, Sideric, Typic
	Alaquods	Aeric, Alfic, Alfic Arenic, Arenic, Arenic Ultic, Arenic Umbric, Duric, Grossarenic, Histic, Lithic, Typic, Ultic
	Fragiaquods	Argic, Histic, Plagganthreptic, Plaggeptic, Typic
	Placaquods	Andic, Typic
	Duraquods	Andic, Histic, Typic
	Epiaquods	Alfic, Andic, Histic, Lithic, Typic, Ultic, Umbric
	Endoaquods	Andic, Argic, Histic, Lithic, Typic, Umbric
Cryods	Placocryods	Andic, Humic, Typic
	Duricryods	Andic, Aquandic, Aquic, Humic, Oxyaquic, Typic
	Humicryods	Andic, Aquandic, Aquic, Lithic, Oxyaquic, Pergelic, Typic
	Haplocryods	Andic, Aquandic, Aquic, Entic, Lithic, Oxyaquic, Pergelic, Typic
Humods	Placohumods	Andic, Cryic, Typic
	Durihumods	Andic, Typic
	Fragihumods	Typic
	Haplohumods	Andic, Arenic, Arenic Ultic, Entic, Grossarenic, Grossarenic Entic, Lithic, Plagganthreptic, Plaggeptic, Typic, Ultic
Orthods	Placorthods	Typic
	Durorthods	Andic, Typic
	Fragiorthods	Alfic, Alfic Oxyaquic, Aquentic, Aquic, Cryic, Entic, Humic, Oxyaquic, Plagganthreptic, Plaggeptic, Typic, Ultic
	Alorthods	Alfic, Arenic, Arenic Ultic, Entic, Entic Grossarentic, Grossarentic, Grossarentic Entic, Oxyaquic, Plagganthreptic, Plaggeptic, Typic, Ultic
	Haplorthods	Alfic, Alfic Oxyaquic, Andic, Aqualfic, Aquentic, Aquic, Duric, Entic, Entic Lithic, Fragiaquic, Fragic, Humic, Lamellic, Lamellic Oxyaquic, Lithic, Oxyaquic, Oxyaquic Ultic, Typic, Ultic

Source: Soil Survey Staff. 1999. Soil taxonomy. USDA-NRCS agriculture handbook no. 436. U.S. Government Printing Office, Washington, DC.

oxidized and precipitated in the B horizon. This precipitated Fe can adsorb soluble organic matter, but not necessarily form organometallic complexes (Hseu et al., 1999, Soil Survey Staff, 1999). The iron is usually present as ferrihydrite and poorly crystalline goethite. Placic horizons typically form in soils at landscape positions where conditions change from reducing to oxidizing. The reducing conditions, necessary for placic horizon genesis, are associated with (1) high rainfall and cool temperatures, typical of perhumid climates (and often also associated with Histosols) and (2) a slowly permeable subsurface layer (Lapen and Wang, 1999).

33.9.7.1 Taxonomic Groupings of Spodosols

In a review of Spodosols, Rourke et al. (1988) commonly referred to a "warm" (thermic, hyperthermic, or isohyperthermic temperature regimes) vs. "cold" (frigid, mesic, or isomesic temperature regimes) Spodosol dichotomy. Although this level of distinction is not taxonomically observed at the suborder or great group level, we agree it is a useful way of considering Spodosols because, globally, these soils are mainly found in two distinctly different climatic settings: subtropical/tropical and cool-cold midlatitude locations. In subtropical and tropical locations, they tend to be located on wet sites (Goldin and Edmonds, 1996), whereas in cooler locations Spodosols occur in all landscape positions. Most "warm" Spodosols are Aquods, Orthods, and Humods; many of these soils lack a Bs horizon and have Bh horizons instead (Rourke et al., 1988). "Cool" Spodosols tend to have thinner profiles, and Bs, Bh, and/or Bhs horizons. Table 33.44 provides a complete listing of the suborders, great groups, and subgroups of Spodosols. Figure 33.64 shows the typical profile morphologies for the most common Spodosol subgroups (Soil Survey Staff, 1999).

33.9.7.1.1 Orthods

Orthods are the most common Spodosol suborder (Rourke et al., 1988). They typically have A–E–Bhs–Bs–BC–C horizon

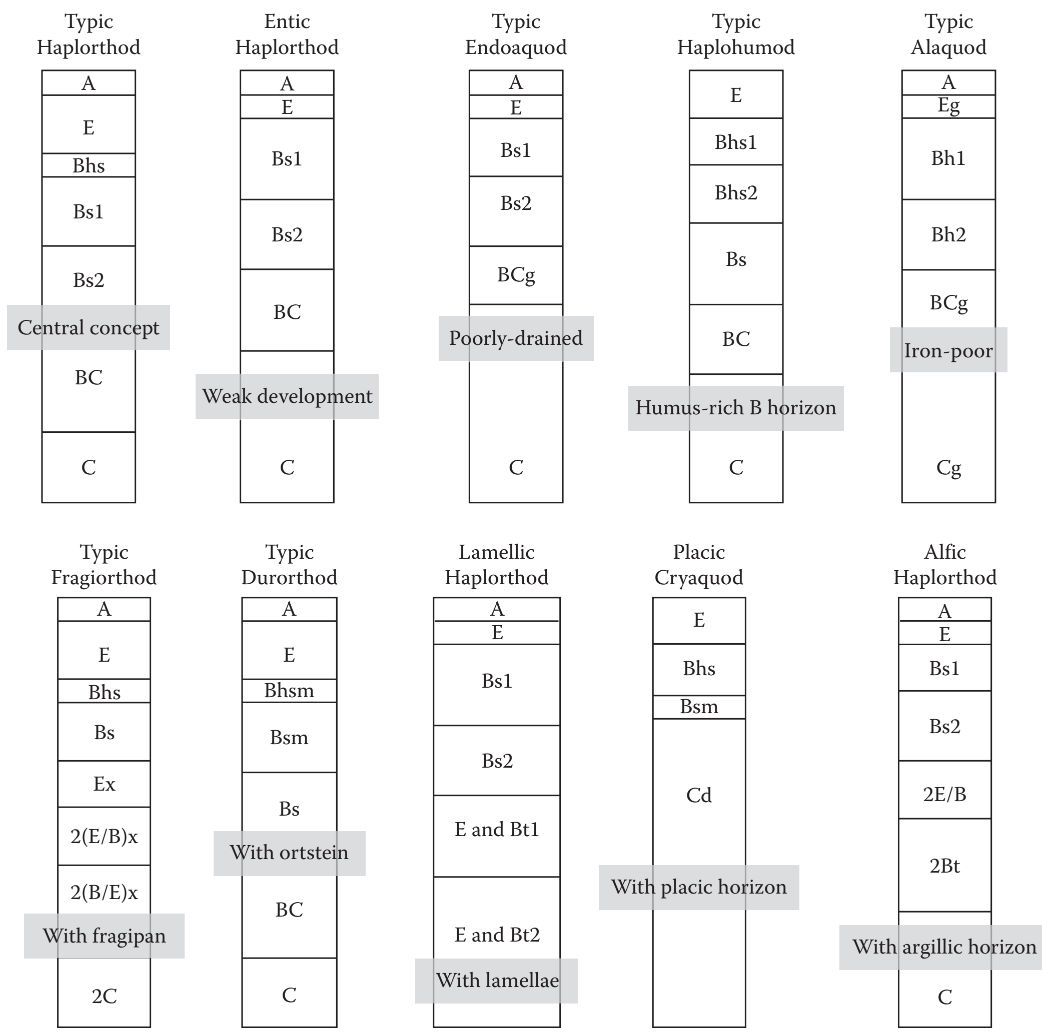


FIGURE 33.64 Typical horizonation patterns in common Spodosol profiles, illustrating the variations found naturally in extragrade and intergrade subgroups. Natural undulations in horizon boundaries, as well as O horizons, have been omitted to keep the diagrams simple and straightforward.

sequences, although the Bhs horizon is usually lacking in weakly developed pedons (Base and Brasher, 1990; Schaetzl, 2002).

33.9.7.1.2 Aquods

Aquods have a shallow, often fluctuating, water table. They are the most prevalent Spodosol suborder in subtropical and tropical regions and are a common admixture in other Spodosol landscapes in low-lying landscape positions.

33.9.7.1.3 Humods

Humods have high organic matter contents in the B horizon, usually concentrated in the upper B. In mesic and colder climates, Humods are most common in areas with very humid climates and lush vegetative growth.

33.9.7.1.4 Cryods

Cryods are Spodosols with a cryic soil temperature regime and include soils that have a mean annual temperature <8°C, but lack permafrost.

33.9.8 Management of Spodosol Landscapes

Many Spodosol landscapes are in natural or managed forest systems, where harvest of forest products, recreation, and tourism are the major economic activities. The cool climate of these landscapes, coupled with low soil fertility and sandy textures, tends to limit agricultural options on Spodosols. Particularly, restrictive are the low pH values of these soils, as well as low CEC and available P levels. Many Spodosols adsorb P and K, thereby restricting the ability of plants to obtain these nutrients (Bartlett and McIntosh, 1969; Bartlett, 1972; Villapando and Graetz, 2001). Some types of agriculture have been successful on Spodosols, particularly potatoes, small grains such as rye and barley, some hay crops, and irrigated berries, in some areas, especially where the soils are sandy loam or finer in texture. Blueberries are particularly well suited to cool, acid, sandy Aquods. A long growing season and climatic suitability for high value crops, for example, citrus, warm season vegetables, etc., makes it economically feasible to manage subtropical Spodosols (mainly Aquods) for intensive agriculture or horticulture (Harris et al., 2010). In these agricultural systems, management for some crops requires artificial drainage. Managing Aquods to minimize environment impacts of agricultural P is a challenge due to the minimal P retention capacity of their A and E horizons (Harris et al., 1996) that can result in P transport to water bodies (Allen, 1987; Mansell et al., 1991).

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33.10 Aridisols

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33.10.1 Concept of Aridisols

Vast areas of the Earth, roughly one-third of the total land surface, are in arid regions. Such aridity results from their position relative to mountain ranges that scavenge moisture, their latitude around 30° where convective Hadley cells create enduring high pressures, or their great distances from large bodies of water (Ahrens, 1991; Figure 33.65). Rainfall in these dry climates is insufficient to maintain perennial streams, and the soils typically have little organic matter, contain pedogenic carbonate, and because their soil profiles are of the "nonflushing type" (Rode, 1962), contain soluble minerals otherwise leached from humid soils (Southard, 2000). The dominant soil order in arid regions is the Aridisol order, which, with the exception of the Entisol order, covers more of the Earth's surface than any soil order in the *Soil Taxonomy* system (Wilding, 2000).

The central idea of Aridisols is that they do not have water available to mesophytic plants for long periods. This lack of soil moisture may be exacerbated by low water-holding capacity due to a shallow soil depth or restricted infiltration or by low osmotic potential due to salinity. The lack of water also inhibits soil formation.

33.10.1.1 Aridisols vs. Desert Soils

As the order name implies, Aridisols are associated with arid climates, but not all soils in arid regions are Aridisols. In fact, Aridisols occupy less than half of the arid land area (Figure 33.65). Other soil orders in arid climates are the Entisols, Vertisols, Oxisols, and Andisols (Table 33.45). The rationale for having Entisols in arid climates was to preserve the *azonal* concept, which identified soils lacking genetic horizons regardless of the bioclimatic zone in which they occur (Smith, 1986). The rational for having Oxisols in arid climates was that oxic properties, even if irrigated, would remain limiting and the management difficulties would be similar to other Oxisols; thus, they should grouped together with other Oxisols despite their occurrence in dry climates (Smith, 1986). Similar reasons were presented for the shrink–swell properties of Vertisols and the andic properties of Andisols. In some minor areas within the aridic moisture regime, there are soils with a mollic epipedon. These soils are grouped with the Mollisols rather than Aridisols (Smith, 1986; Ahrens and Eswaran, 2000).

33.10.1.2 History of the Aridisol Order

The rationale for the Aridisol order of *Soil Taxonomy* was to separate on a soil map the "sown from the unsown"; that is, the land that can be cultivated from the land that can only be grazed (Smith, 1986). Using climate as the basis for a soil order is somewhat unique to the *Soil Taxonomy* system. Other classification

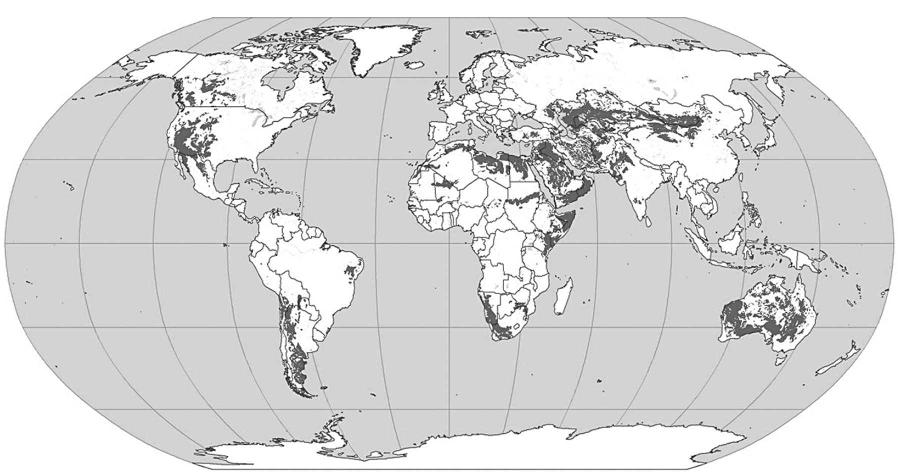


FIGURE 33.65 Global distribution of Aridisols. (Courtesy of USDA-NRCS, Soil Survey Division, World Soil Resources, Washington, DC, 2010.)

TABLE 33.45 Global Extent of Aridisols (km^2) and Vertisols, Oxisols, Andisols, and Entisols Occurring within the Aridic Soil Moisture Regime

Soil Order	Suborder	Africa	Asia	Australia/ Oceania	Europe	South America	Central America	North America	Caribbean	Global
Aridisols										14,264,316
	Cryids	549	398,041	—	100	142,088	—	516,759	—	1,057,537
	Salids	138,542	954,801	69,044	7,993	28,304	—	9,988	31	1,208,703
	Durids	—	—	—	—	—	—	—	—	—
	Gypsids	308,871	298,414	—	1,806	—	—	—	—	609,091
	Argids	292,435	1,107,480	1,395,126	66,339	471,738	2,039	1,012,842	1,446	4,349,445
	Calcids	1,598,740	1,988,228	538,948	1,053	96,171	57	178,254	52	4,401,503
	Cambids	715,612	1,049,670	294,538	29,594	388,308	486	158,233	1,596	2,638,037
Vertisols	Torrerts	161,019	54,772	508,848	—	5,204	133	35,705	1,337	767,018
Oxisols	Torrox	7,481	—	3,527	—	13,461	—	—	1,316	25,785
Andisols	Torrands	942	134	—	—	64	—	103	—	1,243
Entisols	In Aridic moisture regimes									11,266,409
								Total arid soils		26,324,771
								Total ice-free land		130,268,185

Sources: Soil Survey Staff. 2010. Keys to soil taxonomy. 11th edn. USDA-NRCS. U.S. Government Printing Office, Washington, DC. Data courtesy of USDA-Natural Resources Conservation Service, Soil Survey Division, World Soil Resources, 2009.

systems, such as the World Reference Base, do not have an equivalent to the Aridisol order (IUSS Working Group WRB, 2006). Yet the delineation based on climate was inherited from the concept of *zonal* soils, the idea that certain kinds of soils are associated with certain kinds of bioclimatic zones—a concept developed by N.M. Sibirtsev in the late 1800s (Sibirtsev, 1898; Buol et al., 1980).

By the mid-1930s, soils in the arid zone of the United States were mapped as "northern gray desert soils" and "southern gray desert soils" (Marbut, 1935). With the publication of a revised classification system in 1938, the northern gray desert soils were changed to "sierozems" and the southern desert soils, which were associated with creosotebush, were changed to "red desert soils" (Baldwin et al., 1938). With the publication of *Soil Taxonomy* (1975), these soils became Aridisols with two suborders: Argids, if an argillic or natric horizon was present, or Orthids if not (Witty, 1990). In the 1980s and 1990s, the International Committee on Aridisols (ICOMID) developed seven suborders based on important diagnostic horizons and temperature: Cryids, Salids, Durids, Gypsids, Argids, Calcids, and Cambids.

33.10.1.3 Concept of the Aridic Moisture Regime

The concept of the aridic moisture regime is not totally synonymous with the definition of *arid*, which has several definitions. *Arid* in some definitions is based on precipitation alone, as in the arid-semiarid boundary at 250 mm of annual rainfall (Bull, 1991). In other definitions, the arid boundary is based on a combination of precipitation and temperature, as in the aridity index of 10 (Schmidt, 1979). Still, other definitions, such as the Köppen system, combine precipitation, temperature, and seasonality of precipitation (Strahler and Strahler, 1987). In contrast, the soil moisture regime is based on the number of days a soil is dry (i.e., soil moisture held at tensions less than −1500 kPa). Soil moisture is used because the availability of soil moisture to plants is affected by runoff–runon, infiltration, and available water-holding capacity rather than annual precipitation per se. Generally speaking, however, most soils that have an aridic moisture regime do occur in areas of arid climates (Soil Survey Staff, 1999). Exceptions include semiarid soils with physical properties that keep them dry, such as physical crusts that prevent infiltration, steep slopes where runoff is high, low water-holding capacity in shallow soils, or low osmotic potential in strongly saline soils.

The aridic moisture regime is based on the duration of dryness in the *soil moisture control section* during the period the soil is warm enough for plant growth. The soil moisture control section will vary among soils because it is based on how deep water infiltrates. The top boundary of the soil moisture control section is the depth reached by the wetting front when 2.5 cm (1 in.) of water is applied for 24 h. The bottom boundary is the depth reached by the wetting front when 7.5 cm (3 in.) of water is applied for 48 h. Because these depths are largely a function of soil texture, finer-textured soils, such as those with fine-loamy, coarse-silty, fine-silty, or clayey textural classes, have moisture control sections that lie between approximately 10 and 30 cm (Herbel and Gile, 1973). In contrast, coarse-loamy soils have moisture control sections roughly between 20 and 60 cm, and sandy textural classes have boundaries between approximately 30 and 90 cm depending on coarse fragments, which deepen these limits for all textures.

By definition, a soil has an aridic moisture regime if that soil's control section is dry more than half the year (in normal years) when the temperature at 50 cm is above 5°C. The criterion of "above 5°C" was added because cold temperatures, rather than dryness, can be the limiting factor to plant growth (Smith, 1986). Additionally, the soil moisture control section is "Moist in some or all parts for less than 90 consecutive days when the

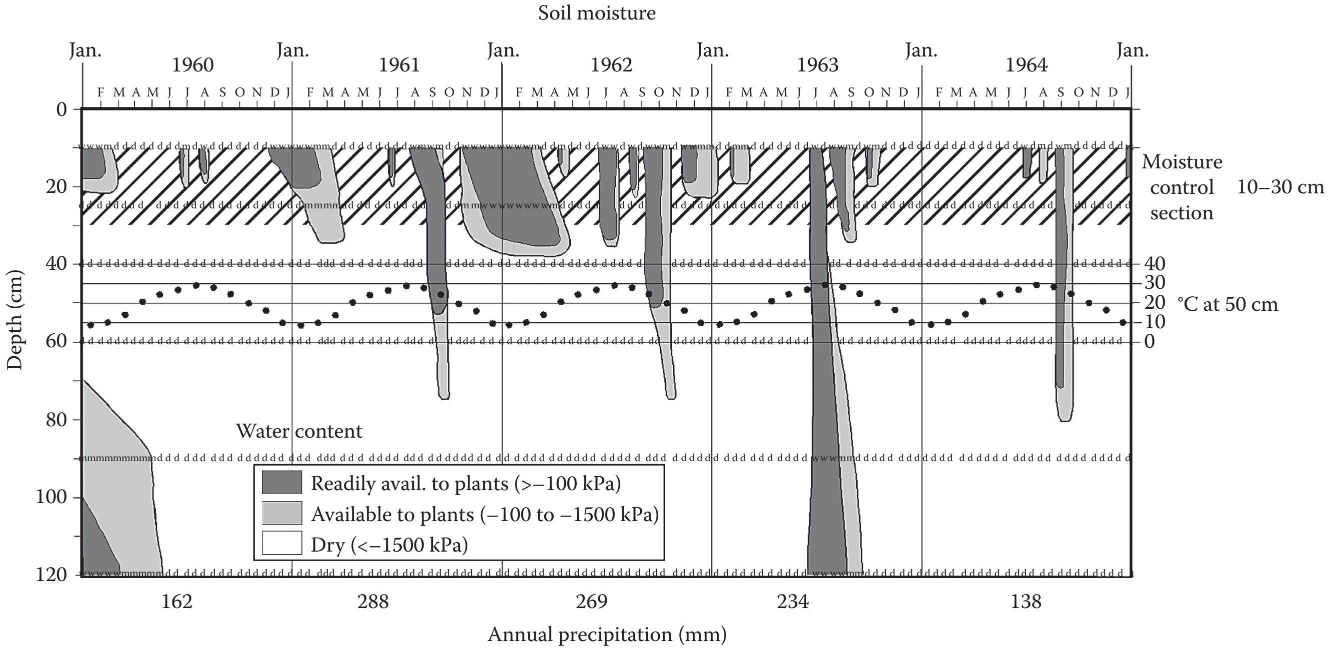


FIGURE 33.66 Soil with an aridic soil moisture regime. The moisture control section (10–30 cm) is dry for over 50% of the time in "normal years" (Soil Survey Staff, 1999). The year 1962, however, was not a normal year and does not meet 50%-dry criterion. Neither did it meet the criterion that the soil cannot be moist for more than 3 months (90 consecutive days). The temperature at 50 cm is above 5°C and 8°C (Soil Survey Staff, 1999). Moisture data from Herbel et al. (1994). Temperature data from Gile and Grossman (1979). Soil is a fine, mixed, superactive, thermic Ustic Calciargid. (From Soil Survey Staff. 1999. Soil taxonomy—A basic system of soil classification for making and interpreting soil surveys. USDA agriculture handbook no. 436. USDA-SCS. U.S. Government Printing Office, Washington, DC.)

soil temperature at 50 cm is above 8°C" (Soil Survey Staff, 1999). In other words, if a soil is moist, "not dry" (i.e., water held at tensions greater than −1500 kPa), for more than 3 months (90 consecutive days), the soil is too moist to have an aridic moisture regime. A graphical illustration of an aridic soil moisture regime is shown in Figure 33.66.

33.10.2 Geographical Distribution

The total ice-free land area on Earth is 130,268,185 km^2. Of that area, 20% or 26,324,771 km^2 is occupied by soils that occur within the aridic moisture regimes. Aridisols occupy about half (54%) of this area (Table 33.45; Figure 33.65). The rest is occupied by Entisols (43%), Vertisols (3%), Oxisols (0.1%), and minute amounts of Andisols (0.005%) with even smaller areas of Gelisols and Mollisols not shown in Table 33.45. Of the total extent of Aridisols, most (41%) occur in Asia, stretching eastward from the Middle East to northern China (Figure 33.65). Africa contains the next largest extent of Aridisols (21%), followed by Australia (16%), North America (13%), South America (8%), and 1% or less in Europe and Central America. In Asia, most suborders are Calcids, Argids, and Cambids that occur in northern China, Mongolia, Kazakhstan, Iran, Afghanistan, Pakistan, Iraq, and Saudi Arabia, with smaller amounts of Salids in Uzbekistan, Turkestan, and Iran, and Gypsids in localized areas throughout Asia, especially Syria and Turkey. Extensive areas of Cryids exist in the higher elevations of China and Mongolia. Substantial areas of Cryids also exist in North and South America where they occur at both high elevations and high latitudes. Africa, Australia, North and South America all have expansive areas of Calcids, Argids, and Cambids with smaller areas of Salids and Gypsids, which also occur in a few areas of southern Europe (Eswaran and Zi-Tong, 1991).

Aridisols in mountainous regions form complex patterns with semiarid soils that have ustic and xeric moisture regimes at higher elevations and in topographic low-lying areas that receive run-on water. These topographically controlled soil patterns are also influenced by seasonality of precipitation. For example, in North America, the most southerly Aridisols occur in the subtropical high pressure belt that receives most of its precipitation in the summer from subtropical convective thunderstorms and border the ustic soil moisture regime. Farther north, Aridisols of the intermountain west occur in more continental climates, but precipitation is still dominated by summer thunderstorms driven by orographic lifting in the Rocky Mountains. To the west, Aridisols of the Great Basin region receive most of their precipitation in the winter from Pacific frontal storms and have Aridisols that border the xeric moisture regime.

33.10.3 Diagnostic Horizons and Classification

33.10.3.1 Epipedons

Virtually, all Aridisols have the ochric epipedon, which by definition is too thin, too dry, too light in color, contains too little organic carbon, has too high an *n* value or melanic index, or is too hard to qualify as any of the other epipedons (Soil

TABLE 33.46 Abbreviated Definitions of Diagnostic Subsurface Horizons in Aridisols

Horizon	Properties
Argillic	Clay accumulation relative to overlying horizon, evidence of clay translocation, minimum thickness of 7.5 cm if loamy or clayey, 15 cm if sandy (Bt)
Calcic	$CaCO_3$ accumulation, generally ≥15 cm thick, ≥15% $CaCO_3$ equivalent, ≥5% more $CaCO_3$ than underlying horizon. In coarse-textured soil > 5% $CaCO_3$ equivalent (Bk or Bkk)
Cambic	Evidence of alteration, reddening, development of soil structure, loss of carbonates (Bw, B, or Bq)
Duripan	Silica-cemented, does not slake in water or acid, evidence of opal accumulation (Bqm)
Gypsic	Gypsum accumulation, ≥15 cm thick, ≥5% gypsum, ≥1% visible secondary gypsum, thickness in centimeter multiplied by % gypsum has product ≥ 150 (By or Byy)
Petrocalcic	$CaCO_3$-cemented, ≥10 cm thick, or ≥1 cm thick if a laminar cap is underlain directly by bedrock, does not slake in water but dissolves in acid, very hard or harder when dry (Bkm or Bkkm)
Petrogypsic	Gypsum-cemented, ≥10 cm thick, ≥5% gypsum and thickness in cm multiplied by % gypsum has product ≥50, roots cannot enter except in vertical fractures that have a horizontal spacing of ≥10 cm (Bym or Byym)
Natric	Properties of argillic horizon, plus columnar or prismatic structure, or degraded blocky structure, sodium adsorption ratio (SAR) ≥ 13 or exchangeable sodium percentage (ESP) ≥ 15 (Btn)
Salic	Accumulation of salts more soluble than gypsum, ≥15 cm thick, electrical conductivity (EC) ≥ 30 dS/m, thickness in centimeter multiplied by EC has product ≥ 900 (Az, Bz, or Cz)

Sources: Soil Survey Staff. 1999. Soil taxonomy—A basic system of soil classification for making and interpreting soil surveys. USDA agriculture handbook no. 436. USDA-SCS. U.S. Government Printing Office, Washington, DC; Soil Survey Staff. 1975. Soil taxonomy: A basic system of soil classification for making and interpreting soil surveys. Agriculture handbook no. 436. USDA-SCS. U.S. Government Printing Office, Washington, DC, 1975.

Horizon designations typically associated with diagnostic horizons are given in parentheses.

Survey Staff, 1999). Dark anthropic epipedons can also occur in Aridisols, but are very rare and localized. If irrigated for long time periods, a mollic-like epipedon can form, but is grouped with the anthropic epipedons rather than the mollic for taxonomic purposes (Soil Survey Staff, 1999). Surface horizons of Aridisols are uniquely different from surface horizons of more humid soils because of their desert pavement, desert varnish, vesicular A horizon, or biological soil crusts.

33.10.3.2 Subsurface Horizons

Subsurface horizons include the cambic, argillic, natric, calcic, petrocalcic, gypsic, petrogypsic and salic horizons, and the duripan. None of these horizons is unique to Aridisols, although the occurrence of salic, gypsic, and petrogypsic horizons in other soil orders is rare. A summary of the properties of these horizons is given in Table 33.46. Many combinations of these horizons are possible in Aridisols depending on the interaction of the five soil-forming factors. As an example, Figure 33.67 shows a petrocalcic, calcic, and argillic horizons, as well as the ochric epipedon in two Aridisols in the southwestern United States. Despite being in arid climates, Aridisols can have prominent pedogenic horizons that display advanced stages of soil development (cf., Huenneke and Noble, 1996).

33.10.3.3 Suborders, Great Groups, and Subgroups

The nine subsurface horizons (Table 33.46) and the cryic soil temperature regime are used to differentiate seven suborders, presented below in the sequence they are keyed out in *Soil Taxonomy*: (1) Cryids (Aridisols in cold areas), (2) Salids (accumulations of salts more soluble than gypsum), (3) Durids (accumulations of and cementation by secondary silica), (4) Gypsids (accumulations of gypsum), (5) Argids (accumulations of illuvial clay), (6) Calcids (accumulations of pedogenic carbonates), and (7) Cambids (weak translocation or transformation of subsoil material). The Cryids occur at high latitudes and high elevations and are common in the intermountain western region of the United States (Hipple et al., 1990). The Argids, Durids, and Petrocalcids (petrocalcic horizon) occur on the oldest geomorphic surfaces, usually late Pleistocene and older, as on the least eroded parts of dissected alluvial fans. Cambids and other Calcids are found on geologically younger side slopes and surfaces of intermediate age, usually latest Pleistocene or younger. Gypsids and Salids often occur near playa margins, where salts are concentrated at or near the soil surface by upward flux from a water table. Less commonly, they occur in association with Cambids and Calcids where parent materials are saline or where gypsum is added with eolian material (Reheis, 1987; Eswaran and Gong, 1991).

The keying sequence follows, for the most part, properties that have the greatest constraints to use and management (Ahrens and Eswaran, 2000). The cold climate of Cryids, for example, presents a limitation that cannot be overcome by reclamation efforts. Similarly, the Salids present special problems because drainage and high-quality water are required before productive agriculture is possible. These 7 suborders are subsequently subdivided into 28 great groups, which in turn are subdivided into more than 269 subgroups (Table 33.47). The great groups are differentiated on the basis of a number of other properties, including aquic conditions, cementation of horizons by gypsum or carbonates, and presence of diagnostic subsurface horizons not identified at the suborder level, generally at depths between 100 and 150 cm. Subgroups are differentiated by a large number of properties. Some common criteria are shallow lithic contacts, presence of nodules and concretions, soil properties that grade toward Andisols or Vertisols, and soil moisture regimes that border on ustic and xeric.

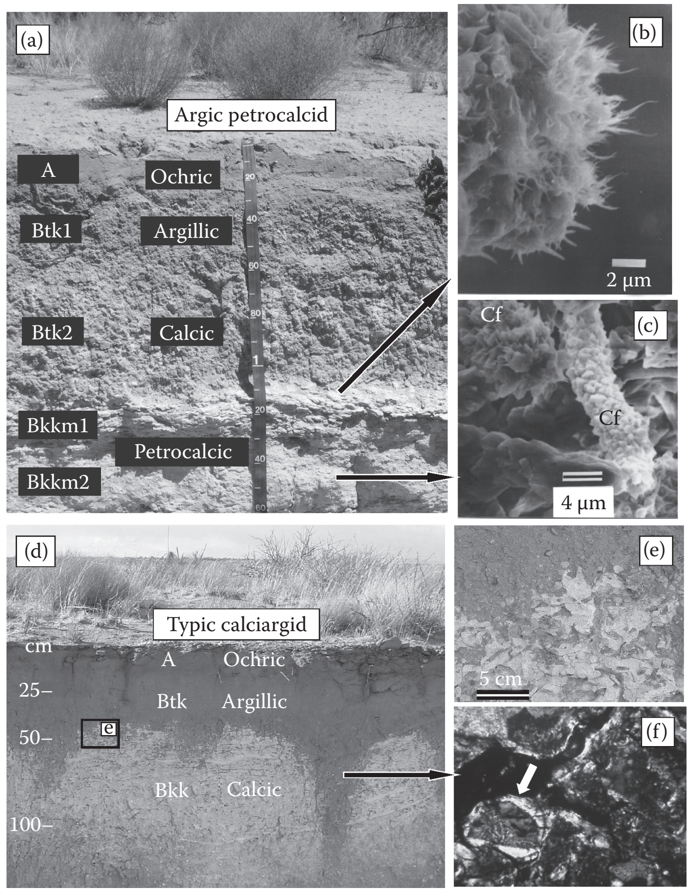


FIGURE 33.67 Morphological and micromorphological features of two Aridisols in southern New Mexico. (a) Illustration of a petrocalcic horizon containing (b) authigenic palygorskite fibers radiating into pore space and (c) calcified fungal hyphae filaments (cf.) suggesting biomineralization of calcite by soil microbes. (d) Illustration of the degradation of a calcic horizon indicated by K-fabric crosscut by noncalcareous material (e) and a vertical pipe containing grain argillans suggesting clay illuviation (f).

33.10.4 Properties

33.10.4.1 Physical Properties

Physical properties of Aridisols can vary dramatically within a local landscape as a consequence of differences in the five soil-forming factors. Particle size distribution, for example, in an area 32 × 32 km^2 (400 mi^2) area in southern New Mexico contains soils with nine particle size classes: sandy-skeletal, loamy-skeletal, clayey-skeletal, sandy, coarse-loamy, fine-loamy, fine-silty, fine, and very fine (Gile et al., 2003). In the same area, bulk density values ranges from a high of 2.19 g cm^{-3} in the laminar zone of a petrocalcic horizon to a low of 0.99 g cm^{-3} in a gravelly loam B horizon. Infiltration is highly variable as well. When a silty soil surface becomes barren upon the loss of desert grasses and becomes sealed via a physical crust, the amount of water entering the soil is significantly reduced compared with neighboring soils still coved with grasses (Herbel and Gibbens, 1989; Bhark and Small, 2003). Consequently, positive feedback loops develop (Ludwig et al., 2005; Monger and Bestelmeyer, 2006) that exacerbate desertification and create heterogeneous vegetative patterns, such as banded vegetation and localized "islands of fertility" (Schlesinger et al., 1990; Tongway et al., 2001; Rango et al., 2006).

33.10.4.2 Chemical Properties

Organic C and nitrogen contents in Aridisols are among the lowest for all soils. Maximum concentration of organic C in ochric epipedons is often <1% on a mass basis due to low biomass production and rapid decomposition. In Aridisols, as in other soil orders, organic C is inversely related to mean annual soil temperature. Thus, in cool and cold arid regions (frigid and cryic soil temperature regimes), organic C contents rival those of warmer semiarid to subhumid climates (Southard, 2000). Nitrogen occurs mostly in organic matter and is quickly nitrified after the organic matter is decomposed. The NO_3^- produced is susceptible to some leaching, but more importantly, to runoff during overland flow and accumulation in playa environments where denitrification can occur (Schlesinger et al., 1990, 2006).

Soluble salts have accumulated in some, but not all, Aridisols as the result of limited leaching and high evaporation near the soil surface. The presence of salts in Aridisol profiles reflects their high solubility and indicates their relative susceptibility

TABLE 33.47 Suborders, Great Groups, and Subgroups of Aridisols

Suborders	Great Groups	Subgroups
Cryids	Salicryids	Aquic, Typic
	Petrocryids	Xereptic, Duric Xeric, Duric, Petrogypsic, Xeric, Ustic, Typic
	Gypsicryids	Calcic, Vitrixerandic, Vitrandic, Typic
	Argicryids	Lithic, Vertic, Natric, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
	Calcicryids	Lithic, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
	Haplocryids	Lithic, Vertic, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
Salids	Aquisalids	Gypsic, Calcic, Typic
	Haplosalids	Duric, Petrogypsic, Gypsic, Calcic, Typic
Durids	Natridurids	Vertic, Aquic Natrargidic, Aquic, Natrixeralfic, Natrargidic, Vitrixerandic, Vitrandic, Xeric, Typic
	Argidurids	Vertic, Aquic, Abruptic Xeric, Abruptic, Haploxeralfic, Argidic, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
	Haplodurids	Aquicambidic, Aquic, Xereptic, Cambidic, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
Gypsids	Petrogypsids	Petrocalcic, Calcic, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
	Natrigypsids	Lithic, Vertic, Petronodic, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
	Argigypsids	Lithic, Vertic, Calcic, Petronodic, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
	Calcigypsids	Lithic, Petronodic, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
	Haplogypsids	Lithic, Leptic, Sodic, Petronodic, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
Argids	Petroargids	Petrogypsic Ustic, Petrogypsic, Duric Xeric, Duric, Natric, Xeric, Ustic, Typic
	Natrargids	Lithic Xeric, Lithic Ustic, Lithic, Xeretic, Ustertic, Vertic, Aquic, Durinodic Xeric, Durinodic, Petronodic, Glossic Ustic, Haplic Ustic, Haploxeralfic, Haplic, Vitrixerandic, Vitrandic, Xeric, Ustic, Glossic, Typic
	Paleargids	Vertic, Aquic, Arenic Ustic, Arenic, Calcic, Durinodic Xeric, Durinodic, Petronodic, Ustic, Petronodic, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
	Gypsiargids	Aquic, Durinodic, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
	Calciargids	Lithic, Xerertic, Ustertic, Vertic, Aquic, Arenic Ustic, Arenic, Durinodic Xeric, Durinodic, Petronodic Xeric, Petronodic Ustic, Petronodic, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
	Haplargids	Lithic Ruptic-Entic, Lithic Xeric, Lithic Ustic, Lithic, Xerertic, Ustertic, Vertic, Aquic, Arenic Ustic, Arenic, Durinodic Xeric, Durinodic, Petronodic Ustic, Petronodic, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
Calcids	Petrocalcids	Aquic, Natric, Xeralfic, Ustalfic, Argic, Calcic Lithic, Calcic, Xeric, Ustic, Typic
	Haplocalcids	Lithic Xeric, Lithic Ustic, Lithic, Vertic, Aquic Durinodic, Aquic, Duric Xeric, Duric, Durinodic Xeric, Durinodic, Petronodic Xeric, Petronodic Ustic, Petronodic, Sodic Xeric, Sodic Ustic, Sodic, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
Cambids	Aquicambids	Sodic, Durinodic Xeric, Durinodic, Petronodic, Vitrixerandic, Vitrandic, Fluventic, Xeric, Ustic, Typic
	Petrocambids	Sodic, Vitrixerandic, Vitrandic, Xeric, Ustic, Typic
	Anthracambids	Typic
	Haplocambids	Lithic Xeric, Lithic Ustic, Lithic, Xerertic, Ustertic, Vertic, Durinodic Xeric, Durinodic, Petronodic Xeric, Petronodic Ustic, Petronodic, Sodic Xeric, Sodic Ustic, Sodic, Vitrixerandic, Vitrandic, Xerofluventic, Ustifluventic, Fluventic, Xeric, Ustic, Typic

Source: Soil Survey Staff. 2010. Keys to soil taxonomy. 11th edn. USDA-NRCS. U.S. Government Printing Office, Washington, DC.

to downward leaching vs. upward movement of groundwater due to evaporation. Typically, well-drained Aridisols have chromatographic-type accumulations of the most soluble salts (chlorides and sulfates) at greater depths than those that are less soluble (gypsum and calcium carbonate), whereas more poorly drained Aridisol have soluble salts closest to the surface. The distribution of these salts within the soil profile is reflected in the electrical conductivity (EC) (Table 33.48; Figure 33.68).

Exchangeable sodium and pH can be uniquely higher in Aridisols than in other soils. The high accumulation of Na on the cation exchange complex is a common phenomenon in Aridisols (exchangeable Na is related to Na in solution and is reflected in the Na adsorption ratio [SAR]) (Table 33.48). Hydrolysis of exchangeable Na, particularly when electrolyte concentrations are low (i.e., little or no soluble salts), can produce alkaline soil reactions near pH 10. Swelling and dispersion of soil clays can cause plugging of soil pores and reducing permeability (Levy, 2000). Clay dispersion also enhances the mobility of clay particles in the soil solution and increases the accumulation of clay in subsoil horizons, particularly for natric horizons.

33.10.4.3 Mineralogical Properties

Mineral diversity is high in Aridisols compared with other soil orders for several reasons. First, Aridisols are widely distributed throughout a wide range of parent materials over vast areas of Earth's surface (Figure 33.65; Nettleton and Peterson, 1983; Allen, 1990). Second, in young Aridisols, there has been relatively little weathering of inherited minerals since the arid

TABLE 33.48 Selected Properties[a] of Representative Aridisols

Depth (cm)	Horizon	Sand (%)	Silt (%)	Clay (%)	OC (%)	N (%)	CCE (%)	BD (g cm^{-3})	pH	EC (dS m^{-1})	SAR
Xeric Calcicryid											
0–10	A	32	56	12	2.98	0.275	15	0.95	7.7	1.7	tr
10–18	Bk1	30	55	15	1.83	0.189	80		8.2	1.0	1
18–33	Bk2	34	44	22	0.78	0.078	41	1.17	7.8	12.9	18
33–58	Bk3	35	41	24	0.37	0.037	73	1.27	8.0	19.0	21
58–84	Bkq1	31	34	35	0.38		54	1.11	7.9	19.6	19
84–107	Bkq2	31	30	39	0.46		88		7.7	30.2	19
107–152	Bkq3	38	32	30	0.23		62		7.8	24.3	18
Typic Haplosalid											
0–11	A	1	50	49	0.69	0.064	4		8.4[b]	22.5	67
11–15	Cyz	1	44	55	0.38	0.039	2		8.7	116.7	281
15–20	Czy	2	55	43	0.60	0.063	3		8.8	122.5	553
20–36	Cz	10	53	37	0.44	0.046	4		8.4	83.8	201
36–56	C1	17	54	29	0.27	0.030	6		8.4	66.7	110
56–76	C2	2	58	40	0.26	0.033	13		8.6	63.0	110
76–105	C3	4	78	18	0.18	0.019	11		8.6	37.7	74
105–150	C4	4	53	43	0.21	0.028	13		8.6	38.3	34
Typic Haplodurid											
0–3	A1	86	12	2	0.11	0.013	3		8.2[b]	0.4	1
3–10	A2	81	14	5	0.08	0.015	5		8.3	0.4	1
10–41	A3	75	14	11	0.15	0.021	7		8.3	0.3	1
41–58	Bqkm1	87	7	6	0.20	0.019	14		8.1	0.4	4
58–76	Bqkm2	81	13	6	0.03		9		8.4	0.4	10
76–94	C	87	9	4	0.01		3		8.5	0.5	15
94–140	Cqkm	84	12	4	0.01		14		8.3	1.0	7
140–175	C′	90	8	2			4		8.4	0.9	6
Typic Calcigypsid											
0–0.5	A1	31	43	26	1.27	0.112	12		7.9	2.5	
0.5–5	A2	28	42	30	0.96	0.094	12		7.9	3.8	3
5–13	A3	29	42	29	0.93	0.088	12	1.37	7.9	4.0	3
13–33	Bk1	25	40	35	0.63	0.066	19		7.8	9.2	9
33–56	Bk2	22	40	38	0.63	0.068	26	1.22	7.4	29.1	8
56–84	Cy1	20	30	50	0.26	0.025	10	1.30	7.6	21.3	
84–115	Cy2	10	50	39	0.14		8		7.7	17.4	2
115–140	Cy3	15	41	44	0.16		14		7.8	16.6	6
140–180	Cy4	14	43	43	0.18		11		7.9	21.3	7
Typic Haplargid											
0–3	A1	58	28	14	0.98	0.095	—	1.49	6.9[b]	0.5	
3–12	A2	60	23	17	0.26	0.038	—	1.50	6.6	0.5	
12–23	BAt	57	22	21			—	1.63	6.9	0.3	
23–46	Bt	46	24	30			—	1.47	7.5	0.2	
46–66	Btk1	40	31	29			3	1.39	7.9	0.2	
66–84	Btk2	33	31	36			8	1.41	8.0	0.2	
84–110	Btk3	31	39	29			15	1.34	8.1	0.3	
110–130	BCtk1	34	40	26			16	1.45	8.1	0.2	
130–160	BCtk2	33	44	23			18	1.60	8.3	1.0	7
160–180	BCtk3	27	49	24			12	1.42	8.1	2.7	11
180–230	C	32	49	19			9	1.37	7.9	6.8	10

(*continued*)

TABLE 33.48 (continued) Selected Properties[a] of Representative Aridisols

Depth (cm)	Horizon	Sand (%)	Silt (%)	Clay (%)	OC (%)	N (%)	CCE (%)	BD ($g\ cm^{-3}$)	pH	EC ($dS\ m^{-1}$)	SAR
Argic Petrocalcid											
0–5	A	85	5	10	0.25		—				
5–18	BAt1	87	4	9	0.13		—	1.86[c]			
18–25	BAt2	82	5	13	0.17		—	1.78			
25–36	Bt	80	5	15	0.14		1	1.70			
36–48	Btk	77	6	17	0.24		16	1.53			
48–74	Bkm1	68	9	23	0.15		75				
74–100	Bkm2	71	11	18	0.05		65	1.68			
100–150	Bkm3	75	6	19	0.05		51	1.66			
Typic Haplocambid											
0–6	A	47	43	10	0.52		tr	1.41	8.0		
6–15	Bw1	44	45	12	0.31		tr	1.26	8.3		
15–33	Bw2	46	43	11	0.27		tr	1.22	8.5		
33–48	Bk1	44	44	12	0.35		4	1.22	8.7	0.7	5
48–60	Bk2	65	29	6	0.28		tr	1.24	8.7	1.7	10
60–87	Bk3	49	44	7	0.22		tr	1.17	8.2	7.8	15
87–135	C1	39	52	9	0.22		tr	1.24	7.9	13.3	14
135–150	C2	66	25	9	0.12		tr	1.30	7.8	11.8	12
150–175	C3	62	29	9	0.12		tr	1.30	7.7	11.1	11
175–205	C4	64	22	14	0.10		tr	1.30	7.9	10.6	9

Source: Data from Soil Survey Staff. 1975. Soil taxonomy: A basic system of soil classification for making and interpreting soil surveys. Agriculture handbook no. 436. USDA-SCS. U.S. Government Printing Office, Washington, DC. USDA-NRCS, 1997c.

[a] OC, organic carbon; N, nitrogen; CCE, calcium carbonate equivalent; BD, bulk density at 33 kPa water potential; EC, electrical conductivity; and SAR, sodium adsorption ratio.

[b] pH of 1:1 soil:water mixture, unless noted by "b," then pH of saturated paste.

[c] Bulk density at 33 kPa water potential, unless noted by "c," then air-dry.

climate provides little surplus water for mineral hydrolysis or for leaching of weathering products. Third, the dry climate enables soluble minerals to accumulate that would otherwise be flushed from soils of more humid climates. Fourth, older Aridisols span multiple glacial–interglacial cycles of the Pleistocene; thus, the mineralogy of wetter paleoclimates is superimposed on the mineralogy of arid recent climates, such as the kaolinitic soils in regions of West Africa and Australia.

Consequently, family mineralogic classes for Aridisols in *Soil Taxonomy* include the following categories: Mixed, Smectitic, Carbonatic, Gypsic, Siliceous, Glassy, Illitic, and Kaolinitic (Southard, 2000). Mineralogy of the sand and silt fractions generally reflects relatively little weathering with a silicate mineral suite that is dominated by quartz, feldspar, amphiboles, and micas, with a few Aridisols containing abundant inherited zeolites (Allen, 1990; Ming and Boettinger, 2001). Mineralogy of the clay fraction is usually dominated by smectites (usually montmorillonite), as reflected by the family mineralogical classes of *Soil Taxonomy* (Southard, 2000). However, even within a small region, the suite of clay minerals can be diverse. Analysis of 154 horizons from 72 pedons revealed the presence of illite, kaolinite, smectite, palygorskite, sepiolite, chlorite, interstratified mica-montmorillonite, and vermiculite (Monger and Lynn, 1996). In these Aridisols, illite and kaolinite were found to be ubiquitous and their concentrations changed little with depth or age, unlike smectite whose concentration formed a bulge in B horizons and became more abundant with age. Palygorskite and sepiolite occurred in well-developed petrocalcic horizons, whereas chlorite, interstratified mica-montmorillonite, and vermiculite were more common in soils formed in igneous parent material than in soils derived from limestone.

Mineralogy of Aridisols may have significant contents of relatively soluble, nonsilicate minerals, including calcite ($CaCO_3$), gypsum ($CaSO_4 \cdot 2H_2O$), halite (NaCl), thenardite (Na_2SO_4), mirabilite ($Na_2SO_4 \cdot 10H_2O$), epsomite ($MgSO_4 \cdot 7H_2O$), nahcolite ($NaHCO_3$), and trona ($Na_2CO_3 \cdot 10H_2O$) (Doner and Lynn, 1989; Buck et al., 2006).

33.10.5 Pedogenesis of Aridisols

The most distinguishing characteristic about soil formation in Aridisols is the limited water available for plant growth, mineral weathering, and leaching of soluble weathering products. A typical water budget is dominated by a long period of deficit during the summer when potential evapotranspiration greatly exceeds precipitation followed by partial recharge of water-holding capacity, then rapid utilization of that stored water during the brief growing season. The lack of a surplus of precipitation over evapotranspiration and water-holding capacity prevent wholesale leaching of weathering products from Aridisols. Nonetheless, many Aridisols have well-developed soil profiles due to (1) the concentration of the limited

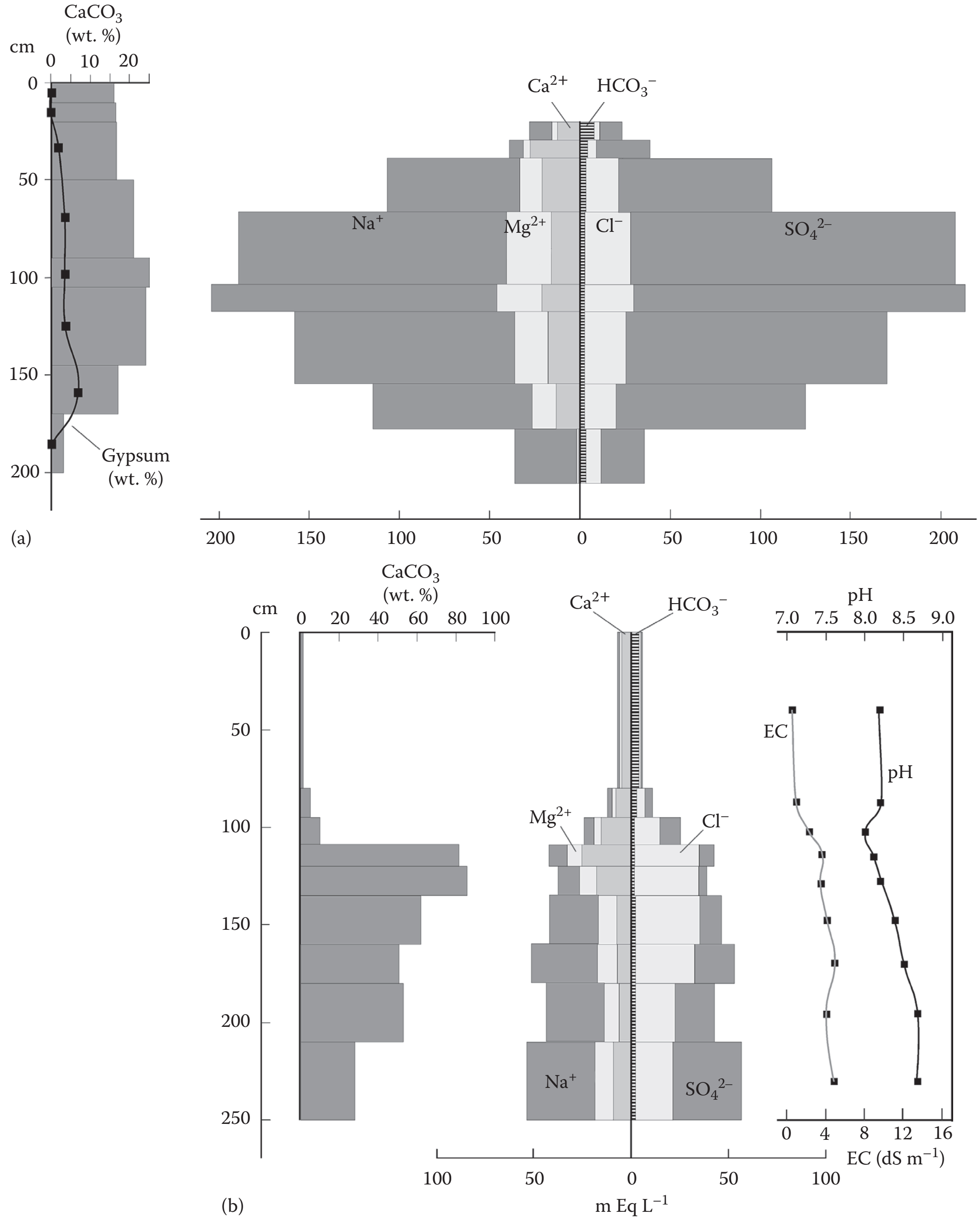


FIGURE 33.68 Salinity profiles, pH, EC, $CaCO_3$, and gypsum content illustrating chemical properties of two Aridisols in southern New Mexico. (a) Vertic Haplocambid formed in the clayey sediments of a playa with high soluble salt concentrations and (b) a neighboring Argic Petrocalcid formed sandy alluvium on a basin floor.

water in a relatively small volume of the upper soil profile, (2) occasional periods of relatively high precipitation in the winter months when deeper leaching could occur, and (3) wetter paleoclimates. The depth of leaching and horizon development may reflect the rainfall of the extreme years, rather than that of average years (Archibold, 1995).

Working in concert with the effects of soil moisture is the effects of soil temperature. Soil temperature regimes of Aridisols range from cryic to isohyperthermic (Soil Survey Staff, 1999). Although the rate of chemical weathering increases as temperature increases (Buol et al., 2002), Aridisols in cooler temperatures retain water longer and dissolve more CO_2 (Breecker et al., 2009). Most Aridisols develop in sediments transported by wind or water on relatively low gradient landscapes. Lithology of the sediments varies widely. Truncation of steep landforms and slow weathering rates retard the formation of subsurface soil horizons and result in Entisols on most erosional landforms (Nordt et al., Section 33.4). Some Aridisols have formed in residual rock material on erosional landforms, for example, on old and highly weathered granitic pediments in the Mojave Desert (Boettinger and Southard, 1991, 1995) and on stable, steep hills where episodic erosion has been minimal (Nettleton and Peterson, 1983).

The four generalized soil-forming processes of additions, transformations, translocations, and losses (Simonson, 1959) are operating in Aridisols, as in more humid soils, but at different rates. Of particular importance, as discussed in the sections below, are eolian accretions and rain chemistry (additions), neoformation of silicate clay, silica, carbonate, and gypsum (transformation), and illuviation of silicate clay (translocations).

33.10.5.1 Atmospheric Additions

Because Aridisols that are of the "nonflushing type" (Rode, 1962), their profiles are cumulative repositories of airborne particles and rain-borne cations and anions. These atmospheric additions have major impacts on the physical and chemical properties and provide a historical record of paleoenvironmental changes (e.g., Chadwick and Davis, 1990; Reynolds et al., 2001; Graham et al., 2008).

33.10.5.1.1 Eolian Accretions

Eolian accretions in many Aridisols are well documented (Yaalon and Ganor, 1973; Gile and Grossman, 1979; Reheis, 1987; Reheis et al., 1995; Valentine and Harrington, 2006). The role of eolian dust is particularly important in many Holocene Aridisols, where much of the carbonate, gypsum, and opaline silica are derived from an external source (Harden et al., 1991). In some cases, eolian material constitutes nearly the entire fine-earth fraction and accounts for the nearly stone-free vesicular horizons below many desert pavements (McFadden et al., 1987). Eolian accretion in Aridisols is enhanced if they have a surface structure that can trap dust. For example, landscapes derived from rock types that tend to weather to soils with few surface clasts are less likely to have significant eolian accretion because an effective dust trap is absent (Boettinger and Southard, 1991).

Eolian accretion of carbonate dust is often required for the development of thick petrocalcic horizons in soils formed in noncalcareous parent materials given the limited amount of water necessary for silicate mineral weathering (Gile et al., 1966; Birkeland, 1984). Ruhe (1967), for example, calculated that there was insufficient calcium in rhyolite alluvium to account for the large amount of calcium contained in the calcic and petrocalcic horizons formed in rhyolitic alluvium. Moreover, even if the rhyolitic parent material had contained enough calcium, the rocks were only slightly weathered and therefore seemed probable that the source of calcium was atmospheric, especially since depth to groundwater was over 100 m. Eolian accretion of carbonate dust, however, is not prerequisite to the formation of all petrocalcic and calcic horizons. Aridisols formed on limestone bedrock or in limestone alluvium may have some accumulation of calcareous eolian dust, but its impact in terms of carbonate source may be of lesser importance in these soils (Rabenhorst and Wilding, 1986).

33.10.5.1.2 Rainwater Chemistry

Rainwater typically contains various amounts of Ca^{2+}, Mg^{2+}, K^+, Na^+, SO_4^{2-}, and NO_3^-. An important consequence of this impure rain is that it can be an important source of nutrients for arid ecosystems (Schlesinger, 1997). Another important consequence is that calcium in rain, like calcium in dust, is a potential source of calcium for the formation of petrocalcic horizons. For example, a 10-year analysis of dust trap data revealed that the calcareous dust, combined with water-soluble calcium extracted from the dust, could generate 0.4 g of soil $CaCO_3$ per meter squared per year (Gile and Grossman, 1979). Rain, however, could generate 1.5 g m^{-2} $year^{-1}$, assuming 200 mm of annual rainfall and a Ca^{2+} concentration of 3 mg L^{-1} (Gile et al., 1981).

33.10.5.2 Surface Features

Pedogenic processes in Aridisols produce unique surface features not found in soils of more humid regions (Cooke et al., 1993). Of special interest are (1) desert pavement, (2) desert varnish, (3) vesicular horizons, and (4) biological soil crusts. In each case, the open surface exposure to sun, wind, rain, temperature fluctuations, and lack of vascular plant roots are involved with their formation.

33.10.5.2.1 Desert Pavement and Varnish

Desert pavement, a surficial layer of rock fragments, usually one or two fragments thick, occurs on many Aridisols (Figure 33.69a and b). The origin of desert pavement has been attributed to loss of fine particles by wind deflation, water erosion, splitting of stones, vertical sorting of coarse fragments by repeated wetting and drying, and by rafting of surface fragments by infiltrating eolian material (Springer, 1958; Cooke, 1970; McFadden et al., 1987). In the case of rafting of surface fragments (Figure 33.69b), the pavement itself serves as a dust trap (Yaalon and Ganor, 1973; Gile, 1975a; Peterson, 1977). The clasts shown in Figure 33.69a, for example, act as a dust trap to catch eolian materials derived from nearby playa and alluvial fans. The effectiveness of desert pavement decreases as surface fragments become more interlocked. When the land surface is nearly completely covered by pavement, the surface neither retains much new eolian material nor loses much material to erosion, unless disturbed.

Desert varnish is a yellowish- or reddish-brown to nearly black coating on desert pavement rock fragments. The varnish is commonly black on pebble surfaces aboveground, but reddish-brown, red, or yellowish red belowground in soils with argillic horizons (Gile and Grossman, 1979). Its composition includes 2:1 silicate clay minerals and oxyhydroxides of Fe and Mn. These constituents are probably of eolian origin and appear to be microbially precipitated (Dorn and Oberlander, 1981; Broecker and Liu, 2001). The varnish occurs most frequently on stable, mafic and various other rock fragments and is generally thickest (as much as 100 μm thick) and darkest on Pleistocene pavements (Nettleton and Peterson, 1983; Garvie et al., 2008). Desert varnish can form in little more than a decade and may serve as an armor that protects microbial weathering fronts from harsh

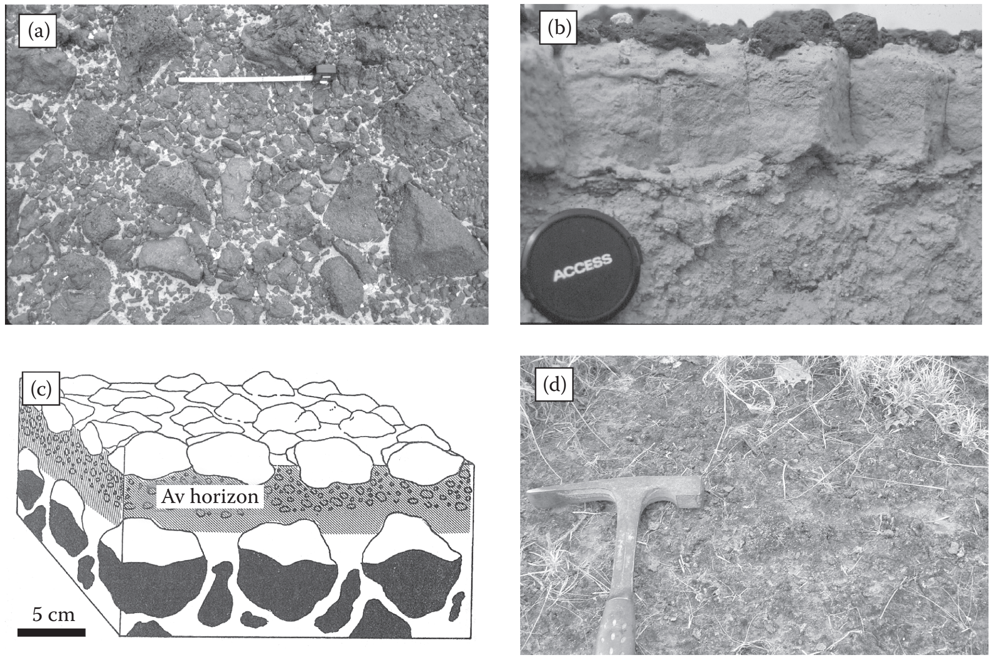


FIGURE 33.69 (a) Desert varnished on clasts interlocking to form a nearly continuous desert pavement on a Pleistocene basalt flows in the Mojave Desert, California. Tape length is about 30 cm. (b) Vesicular surface horizon (an Av horizon) rafting atop desert-varnished basalt pebbles of a desert pavement in the Cima volcanic field in the Mojave Desert. Lens cap is about 60 mm. (c) Diagram of desert pavement underlain by a thin, vesicular A horizon. (From Gile, L.H., and R.B. Grossman. 1979. The desert project soil monograph. USDA-SCS, Washington, DC.) (d) Biological soil crust covering an otherwise bare inter-vegetative surface of a silty soil in the Chihuahuan Desert, New Mexico.

desert environments of irradiation, extreme temperature fluctuations, and low moisture conditions (Krumbein and Jens, 1981). Comparison of desert varnish on exposed stones of adjacent geomorphic surfaces, the same age shows desert varnish to be absent on limestone (Gile and Grossman, 1979), unless the limestone contains exposed deposits of chert or silicified veins.

33.10.5.2.2 Vesicular A Horizons

Beneath the pavement, especially in silty eolian material, a horizon may form with many fine, discontinuous vesicular pores. These horizons are widespread feature of Aridisols and are generally referred to as an Av horizon ("v" in this context is not to be confused with "v" used to describe horizons with plinthite) (Figure 33.69c). The vesicles result from repeated saturation and desiccation and progressive enlargement of air bubbles trapped beneath the thin, saturated surface layer (Springer, 1958; Miller, 1971). Av horizons coevolve with the formation of desert pavement (Anderson et al., 2002). The rate at which vesicular A horizons form can be quite rapid based on the observation that well-developed vesicular horizon formed in tire tracks known to be less than 2 years old (Gile and Grossman, 1979). The discontinuous porosity of Av horizons, in addition to their platy structure, leads to low infiltration rates and shallow leaching that affect ecologic function (Turk and Graham, 2009).

33.10.5.2.3 Biological Soil Crusts

A continuum of crust types occupies surfaces of many Aridisols depending on climate, soil properties, and disturbance history (Johnston, 1997; Belnap and Lange, 2001). On one end is the abiotic physical crust; on the other end are mosses. Soil biological crusts (a.k.a., microbiotic or cryptogamic crusts) include the cyanobacterial crusts, green algae crusts, lichen crusts, and mosses. Abiotic physical crusts are widespread, but not present everywhere, and appear as a surficial layer of uncemented, coherent fine earth that can be broken free from underlying soil. The crust is generally less than 10–20 mm thick and is often platy in the upper part and massive below. Water infiltration into physical crust is slow, in contrast to the rapid infiltration into uncrusted coppice sand dunes. Physical crusts form by raindrop impacts, localized sheet erosion, and by repeated wetting and drying of loamy soil material (Nettleton and Peterson, 1983).

Biological soil crusts occupy many habitats. They occur on top of the soil surface, within soil, beneath translucent stones, within rock fissures and cracks, and within the airspaces of porous rocks. Functions of biological soil crusts include stabilizing soil surfaces (Belnap and Gillette, 1998), increasing mineral weathering, fixing nitrogen (Malam Issa et al., 2001; Belnap, 2002), modifying evaporation (Verecchia et al., 1995), decreasing seed germination until periodically disturbed (Prasse and Bornkamm, 2000), influencing populations of other soil microorganisms, such as nematodes (Zhi et al., 2009), and affecting surface hydrology (Belnap, 2006). Biological soil crusts can decrease as well as increase runoff depending on the crust type and soil setting. In some sandy soils, cyanobacterial crusts increases runoff as the result of hydrophobic properties and pore plugging, in contrast to silty soils where the roughness of lichen crusts slows overland flow and increases infiltration (Kidron et al., 2003; Kidron, 2007).

33.10.5.3 Weathering and Silicate Clay Formation

Physical weathering in Aridisols in the form of disintegration of rock results from many processes, most prominent of which are thermal expansion and contraction, ice crystal growth, root growth in cracks, and salt crystal growth (Birkeland, 1999). Chemical weathering, despite being hampered by long periods of dryness, can be substantial depending on soil age. In Holocene Aridisols of North America chemical weathering is minimal due to the aridity of the Holocene climate (Antevs, 1955). In other regions, such as South Africa, mid-Holocene climates during the Altithermal were wetter and warmer (Linacre and Geerts, 1997). But even in North America, ferrugination (reddening) in arid Holocene soils has been sufficient to create cambic horizons, thus converting Torriorthents into Haplocambids (Gile et al., 1995). Aridisols on Pleistocene landscapes in North America show clear signs of weathering, such as the transformation of primary silicate minerals to Ca- and Mg-rich clays (Harden et al., 1991), interstratified mica-vermiculite (Eghbal and Southard, 1993a), vermiculite and Al-rich smectite (Boettinger and Southard, 1995), and Ca and silica in calcareous duripans (Boettinger and Southard, 1991; Eghbal and Southard, 1993c). In addition, the deeply weathered granitic pediments of the Mojave Desert, alteration of biotite to vermiculite are extensive. These alteration and neosynthesis of Al-rich smectites from feldspar weathering are the main sources of phyllosilicates in the clay fraction (Boettinger and Southard, 1995). Similar evidence of silicate clay neoformation during more humid conditions of the Pleistocene was found in the granitic soils in the Chihuahuan Desert (Monger and Lynn, 1996), as discussed in Section 33.10.4.3.

The pedogenic formation of the fibrous clays, palygorskite [ideal structure: $Si_8Mg_5O_{20}\cdot(OH)_2(OH_2^-)_4\cdot 4H_2O$], and sepiolite [ideal structure: $Si_{12}Mg_8O_{30}(OH)_4(OH_2)_4\cdot 8H_2O$] (Singer, 1989) has been documented in the alkaline, Si-preserving environment of petrocalcic horizons. Scanning electron microscopy of palygorskite fibers radiating into pore spaces has been used as evidence of in situ formation, rather than illuviation of palygorskite (Figure 33.67b; Monger and Daugherty, 1991a).

33.10.5.4 Pedogenic Silica

Pedogenic silica in Aridisols accumulates as opal ($SiO_2\cdot nH_2O$) (Flach et al., 1969; Chadwick et al., 1987a), opal-CT (Monger and Kelly (2002)), and possibly as chalcedony and microcrystalline quartz (Flach et al., 1973; Smale, 1973; Milnes et al., 1991). Incipient cementation produces horizons that are very hard when dry and brittle when moist. More complete cementation of the soil by opal produces a duripan. Pedogenic silica, as with pedogenic carbonate, has stages whereby progressively more complex silica accumulations occur on progressively older geomorphic surfaces (Taylor, 1986). But unlike pedogenic carbonate, which cements by engulfing soil particles like mortar between bricks, pedogenic silica cements by bonding soil particles at contacts like glue (Chadwick and Nettleton, 1990). Relatively small amounts of silica (about 10% by weight) cause significant cementation of clay particles (Nettleton and Peterson, 1983), which can significantly change water retention characteristics and CEC. Cementation by silica tends to be less disruptive of soil fabric than cementation by carbonate.

Climate is an important factor for the accumulation of pedogenic silica. It must provide enough water for weathering, but not enough for intensive leaching (Flach et al., 1973). This condition is met in many arid and semiarid climates. The second condition is a readily available source of $H_4SiO_4°$. This condition is met in many areas where easily weatherable volcanic ash and ultramafic materials rapidly release $H_4SiO_4°$ (Alexander et al., 1994; Alexander, 1995).

Secondary silica in the form of opaline veins and stringers is a frequent feature in many highly indurated petrocalcic horizons (Franks and Swineford, 1959). The source of this silica may be derived by weathering of siliceous rocks (Boettinger and Southard, 1991), volcanic ash (Chadwick et al., 1987b), or eolian dust (Harden et al., 1991). It may also originate from the replacement of quartz and feldspar grains by pedogenic calcite. As revealed by thin section petrography, the original shapes of some silicate grains are gone, occupied instead by calcite—hence, the term "replaced" (Swineford et al., 1958; Reheis, 1988; Monger and Daugherty, 1991b). Replacement of silicate grains has been attributed to *force-of-crystallization* (Maliva and Siever, 1988a, 1988b). This process, like pressure solution in metamorphic rocks, is based on the principle that minerals are more soluble at contact points when under stress. The force-of-crystallization hypothesis in petrocalcic horizons is postulated to occur as calcite progressively crystallizes and engulfs detrital grains. It accounts for why some surface pits in silicate grains match the shapes of calcite crystals impacted against them, and it eliminates the need to have unverifiable high pH to account for silica dissolution (Maliva and Siever, 1988b). Silica released by force-of-crystallization would not only precipitate as opaline phases, but would also be a source of Si for authigenic palygorskite and sepiolite, which are common clay minerals in petrocalcic horizons (Vanden Heuvel, 1966; Monger and Daugherty, 1991a; Reheis et al., 1992).

33.10.5.4.1 Duripans and Silcretes

Most duripans in Aridisols have a considerable content of $CaCO_3$, as much as 70% by weight (Southard et al., 1990), but at least half the volume of a duripan does not slake in acid, as petrocalcic horizons do. Many of these duripans have indurated laminar layers as much as 2 cm thick that overlie massive and less strongly cemented material that is highly calcareous. These duripans can be distinguished from petrocalcic horizons only by acid treatment. In many cases, only the upper laminar layer survives the acid treatment intact.

Duripans in Aridisols commonly occur beneath or within an argillic horizon (Flach et al., 1973). In soils with both an argillic horizon and underlying calcic horizon, duripans often straddle the two horizons, superimposed on the lower portion of the argillic and the upper portion of the calcic, thus, forming a chromatographic column of illuviated constituents (Chadwick et al., 1987a). Silica in duripans is typically some form of opal or chalcedony that is finely distributed throughout the matrix, or as coatings, diffuse isotropic flocs, nodules, glaebules, and stringers (Flach et al., 1973; Boettinger and Southard, 1990; Monger and Adams, 1996). High pH promotes duripan formation in granitic terrain and, in some cases, precludes the requirement for volcanic ash as a source of silica (Boettinger and Southard, 1990).

Silcretes or duricrust in Aridisols is more indurated and generally older than duripans (Jackson, 1957). Their occurrence in the stratigraphic record may stretch back to the Precambrian (Summerfield, 1983). Silcretes develop when sediments become silicified in areas (1) where drainage of silica-rich groundwater is poor and has a fluctuating water table, (2) where silica-rich solutions evaporate, and (3) where upward moving silica-rich solutions meet downward percolating less alkaline or more salty waters (Smale, 1973; Summerfield, 1983; Milnes et al., 1991). Mineralogically, silcretes are a combination of opal-A, opal-CT, and microcrystalline quartz (microquartz) depending on the concentration of $H_4SiO_4°$ (Milnes et al., 1991). Like duripans, high $H_4SiO_4°$ concentrations (>60 mg L^{-1}) favor opal-A; intermediate concentrations (20–60 mg L^{-1}) favor opal-CT; and low $H_4SiO_4°$ concentrations (<20 mg L^{-1}) favor microquartz (Chadwick et al., 1987b).

33.10.5.4.2 Biogenic Silica

Silica cycling by plants, especially grasses, is an important component of Aridisols. Desert plants, especially grasses, use silica for structural stability, mineral nutrition, and antiherbivore defense mechanisms (Iler, 1979; Massey et al., 2006). When silica-containing plants die and decompose, they release silica into the soil as phytoliths. Phytoliths are ubiquitous constituent of not only Aridisols, but most soils (Wilding et al., 1977; Meunier et al., 1999), ranging from 8 to 300 kg ha^{-1} depending on biomass production and mean annual precipitation (Pease, 1967; Kelly, 1989). Except for certain silica-rich geological environments, such as hot springs, most soil opal has a biogenic origin that exerts a major control on silica in the soil solution (Alexandre et al., 1997). The marine record indicates that biogenic silica underwent a spectacular expansion during the middle Cenozoic when grasses proliferated across terrestrial environments (Lowenstam and Weiner, 1989). Vertically in soils, phytoliths are most abundant in surface horizons and decreases with depth, unless buried paleosols are encountered. In paleosols, phytolith morphology and isotopic composition can be a useful method for providing information about ecological and archaeological changes (Wilding and Drees, 1974; Kelly et al., 1991; Rosen and Weiner, 1994; Schaetzl and Anderson, 2005).

33.10.5.5 Pedogenic Carbonate

Several names have been given to pedogenic carbonate: caliche, calcrete, croute calcaire, tosca, caprock, crust, calcic horizons, and petrocalcic horizons (Gile, 1961; Goudie, 1973; Dregne, 1976; Soil Survey Staff, 1999). Although some pedgenic carbonate has been reported to be dolomite (e.g., Capo et al., 2000), the vast majority is calcite (e.g., Doner and Lynn, 1989; Kraimer et al., 2005). Pedogenic carbonate is a subset of *soil carbonate*, which is the total amount of carbonate measured in the soil, typically by acid dissolution methods, and reported as calcium carbonate equivalent (Soil Survey Staff, 1996). Soil carbonate is a broad category containing *primary carbonate* (if present) and *pedogenic carbonate* (if present). Primary carbonate (also termed geogenic, lithogenic, inherited, or nonpedogenic) is $CaCO_3$ originally precipitated in aquatic environments as limestone. Pedogenic carbonate (also termed secondary or authigenic carbonate) is $CaCO_3$ developed in soil and can occur as either authigenic (in situ) $CaCO_3$ or allogenic (ex situ) $CaCO_3$, where allogenic $CaCO_3$ is derived from preexisting soils upslope or downwind. The age of pedogenic carbonate spans vast time periods, ranging from carbonates in Paleozoic paleosols (Mora et al., 1996) to carbonate formed on artifacts (Pustovoytov, 2002) and living plant roots (Monger et al., 2009).

Of the global amount of soil carbonate, Aridisols contain the largest proportion (48%), followed by Entisols (28%), and Mollisols (12%) (Eswaran et al., 2000). At the suborder level, however, Orthents contain the most (21.4%) because many Orthents are formed by the truncation of Aridisols (see Nordt et al., Section 33.4). Other suborders containing notable amounts of soil carbonates are Calcids (17.4%), Argids (14.2%), Cambids (10.0%), Ustolls (9.9%), Psamments (3.5%), Ustalfs (3.1%), Fluvents (2.8%), Gypsids (2.4%), and Salids (2.2%) (Eswaran et al., 2000).

Calcium carbonate is present in almost all Aridisols, except for some Aridisols on stable platforms in West Africa and Australia (Southard, 2000). Inputs of Ca from mineral weathering or from the atmosphere, either dissolved in rainwater or as particles of Ca-bearing minerals, coupled with relatively low solubility, cause $CaCO_3$ to accumulate. Once in soil solution, Ca can react with HCO_3^- from rainwater or produced via soil CO_2 (from respiration) with water. Loss of water by evaporation, or reduction of the partial pressure of CO_2 (P_{CO_2}), drives the precipitation of $CaCO_3$ from solution.

Aridisols on Holocene or latest Pleistocene landscapes in the southwestern United States, for example, are often calcareous in all horizons because the relatively dry conditions of the Holocene prevented leaching of carbonate from surface horizons (e.g., Typic Haplodurid and Typic Calcigypsid in Table 33.48). In contrast, Aridisols on older Pleistocene landscapes are often carbonate free in the upper horizons, presumably due to more effective leaching during the cooler, moister Pleistocene (e.g., Typic Haplargid and Argic Petrocalcid in Table 33.48). Aridisols of West Africa are acidic and lack carbonates because they form on old preweathered platform surfaces or from more recent polycycled sediments derived therefrom.

Soil taxonomy systems have used pedogenic carbonate as an important feature to classify soils (e.g., Pedocal vs. Pedalfer, Marbut, 1928). Pedogenic carbonate has also provided important clues about paleoclimate, paleoecology, paleoatmospheric composition, relative ages of geomorphic surfaces, landscape evolution, carbon sequestration, and vegetation patterns in arid landscapes (e.g., Mora et al., 1996; Ekart et al., 1999; Monger and Martinez-Rios, 2001; Duniway et al., 2007).

33.10.5.5.1 Pedogenic Carbonate Forms

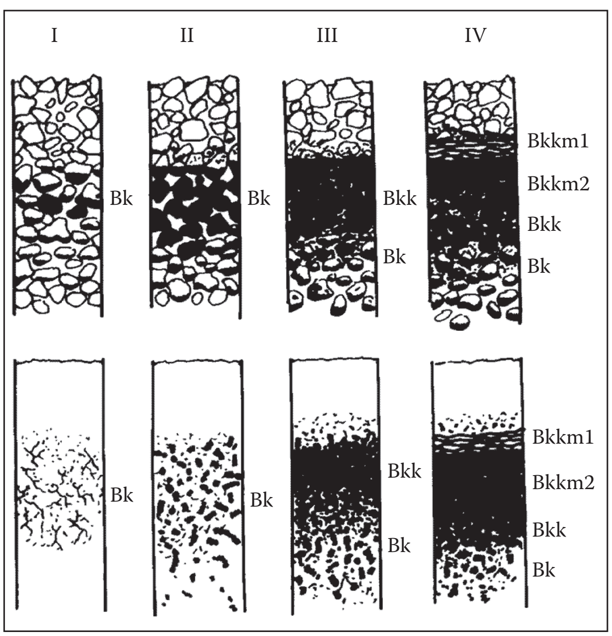
Forms of pedogenic carbonate accumulation have been described as filamentary, concretionary, cylindroidal, nodular, platy, blocky, bedded, massive, veined, and flaky, as well as root casts, bands, joint fillings, coatings, pendants, beds, plugged horizons, laminar horizons, laminae, pisoliths, and ooliths (Gile, 1961; Harden et al., 1991; Eghbal and Southard, 1993b; Soil Survey Division Staff, 1993). These carbonate forms range from nonindurated, which slake when placed in water, to very strongly indurated, which do not slake in water and cannot be scored with a knife. On the low end of carbonate accumulation, horizons with carbonate filaments can contain as little as 1% $CaCO_3$ and, depending on texture, have typical bulk densities (e.g., 1.68 g cm^{-3}) and high infiltration rates (e.g., 12.4 cm h^{-1}, Gile, 1961). On the high end of carbonate accumulation, laminar horizons can contain over 90% $CaCO_3$ with high bulk densities (e.g., 2.22 g cm^{-3}) and low infiltration rates (e.g., 0.1 cm h^{-1}).

Pedogenic carbonate is an important indicator of soil age because progressively older geomorphic surfaces contain progressively greater amounts of carbonate. It was observed that with time, four morphogenetic stages of carbonate develop, progressing from having no visible carbonate to having stage I filaments, then stage II nodules, a stage III plugged horizon, and eventually a stage IV laminar horizon developed atop the plugged horizon (Gile et al., 1966; Figure 33.70). More recently, two additional stages of carbonate accumulation have been added: stages V and VI (Machette, 1985). Stage V contains thick laminae (>1 cm) and pisolites, and stage VI contains multiple generations of laminae, breccia, and pisolites. The time required to reach a certain morphogenetic stage depends on soil texture. Gravelly soils pass through the stages more quickly than fine-textured soils because gravelly soils have lower surface area and less pore space. Associated with an increase in pedogenic carbonate is a large expansion of the volume of the calcic and petrocalcic horizons as carbonate precipitates and forces matrix grains apart. Cementation of horizons occurs when carbonate content reaches 25%–60% (Machette, 1985).

Until about 1965, horizons with carbonate accumulation were designated as C horizons with a "ca" suffix (or "cam" suffix if indurated) following the procedure of the *Soil Survey Manual* (Soil Survey Division Staff, 1951). Subsequent work showed that carbonate accumulation was a pedogenic process (Gile et al., 1965). Therefore, horizons with significant amounts of pedogenic carbonate were afterward designated B horizons,

Schematic diagram of diagnostic carbonate morphology for the stages of carbonate accumulation in the two morphogenetic sequences (left below). Stages are described (right below), including stages V and VI not shown in diagram

Carbonate morphogenetic stages



Definitions of carbonate morphogenetic stages

Stage	Diagnostic carbonate morphology: Gravelly sequence[a]	Diagnostic carbonate morphology: Nongravelly sequence
I	Thin pebble coatings, continuous in last part of stage.	Few[b] filaments, masses, or faint coatings
II	Continuous pebble coatings, some interpebble masses.	Few to common nodules or masses.
III	Many interpebble masses; plugged horizon[c] develops in last part of stage.	Many nodules and internodular masses; plugged horizon develops in last part of stage.
IV	Laminar horizon overlying plugged horizon.	Laminar horizon overlying plugged horizon.
V	Very strongly cemented to indurated with >1 cm thick laminae; may contain pisoliths and have vertical faces and fractures coated with laminated carbonate.	
VI	Indurated with multiple generations of recemented laminae, breccia, and pisoliths.	

[a] Morphologies are best expressed where "nongravelly" materials contain less than about 20% by volume of rock fragments (fragments 2 mm or larger in diameter), and "gravelly" materials contain more than about 60% by volume of rock fragments. Materials that have between 20% and 60% by volume of rock fragments have intermediate morphologies.

[b] Few <2, common 2 to <20, many = 20 or greater percent of area covered.

[c] Plugged horizon contains 50% or more pedogenic carbonate (by vol).

FIGURE 33.70 Illustration and definitions of the morphogenetic stages of pedogenic carbonate accumulation in gravelly soils and soils without gravels. Stages V and VI are not shown in diagram. (From Gile, L.H., F.F. Peterson, and R.B. Grossman. 1966. Morphological and genetic sequences of carbonate accumulation in desert soils. Soil Sci. 101:347–360; Machette, M.N. 1985. Calcic soils of the southwestern United States, p. 1–21. *In* D.L. Weide (ed.) Soils and Quaternary geology of the southwestern United States. Geological Society of America. Special Paper 203.)

not C horizons. The prominence of carbonate horizons in arid and semiarid regions led to the proposed K horizon as a master horizon, such as the O, A, B, C, and R horizons at that time. The K horizon was based on the presence of K-fabric, which is defined as "fine-grained authigenic carbonate that coats or engulfs skeletal pebbles, sand, and silt grains as an essentially continuous medium" (Gile et al., 1965). The K horizon, by definition, contained 90% or more K-fabric. The K horizon was not adopted by the National Cooperative Soil Survey, except that the suffix "k" eventually replaced the "ca" suffix. In other earth science disciplines, however, the K horizon has been used (e.g., Machette, 1985; Birkeland, 1999). Recently, the "kk" suffix, which corresponds to the stage III plugged horizon or higher of the carbonate morphogenetic stages, has been formally adopted by the National Cooperative Soil Survey (Soil Survey Staff, 2010).

33.10.5.5.2 Origin of Pedogenic Carbonate

Models of pedogenic carbonate formation can broadly be classified into four groups (Monger and Wilding, 2002): (1) the *per ascensum*-groundwater model in which carbonate develops from upward rising groundwater via capillarity, (2) the *per descensum*-leaching model in which pedogenic carbonate develops from downward redistribution of calcareous parent material, (3) the *per descensum*-dust model in which pedogenic carbonate develops from atmospheric additions of dust and calcium in rainwater, and (4) the in situ-biogeochemical model in which carbonates develops from biological weathering, Ca upward translocation by hyphae and roots, and precipitation of carbonates within the soil profile via both inorganic and biogenic precipitation. The formation of pedogenic carbonate in any particular soil may involve several of these models working together at different magnitudes.

The *per ascensum*-groundwater model accounts for pedogenic carbonate formation as the result of precipitation of carbonate minerals from groundwater containing high concentrations of Ca^{2+} and HCO_3^- drawn up from shallow water tables by capillarity (Sobecki and Wilding, 1983). This model includes groundwater calcretes. These differ from pedogenic calcretes, that is petrocalcic horizons, in two ways. The groundwater calcretes have coarse (often poikilotopic) interlocking carbonate crystals that engulf detrital grains in contrast to pedogenic calcretes that have fine silt-size carbonate crystals. Second, groundwater calcretes impregnate sediments in a manner that preserves the original sedimentary structure in contrast to pedogenic calcretes that obliterates sedimentary structure.

The *per descensum*-leaching model accounts for pedogenic carbonate formation as the result of dissolution of indigenous carbonates from upper horizons, downward transport of ions in solution, and precipitation of secondary carbonates from soil solution in lower horizons. Biogenic CO_2 raises P_{CO_2} to levels as high as 1.2%, and the relative rates of CO_2 production and loss to the atmosphere by diffusion generally cause P_{CO_2} to increase with soil depth (McFadden et al., 1991). Plant roots take up water preferentially over dissolved HCO_3^-, causing $CaCO_3$ to precipitate due to desiccation. This process occurs in the upper profile of many calcareous loess and alluvial deposits and was invoked to explain the correlation of progressively shallower carbonate horizons in progressively drier climates (Jenny and Leonard, 1934). Later, this model was used as the basis for calculating the number of wetting fronts required to leach carbonates to a particular depth (Arkley, 1963). In both cases, it was assumed that carbonate was uniformly distributed in parent material at the beginning of pedogenesis.

The *per descensum*-dust model was developed in recognition that pedogenic carbonate also occurs in soils that do not have indigenous calcareous parent materials. In southern New Mexico, for example, prominent calcic and petrocalcic horizons occur in soils with siliceous and only-slightly weathered parent materials. In this case, atmospheric additions were judged to be the source of pedogenic carbonates (Gile et al., 1966). Initially, calcareous dust was considered to be the source. Later, it was recognized that Ca^{2+} in rain was an additional, and even larger, source of Ca^{2+} for reaction with soil HCO_3^- to form pedogenic carbonates (Section 33.10.5.1.2). The depth and rate of $CaCO_3$ accumulation in many Aridisols illustrate the importance of varying eolian input and P_{CO_2} (McFadden et al., 1991). Building on these *per descensum* concepts, compartmental models have been constructed that compute the depth, amount, and distribution of pedogenic carbonate as a function of climate and time (Arkley, 1963; McFadden and Tinsley, 1985; Hirmas et al., 2009; Section 33.10.6).

The in situ-biogeochemical model focuses on the biogenic activity of plants and microorganisms in harvesting Ca from parent material and transporting it to the land surface. This process occurs in all soils, but its effect is more apparent in arid and semiarid soils because of their nonflushing nature and nonacidic pH. The in situ-biogeochemical model operates in both noncalcareous and calcareous parent materials. In noncalcareous parent materials, evidence that various plants play a role in carbonate formation comes from chemical analysis of soil profiles showing that plants transfer Ca^{2+} to the land surface from subsoil, rock, or groundwater (Goudie, 1973). Depending on the chemical environment, a portion of this Ca^{2+} is available for precipitation with HCO_3^- derived from CO_2 generated by root and microbial respiration. Moreover, references in the Russian literature note carbonate formation in plant tissue followed by its downward translocation with wetting fronts (Labova, 1967). Evidence that microorganisms precipitate carbonates is based on microscopy of calcified fungal hyphae (West et al., 1988) and in vitro laboratory experiments (e.g., Philips et al., 1987; Monger et al., 1991). Termites may also precipitate carbonate in large termite mounds of Africa and southeast Asia (Thorp, 1949), but not in small sheaths common in North American deserts (Liu et al., 2007). In formed on limestone bedrock, pedogenic carbonate forms via the progressive transformation of limestone into micritic pedogenic carbonate as a result of short-range carbonate dissolution and reprecipitation processes (Rabenhorst and Wilding, 1986), which is a unique process of pedogenic carbonate accumulation because the content of pedogenic carbonate is less than the carbonate content of the original limestone.

33.10.5.6 Pedogenic Gypsum and Soluble Salts

Soils containing gypsum ($CaSO_4 \cdot 2H_2O$) in the form of gypsic or petrogypsic horizons are estimated to cover some 207 million ha around the world (Eswaran and Zi-Tong, 1991). These soils can be grouped into two categories: *gypsiferous*(in which the suffix-ferous means "yielding" or "containing") or *gypseous* (in which the suffix-ous means "abounding in" or "consisting of"; Herrero and Porta, 2000). Unlike pedogenic carbonates that can form in any soil given a source of free calcium, bicarbonate, and nonacidic pH, gypsum formation is restricted to a parent material source of gypsum.

Soils with gypsum have unique properties because of gypsum's softness (2 on the Mohs scale), its solubility (2.6 g L^{-1} at 25°C), and its ability to corrode iron and disintegrate concrete. The disintegration of concrete occurs when sulfate reacts with Na to form mirabilite or thenardite or with Ca and Al to form ettringite [$Ca_6Al_2(SO_4)_3(OH)_{12} \cdot 26H_2O$)] and thaumasite [$Ca_3Si(CO_3)(SO_4)(OH)_6 \cdot 12(H_2O)$] (Herrero et al., 2009). Crystallization of these sulfate minerals increases the volume of solids, thus leading to disintegration.

Gypseous and gypsiferous soils are not necessarily salt-affected soils because gypsum does not significantly increase osmotic potential or ionic toxicity. In fact, gypsum has long been used as a source of calcium and sulfur nutrients, and a material used to reclaim sodic soils (Shainberg et al., 1989). Native vegetation on gypseous soils is distinguished from vegetation of saline soils by having more abundant biomass and greater number species. However, gypseous and gypsiferous can also be saline (Herrero et al., 2009). Salts more soluble than gypsum (soluble salts) most commonly accumulate in soils where there is a source of easily dissolvable minerals or ions in parent material (Gumuzzio et al., 1982), a source of saline water, and a climate with relatively high evapotranspiration rates. Pedogenic accumulation of soluble salts is often associated with gypsum and/or carbonates, depending on the initial ionic composition of the evaporating solution (Boettinger and Richardson, 2001).

Gypsic and salic horizons are typical of basins with playas (Driessen and Schoorl, 1973) where leaching is limited, water tables are high, and where eolian salts are recycled between playas and surrounding terrain. These salts in soil profiles are often seasonally dynamic due to their high solubility and the ease with which they are moved downward by leaching during the wet season and upward by evapotranspiration during the dry season (Eghbal et al., 1989). Aridisols with salic and gypsic horizons may also occur where there is throughflow and seeps (e.g., Buck et al., 2006) and artesian springs in arid to semiarid climates (Boettinger and Richardson, 2001).

33.10.5.7 Illuvial Clay and the Argillic Horizon in Aridisols

The processes by which clay accumulates in Aridisols are similar to those in other soil orders and include disaggregation, dispersion, and illuviation of clays in eolian dust, in situ weathering of primary minerals, illuviation of dispersed clay from surface horizons, and neosynthesis of clay minerals from soil solution (Nikiforoff, 1937; Brown and Drosdoff, 1940; Agriculture Experiment Station-Soil Conservation Service, 1964). Argillic and natric horizons in Aridisols can range in texture from sand to clay with thickness from 7.5 to 75 cm. Argillic and natric horizons generally begin at shallow depths (4–25 cm below the soil surface), which are shallower than argillic and natric horizons in soils of more humid regions. Argillic horizons typically are just above or extend into calcic horizons.

Mineral weathering that produced clays susceptible to dispersion and translocation is primarily a relict of wetter Pleistocene climatic conditions (Nettleton and Peterson, 1983). Wetter conditions of the Pleistocene pluvials were probably also required to dissolve and leach carbonates from surface horizons, the removal of Ca being a prerequisite for dispersion and translocation of the clays (Goss et al., 1973; Southard, 2000). This was shown by the absence of argillic horizons in Aridisols with abundant fragments of limestone in contrast to neighboring Aridisols of the same age formed in igneous alluvium (Gile and Hawley, 1972). The explanation apparently lies in the flocculating effect that carbonates have on clay movement, although clay illuviation has been reported in some calcareous soils (Goss et al., 1973; Pal et al., 2003). The absence of argillic horizons in many Aridisols can be the result of obliteration by landscape dissection (Nordt et al., Section 33.4), engulfment by carbonate (Gile et al., 1969), or faunal mixing (Gile, 1975a,b).

Many argillic horizons in Aridisols appear to be illuvial, even though they lack clay skins (argillans) on peds. This conclusion has been based on several lines of evidence: (1) Soils have thin, grayish E-like horizon with less clay than the underlying reddish-brown or red B horizon; (2) the argillic horizons when traced laterally have prominent clay skins in pipes that penetrate calcic horizon (Figure 33.67f); (3) reddish coatings of silicate clay occur on and in cracks in the tops of petrocalcic horizons and bedrock that underlie argillic horizons without clay skins at shallow depths; and (4) distinct linear bodies of oriented clay occur within peds, interpreted as former clay skins (Buol and Yesilsoy, 1964; Gile and Grossman, 1968; Smith and Buol, 1968; Nettleton et al., 1969, 1975; Gile et al., 1981).

33.10.6 Links to Past Climates

Aridisols are some of the oldest, if not the oldest soils in the world. They range from Cretaceous age or older in Australia (Pillans, 2005), through the Miocene (Hawley, 2005), the Pliocene (e.g., Gardner 1972; Mack et al., 1996), and early Pleistocene in many regions of the world (e.g., Nettleton and Peterson, 1983; Machette, 1985; Alsharhan et al., 1998; Kapur et al., 2000; Matmon et al., 2009). Such ancient soils bear the marks of many paleoclimatic changes at both the landscape scale and profile scale.

At the landscape scale, periods of landscape stability (soil formation) alternate with periods of landscape instability (i.e., erosion and sedimentation), as described by Ruhe (1962):

> By analogy with present condition … and on a theoretical basis … Past pluvial environments in present arid regions, correlative of glacial episodes elsewhere, should have resulted in greater vegetative cover on relatively stable landscapes … and soil formation. Interpluvial environments, as at present, should have resulted in increased aridity, lesser vegetative cover, unstable landscapes subject to severe erosion and to sediment transport.

Because shrublands have much bare ground, erosion increases. In contrast, during wetter climates, grasslands and woodlands (which have little bare ground) increase causing erosion to decrease (Langbein and Schumm, 1958). Such cycles commonly produce stacked sequences of buried paleosols in intermontane basins. Along floodplains, glacial–interglacial climate cycles commonly produce stepped sequences of geomorphic surfaces (Figure 33.71a). In the American Southwest, the model is a follows: (1) Rivers downcut during glacial periods when greater precipitation gives rise to more water in river channels and denser vegetative cover across landscapes, (2) during waning glacial and early interglacial times, river valleys backfill (partially) as a result of less water in the river channels and greater amounts of sediment supplied by erosion from sparsely vegetated landscapes, and (3) aggradation ceases and base levels stabilize during interglacial times, until the cycle began again when downcutting is renewed by the waxing phase of the next glacial cycle (Hawley et al., 1976).

At the pedon scale, mineralogy is a function of climate. Highly weathered minerals can be relicts from a wetter climate

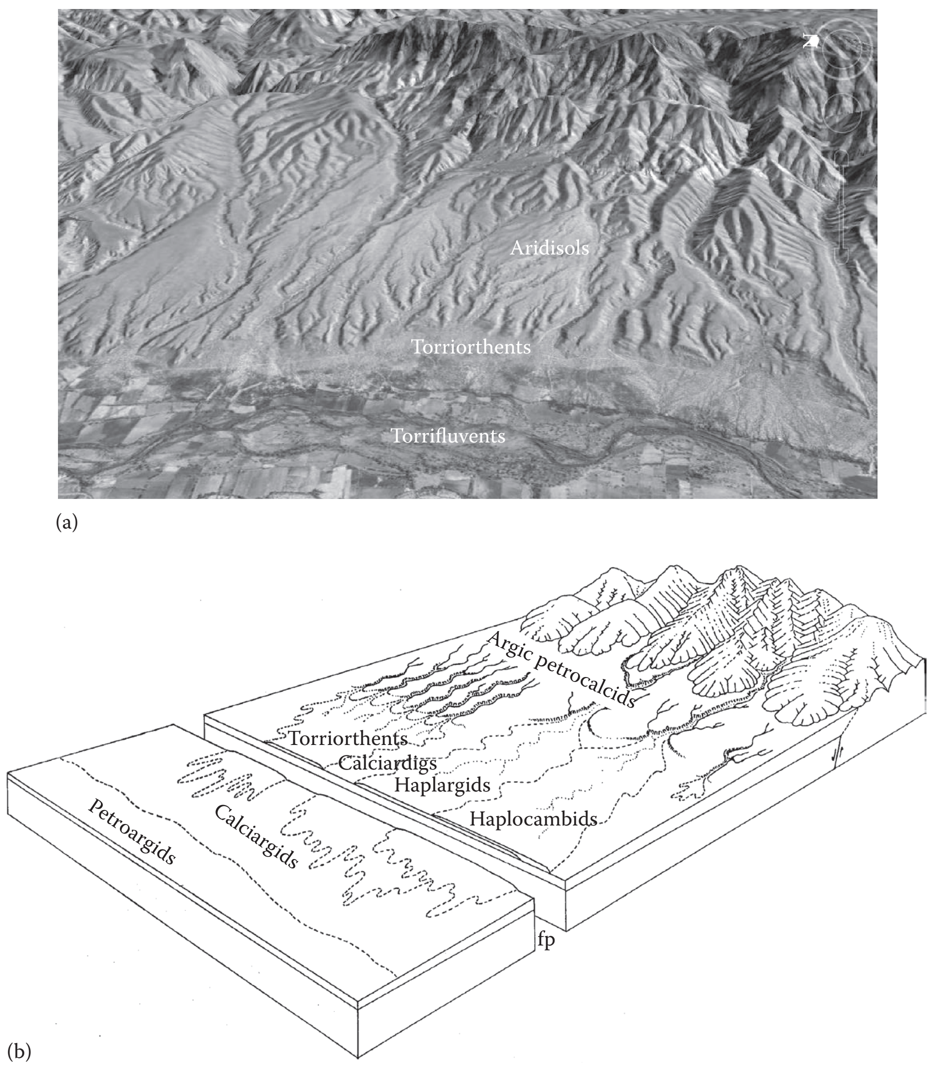


FIGURE 33.71 (a) Aridisols and Entisols (Torriorthents) on stepped geomorphic surfaces bordering a river flood plain (Torrifluvents) in Sonora, Mexico. The stepped surfaces record climate-controlled entrenchment and partial backfilling. Older land surfaces are topographically higher and can persist for hundreds of thousands of years, especially if armored by petrocalcic horizons. (b) Landscape relationships of different types of Aridisols in the Basin and Range Province, western United States. (Modified from Peterson, F.F. 1980. Holocene desert soil formation under sodium salt influence in a playa-margin environment. Quat. Res. 13:172–186.)

(Section 33.10.4.3). In contrast, soluble minerals reflect dry climates. Pedogenic carbonate is particularly valuable for providing evidence of climate change in at least three forms: its presence or absence, its depth, and its carbon and oxygen isotopes signatures.

Pedogenic carbonate is a feature of aridity (Mack, 1992; Guo et al., 2006). Although it can exist in humid climates under certain circumstances of high water tables and calcareous parent materials (Sobecki and Wilding, 1983), an analysis of 1168 soil profiles with carbonate showed that 95% existed in climates where the annual precipitation is less than 760 mm (Royer, 1999). A typical value of 500 mm (20 in.) has been used as the general boundary between soils with carbonate and soils without carbonate (Birkeland, 1999). The inference, therefore, can be made that paleosols containing pedogenic carbonate formed in dry climates.

Depth of pedogenic carbonate is commonly, but not always, proportional to annual rainfall. In general, greater precipitation corresponds to greater depths carbonate horizons (Jenny and Leonard, 1934; Arkley, 1963; Gile, 1975c). However, erosion, runoff, and run-on can confound this general relationship, as does the seasonal distribution of precipitation. Arid climates with winter rainfall often produce deeper leaching of carbonates than those with a similar amount of summer rainfall (Yaalon, 1983). It has been shown using the USDA-NRCS database that no statistically significant correlation exists between carbonate depth and rainfall, especially for shallow carbonate in arid and semiarid climates (Royer, 1999). Still, within a local region where erosion, runoff, and run-on can be held constant, carbonate depth is likely to be deeper in wetter environments than in drier environments.

Deeper wetting fronts in the Pleistocene are probably responsible for vertical, karst-like pipes that cross-cut petrocalcic and calcic horizons (Figure 33.67d). Similarly, carbonate filaments in B horizons overlying petrocalcic and calcic horizons in soils of Pleistocene age are probably the result of an upward shift in the depth of wetting during subsequent drier climates based on depths of carbonates in soils of Holocene age and radiocarbon dates of the carbonate crystals themselves (Gile et al., 1981; Monger, 2003). Other evidence for climatically driven shifts in carbonate depth includes engulfment of argillic horizons by calcic horizons as the depth of wetting shifts upward with increasing aridity (Nettleton and Peterson, 1983), and micromorphologic evidence of episodic deposition of carbonates, opal, and clay in argillic horizons and duripans (Eghbal and Southard, 1993c).

Simulation models are also useful for understanding links between carbonate depth and climate. Such models have a vertical sequence of compartments, each with a specified texture, bulk density, water-holding capacity, mineralogy, CO_2 content, ionic strength, and temperature (McFadden and Tinsley, 1985). Similarly, models by Marion et al. (1985) and later models by Marion and Schlesinger (1994) and Hirmas et al. (2009) have used stochastic precipitation, evapotranspiration, chemical thermodynamics, soil parameterization, and soil water movement to simulate carbonate distribution. The depth of carbonate for a soil in New Mexico, for example, was modeled as being about 15 cm deeper during wetter Pleistocene climates than its present arid conditions (Marion et al., 1985).

Carbon ($^{13}C/^{12}C$) and oxygen ($^{18}O/^{16}O$) isotopes in pedogenic carbonate contain information about paleoclimate in two ways: paleoecological and paleotemperature. First, $^{13}C/^{12}C$ ratios are linked to the relative abundance of C_4 to C_3 plants (i.e., plants that use either the C_4 or C_3 photosynthetic pathways; Cerling, 1984; Amundson et al., 1988; Ehleringer, 1988; Quade et al., 1989; Monger et al., 2009). Aridisols are associated with both groups: C_4 plants are warm-season desert grasses (although one shrub type is included, *Atriplex* spp.) and C_3 plants are desert shrub species (Sage et al., 1999). Pedogenic carbonate formed in a pure C_3 ecosystem has $\delta^{13}C$ values of around −12‰ to −9‰, in contrast to pedogenic carbonate formed under pure C_4 vegetation which has $\delta^{13}C$ values of around 1‰–3‰ (Cerling, 1999). Aridisols are also associated with a third group—the CAM plants (crassulacean acid metabolism)—which include the cacti and have isotopic values intermediate between C_4 and C_3 plants.

Second, oxygen isotopes have been used to make inferences about paleotemperatures and rainwater sources (Amundson et al., 1996; Liu et al., 1996). With higher mean annual temperatures, rainwater becomes preferentially enriched in ^{18}O, which is preserved in pedogenic $CaC^{18}O_3$ (e.g., Cerling and Quade, 1993). The isotopic signature of rain, however, is complicated because it is influenced by several factors: (1) latitudinal location (isotopically lighter rain occurs poleward); (2) seasonality (winter rain is isotopically lighter and percolates deep in soil); (3) mountain ranges and distance from the coast (isotopically lighter rain at higher altitudes and greater distance from coast); (4) the amount of rain (rain becomes lighter as the amount of rain increases, this includes both monthly rainfall and individual rainfall events); and (5) storm trajectories (source area, including vapor from land, yields different $\delta^{18}O$ values; Grootes, 1993).

33.10.7 Agriculture

Agricultural use of Aridisols is limited chiefly by the lack of irrigation water. Center pivot irrigation, however, has brought vast areas of Aridisols on upland landforms into agricultural production (e.g., Navajo Indian and Saudi Arabia irrigation projects). Where irrigation systems have been developed, most Aridisols present some problems and generally are less well suited for irrigation than are Entisols, such as Torriorthents and Torrifluvents, which are commonly associated with Aridisols (Figure 33.71a). Land leveling to allow flood, furrow, or sprinkler irrigation often exposes duripans, calcic, petrocalcic, natric, or argillic horizons.

Under irrigation, control of salinity in the rooting zone is critical to the long-term success of any crop management system. Flushing of salts by an amount of water in excess of the crop evapotranspiration demand (a leaching fraction) helps reduce salinity in the root zone. The excess water and dissolved salts must be removed by subsurface drainage. The presence of shallow subsurface horizons with slow permeability (duripans, natric or petrocalcic horizons) compounds the irrigation and drainage management problems for crop production, as well as for septic tank leach fields and irrigation of lawns in urbanized areas. Soil subsidence, due to solution of gypsum, and corrosion of concrete are common problems where soils containing gypsum are irrigated (Nettleton et al., 1982; Section 33.10.5.6). Other than N, most major plant nutrients are abundant in Aridisols, although micronutrient availability is often limited by low solubility at alkaline pH. Careful management of P is also often needed due to sorption of P by $CaCO_3$.

33.10.8 Ecosystem Services

The term *ecosystem services* refer to ways humans benefit economically from processes and resources that are supplied by natural ecosystems for free, such as clean drinking water and pollination of crops (Withgott and Brennan, 2009). Aridisols provide many ecosystem services. Three prominent categories include *provisioning*, *regulating*, and *cultural* ecosystem services.

33.10.8.1 Provisioning Ecosystem Services of Aridisols

Grazing by livestock (mainly cattle, sheep, and goats) and wildlife is the common use of Aridisols where water is not available for irrigation. More than half of the Earth's land surface is grazed (Follett et al., 2001); most of which occurs in the dry regions (e.g., Sobecki et al., 2001). In the United States, for example, there are approximately 308 million ha of rangeland, about 31% of the total area (Havstad et al., 2007). Most vegetation on Aridisols, however, is dominated by drought-escaping annual grasses, forbs, and drought-enduring evergreen shrubs (Archibold, 1995). These have some grazing value, but less than the grazing value of Aridisols transitional to xeric or ustic soil moisture regimes, or frigid or cryic soil temperature regimes, that often have a significant component of perennial grasses. Forage on Aridisols is highly variable (Eckert et al., 1978; Schlesinger and Jones, 1984), ranging up to 10-fold differences in forage production (Ludwig, 1987). Still, huge amounts of beef have historically been generated on indigenous vegetation of Aridisols (Fredrickson et al., 1998). Overgrazing on some Aridisols has been accompanied by a drastic change in the structure of vegetation giving rise to geomorphic feedbacks, such as the formation of coppice dunes (Gile, 1966; Otterman and Gornitz, 1983; Schlesinger et al., 1990; Reid et al., 1993). In the past few decades in the United States, food production from the livestock industry using Aridisols has declined because of poor economic returns, increased regulation, an aging rural population, and increasing use of drylands for diverse purposes, such as exurban development (Havstad et al., 2007).

Medicinal plants are another provisioning service associated with Aridisols. Many desert plants, especially evergreen perennials, have evolved chemical defenses that may be prime sources of anticancer compounds (Donaldson and Cates, 2004). For example, extracts from a screen of 63 Sonoran desert plants revealed that 42% of evergreen species, 37% of woody perennials, 36% herbaceous perennials, and 23% of annuals showed greater than 50% inhibition against HeLa cells. Other potential sources of anticancer compounds are found in the rhizosphere of plant roots growing in Aridisols (He et al., 2004) and in plants with genetic variability among the populations unique to Aridisol habitat, such as *Anemopsis californica* (Medina-Holguín et al., 2007, 2008).

33.10.8.2 Regulating Ecosystem Services of Aridisols

Aridisols are important in regulating air quality at both the local and global scales. At the local scale, dust storms can reduce visibility on highways and aggravate respiratory problems in areas where Aridisols have been disturbed by construction or agriculture (e.g., Bar-Ziv and Goldberg, 1974). Globally, Aridisols are important sources of atmospheric dust that not only affect air quality, but also influence rain chemistry, ocean fertilization, and albedo (Schlesinger, 1997; Kaufman et al., 2005). The key variable linking Aridisols and air quality are ground cover in the form of desert pavement, biological soil crusts, or vascular plants. When ground cover is removed or disturbed, soil particles, especially in the very fine sand and silt fractions, are vulnerable to entrainment and movement by wind (e.g., Belnap and Gillette, 1998).

Aridisols play a similarly important role in regulating water quality. While forested mountains are the source of most water in regions with Aridisols, a significant proportion is also generated by lower elevation rangelands during infrequent and often torrential rainfall events (Havstad et al., 2007). Aridisols at lower elevations that are occupied by shrublands generally produce more runoff than those occupied by grasslands (Schlesinger et al., 2000; Wilcox and Thurow, 2006). As overland flow moves across shrublands, it carries more sediment than in grasslands. Moreover, soil moisture is more heterogeneous in shrublands than in grasslands because soil moisture is deeper beneath rills and gullies (Dick-Peddie, 1993). As with air quality, a highly disturbed landscape of Aridisols will give rise to lower quality water than neighboring nondisturbed Aridisols.

Carbon sequestration by soil is an important ecosystem service of global significance. Aridisols, however, have low contents of organic carbon, only 4% of the total, but contain the greatest amount of soil inorganic carbon (SIC, calcium carbonate), 48% of the total (Eswaran et al., 2000). Soil inorganic carbon is greatest in Aridisols because, as discussed earlier (Section 33.10.5.5), its occurrence is mainly limited to dry climates. Of the global 946 Pg of SIC, arid regions contain 77.8%, followed by semiarid regions (14.2%), Mediterranean climates (5.4%), permafrost (1.9%), humid (0.5%), and perhumid areas (0.2%) (Eswaran et al., 2000). Although the global amount of SIC is large, containing more carbon than the amount in terrestrial vegetation (560 Pg) or the atmosphere (820 Pg), the rate at which SIC forms is relatively low: 0.3–15 g $CaCO_3$ m^{-2} $year^{-1}$ (Machette, 1985; Eghbal and Southard, 1993b; Reheis et al., 1995; Monger and Gallegos, 2000). The loss of CO_2 from SIC also appears to be low even when petrocalcic horizons are exposed at the land surface by erosion (Serna-Perez et al., 2006). However, because some SIC is linked to microbial and root precipitation, labile pools of SIC may exist, but presently remain unmeasured.

33.10.8.3 Cultural Ecosystem Services of Aridisols

Early civilizations arose on Torrifluvents and neighboring Aridisols, such as those in Sumeria in the fourth millennium

B.C. when irrigated agriculture in that arid climate encouraged stable settlements, led to surplus food, freed people to pursue specialized trades and develop social order and cultural creativity (Durant, 1935). Similar cultural developments arose in arid and semiarid climates along the Indus River of ancient India, Hoang-Ho (Yellow River) of ancient China, and in dryland climates of the western Hemisphere, such as those occupied by the Inca, Anasazi, and Hohokan cultures. Today, arid climates are well known for their archaeological preservation (e.g., Egypt) as well as other cultural ecosystem services including wildlands, solitude, recreation, esthetic experiences, and places for scientific discovery.

33.10.9 Desertification

Desertification refers to the broad set of environmental degradation processes that results in a persistent decrease in the productivity of drylands—a condition that impoverishes an estimated 1 billion people (Verstraete et al., 2009). Such degradation typically involves the loss of productive grasslands and their replacement by shrublands accompanied by erosion. In addition to erosional degradation, desertification of Aridisols also includes chemodesertification in which improper irrigation causes salt accumulation (Dregne, 1983). Desertification of grazingland Aridisols can result in the spatial distribution of plants becoming very heterogeneous (Eckert et al., 1978; Schlesinger and Jones, 1984), commonly giving rise to islands of fertility (Schlesinger et al., 1990). These islands tend to be relatively stable, resistant to erosion (Otterman and Gornitz, 1983) and often trap eolian very fine and fine sand to form coppice dunes (Gile, 1966; Reid et al., 1993). These coppice dunes can range in height from a few centimeters around the base of shrubs to 1 m or more, in which case they dominate the landscape microtopography.

Desertification involves ecogeomorphic thresholds. When these thresholds are exceeded, the system is slow to recover in large part because the soil resource has changed (Bestelmeyer et al., 2009). Forces that drive desertification are coupled environmental and human systems, such as short-term climate variability and exploitation of scarce resources in areas remote from centers of political power (Reynolds et al., 2007). Recent proposals to combat desertification involve the effort to view desertification as (1) a provider of ecosystem services, (2) a coupled human ecological system, and (3) a system that must be viewed at multiple scales in order to understand its causes and consequences (Verstraete et al., 2009).

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33.11 Alfisols

C.T. Hallmark
D.P. Franzmeier

33.11.1 Introduction

The central concept of Alfisols embraces soils of the semiarid to humid climates with light-colored surfaces and subsoils moderately rich in basic cations, formed, at least in part, by movement of clay from overlying horizons into the subsoil. Most Alfisols developed under deciduous forest vegetation in humid climates and grass or savannah vegetation in drier regimes. In loamy and silty Alfisols of the humid regions, fragipans are not uncommon; in drier climates, incomplete leaching of salts and carbonates can result in concentrations of secondary carbonates, gypsum, amorphous silica, or enrichment of exchangeable Na. As the processes of formation of Alfisols require both time and energy, most are found on geomorphic surfaces that are of Pleistocene age or older. They form on a wide variety of parent materials.

In previous classification systems in the United States, most Alfisols were in the Gray–Brown Podzolic great group, but others were in the Low-Humic Gley, Noncalcic Brown, Reddish-Brown Lateritic, Reddish Brown, Reddish Chestnut, Gray Wooded and Planosol great groups. Further, natric great groups or subgroups of Alfisols also include salt-affected soils that were previously classified as Solonetz and Soloth soils (Soil Survey Staff, 1975).

33.11.2 Distribution

Alfisols occur on all continents except Antarctica (Figure 33.72). They occupy about 9.6% of the ice-free Earth's surface (Table 33.1). They are found between the latitudes of 65°N and 45°S and concentrated in the Northern Hemisphere, specifically North America, Europe, and east Central Asia between latitudes of 30°N and 65°N; also, significant occurrences are found in sub-Saharan Africa, India, South America, and Australia. Cryalfs are found within the colder reaches such as Northern Europe and into Western Russia. Ustalfs are common in sub-Saharan Africa, Eastern Brazil, the eastern portion of India, and Southeastern Australia. Udalfs are prominent in Central Europe and Eastern Australia, while Xeralfs are found in countries bordering the Mediterranean Sea.

Within the United States, Cryalfs are found in the north central states and at higher elevations in the western states. Udalfs are prominent in the eastern portion of the Midwest, along the Mississippi and Ohio River valleys while Ustalfs occur to the south and west of the Great Plains as well as the southern portion of the High Plains. Xeralfs are restricted to the west coast states.

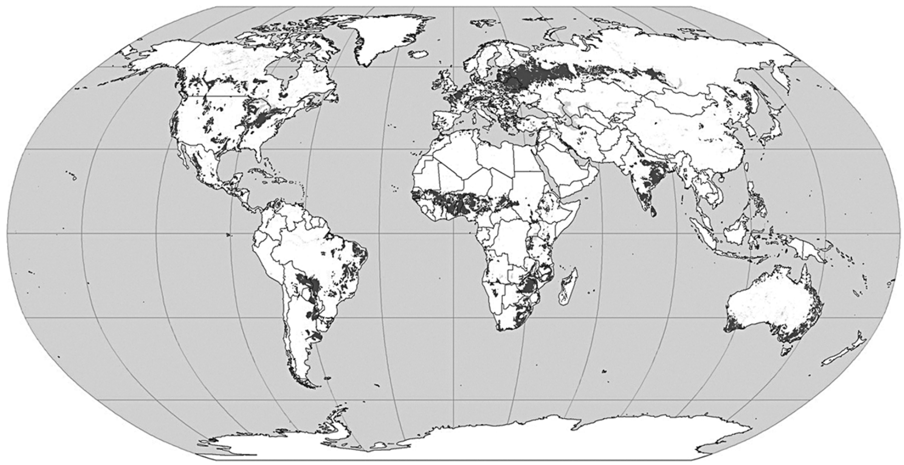

FIGURE 33.72 Global distribution of Alfisols. (Courtesy of USDA-NRCS, Soil Survey Division, World Soil Resources, Washington, DC, 2010.)

33.11.3 Factors of Soil Formation

33.11.3.1 Parent Material

Alfisols form on a wide variety of parent materials, which includes glacial till (Borchardt et al., 1968; Smeck et al., 1968; Smith and Wilding, 1972; Ranney et al., 1975; Franzmeier et al., 1985; Weisenborn and Schaetzl, 2005), loess (Bouma et al., 1968; Grossman and Fehrenbacher, 1971), eolian sands (Davis, 1970; Allen et al., 1972; Miles and Franzmeier, 1981), residuum from limestone (Reynders, 1972), sandstone (Anderson et al., 1975; Stahnke et al., 1983; Chittleborough, 1989; Delgado et al., 1994), siltstones and shales (Gallez et al., 1975; Delgado et al., 1994), acid igneous and metamorphic rock (Goss and Allen, 1968; Ranney and Beatty, 1969; Yerima et al., 1989; Kooistra et al., 1990), mafic igneous and metamorphic rock (Gallez et al., 1975; Bhattacharyya et al., 1993; Juo and Wilding, 1996), lacustrine sediments (Borchardt et al., 1968; Ranney and Beatty, 1969), colluvium derived from acid shales and siltstones (Bailey and Avers, 1971), basalt (Weitkamp et al., 1996), coastal plain sediments (Vepraskas and Wilding, 1983; Juo and Wilding, 1996), and alluvium (Parsons et al., 1968; Ranney et al., 1975; Chittleborough et al., 1984; Klich et al., 1990; Delgado et al., 1994; Shaw et al., 2003; Calero et al., 2008). The literature is replete with examples of Alfisols developing from two or more parent materials, for example, loess over till (Allan and Hole, 1968; Foss and Rust, 1968; Ransom et al., 1987), loess over basalt (Blank and Fosberg, 1991), loess over shale (Ranney et al., 1975), loess over sandstone or limestone (Aide and Marshaus, 2002) and colluvium over residuum (Weitkamp et al., 1996). Uniformity of parent material also continues to be a question in evaluating the influence of lithogenic inherited vs. pedogenic and translocated clay in the argillic horizon (Ruhe, 1984a; Schaetzl, 1998). In eolian sands, the argillic (zone of clay accumulation) horizon usually consists of a sequence of thin bands called lamellae that contain somewhat more clay than the soil material between the bands.

Lotspeich and Coover (1962) stated that "texture of the parent material controls the texture of the soil because texture is a nearly permanent characteristic of soil," particularly in arid and semiarid regions. While this may be true in Alfisols for which the parent materials have undergone previous cycles of weathering and are rich in quartz and resistant minerals (Anderson et al., 1975), many examples are available where weathering gives rise to textures and a pedogenic mineral suite markedly different from the parent material (Paeth et al., 1971; Gallez et al., 1975).

33.11.3.2 Organisms

Most Alfisols formed primarily under hardwood forest vegetation (Soil Survey Staff, 1999), but some suborders and great groups support a significant component of grass vegetation. This is illustrated in Figure 33.73 where soil classes with abbreviated vegetative descriptions from the literature are arranged in a temperature–moisture regime matrix. Aqualfs and Udalfs support forest vegetation, although dominant species and composition of forests change in response to climatic conditions. Ustalfs and Xeralfs support moderate to dominant grassland vegetation, commonly described as savanna. The forest vegetation of moister Cryalfs yields to a greater composition of grasses as Cryalfs occupy dryer climates. It is hazardous to assume that present or presettlement vegetation (Figure 33.73) represents the native vegetation of the soil as both evolve together with climate changes over time modifying vegetative communities (Wells, 1970). Even within broad vegetative zones, individual species may influence soil properties over relatively short distances. Gersper and Holowaychuk (1970a, 1970b) showed precipitation stemflow down the trunk of American beech (*Fagus grandifolia*) trees thickened E (A2) horizons, increased low chroma mottling of B horizons, and decreased clay content, pH, exchangeable Ca, Mg, and K, CEC, base saturation, and free Fe oxides in the B horizons near the stem. Davis et al. (1995) believe available P that was correlated with microelevational changes on a cultivated Psammentic Paleustalf in Niger is related to location of bushes and trees in the savanna or to cultivation.

Fauna dwelling in Alfisols has also been credited with affecting soil formation. Nielsen and Hole (1964) found earthworm population masses of 2,218 kg ha^{-1} under a deciduous forest

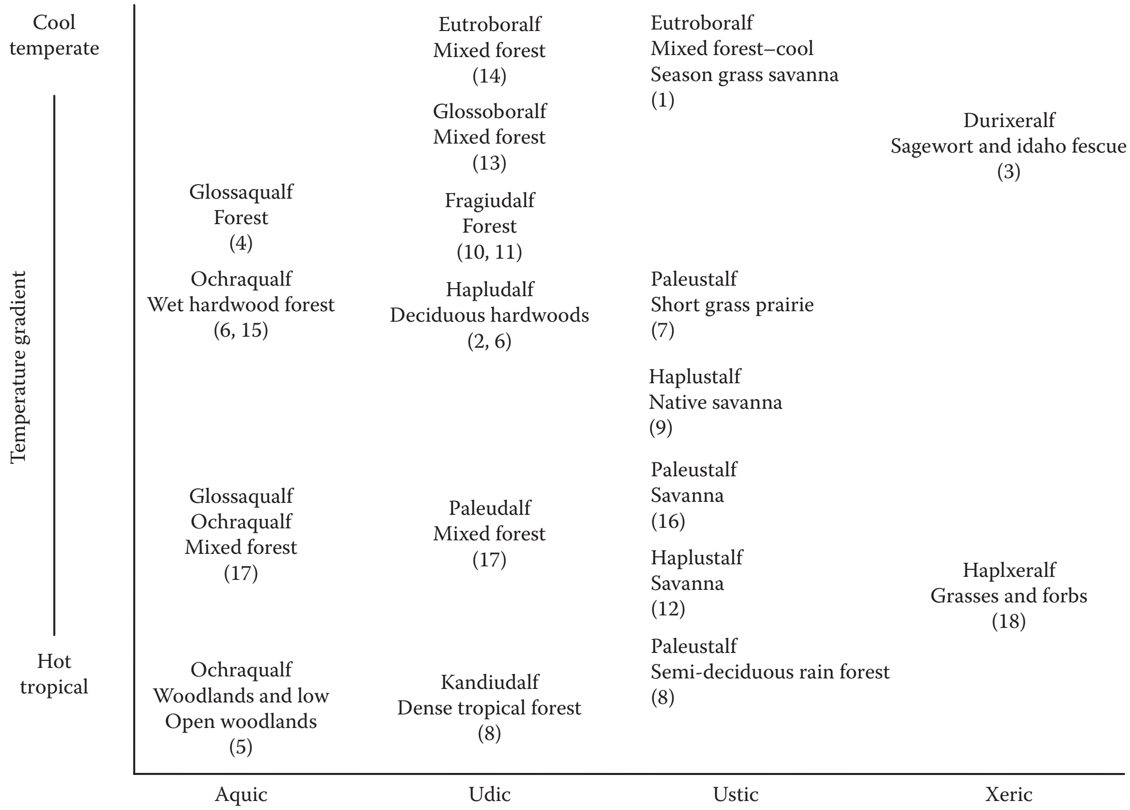

FIGURE 33.73 Generalized native vegetation for selected Alfisols as a function of temperature and moisture regime. Sources are as follows: (1) Anderson et al. (1975); (2) Bailey and Avers (1971); (3) Blank and Fosberg (1991); (4) Bouma and van Schuylenborgh (1969); (5) Coventry and Williams (1984); (6) Cremeens and Mokma (1986); (7) Davis (1970); (8) Gallez et al. (1975); (9) Goss and Allen (1968); (10) Lozet and Herbillon (1971); (11) Miller et al. (1971); (12) Moberg and Esu (1991); (13) Ranney and Beatty (1969); (14) Schaetzl (1996); (15) Smith and Wilding (1972); (16) Stahnke et al. (1983); (17) Vepraskas and Wilding (1983); (18) Weitkamp et al. (1996).

and earthworm middens numbering 274,000 ha^{-1}. Essentially, all annual leaf fall was moved to the middens by earthworms. Wiecek and Messenger (1972) described calcite spheroids in acid A horizons of Hapludalfs under forest vegetation that are produced in earthworm calciferous glands and were responsible for close-range variability in soil pH of A horizons. Hugie and Passey (1963) describe burrowing activity of western cicadas, crediting them with development of cylindrical structure as they burrowed through loamy soils. Their numbers were greatly reduced on coarse- and fine-textured materials. Ant, termite, and crustacean activity have also been described (Thorp, 1949). Crayfish activity in Ochraqualfs (formally) of the Texas coast has been so intense that up to 75% of the subsoil volume comprise their krotovinas prompting the establishment of a new great group (Vermaqualfs) to accommodate these and similar soils highly modified by fauna (Vepraskas and Wilding, 1983). Arkley and Brown (1954) observed sufficient rodent activity to suggest that mima mounds in western states are the result of pocket gophers.

33.11.3.3 Climate

In the process of clay eluviation, percolating water has mobilized silicate clays in upper horizons, carried them downward in the pedon with final deposition in subsoil horizons. This occurs most readily in climates in which precipitation exceeds evapotranspiration for at least several months each year. Consequently, Alfisols occur in semiarid to humid climates. In temperate and cool climates, Alfisols tend to form a belt between the Mollisols of the grasslands and the Spodosols and Inceptisols of very humid climates (Soil Survey Staff, 1999). In warmer climates, they often lie between the Aridisols of arid regions and the Inceptisols, Ultisols, and Oxisols of warm, humid climates.

Within the Alfisol areas, several soil properties vary with climate. Early works of Jenny (1935) and Jenny and Leonard (1934) serve to broadly illustrate soil changes over long transects where gradients of temperature and precipitation exist. When crystalline rocks of similar composition and geomorphic age weather across a temperature gradient, soil clay content in the upper solum increases with mean annual temperature. Along the 11°C annual isotherm from eastern Colorado to western Missouri, they found that with increased annual precipitation, depth to free carbonates, N and clay contents increased, and pH decreased. Partitioning climatic influence over long distances is difficult at best. Climate implies both temporal and spatial fluxes in a number of parameters including temperature, precipitation, radiation, humidity, and wind speed and direction. Locally, relief, slope, and aspect can provide a microclimate somewhat dissimilar to the broader regional climate. Cooper (1960) illustrated the role of slope aspect on microclimate in Michigan Udalfs where he found south-facing slopes had higher light intensities, maximum air temperature, evaporation rate, and soil temperature and lower soil moisture than north-facing slopes. The B horizons of soils on the southern aspect were redder and finer in texture, and sola were thinner than those of the northern exposure. Torrent and Nettleton (1979) proposed a textural index (based upon fine silt:total silt ratios) to assess chemical weathering. Using soils, primarily Udalfs and Aqualfs developed from loess along the Mississippi Valley from Minnesota to Mississippi, they related the textural index describing weathering of silts to clays with increasing mean annual temperature. Subsequently, Ruhe (1984a), utilizing a smaller data set of Hapludalfs along a similar Minnesota to Mississippi transect, interpreted the trends as sedimentation dependent. In a companion study on the same Hapludalfs, Ruhe (1984b) related the more intensive and deeper leaching of bases in the southern sector to both historic and paleoclimate differences in annual and effective precipitation.

33.11.3.4 Relief

Alfisols occur mostly on hillslopes that are of moderate gradient and are linear or convex in cross section. On steeper slopes, clay translocation is limited because the soil is very shallow or stony or because material moves downslope, resulting in Inceptisols or Entisols. In depressional areas, downward water movement might be limited, or soil organic matter (SOM) content of upper horizons is high enough for the soils to be Mollisols.

Relief is responsible for the distribution of Aqualfs in many settings where water moves to lower topographic positions either overland or by throughflow. In these lower positions, epiaquic (perched water table) or endoaquic conditions may occur. Cremeens and Mokma (1986) described soil hydrosequences in Michigan where well- and moderately well-drained Hapludalfs yielded to somewhat poorly drained Ochraqualfs (later reclassified to Endoaqualfs) and poorly drained Haplaquepts and Argiaquolls. Argillans (clay films) and parameters related to movement of clay from eluvial to illuvial zones were greatest in the better drained soils. Working with seasonally saturated Alfisols in a toposequence on the Texas Coastal Plain, Vepraskas and Wilding (1983) found soil pH generally increased as did base saturation and exchangeable Na levels from upper to lower topographic positions. Also, smectite was the predominant clay mineral in the upper sola of the soils in the lower positions while kaolinite was dominant in higher positions. Cremeens and Mokma (1986) also found smectite in the wettest sites of their study with its absence in the better drained soils. Goss and Allen (1968) found appreciable accumulation of exchangeable Na downslope in a Natrustalf and evidence of accumulation of smectite due to relief.

33.11.3.5 Time

Alfisols in areas with moderate and cool temperatures are mostly on late Pleistocene surfaces as most of these surfaces were affected by glaciation. In warmer areas, they may be on older surfaces. Usually, there has not been enough clay translocation for the soils to be classified as Alfisols on surfaces younger than late Pleistocene. On old surfaces, leaching of bases commonly has progressed to the point where the soils are classified as Ultisols instead of Alfisols.

The length of time for development of soil and specific soil properties has long been of scientific interest (Jenny, 1941). In recent years, interest in soil age has been increasing as evidenced by the growing volume of literature (Bilzi and Ciolkosz, 1977; Bockheim, 1980; Little and Ward, 1981; Meixner and Singer, 1981; Harden, 1982; Muhs, 1982; Chittleborough et al., 1984; Catt, 1991; Markewich and Pavich, 1991; Delgado et al., 1994; Shaw et al., 2003; Calero et al., 2008). Part of this renewed interest is due to application of soil interpretation to subjects of current interest such as climate change, C sequestration rates, landscape stability, and archaeology. Improved and new methodologies to directly determine soil age such as radiocarbon assay, luminescence, amino acid analysis, paleomagnetic measurement, and U–Th disequilibrium series have augmented dendrochronology and recorded historic events (mudflows, volcanic eruptions, etc.) and indirect methods of correlation of stratigraphic units. As soil (and landscape) age increases, so does the concern for significant changes in other factors of state, particularly climate and its effect on vegetation. The time required for the development of Alfisols is incumbent upon the formation of argillic and kandic horizons. Bilzi and Ciolkosz (1977) believe that 2,000–12,000 years are required to develop fragipan-like features and argillic horizons in alluvium in Pennsylvania under udic soil moisture and mesic soil temperature regimes. In northeast Iowa on earthen mounds of varying age built by Indians, Parsons et al. (1962) found that soil genesis had taken place with A horizon properties remaining relatively constant with increasing age beyond 1000 years, E horizon platy structure developing within 1000 years (maximum expression required >2500 years) and the blocky structure plus clay translocation sufficient to meet argillic horizon definitions in the B horizon occurring between 1000 and 2500 year BP. In comparison, the nearby Fayette soil (Typic Hapludalf) had developed over a 14,000 year period. In preconditioned material used by Indians to construct mounds and ridges in Louisiana, argillic horizons in Ultisols and Alfisols had developed in 5000–5400 years (Saunders et al., 1997). Shaw et al. (2003) found Alfisols on stream terrace surfaces dating to 5500 year BP in Alabama.

In the Midwest, most Alfisols are found on surfaces affected by late Wisconsinian glaciation (12,000–25,000 year BP). In the Upper Midwest, many Alfisols formed in late Wisconsinian till deposited by glacier ice, and in outwash and lacustrine deposits from the glacier when it melted. Also, much of the Midwest and the lower Mississippi Valley was blanketed with late Wisconsinian age loess that was calcareous when it was deposited. Soils formed in it are relatively high in basic cations. This loess also covered surfaces on which soils had formed previously and had been eroded. These buried soils (paleosols) were enriched with basic cations (especially Ca and Mg) that were leached from the loess and translocated downward.

TABLE 33.49 Ages of Ustalfs and Xeralfs Included in Chronosequence Studies

Soil Class	Age, Years BP	MAP (mm)	Location	Source
Typic Haploxeralf	250,000	310	California	Meixner and Singer (1981)
Typic Haploxeralf	40,000	410		Harden (1982)
Typic Haploxeralf	130,000			
Typic Haploxeralf	600,000			
Typic Natrixeralf	60–107,000	170		Muhs (1982)
Typic Natrixeralf	120–134,000			
Typic Natrixeralf	375–460,000			
Ultic Haplustalf	3,740	420	New South Wales, Australia	Chittleborough et al. (1984)
Rhodic Paleustalf	9,220			
Ultic Paleustalf	26,700			
Ultic Paleustalf	27,040			
Ultic Paleustalf	29,000			
Paleustalf	42,500	—	Victoria, Australia	Little and Ward (1981)
Paleustalf	210,000			
Paleustalf	227,000			
Paleustalf	763,000			

All soils developed from alluvium. Dating methods in the studies include amino acid, U-series, uplift rates, and stratigraphic correlations.

Alfisols of ustic and xeric soil moisture regimes and developed from alluvium have been included in a number of chronosequence studies (Table 33.49). The annual precipitation ranged from 170 to 420 mm. The youngest Alfisol (Ultic Haplustalf) was 3,740 years BP and the oldest (Paleustalf) was 763,000 years BP, while the youngest Paleustalf was 9,220 years BP. Much older Haploxeralfs were noted in the California studies, but in the xeric moisture regime sufficient development of the argillic horizon to permit classification as a Palexeralf had not occurred.

33.11.4 Morphology

The morphology of Alfisols ranges greatly across the broad spectrum of conditions under which they occur. In all Alfisols, however, there is an increase in clay content from the A and E horizons near the surface to the B horizons below. In most Alfisols, this clay increase meets the concept of an argillic horizon where the illuvial Bt (t from ton Gr, clay) horizon shows evidence of oriented clays. In older soils dominated by low activity clays (low CEC of the clay fraction), evidence of oriented clay may be absent but clay increase is sufficient for a kandic horizon.

Where some of the clay in the Bt horizon is oriented, that is, stacked like a pile of papers or playing cards, the oriented clay may be on ped surfaces where it forms clay films (thin layers of oriented clay) that often cover and may be a different color from sand grains in the ped (see Chapter 30 for electron micrograph of a clay film). In sandy soils, the oriented clay forms bridges between adjacent sand grains and may line root channels. Often, this clay mentioned above actually contains a fair amount of silt. In many Alfisols, the transitional EB or BE horizons have silt coatings instead of clay on the outside of peds. In these, and some other Alfisols, especially those formed in calcareous parent materials, the best developed clay films are in the lower Bt horizon. Apparently, clay is stripped from the ped coatings in the upper B or BE horizon and is deposited on ped surfaces deeper in the profile, illustrating that clay translocation is a dynamic process, and that the argillic horizon is moving down in the profile.

Before a soil can be classified in *Soil Taxonomy*, germane diagnostic horizons and soil properties (defined in Chapter 31) must be identified. Several are essential or common in Alfisols, and their occurrence in the various suborders of Alfisols is shown in Table 33.50.

TABLE 33.50 Occurrence of Soil Horizons or Properties in Soils of the Suborders of Alfisols

	Soil Horizon or Property									
Suborder	Argillic Horizon	Natric Horizon	Kandic Horizon	Glossic Horizon	Fragipan	Duripan	Plinthite	Very Red Argillic	Very Deep Argillic	Other
Aqualfs	X	X	X	X	X	X	X			Biotic activity, very cold[a]
Cryalfs	X			X					X	
Udalfs	X	X	X	X	X			X	X	Fe nodules
Ustalfs	X	X	X			X	X	X	X	Calcic/Petrocalcic
Xeralfs	X	X			X	X	X	X	X	Calcic/Petrocalcic

[a] Also: abrupt clay increase at upper boundary of the argillic horizon, and perched water table.

33.11.5 Suborders: Their Morphology, Distribution, and Use

The Alfisol order is subdivided into five suborders, one based on temperature and four based on soil moisture regime. Cryalfs are the Alfisols in cold climates. Of the four suborders based on soil moisture regime, three (Udalfs, Ustalfs, and Xeralfs) follow broad geographical climatic belts. The distribution of Aqualfs depends on local landscape and parent material factors. A complete listing of the great groups and subgroups in each suborder appears in Table 33.51.

33.11.5.1 Aqualfs

These are the wet Alfisols in which the water table is at or near the surface (50 cm) for at least several weeks each year, but may be very deep during the dry season. The water table may be continuous to the aquifer, or it may be held up by a layer with low permeability in or just beneath the soil surface. Their location in the landscape is controlled mainly by local conditions of parent material and relief. They occur as small areas in larger areas of Cryalfs, Udalfs, Ustalfs, Xeralfs or other soil suborders, but they are more common in zones of Udalfs than Ustalfs. They may have several kinds of diagnostic horizons (Table 33.50). Nearly all Aqualfs had forest vegetation at some time in the past (Soil Survey Staff, 1999). Most are cultivated after they were drained by underground tiles or ditches. Rice is a common crop in warmer regions.

Aqualfs display quantifiable redoximorphic features (gray colors) immediately below a depth of 25 cm from the mineral soil surface. In the absence of quantifiable redoximorphic features, a positive reaction to active Fe^{2+} (α,α-dipyridyl) can be used. In this process (gleying), soil horizons are saturated with water, microbes deplete available O_2, causing Fe(III) in brownish and reddish Fe oxide minerals to be reduced. The resulting Fe^{2+} is released into the soil solution, much of which is leached from the soil. The result is that the horizon, or much of it, takes on the gray color of the silicate minerals. Gleyed horizons are designated with a g as Btg.

Gleying in Aqualfs may be noted in the upper 25 cm of the soil. Gleyed soils are generally found on lower topographic positions receiving additional water, or on broad, level interfluves where surface drainage may be very slow. The A horizon, commonly overlain by O horizon material in varying stages of decomposition, is usually either too thin or too low in organic C to meet requirements of a mollic epipedon. A few may qualify as an umbric or mollic epipedon, but have insufficient bases in the subsoil to be classified as Mollisols. Most Aqualfs have well-developed E horizons, although the underlying Bt and Btg horizons often show only minimally expressed clay films. Boundaries between E and Bt and Btg horizons often show evidence of mixing by fauna (i.e., crayfish) or tongues and stringers of E horizon material between structural units of Bt and Btg material (glossic material).

33.11.5.2 Cryalfs

Cryalfs are restricted to frigid and cryic temperature regimes and generally occur in more mountainous and sloping landscapes that are generally well drained. They formed in North America, eastern Europe, and Asia above 49°N latitude and in some high mountains of lower latitude where they occur at lower altitudes than Spodosols or Inceptisols. Most have been under coniferous forest because of the cool, short growing season. Commonly, parent materials are residuum or colluvium from local bedrock. Most are not cultivated, and thus, have thin to moderately thick (2–10 cm) O horizons in varying stages of organic matter decomposition. Many lack A horizons, but when present, they are thin (5 cm) with granular or subangular blocky structure. Most have E horizons of platy or subangular blocky structure that are relatively thick (15–30 cm). The Bt horizons range from thin to thick (15–135 cm), display a wide range of clay film features, and commonly have angular or subangular blocky structure. Many have lithic or paralithic contacts, and most contain significant quantities of coarse fragments throughout the solum.

33.11.5.3 Udalfs

These are the freely drained Alfisols of humid regions, extensive in the United States and Europe. They formed under forest vegetation at some time during their development. The forests were mainly deciduous, but in colder regions, may have been mixed coniferous and deciduous. Many Udalfs have been cleared and intensively farmed; in some that have been severely eroded, the Bt horizon is immediately below or may be incorporated into the Ap horizon (Soil Survey Staff, 1999). Many have fragipans, and some have natric and other diagnostic horizons (Table 33.50). Corn and soybeans are common crops on Udalfs.

In well-drained, temperate humid environs under forest, uncultivated Udalfs have thin (2–5 cm) O horizons with organic matter in various stages of decomposition. Most also have a thin (10–15 cm) A horizon commonly of granular structure darkened by organic matter and underlain by a thin (10–15 cm), light-colored E horizon with platy structure, a zone of maximum eluviation of silicate clays. Because many Alfisols are cultivated, the A and E horizons have been mixed by plowing producing a light color. The intense eluviation in E horizons results in a concentration of sand and silt, and in extreme cases, all or most coloring compounds have been removed leaving skeletal grains whose color dominates the horizon. Often there is a transitional EB or BE horizon between the E and Bt horizons. As compared with the thinner A and E horizons, the Bt horizon is thick, possesses subangular or angular blocky structure, and contains significantly more silicate clay. At least a portion of the silicate clay occurs as clay films on structural or pore surfaces. In the Midwest where most of the parent material is calcareous and associated with the last glaciation, complete leaching of carbonates from the solum has occurred while C horizons remain calcareous.

33.11.5.4 Ustalfs

These are the Alfisols of subhumid to semiarid regions. Moisture moves through these soils to deeper layers only in occasional years (Soil Survey Staff, 1999). If there are carbonates in the parent material or in dust, the soil may have a carbonate-enriched horizon (calcic) below the argillic. Original vegetation may have been xerophytic trees, savanna, or grassland. Dryland crops for

TABLE 33.51 Listing of Suborders, Great Groups, and Subgroups in the Alfisols Order (Ref. Soil Survey Staff, 2010)

Suborder	Great Group	Subgroup
Aqualfs	Cryaqualfs	Typic
	Plinthaqualfs	Typic
	Duraqualfs	Typic
	Natraqualfs	Vertic, Vermic, Albic Glossic, Albic, Glossic, Mollic, Typic
	Fragiaqualfs	Vermic, Aeric, Plinthic, Humic, Typic
	Kandiaqualfs	Arenic, Grossarenic, Plinthic, Aeric Umbric, Aeric, Umbric, Typic
	Vermaqualfs	Natric, Typic
	Albaqualfs	Arenic, Aeric Vertic, Chromic Vertic, Vertic, Udollic, Aeric, Aquandic, Mollic, Umbric, Typic
	Glossaqualfs	Histic, Arenic, Aeric Fragic, Fragic, Aeric, Mollic, Typic
	Epiaqualfs	Aeric Chromic Vertic, Aeric Vertic, Chromic Vertic, Vertic, Aquandic, Aeric Fragic, Fragic, Arenic, Grossarenic, Aeric Umbric, Udollic, Aeric, Mollic, Umbric, Typic
	Endoaqualfs	Aquandic, Chromic Vertic, Vertic, Aeric Fragic, Fragic, Arenic, Grossarenic, Udollic, Aeric Umbric, Aeric, Mollic, Umbric, Typic
Cryalfs	Palecryalfs	Andic, Vitrandic, Aquic, Oxyaquic, Xeric, Ustic, Mollic, Umbric, Typic
	Glossocryalfs	Lithic, Vertic, Andic, Vitrandic, Aquic, Oxyaquic, Fragic, Xerollic, Umbric Xeric, Ustollic, Xeric, Ustic, Mollic, Umbric, Eutric, Typic
	Haplocryalfs	Lithic, Vertic, Andic, Vitrandic, Aquic, Oxyaquic, Lamellic, Psammentic, Inceptic, Xerollic, Umbric Xeric, Ustollic, Xeric, Ustic, Mollic, Umbric, Eutric, Typic
Ustalfs	Durustalfs	Typic
	Plinthustalfs	Typic
	Natrustalfs	Salidic, Leptic Torrertic, Torrertic, Aquertic, Aridic Leptic, Vertic, Aquic Arenic, Aquic, Arenic, Petrocalcic, Leptic, Haplargidic, Aridic Glossic, Aridic, Mollic, Typic
	Kandiustalfs	Grossarenic, Aquic Arenic, Plinthic, Aquic, Arenic Aridic, Arenic, Aridic, Udic, Rhodic, Typic
	Kanhaplustalfs	Lithic, Aquic, Aridic, Udic, Rhodic, Typic
	Paleustalfs	Aquertic, Oxyaquic Vertic, Udertic, Vertic, Aquic Arenic, Aquic, Oxyaquic, Lamellic, Psammentic, Arenic Aridic, Grossarenic, Arenic, Plinthic, Petrocalcic, Calcidic, Aridic, Kandic, Rhodic, Ultic, Udic, Typic
	Rhodustalfs	Lithic, Kanhaplic, Udic, Typic
	Haplustalfs	Lithic, Aquertic, Oxyaquic Vertic, Torrertic, Udertic, Vertic, Aquic Arenic, Aquultic, Aquic, Oxyaquic, Vitrandic, Lamellic, Psammentic, Arenic Aridic, Arenic, Calcidic, Aridic, Kanhaplic, Inceptic, Calcic Udic, Ultic, Calcic, Udic, Typic
Xeralfs	Durixeralfs	Natric, Vertic, Aquic, Abruptic Haplic, Abruptic, Haplic, Typic
	Natrixeralfs	Vertic, Aquic, Typic
	Fragixeralfs	Andic, Vitrandic, Mollic, Aquic, Inceptic, Typic
	Plinthoxeralfs	Typic
	Rhodoxeralfs	Lithic, Vertic, Petrocalcic, Calcic, Inceptic, Typic
	Palexeralfs	Vertic, Aquandic, Andic, Vitrandic, Fragiaquic, Aquic, Petrocalcic, Lamellic, Psammentic, Arenic, Natric, Fragic, Calcic, Plinthic, Ultic, Haplic, Mollic, Typic
	Haploxeralfs	Lithic Mollic, Lithic Ruptic-Inceptic, Lithic, Vertic, Aquandic, Andic, Vitrandic, Fragiaquic, Aquultic, Aquic, Natric, Fragic, Lamellic, Psammentic, Plinthic, Calcic, Inceptic, Ultic, Mollic, Typic
Udalfs	Natrudalfs	Vertic, Glossaquic, Aquic, Typic
	Ferrudalfs	Aquic, Typic
	Fraglossudalfs	Andic, Vitrandic, Aquic, Oxyaquic, Typic
	Fragiudalfs	Andic, Vitrandic, Aquic, Oxyaquic, Typic
	Kandiudalfs	Plinthaquic, Aquic, Oxyaquic, Arenic Plinthic, Grossarenic Plinthic, Arenic, Grossarenic, Plinthic, Rhodic, Mollic, Typic
	Kanhapludalfs	Lithic, Aquic, Oxyaquic, Rhodic, Typic
	Paleudalfs	Vertic, Andic, Vitrandic, Anthraquic, Fragiaquic, Plinthaquic, Glossaquic, Albaquic, Aquic, Oxyaquic, Fragic, Arenic Plinthic, Grossarenic Plinthic, Lamellic, Psammentic, Arenic, Grossarenic, Plinthic, Glossic, Rhodic, Mollic, Typic
	Rhodudalfs	Typic
	Glossudalfs	Aquertic, Oxyaquic Vertic, Vertic, Aquandic, Andic, Vitrandic, Fragiaquic, Aquic Arenic, Aquic, Arenic Oxyaquic, Oxyaquic, Fragic, Arenic, Haplic, Typic
	Hapludalfs	Lithic, Aquertic Chromic, Aquertic, Oxyaquic Vertic, Chromic Vertic

Ustalfs include drought tolerant selections such as sorghum, wheat, cotton, and millet. Ustalfs, which are common in the southern Great Plains, do not have fragipans, but may have several other kinds of diagnostic horizons Table 33.50.

In the United States, Ustalfs tend to be on older landforms than Udalfs and have thicker sola; many are in the Paleustalf great group because of either an abrupt textural change between E and Bt horizons or a thick Bt horizon enriched in clay throughout. There is less tendency for E horizons to display platy structure, while with less rainfall, E horizons are often absent. The Bt horizons often have prismatic parting to angular or subangular blocky structure. Secondary carbonates commonly accumulate to form calcic horizons in the lower portion of the argillic horizon or just below it in the dryer Ustalfs.

33.11.5.5 Xeralfs

These occur in regions that have a Mediterranean climate, where precipitation occurs during the cool season and the summers are hot and dry. In some winters, water moves through the entire soil profile to deeper layers. Most Xeralfs border the Mediterranean Sea or lie to the east of an ocean, such as those in the western United States and Australia. Where there is no irrigation, grains are common crops but with irrigation, a variety of crops can be grown. Grapes and olives are grown in the warmer areas (Soil Survey Staff, 1999). Occurrence of E horizons is uncommon, and A horizons with weak subangular blocky structure to massive condition tend to range from 15 to 25 cm thick. Most argillic horizons are relatively thin (15–75 cm) with significant quantities of coarse fragments throughout the pedon.

33.11.6 Soil Formation Processes

33.11.6.1 Clay Translocation

Alfisols and Ultisols form mainly by eluviation of silicate clay from A and E to Bt horizons. In Alfisols, however, the base saturation of deep subsoil layers is greater than it is in Ultisols. Alfisols are similar to the Gray–Brown Podzolic soils of previous soil classification systems. The name Alfisol was derived from the term Pedalfer, a coined word used in previous classifications derived from ped, as in pedology (the science of natural soils), Al, aluminum, and Fe, iron. The implication is that in Pedalfers, Al and Fe accumulate in the subsoil, in contrast to Pedocals, in which Ca accumulates.

The concept of Gray–Brown Podzolic soils originated at a time when soil colloids (clays) were thought to be mainly amorphous. Then, soil scientists determined total chemical composition of the colloidal fraction and interpreted the results by examining SiO_2/Al_2O_3 and $SiO_2/(Fe_2O_3 + Al_2O_3)$ ratios of the various horizons. These ratios showed that both Podzols (Spodosols) and Gray–Brown Podzolic soils had B horizons enriched in Fe and Al, but that Podzol B horizons were also enriched in organic C. This led Marbut (1935) to conclude that the differences in the processes leading to the two kinds of soil were of degree, not kind, and that Gray–Brown Podzolic soils were less developed than Podzols. The word Podzolic is a descriptor previously used for many Alfisols.

Hendricks and Fry (1930) reported two key discoveries that laid the groundwork for the current understanding of the processes by which Alfisols form. First, they showed that soil colloids are mainly crystalline silicate minerals, and that when clay suspensions are allowed to dry, the plate-shaped minerals lay flat to form masses that have a regular crystallographic, or at least optical, arrangement and are large enough to study with a polarizing microscope. This discovery led to an understanding that in Alfisols, the material translocated is mainly silicate clay minerals (with Fe oxides on their surfaces), but in Spodosols, it is mainly amorphous complexes of organic matter with Al and usually Fe. Second, they provided the basis for interpreting what are now called clay films.

These ideas were gradually applied to studies of soil genesis. Jenny and Smith (1935) produced early stages of argillic horizon formation in columns of pure quartz sand by first coating the sand grains with electropositive iron hydroxide and then passing dispersed clay through the column. Brown and Thorp (1942) noted that the B horizons in the Miami soil (Hapludalf) from Indiana and similar soils contained much more clay than did the A and E horizons. They concluded, however, that the clay increase in the B horizons was only partly due to illuviation. Although the types of clay were identified by x-ray diffraction and thermal methods, soil formation processes were still discussed mainly in terms of $SiO_2/(Fe_2O_3 + Al_2O_3)$ ratios. Frei and Cline (1949) concluded that the apparent loss of clay from A and E horizons and the apparent gain in clay in B horizons of a New York soil could be accounted for in at least three possible ways: (1) clay migrates downward as a sol in percolating water; (2) clay is synthesized in the B horizon from soluble weathering products that were released there or moved down in the profile; or (3) the apparent increase in clay was really due to loss of other materials, such as $CaCO_3$. They showed, too, that there was a strong concentration of clays with a high degree of optical continuity on peds in the B horizon of a Gray–Brown Podzolic soil, but in the upper B horizon, these coatings were highly degraded. According to their interpretation, clay that was mobilized in A and E horizons and stripped from ped surfaces in the upper B horizon, moved down the profile in suspension, and was deposited on ped surfaces as films of oriented clay particles. Thorp et al. (1959), in another study of the Miami soil of Indiana, showed that the small particle size of clays, their association with organic compounds, and shrink/swell cycles in upper horizons all enhanced their mobility.

In summary, the movement of clay from A and E horizons to Bt horizons can be viewed as the net result of three subprocesses: dispersion, transport, and deposition (Jenny, 1980). Dispersion, the release of individual clay particles from aggregates, is favored by the replacement of exchangeable Ca^{2+} with Na^+, by adsorption of humus molecules on clay particles, and by mechanical disruption of aggregates, which may be caused by wetting and drying, freezing and thawing, or by mechanical disruption by faunal activity. Small particles move more readily than large particles, so that illuvial clay, as in clay films, contains more fine than immobile clay, such as that in ped interiors. Little clay is dispersed in calcareous soil materials because Ca^{2+} ions on exchange sites favor clay flocculation. Clay that is mobilized in A and E horizons moves as a percolating sol or as slurry creep along ped surfaces. Deposition is favored when Na^+ is replaced by Ca^{2+}, resulting in clay flocculation by an increase in pH, or by reaction of the silicate clay with oxides such as Fe oxide. When the water of a slurry enters into a ped, the clay it contains is deposited on the ped surface, like passing the slurry through filter paper. In some cases, the slurry moves down to the wetting front in the subsoil, and clay is deposited as water evaporates or is taken up by plants. Texture discontinuities may also arrest the percolation of suspensions.

Because many Alfisols formed in calcareous materials, $CaCO_3$ must be dissolved and the weathering products leached from the soil before much clay can move. Usually H^+ replaces a portion of the Ca^{2+} in these Alfisols, causing leached horizons to become acidic. The shallowest depth at which carbonate minerals are found, as detected by dilute HCl, serves as an index of the intensity and length of soil formation.

It should be stressed that although illuviation of clays is diagnostic for the argillic horizon, much or even most clay in the Alfisol Bt horizons originates from other sources to include inheritance from parent material, clay neoformation, in situ weathering of lithogenic minerals, and dissolution of silt- and sand-sized carbonates (Smeck and Wilding, 1980).

33.11.6.2 Natric Horizons

Natric horizons are argillic horizons high in exchangeable Na^+. A natric horizon in an Alfisol seems to be an oxymoron; if rainfall has been high enough to move sufficient clay for the horizon to qualify as an argillic horizon, Na^+ should have long since been leached from the soil. However, natric horizons do exist in Alfisols indicating that there are circumstances in which they form. Some might be caused by a change in climate where the argillic horizon was formed during a previous wetter period, but rainfall is now less, and the leaching of Na^+ is reduced. Other soils with natric horizons occur in small areas where Na^+ has accumulated. Although the original source of much of the Na is feldspar minerals, Na^+ moves about in the landscape once it is released into solution. Often, soils with natric horizons are low in the landscape because solutions containing Na^+ move downslope either on the soil surface or through subsurface horizons, thus concentrating Na^+ in the low areas. In some soils, Na^+ is retained in the soil because the hydraulic conductivity of B horizons is very low, and leaching of Na^+ is retarded. This seems to be a self-perpetuating process; Na^+ accumulates because permeability is low, while Na^+ promotes dispersion, further slowing permeability. Furthermore, some soils with natric horizons were formerly saline soils. When soluble salts are leached from a soil, Na^+ may remain behind on the exchange sites.

33.11.6.3 Fragipan Formation

Fragipans occur in many Alfisols of the humid temperate regions, especially those of silty and loamy textures. These subsurface horizons restrict the entry of water and roots into the soil matrix (prisms) and often underlie argillic horizons. Commonly, the fragipan has a relatively low content of organic matter and high bulk density relative to horizons above. Most fragipans consist of prisms more than 10 cm across with light-colored coatings on prism faces. These light-colored coatings form a polygonal pattern as shown in Figure 33.74. Material within prisms has brittle failure, a tendency for a clod to rupture suddenly rather than undergo slow deformation when pressure is applied. Fragments of the fragipan slake (fall apart) when submersed in water.

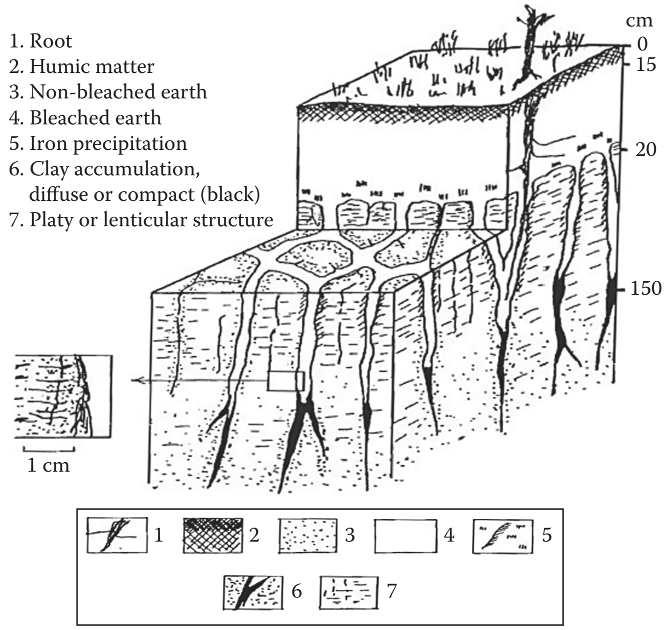

FIGURE 33.74 Schematic of a fragipan in an imperfectly drained silty soil. (Reprinted from van Vliet, B., and R. Langohr. 1981. Correlation between fragipans and permafrost with special reference to silty deposits in Belgium and northern France. Catena 8:137–154. Copyright with permission from Elsevier Science.)

There are many theories on how fragipans form, summarized in two reviews (Smalley and Davin, 1982; Smeck and Ciolkosz, 1989). One reason for the diversity in ideas is that there may be several kinds of horizons included under the name fragipan, each with its own set of processes of formation. Most soil scientists believe that they have formed, or are forming, under present day conditions. The main evidence that fragipans are genetic horizons related to the current land surface is that they are roughly parallel to the soil surface, and the upper boundary has a relatively narrow range of about 50–100 cm depth. Some scientists, however, believe that they formed much earlier. The following are some of the various theories of fragipan formation.

33.11.6.3.1 Paleosols

According to this theory, fragipans formed in earlier times and were buried by younger deposits in which fragipans are not forming. In western Tennessee, Buntley et al. (1977) observed that fragipans are mainly in older loess deposits that were buried by younger Peorian loess in which no fragipans form.

33.11.6.3.2 Rapid Initial Development

Some workers believe that fragipans formed quickly in their life history. Fragipan formation has been related to periglacial conditions (Payton, 1992) and to permafrost (FitzPatrick, 1956; van Vliet and Langohr, 1981). The polygonal pattern formed by coatings on the coarse prismatic structure is similar to that found in periglacial regions. Others present evidence that the structure of loess collapses when it first dries after deposition (Bryant, 1989), and that this denser material becomes the fragipan.

Much glacial till is very dense because it was compacted by the mass of glacier ice. Subsoil horizons that form in dense till may inherit some of their strength from the parent material (Lindbo and Veneman, 1989), but they might also develop fragipan properties through soil-forming processes. In these soils, which are common from eastern Ohio to New England, it is difficult to decide whether the dense material should be considered a fragipan or a slightly modified parent material. The great strength of these horizons is due in part to their high density, but it may also be due to cementation by Si or carbonates (McBurnett and Franzmeier, 1997). The Cd horizon was introduced, in part, to describe little altered, unconsolidated, dense till which exclude roots much like fragipans (Soil Survey Staff, 2006).

33.11.6.3.3 Chemical Cementation

Fragipans could also be cemented by Fe-, Al-, or Si-rich materials. Iron-rich materials cause the cementation of ortstein layers, and silica bonds particles together to form duripans. Although duripans form mainly in aridic and xeric climates, in more humid climates, they grade into fragipans (Soil Survey Staff, 1999). The idea that fragipans are cemented by silica was first proposed in the 1930s and 1940s but was discredited and ignored for about 40 years before revitalized in the late 1970s. The proposal requires certain interactions of parent materials and climate (Franzmeier et al., 1989). In addition to undergoing periodic leaching, soils must also become seasonally dry, which commonly occurs in areas with a udic moisture regime under hardwood forest. Also, downward percolation of soil solutions must be arrested, rather than passing quickly through subsoil layers. This could be caused by the depth of penetration of the wetting front, by discontinuities in parent materials (Smeck et al., 1989), or by slowly permeable, deeper subsoil layers. Silicate minerals weather in upper horizons, and the silica released moves down, usually in the winter and early spring, to the subsoil where it remains for a time. When the trees leaf out, water taken up by roots concentrates silica in solution eventually causing it to precipitate. Compared with grasses, hardwood trees take up relatively little Si, and thus have the potential to concentrate Si in the soil. The precipitated silica bridges between clay particles, or more likely between Fe oxide minerals adsorbed on silicate clay surfaces, causing weak cementation. The greater Si uptake in grasses may be the reason that fragipans do not form under prairie vegetation in humid climates (Franzmeier et al., 1989). The precipitation of Si must be somewhat irreversible, or else any cementation that formed in the summer would be destroyed in the winter, and no net fragipan formation would take place. Fragipans do not form readily on steep slopes, in deep loess, or in loess over more permeable materials, such as outwash. In these soils, the soil solution moves laterally downslope or through the profile, preventing the accumulation of Si-rich soil solutions.

Most of these theories of fragipan formation individually have deficiencies. Perhaps, various combinations of them can explain how a particular soil forms.

33.11.7 Selected Physical, Chemical, and Mineralogical Properties

An increase in clay content between the E and/or A horizon(s) and underlying Bt horizons occurs in all Alfisols; however, this textural contrast ranges from little to great. This is illustrated (Figure 33.75) by the change in clay content with depth for a Psammentic Haploxeralf from California (Torrent et al., 1980), a Typic Hapludalf from Illinois (Grossman and Fehrenbacher, 1971) and a Udertic Paleustalf from Texas (Klich et al., 1990). The Psammentic Haploxeralf has a minimal argillic horizon, with a clay increase of about 4% above that of eluvial horizons. By contrast, the Paleustalf shows a clay increase of more than 40% between the surface horizon and the finest textured portion of the argillic horizon. The Hapludalf is intermediate in textural contrast.

Alfisols with Bt horizons dominated by fine clay (<0.0002 mm, usually smectitic) display significant volume change between wet and dry states. This volume change is reported as the coefficient of linear extensibility (COLE) and can be determined from bulk density values associated with moist and dry, natural clods

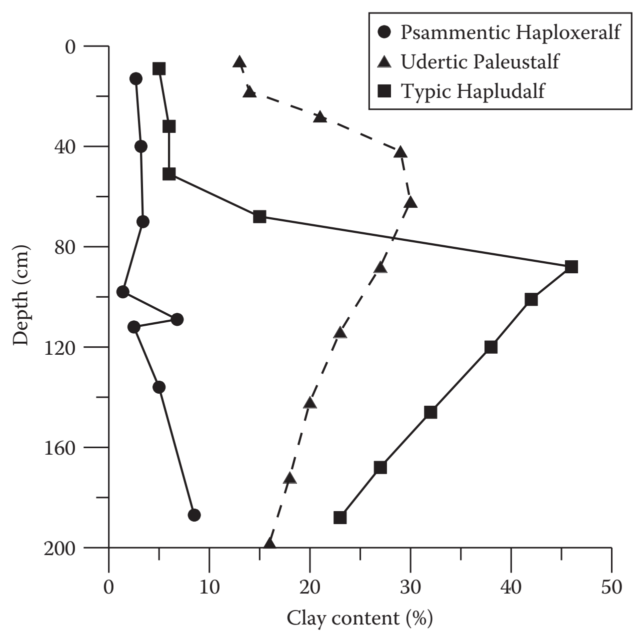

FIGURE 33.75 Clay distribution in three Alfisols of contrasting degrees of argillic horizon development. (From Torrent, J., W.D. Nettleton, and G. Borst. 1980. Clay illuviation and lamella formation in a Psammentic Haploxeralf in southern California. Soil Sci. Soc. Am. J. 44:363–369; Grossman, R.B., and J.B. Fehrenbacher. 1971. Distribution of moved clay in four loess-derived soils that occur in southern Illinois. Soil Sci. Soc. Am. Proc. 35:948–951; Klich, I., L.P. Wilding, and A.A. Pfordresher. 1990. Close-interval spatial variability of udertic paleustalfs in East Central Texas. Soil Sci. Soc. Am. J. 54:489–494.

TABLE 33.52 Mineralogy of the Argillic and Kandic Horizons of Selected Alfisols

			Minerals and/or Silt-Fraction[a]																
			Clay Fraction[a] (<0.002 mm)								Sand and/or Silt-Fraction[b]								
Classification	Location	Parent Material	Ch	Ve	ln	Sm	Mi	Ka	Qz	GH	Ps	Kf	Pl	Mi	Qz	Vg	AP	Ze	Source
Glossoboralfs	Wisconsin	Lacustrine sediments	1	2		2	2	1				12	18	8	55				Ranney and Beatty (1969)
Eutroboralf	New Mexico	Sandstone				2	2	2					18		76	6			Anderson et al. (1975)
Hapludalf	Wisconsin	Thin loess over till	1	2		2	2	2	1			13	14	8	61				Borchardt et al. (1968)
Hapludalf	Oregon	Greenish tuff and breccia	1			3					22		52	3	1	t	6	11	Paeth et al. (1971)
Haploxeralf	Spain	Calcareou salluvium			1	2	3	1				-a-		t	t				Delgado et al. (1994)
Rhodoxeralf		Red sandstone			1	1	3	1				-a-		c					
Haplustalf	Texas	Granite				2	2	1											Goss and Allen (1968)
Natrustalf						2	2	1											
Paleudalf	Texas	Coastal plains sediments				1	1	3+		1									Vepraskas and Wilding
Haplustalf?	India	Schist/quartzite/phyllite				t	2	3	1	1		31			65				Sahu et al. (1990)
Kandiudalf (reclassified)	Nigeria	Olivine basalt						3	t	1									Gallez et al. (1975)
Plinthustalf		Basement complex			t		1	3		1									
Paleustalf	Niger	Eolian sands						3	1										West et al. (1984)

t, trace; Ch, chlorite; Ve, vermiculite; ln, interstratified 2:1 minerals; Sm, smectite; Mi, hydrous mica or illite; Ka, kaolinite; Qz, quartz; GH, goethite and/or hematite; Ps, pseudomorphs of clay minerals; Kf, K feldspar; Pl, plagioclase feldspar; Mi, mica; Qz, quartz; Vg, volcanic glass; AP, amphibole and/or pyroxene; Ze, zeolite; a, abundant; c, common; t, trace; +, chloritized vermiculite also noted.

[a] Relative amounts 3 = >50%; 2 = 10%–50%; 1 = <10%.

[b] Values are for a major fraction of sand or slit.

(Grossman et al., 1968). The Paleustalf in Figure 33.75 has COLE values >0.07 (7% volume change between moist and dry states) for horizons containing 30% or more clay. Because such soils are intergrades to Vertisols and possess similar features such as high subsoil clay content and shrink/swell potential, and slickensides, they present similar problems for use as construction materials (i.e., unstable subgrades for roads and houses).

Texture and clay content affect water-holding and water transmission properties of soils since water-related properties are dependent on void size and continuity. When the size of voids change greatly over a short distance, water movement is restricted. Alfisols with lamella type argillic horizons may exhibit slower drainage and better soil–plant relations due to greater available water retention than similar textured soils without lamella. Under more extreme conditions, strongly contrasting textures of eluvial and argillic horizons may result in water perched above the argillic horizons for several weeks as in Natraqualfs and Albaqualfs. In Alfisols with compact, clayey argillic horizons, root growth may be limited primarily to structural planes due to the high bulk density of ped interiors. Jones (1983) showed that root exclusion due to soil bulk density is a function of clay content.

Although organic C is greatest in the surface horizons, Alfisols tend to have greater CEC in the argillic horizon where clay and smectite contents are greatest. Generally, Alfisols must have a base saturation of 35% or more at the shallower of the following depths: (1) 1.25 m below the top of the argillic or kandic horizon, (2) 1.8 m below the soil surface, (3) 75 cm below the top of a fragipan, or (4) immediately above a densic, petroferric, lithic, or paralithic contact (just above continuous hard or soft bedrock; see Soil Survey Staff, 2010, for details). Most Alfisols have base saturation percentages well above 35%; those that approach low base status are recognized as intergrades to Ultisols by classification in Ultic subgroups. Generally, base saturation (and pH) decreases initially with depth in Alfisols, often reaching a minimum in the E horizons or upper portions of the argillic horizon, then increases again with depth. Alfisols under deciduous forest vegetation have relatively high base saturations in the surface horizon due to cycling of basic cations from depth to the surface in leaf fall. The relationship between soil pH and base saturation (Beery and Wilding, 1971; Blosser and Jenny, 1971; Ranney et al., 1974) is often used in field operations to separate Alfisols from Ultisols, but care must be exercised as this relationship changes with mineralogy.

The mineralogy from a range of Alfisols illustrating variations in parent materials, climates, and degree of development is presented in Table 33.52. Most Alfisols contain substantial quantities of 2:1 clay minerals such as smectite and hydrous mica, and perhaps some vermiculite, chlorite, and kaolinite. However, for more intensely weathered Alfisols (Kandiudalf, Paleudalf, Paleustalf, and Plinthustalf), kaolinite is the dominant clay mineral in the argillic or kandic horizon. Similarly, the most common mineral dominating the sand and silt fraction is quartz (Table 33.52), which comprised >90% of the sand fraction in the Paleustalf and Paleudalf. The Udalf from Oregon, which formed from tuff and breccia, was dominated by plagioclase feldspars. Most Alfisols contain significant quantities of weatherable minerals including K feldspars, plagioclase, and mica in the sand and silt fractions.

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33.12 Ultisols

Larry T. West
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33.12.1 Introduction

Ultisols are a group of soils with an argillic or kandic horizon and low base saturation in lower subsoil horizons. These soils may be found on a variety of parent materials and under a range of climatic conditions, though most have developed under forest vegetation in humid climates. Ultisols are generally considered to be less productive than soils in many of the other orders, but in many regions of the world, they are the most productive soils available. With proper management of organic residues, fallow periods, and/or chemical inputs, the productivity of this resource can be enhanced and maintained. This section presents a brief overview of Ultisol genesis and properties. For a more comprehensive review, the reader is referred to Miller (1983) and West et al. (1997).

33.12.2 Classification: Historical and Current

As soil classification has evolved in the United States, soils currently considered to be Ultisols have been included with the Red soils, Yellow soils, or Lateritic soils (Marbut, 1928), the Yellow Podzolic or Red Podzolic great soil groups (Baldwin et al., 1938), and the Red–Yellow podzolic great soil group (Kellogg, 1949). The order Ultisols was introduced in the *7th Approximation* (Soil Survey Staff, 1960) to include soils with an argillic horizon and base saturation <35% in lower subsoil horizons. The 35% base saturation criteria was placed to separate soils where basic cations are maintained in the root zone through biocycling, and soil amendments are commonly needed for sustained productivity (Ultisols), from soils that replenish a significant amount of basic cations through weathering (Alfisols) (Forbes, 1986). Since first publication of *Soil Taxonomy* (Soil Survey Staff, 1975), the classification of Ultisols has been revised to include kandic horizons as criteria for placement in the order and to include soils in any temperature regime. For a comprehensive definition of Ultisols, the reader is referred to the *Keys to Soil Taxonomy* (Soil Survey Staff, 2010). Table 33.53 lists current orders, suborders, and great groups in Ultisols (Soil Survey Staff, 2010).

33.12.3 Geographic Distribution

Worldwide, Ultisols are currently estimated to cover about 11,054,000 km^2 which is 8.1% of the ice-free global landmass (Table 33.53; Figure 33.76). About 80% of Ultisols are in tropical regions, and about 18% of the tropics is covered by Ultisols (Eswaran, 1993). About 50% of Ultisols are Udults, 35% Ustults, and 12% Aquults with Humults and Xerults comprising the

TABLE 33.53 Listing of Suborders, Great Groups and Subgroups in the Ultisols Order

Suborder	Great Group	Subgroup
Aquults	Plinthaquults	Kandic, Typic
	Fragiaquults	Aerie Plinthic, Umbric, Typic
	Albaquults	Vertic, Kandic, Aerie, Typic
	Kandiaquults	Acraquoxic, Arenic Plinthic, Arenic Umbric, Arenic, Grossarenic, Plinthic, Aerie, Umbric, Typic
	Kanhaplaquults	Aquandic, Plinthic, Aerie Umbric, Aerie, Umbric, Typic
	Paleaquults	Vertic, arenic plinthic, arenic umbric. Arenic, grossarenic, plinthic, aerie, umbric, typic
	Umbraquults	Plinthic, Typic
	Epiaquults	Vertic, Aerie Fragic, Fragic, Arenic, Grossarenic, Aerie, Typic
	Endoaquults	Arenic, Grossarenic, Aerie, Typic
Humults	Sombrihumults	Typic
	Plinthohumults	Typic
	Kandihumults	Andic Ombroaquic, Ustandic, Andic, Aquic, Ombroaquic, Plinthic, Ustic, Xeric, Anthropic, Typic
	Kanhaplohumults	Lithic, Ustandic, Andic, Aquic, Ombroaqic, Ustic, Xeric, Anthropic, Typic
	Palehumults	Aquandic, Andic, Aquic, Plinthic, Oxyaquic, Ustic, Xeric, Typic
	Haplohumults	Lithic, Aquandic, Aquic, Andic, Plinthic, Oxyaquic, Ustic, Xeric, Typic
Udults	Plinthudults	Typic
	Fragiudults	Arenic, Plinthaquic, Glossaquic, Aquic, Plinthic, Glossic, Humic, Typic
	Kandiudults	Arenic Plinthaquic, Aquic Arenic, Arenic Plinthic, Arenic Rhodic, Arenic, Grossarenic Plinthic, Grossarenic, Acrudoxic Plinthic, Acrudoxic, Plinthaquic, Aquandic, Andic, Aquic, Plinthic, Ombroaquic, Oxyaquic, Sombric, Rhodic, Typic
	Kanhapludults	Iithic, Fragiaquic, Arenic Plinthic, Arenic, Acrudoxic, Plinthaquic, Andic, Aquic, Ombroaquic, Oxyaquic, Fragic, Plinthic, Rhodic, Typic
	Paleudults	Vertic, Spodic, Arenic Plinthaquic, Aquic Arenic, Plinthaquic, Fragiaquic, Aquic, Anthraquic, Oxyaquic, Lamellic, Arenic, Plinthic, Psammentic, Grossarenic Plinthic, Plinthic, Arenic Rhodic, Arenic, Grossarenic, Fragic, Rhodic, Typic
	Rhoduldults	Lithic, Psammentic, Typic
	Hapludults	Lithic Ruptic-Entic, Lithic, Vertic, Fragiaquic, Aquic Arenic, Aquic, Fragic, Oxyaquic, Lamellic, Psammentic, Arenic, Grossarenic, Inceptic, Humic, Typic
Ustults	Plinthustults	Haplic, Typic
	Kandiustults	Acrustoxic, Aquic, Arenic Plinthic, Arenic, Udandic, Andic, Plinthic, Aridic, Udic, Rhodic, Typic
	Kanhaplustults	Lithic, Acrustoxic, Aquic, Arenic, Udandic, Andic, Plinthic, Ombroaquic, Aridic, Udic, Rhodic, Typic
	Paleustults	Typic
	Rhodustults	Lithic, Psammentic, Typic
	Haplustults	Lithic, Petroferric, Aquic, Arenic, Ombroaquic, Plinthic, Kanhaplic, Typic
Xerults	Palexerults	Aquandic, Aquic, Andic, Typic
	Haploxerults	Lithic Ruptic-Lnceptic, Lithic, Aquic, Andic, Lamellic, Psammentic, Arenic, Grossarenic, Typic

Source: Soil Survey Staff. 2010. Keys to soil taxonomy. 11th edn. USDA-NRCS. US Government Printing Office, Washington, DC.

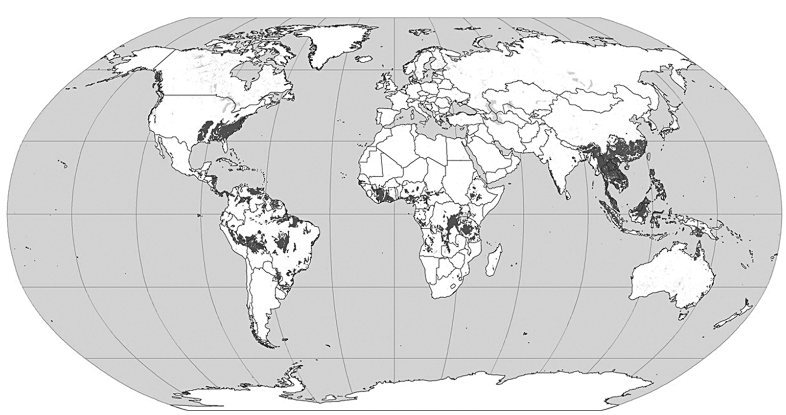


FIGURE 33.76 Global distribution of Ultisols. (Courtesy of USDA-NRCS, Soil Survey Division, World Soil Resources, Washington, DC, 1998.)

TABLE 33.54 Distribution of Ultisols in the Conterminous United States and Puerto Rico

State	Area (km^2)	Percentage of Total
Alabama	97,965	11.4
Arkansas	68,644	8.0
California	8,418	1.0
Delaware	3,421	0.4
Florida	28,106	3.3
Georgia	114,802	13.3
Idaho	17	0.0
Indiana	3,158	0.4
Illinois	44	0.0
Kansas	79	0.0
Kentucky	27,402	3.2
Louisiana	23,395	2.7
Massachusetts	22	0.0
Maryland	16,004	1.9
Missouri	28,534	3.3
Mississippi	49,826	5.8
North Carolina	85,057	9.9
New Jersey	7,617	0.9
New York	579	0.1
Oregon	11,195	1.3
Ohio	7,137	0.8
Oklahoma	16,784	1.9
Pennsylvania	39,375	4.6
Puerto Rico	1,784	0.2
South Carolina	54,720	6.3
Tennessee	44,317	5.1
Texas	24,967	2.9
Virginia	69,910	8.1
Washington	3,574	0.4
West Virginia	25,435	2.9
Total	862,288	100.0

Source: Courtesy of USDA-NRCS, Soil Survey Division, National Soil Survey Center, Washington, DC.

remaining 3% (Table 33.53). In the United States, Ultisols cover about 860,000 km^2 (Table 33.54) and occur in 30 states and Puerto Rico (Figure 33.77). The largest concentration of Ultisols in the United States is in the east central, southeast, and south central parts of the country. Ultisols are also extensive on the older islands of Hawaii, along the west coast, and in unglaciated regions of the northeast and north central parts of the country.

33.12.4 Genesis of Ultisols

Soil as an independent natural body whose properties are a function of five soil-forming factors: local climate, parent materials, organisms, relief, and age (Dokuchaev, 1883; Jenny, 1941), is a unifying philosophy in pedology. As such, Ultisol genesis will be briefly discussed in terms of these factors.

33.12.4.1 Climate

The definition of Ultisols implies two rainfall/evapotranspiration conditions. First, there must be some time during the year when evapotranspiration exceeds precipitation, as this appears to be a prerequisite for the formation of an argillic horizon (van Wambeke, 1991). Second, precipitation must exceed the capacity of the soil to retain water during some time of the year, so that water percolates through the solum to remove basic cations and maintain low base saturation (<35% by sum of cations [pH 8.2] which corresponds roughly to 50% base saturation measured at pH 7.0) in the subsoil (Miller, 1983). Most Ultisols occur in regions with mean annual air temperatures above 6°C. Thus, the climate conducive to the formation of Ultisols is typically humid tropical or warm temperate. As many Ultisols have evolved over long time periods, effects of paleoclimate on Ultisol development cannot be ignored.

33.12.4.2 Parent Material

The definition of Ultisols requires the presence of either an argillic or a kandic horizon, both of which contain silicate clays. Therefore, the parent material of Ultisols must be one that contains either phyllosilicates or primary minerals that can weather to produce them. As the vast majority of rocks meet one or both of these criteria, a wide variety of geologic formations are capable of producing parent materials suitable for Ultisol formation. Residual parent materials that have weathered in situ are usually saprolite. In contrast, transported sediments that form Ultisol parent materials have frequently been preweathered and may have gone through more than one weathering cycle (Fiskell and Perkins, 1970; Allen and Fanning, 1983). In sedimentary landscapes, parent material has a major influence on Ultisol property distribution (Shaw et al., 2004).

33.12.4.3 Biota

Although Ultisols occur within many tropical and subtropical managed ecosystems, forests are the dominant vegetation on most natural Ultisol landscapes. The processes in forest soils appear to accelerate the formation of argillic horizons through the desiccating effect of tree roots that absorb water, but not colloids in suspension. In Africa and South America, vast areas of Ultisols are under savanna vegetation. It cannot be ascertained, however, whether the soils formed under the present vegetation or under forests in one or more humid paleoclimates.

33.12.4.4 Relief

Ultisols occur on almost all landforms. The one characteristic that Ultisol landforms have in common, however, is geomorphic stability over long periods of time. Geomorphic studies of tropical landscapes invariably show that Ultisols occupy geomorphic positions that are younger and less stable than the surfaces where Oxisols occur, but older and more stable than those of soils in other Orders with which they are geographically associated (Beinroth et al., 1974; Lepsch and Buol, 1974; Beinroth, 1981). In temperate regions where Oxisols are absent, Ultisols

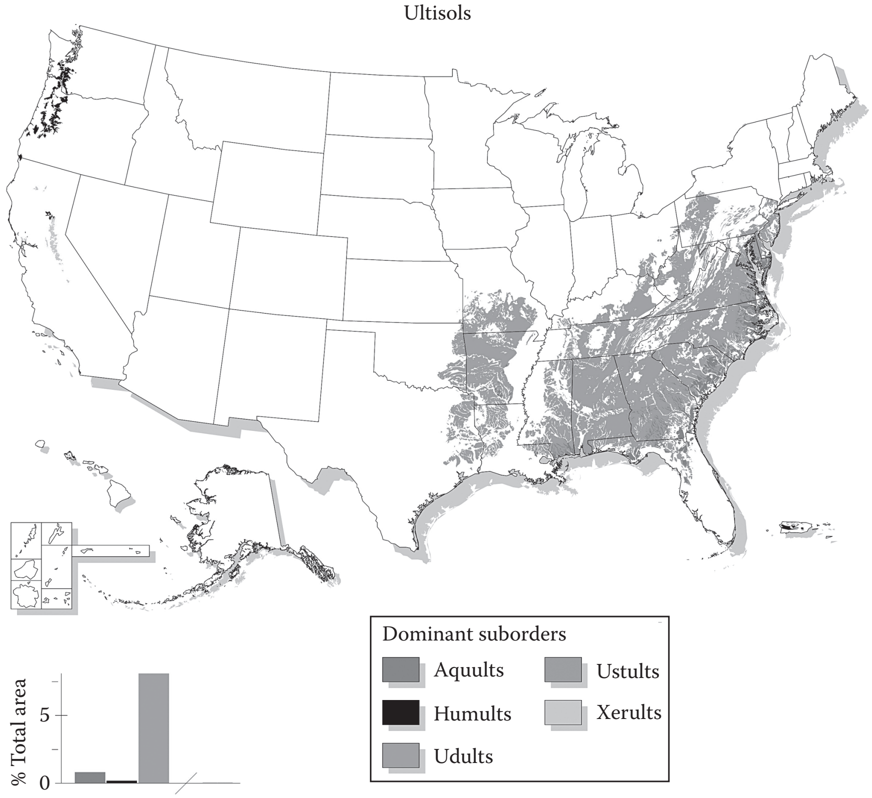


FIGURE 33.77 Distribution of Ultisols in the United States. (Courtesy of USDA-NRCS, Soil Survey Division, National Soil Survey Center, Washington, DC, 2010.)

have developed on the oldest, most stable and highly weathered landscapes such as the central Missouri Ozarks and the coastal plain of the southeastern United States (Scrivner et al., 1966; Daniels et al., 1971).

33.12.4.5 Time

The time required for the formation of Ultisols depends on the other pedogenic factors, notably weatherability, composition, and permeability of the parent materials, landscape stability, and climate fluctuations over time. There is much room for variability in these conditions, and consequently, for the age of Ultisols. The Soil Survey Staff (1999) concluded that formation of the argillic horizon ordinarily requires a few thousand years. This relatively short time may suffice to produce a minimal expression of an argillic horizon. The development of the thick argillic horizons that typify the Pale and Kandi great groups of Ultisols, however, would certainly take longer. As such, age of Ultisols has been reported to range from <12,000 years for soils on preweathered sediments to >15 million years on marine sediments in the coastal plain of the southeast United States (Gamble et al., 1970; Daniels et al., 1971; Miller, 1983).

33.12.5 Morphological Properties

33.12.5.1 Argillic and Kandic Horizons

Other than low base saturation, the one property that distinguishes Ultisols from all but one other soil order (Alfisols, Section 33.11) is the requirement of a clay increase between surface and subsoil horizons in the form of either an argillic or kandic horizon. This clay increase essentially determines the morphology of Ultisols and influences many chemical and physical properties. The depth, thickness, and clay content of the argillic or kandic horizon, however, can vary widely, and this variation has considerable impact on management and interpretation of these soils. Figure 33.78 illustrates the range in depth and clay content of argillic or kandic horizons that can occur in Ultisols.

33.12.5.2 Soil Color

Color of surface horizons in Ultisols is commonly brown, yellowish brown, dark brown, or red as would be expected for soils with relatively low amounts of soil organic C (SOC). However, in cool climates or poorly drained conditions with slow organic matter decomposition, thick, dark, acidic surface horizons (umbric epipedons) are common. Many well-drained Ultisols have reddish Bt horizons because of the presence of hematite as a pigmenting agent. Yellowish-brown subsoil horizons, also common in Ultisols, lack hematite because either the parent material lacked mineral precursors for hematite, environmental conditions precluded hematite formation (Schwertmann and Taylor, 1989), or hematite was preferentially reduced and lost under short periods of saturation leaving behind more stable goethite (Macedo and Bryant, 1989; Dobos et al., 1990). Under extended periods of saturation and reduction, Ultisols will develop redox

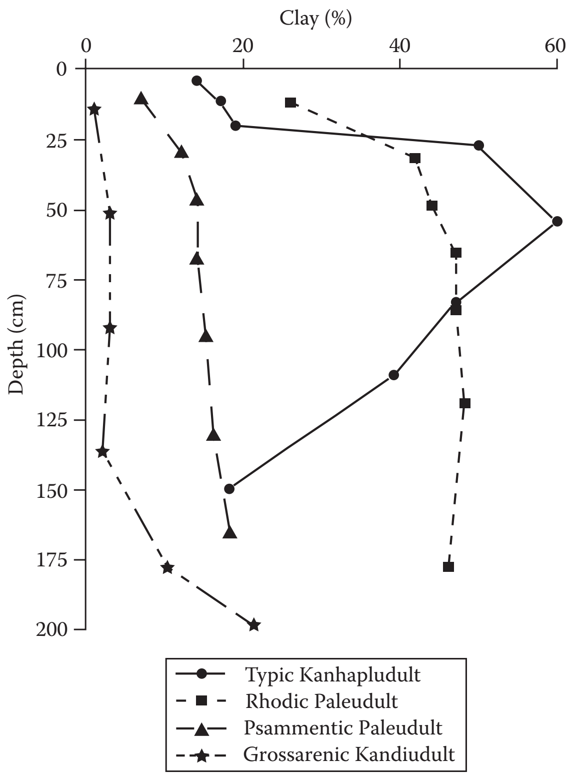


FIGURE 33.78 Clay distribution with depth for four subgroups of Ultisols.

concentrations and depletions or gray matrices as Fe is reduced and mobilized.

33.12.5.3 Plinthite

Plinthite is defined in *Soil Taxonomy* as an Fe-rich, humus-poor mixture of clay with quartz and other diluents (Soil Survey Staff, 2010). This material most often occurs in lower subsoil horizons, is weakly indurated, and is normally low in basic cations and primary silicate minerals other than quartz. Kaolinite is commonly the only clay mineral present with abundance. The reader is referred to van Wambeke (1991) for a complete discussion of the genesis and kinds of plinthite and petroplinthite (or ironstone).

From a pedological perspective, plinthite's relationship with contemporary soil saturation is not well understood, but it does have significance in terms of movement of water through soils. Daniels et al. (1978) suggested that 10% platy plinthite would perch water. A similar amount of nodular plinthite was not considered to be water restrictive, but horizons subjacent to the plinthic horizon did restrict water movement. Movement of water and solutes through horizons containing plinthite is primarily restricted to areas of redox depletions (Carlan et al., 1985; Blume et al., 1987; Shaw et al., 1998). Because of water perching by plinthic or subjacent horizons, plinthite in soils suggests that water and solutes are moving from high to low landscape positions as shallow subsurface flow (Hubbard and Sheridan, 1983; Shirmohammadi et al., 1984).

33.12.6 Mineralogical Properties

33.12.6.1 Clay Minerals

The clay fraction of subsoil horizons of most Ultisols is dominated by kaolinite because of the stability of kaolinite when compared with other phyllosilicates (Jackson et al., 1948; Allen and Fanning, 1983). In addition, gibbsite, K mica, K feldspar, and amorphous silica may transform directly to kaolinite without passing through intermediates (Garrels and Christ, 1965). As such, kaolinite in soils has been shown to be an early weathering product of biotite and K feldspar in acid gneiss and schist parent materials (Robertus and Buol, 1985; Robertus et al., 1986; Norfleet and Smith, 1989).

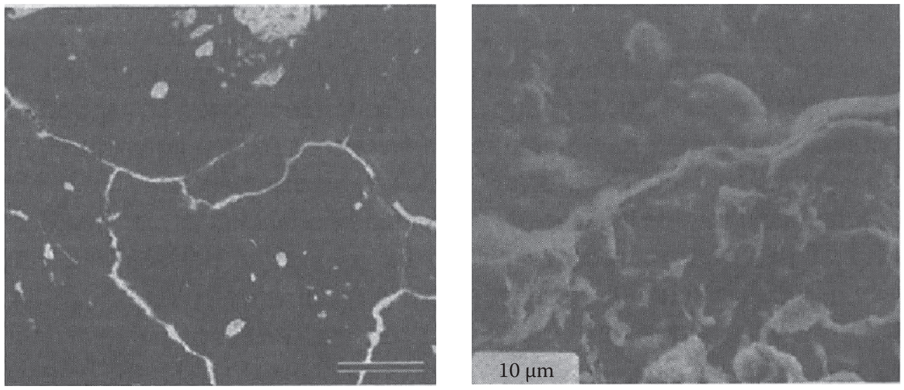
In addition to kaolinite, hydroxy Al-interlayered vermiculite is often abundant in A and E horizons of Ultisols and may be a major component of upper Bt horizons of these soils (Bryant and Dixon, 1964; Fiskell and Perkins, 1970; Carlisle and Zelazny, 1973; Harris et al., 1980; Karathanasis et al., 1983). Studies in the Piedmont and Coastal Plain of Alabama (southeastern United States) indicate that hydroxy Al-interlayered vermiculite is more abundant than kaolinite in the upper solum of many Ultisols (Shaw et al., 2010). Appreciable amounts of mica, smectite, and other 2:1 minerals have also been reported to occur in Ultisols (Nash, 1979; Harris et al., 1984; Karathanasis et al., 1986; Nash et al., 1988). The 2:1 minerals in these Ultisols were inherited from the parent materials instead of forming in situ from primary minerals or other phyllosilicates.

A common but minor mineral component of many Ultisols is gibbsite (Carlisle and Zelazny, 1973; Hajek and Zelazny, 1985). Because this mineral is an end product of silicate mineral weathering (Jackson et al., 1948), minor amounts should be expected to be found in Ultisols. Clay fractions of Ultisols developed from acid gneiss and schist in the southeastern United States, however, have been reported to contain as much as 70% gibbsite formed as a product of K feldspar weathering at the rock–saprolite interface (Losche et al., 1970; McCracken et al., 1971; Norfleet and Smith, 1989). Minor amounts of halloysite and Ti and Mn oxides are commonly present in Ultisols (Figure 33.79).

Iron oxides and oxyhydroxides are also common, but minor components in the clay fraction of Ultisols. Goethite is the most common Fe oxide mineral in most Ultisols, but many well-drained Ultisols have appreciable amounts of hematite. Analyses of 13 horizons from kandic and/or argillic horizons of Ultisols sampled within the Alabama (southeastern United States) Piedmont and Coastal Plain indicated that hematite composed 0%–11.4% of the clay (<2 μm) fraction, while goethite composed 0%–13.7% (Shaw, 2001). Hematite formation is favored by warm temperatures and distinct dry seasons (Schwertmann and Taylor, 1989), which are environmental conditions commonly associated with Ultisols.

33.12.6.2 Sand and Silt Minerals

The sand and silt fraction of Ultisols commonly is dominated by quartz, and many Ultisols have >90% quartz and other resistant minerals in their sand and silt separates. However, weakly to moderately developed Ultisols may have appreciable quantities of mica, feldspars, and other weatherable minerals in their



(a) (b)

FIGURE 33.79 (a) Thin section of micrograph of clay coatings in Btv2 horizon of a Plinthic Kandiudult. Plane-polarized light. Bar length = 0.5 mm; (b) scanning electron micrograph of clay coating in 2Btx1 horizon of a Typic Fragiudult. (Original micrograph from Michael Thompson, Iowa State University, Ames, IW. With permission.)

coarse fraction (Robertus et al., 1986; Norfleet and Smith, 1989; Bockheim et al., 1996).

33.12.7 Physical Properties

33.12.7.1 Bulk Density, Coefficient of Linear Extensibility, Water Retention, and Hydraulic Conductivity (*K*)

Bulk density, COLE, water retention, and hydraulic conductivity of Ultisols vary considerably depending on soil and horizon properties including particle size distribution, SOC content, mineralogy, macroporosity, presence of fragipans, parent material, presence of ironstone and plinthite, and, for surface horizons, management. A horizons of uncultivated Ultisols often have lower bulk density than subsoil horizons, but with tillage and the accompanying loss of organic matter, compaction, and aggregate breakdown, the A horizon may become as dense or more dense than underlying Bt horizons (Table 33.55; Hubbard et al., 1985). Shrink/swell potential for kaolinitic Ultisols commonly is low reflecting the abundance of kaolinite and other low activity minerals (Table 33.53; Hubbard et al., 1985; Southard and Buol, 1988a), but COLE values up to 0.157 have been reported for Bt horizons from Ultisols with smectitic mineralogy (Karathanasis and Hajek, 1985; Griffin and Buol, 1988). However, smectitic Ultisols with high shrink/swell potential often lack wide cracks common to Vertisols and Vertic subgroups of other orders (Karathanasis and Hajek, 1985).

TABLE 33.55 Ranges in Selected Physical Properties of Ultisols

Property	Value	References
Bulk density (Mg m^{-3})		
A horizon	0.54–1.85	Bruce et al. (1983)
Bt horizon	1.05–1.70	Hubbard et al. (1985)
		Quisenberry et al. (1987)
		Southard and Buol (1988a)
		Norfleet and Smith (1989)
COLE (Bt horizons)	0.006–0.157	Karathanasis and Hajek (1985)
		Griffin and Buol (1988)
		Southard and Buol (1988a)
Water retention (cm^3 cm^{-3})		
A horizons		Bruce et al. (1983)
−10 kPa	0.13–0.21	Hubbard et al. (1985)
−30 kPa	0.09–0.18	Quisenberry et al. (1987)
−1500 kPa	0.05–0.09	
Bt horizons		
−10 kPa	0.20–0.42	
−30 kPa	0.17–0.40	
−1500 kPa	0.09–0.33	
Saturated hydraulic conductivity (μm s^{-1})		
Bt horizons	0.01–13.1	Hubbard et al. (1985)
Btv horizons	0.3–1.3	Southard and Buol (1988b)

TABLE 33.56 Mean Pore Area for Dyed and Undyed Areas after Breakthrough of Methylene-Blue for Ultisols from the Piedmont and Coastal Plain of Georgia

	Mean Pore Area[a] (%)	
Horizon	Dyed	Undyed
Typic Kanhapludult (Piedmont)		
A	19.0	3.0
BA	17.7	5.0
Bt	15.8	2.4
Bt	14.4	2.8
Plinthic Kandiudult (Coastal Plain)		
Bt	5.0	1.1
Btv	5.9	0.8
BC	NA[b]	0.9

Sources: Shaw et al., 1997; Franklin, D.H., L.T. West, D.E. Radcliffe, and P.F. Hendrix. 2007. Characteristics and genesis of preferential flow paths in a piedmont ultisol. Soil Sci. Soc. Am. J. 71:752–758.

[a] Area of pores >0.05 mm equivalent circular diameter determined from impregnated polished blocks with image analysis.

[b] The BC horizon had insufficient dyed area for evaluation of pore area.

Ultisols in fine-loamy or fine particle size families would generally be considered capable of retaining and supplying sufficient moisture for plant growth in humid climates (Table 33.55). However, water often becomes a limiting factor for crop growth and yield for Ultisols in sandy or coarse-loamy particle size families, lithic subgroups, or that have thick sandy epipedons (Arenic and Grossarenic subgroups).

Most studies of hydraulic properties of Ultisols have found macroporosity to be the best predictor of saturated hydraulic conductivity (Ks) (O'Brien and Buol, 1984; Southard and Buol, 1988b). Preferential flow paths in Ultisols in the Piedmont and Coastal Plain of Georgia have textures similar to the matrix of the horizon, but the preferential flow zones have a significantly greater number and area of pores >0.05 mm diameter (Table 33.56; Shaw et al., 1998; Franklin et al., 2007). Formation of coarse pores in Ultisols is most often attributed to microfabric alteration and channel formation by roots and burrowing organisms (O'Brien and Buol, 1984; Southard and Buol, 1988b; Shaw et al., 1998; Franklin et al., 2007).

33.12.7.2 Infiltration and Surface Crusting

In many Ultisols, one factor affecting infiltration of water is formation of a surface crust or seal. For 25 Ultisols from the southeastern United States tested under 50–90 mm h^{-1} simulated rainfall, infiltration rate decreased rapidly during the first 10–20 mm of the rainfall event, and the infiltration rate at the end of rainfall was less than 10 mm h^{-1} for all of the soils (Miller and Radcliffe, 1992). In general, Ultisols in this region with sandy loam texture and high amounts of water dispersible clay are most prone to crusting. Soils higher in clay tend to take longer to form a surface crust, and sands and loamy sands form only weakly expressed crusts (Miller and Bahruddin, 1986; Radcliffe et al., 1991; Miller and Radcliffe, 1992; Chiang et al., 1993).

Because energy inputs from rainfall are needed to form surface crusts, crusting is minimal if the soils have residue or vegetative cover and organic matter is maintained in the upper part of the soil. For Ultisols in the Piedmont of Georgia, infiltration rates after 1 h of simulated rainfall were about 40% higher for soils that had been under no-tillage for 5 years than for soils that had been conventionally tilled (Table 33.57; Bruce et al., 1992).

33.12.8 Chemical Properties

The requirement that Ultisols have base saturation <35% in the lower part of the subsoil carries with it certain associated properties including low pH, potentially high Al saturation, appreciable weathering and associated kaolinitic mineralogy, and in many cases, relatively high contents of Fe and Al oxides and oxyhydroxides. This suite of properties has significant impact on the chemical properties of Ultisols and how these properties affect use and management of this soil order.

33.12.8.1 Cation and Anion Exchange Capacity

The CEC of Ultisol horizons depends on the amount and type of clay, Fe, Al, and Mn oxide content, organic matter content,

TABLE 33.57 Comparison of Organic C, Aggregate Stability, Infiltration Rate, and Percentage Moist Days under Conventional and No-Till Cropping Systems

			Infiltration Rate[a]		
Cropping System[a]	Organic C ($g\ kg^{-1}$)	Aggregate Stability ($g\ kg^{-1}$)	Residue ($mm\ h^{-1}$)	Residue Removed ($mm\ h^{-1}$)	Moist Days,[b] % of Growing Season
CTG	10.4	580	36	22	29
NTG	23.3	890	50	46	49

Sources: West, L.T., W.P. Miller, G.W. Langdale, R.R. Bruce, J.M. Laflen, and A.W. Thomas. 1991. Cropping system effects on interrill soil loss in the Georgia Piedmont. Soil Sci. Soc. Am. J. 55:460–466; Bruce, R.R., G.W. Langdale, L.T. West, and W.P. Miller. 1992. Soil surface modification by biomass inputs affecting rainfall infiltration. Soil Sci. Soc. Am. J. 56:1614–1620.

CTG, conventional-tilled grain sorghum (*Sorghum bicolor* (L.) Moench); NTG, no-till grain sorghum with clover winter cover.

[a] Infiltration rate at the end of a 1 h of rainfall simulation at approximately 64 mm h^{-1}.

[b] Moist days are defined as the days with soil moisture tension >0.1 MPa.

soil solution electrolyte concentration, and pH of the horizon. Reported CEC values for Ultisol horizons range from <3 to >20 $cmol_c\ kg^{-1}$ (Sanchez, 1976; Carlisle et al., 1985; Karathanasis et al., 1986). Generally, Ultisols with kaolinitic mineralogy have mostly pH-dependent charge and low CEC, while those with mixed or smectitic mineralogy have a greater proportion of fixed charge and higher CEC values.

Even when the net charge is negative, many Ultisols have an appreciable positive charge and anion exchange capacity (AEC). Gillman and Sumner (1987), in a study of four Ultisols from the Piedmont of Georgia, reported AEC values at the pH of the soil that ranged from 0.1 to 1.2 $cmol_c\ kg^{-1}$, which was attributed to Fe and Al oxides in these soils. For Ultisols from Georgia and South Africa, Grove et al. (1982) reported AEC values ranging from 0.03 to 1.91 $cmol_c\ kg^{-1}$. Measurable AEC has also been reported for Ultisols from tropical Australia and Peru (Gillman and Sumpter, 1986; Gillman and Sinclair, 1987).

33.12.8.2 Acidity and Exchangeable Aluminum

Below a pH of about 5.5, Al released from weathering of primary and secondary minerals is present as part of the exchange complex. An Al saturation of greater than 60% (measured as a percentage of the unbuffered CEC or ECEC) has been reported as the threshold that results in soil solution Al concentrations >1 $mg\ L^{-1}$, which may reduce yield in Al-sensitive crops (Nye et al., 1961; Kamprath, 1970; Sanchez, 1976). Farina and Channon (1991), however, observed yield reductions in maize for Al saturations >25% of the ECEC. Yields are also reduced at low pH due to Mn toxicity, and Ca, Mg, and Mo deficiency (Adams, 1984).

33.12.8.3 Phosphorus

Phosphate is specifically adsorbed by Fe, Al, and Mn oxides and amorphous or poorly crystalline aluminosilicates. Because of eluviation and weathering, surface horizons of most Ultisols have low contents of these components. Thus, P sorption is generally lower in Ultisols than in Oxisols and Andisols (Sanchez, 1976; van Wambeke, 1991). However, differences in crystallinity of Fe oxides among soils may confound this relationship within and among orders (Pratt et al., 1969; Fox et al., 1971).

33.12.9 Biological Properties

33.12.9.1 Organic Matter

Ultisols are commonly a fragile soil resource with low inherent soil quality (Norfleet et al., 2003). However, because these soils often exist in regions with long productive growing seasons and have favorable soil physical properties, Ultisols are often utilized for food, fiber, and timber production. Conservation management of these soils is critical for maximizing and sustaining productivity of agronomic systems. Numerous studies have documented the relationships between management, soil quality, productivity, and SOC in these soils (Reeves, 1997). A common misconception is that Ultisols have low contents of SOC compared with other soil orders. While Ultisols generally have lower amounts of SOC than Mollisols, Oxisols, and Andisols, SOC levels in Ultisols are similar to those in Alfisols both within and across geographic areas (Sanchez, 1976). There is little difference in SOC content between Ultisols in temperate and tropical climates (Buol, 1973; Sanchez, 1976).

Cultivation of Ultisols has decreased SOC levels compared with relatively undisturbed ecosystems, and SOC contents decrease when these soils are converted from forest or pasture to row crop lands. Studies of Georgia (southeastern United States) Ultisols have found that organic matter content in surface horizons decreased from 20.5 to 16.6 $g\ kg^{-1}$ 2 years after conversion from forest to intensive cultivation (Giddens, 1957), while soils in relatively undisturbed longleaf pine (*Pinus palustris* Miller) ecosystems in the Coastal Plain had 64% more SOC (relative) (0–30 cm) than conventional row crop lands (Levi et al., 2010). The use of conservation tillage, cover crops, and residue management (conservation systems) in row crop production can increase Ultisol SOC levels compared with conventional systems (Reeves, 1997). Bruce et al. (1992) reported that conservation systems increased SOC in the upper 15 mm of degraded Ultisols from 10.4 to 23.3 $g\ kg^{-1}$ over a 5 year period. An investigation of 87 sites consisting mostly of Ultisols in the southeastern United States found SOC (0–20 cm) to be highest in pastures (38.9 $Mg\ ha^{-1}$), followed by conservation row crop systems (27.9 $Mg\ ha^{-1}$), and lastly, conventional row crop systems (22.2 $Mg\ ha^{-1}$) (Causarano et al., 2008). In addition, conservation system use in southeastern U.S. cotton (*Gossypium hirsutum* L.) production has been found to increase SOC quantities by 0.48 ± 0.56 $Mg\ C\ ha^{-1}\ year^{-1}$ compared with conventional systems (Schwab et al., 2002; Causarano et al., 2006). Thus, Ultisols can play a significant role in managing SOC sequestration.

Increasing SOC in Ultisols significantly increases water stable aggregates and decreases runoff and interrill soil loss (Table 33.58; West et al., 1991, 1992; Bruce et al., 1992). Similarly, decreased runoff (Truman et al., 2005) and water dispersible clay (Shaw et al., 2003) were found for Alabama (southeastern United States) Coastal Plain Ultisols under conservation management with higher SOC quantities.

33.12.9.2 Biologic Populations and Processes

Populations of microorganisms in Ultisols are not appreciably different from soils in other orders. Surface horizons of most

TABLE 33.58 Interrill Soil Loss and Runoff from Three Ultisols with Varying Tillage and Surface Cover under 50 $mm\ h^{-1}$ Rainfall Intensity for 1 h on 5%–9% Slopes

Tillage	Surface Condition	Interrill Soil Loss ($g\ m^{-2}$)			Total Runoff (mm)		
		Soil 1	Soil 2	Soil 3	Soil 1	Soil 2	Soil 3
Conventional	Crusted	0.46	0.56	0.33	16	23	17
	Tilled	0.72	0.79	0.48	18	23	16
No-till	Bare	0.14	0.19	0.07	3	3	6
	Residue	0.07	0.03	0.02	1	1	1

Source: West, L.T., W.P. Miller, G.W. Langdale, R.R. Bruce, J.M. Laflen, and A.W. Thomas. 1991. Cropping system effects on interrill soil loss in the Georgia Piedmont. Soil Sci. Soc. Am. J. 55:460–466.

Ultisols have acid pH values. Thus, fungi may comprise a greater proportion of the microbial biomass in these soils because of the greater ability of fungi to survive under acid conditions than other microorganisms (Alexander, 1977). Low pH has been reported to reduce rates of nitrification and denitrification in pure cultures (Alexander, 1977). Natural adaptation by microorganisms to local field conditions, however, probably alters the effects of pH and other environmental conditions observed in the laboratory.

33.12.10 Conclusions

Worldwide, Ultisols are a widespread soil resource, which because of their extensive weathering, low base saturation, and associated properties, are often considered to be unproductive. It is true that Ultisols, in general, have lower native fertility than Mollisols and Alfisols, and in areas where these orders are abundant, Ultisols are the less desirable soils. In many regions of the world, however, Ultisols are dominant or are the most productive soils available. Only with a thorough understanding of the genesis, properties, and response to management of these soils can the productivity of this valuable resource be maintained and enhanced.

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33.13 Oxisols

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33.13.1 Introduction

The deep, red and highly weathered soils of the tropics have long fascinated pedologists, particularly those from the temperate region. The uniqueness of these soils lies not only in their properties, but also in their geographic distribution being confined almost exclusively to the tropics. As this account is intentionally concise, ample references are provided for those seeking additional or more detailed information.

33.13.2 Historical Background

In the early literature, the highly weathered soils of the tropics were identified as red soils, red loams, or red earths. In an historical report on travels in South India, Buchanan (1807) described soil material used for construction and called it laterite (Buchanan, 1807). The term was adopted by the pedologic community and is still used today with a wide variety of meanings (Alexander and Cady, 1962; Maignen, 1966). In 1949, a group of scientists proposed the term Latosols (Cline, 1975) that soon became popular and, by the midfifties, the concept of Latosols as highly weathered soils with a low negative charge was firmly established. Several soil classification systems developed during the 1950s and 1960s recognized such soils in a separate class with the nomenclature shown in Table 33.59.

The term Oxisol originated around 1954 (Smith, 1963, 1965) during the development of *Soil Taxonomy* (Soil Survey Staff, 1975). In 1978, an International Committee on the Classification of Oxisols (ICOMOX) was formed (Buol and Eswaran, 1988) to initiate discussions that would lead to an improved classification. Through circular letters and four international soil classification workshops held in Brazil (1976), Malaysia and Thailand (1978), Rwanda (1981), and Brazil (1986), the current state of knowledge of these soils was assembled and documented (Beinroth and Panichapong, 1978; Beinroth and Paramanathan, 1978; Camargo and Beinroth, 1978; Beinroth et al., 1983, 1986).

33.13.3 Concept and Classification of Oxisols

33.13.3.1 Perceptions

The rationale of the taxon of Oxisols is based largely on the concept of Latosols, which was developed to designate all zonal soils having their dominant characteristics associated with low silica:sesquioxide ratios, low base cation exchange capacity, low activity clays, and low content of weatherable minerals (Kellogg, 1949; Cline, 1975). These characteristics were selected because they were thought to manifest the effects of advanced pedogenesis under tropical conditions.

TABLE 33.59 Synonyms for Oxisols in Other Classification Systems

Synonym	Classification System	Sources
Sols ferrallitiques	French	Aubert (1958)
Latosols	Brazilian	Bennema et al. (1959)
Solos ferraliticos	Portuguese	Botelho da Costa (1954, 1959)
Oxisols	Ghana	Charter (1958)
Ferralsols		FAO–UNESCO (1971–1976)
Rotlehm	German	Harrassowitz (1930)
Laterites		Maignen (1966)
Red earths	British	Robinson (1951)
Kaolisols	Belgian	Tavernier and Sys (1965)

33.13.3.2 Definition

Simplified from the *Keys to Soil Taxonomy* (Soil Survey Staff, 2010), Oxisols may be defined as those soils that fail to meet the criteria definitive for Gelisols, Histosols, Spodosols, and Andisols but have either (1) an oxic horizon within 150 cm of the mineral soil surface and no kandic horizon within this depth, or (2) 40% or more clay in the surface horizon and a kandic horizon within 100 cm of the mineral soil surface that meets the weatherable mineral properties of the oxic horizon.

The definition of Oxisols has undergone significant changes since it was first published in *Soil Taxonomy* (Soil Survey Staff, 1975). Consequently, more soils now qualify for Oxisols; this should be kept in mind when consulting the earlier literature. It should also be mentioned that not all intensively weathered soils are Oxisols. Highly weathered soils with a marked clay increase below a coarse-textured surface horizon are by definition excluded from the Oxisols and are classified as kandic great groups of Ultisols and Alfisols (van Wambeke, 1992; Soil Survey Staff, 1998).

33.13.3.3 Diagnostic Criteria

Key to the identification of Oxisols is the presence of either an oxic horizon or a kandic horizon underlying a surface horizon with 40% or more clay. The key properties of the oxic horizon are its charge characteristics and negligible amounts of weatherable minerals. The charge characteristics are defined by the magnitude of the charge (CEC by NH_4OAc at pH 7.0 of <16 $cmol_c\ kg^{-1}$ clay) and the charge estimated by the Effective Cation Exchange Capacity (ECEC < 12 $cmol_c\ kg^{-1}$ clay). The low weatherable mineral requirement of <5% ensures that there are few weatherable minerals that could release plant nutrients. The particle size class of the oxic horizon is sandy loam or finer and its upper boundary is diffuse. To be diagnostic, the oxic horizon must have a minimum thickness of 30 cm and occur within 150 cm of the mineral soil surface.

The kandic horizon (Moormann, 1985) shares some of the properties of the oxic and argillic horizons. Like the argillic horizon, clay increases with depth but the charge characteristics are those of the oxic horizon. To be diagnostic for Oxisols, the kandic horizon must meet the weatherable mineral requirements of the oxic horizon, occur beneath a surface horizon that has 40% or more clay, and have its upper boundary within 100 cm of the soil surface.

Suborders of Oxisols are differentiated on the basis of soil moisture regimes, aquic conditions, the presence of a histic epipedon, and soil color. Defining criteria for great groups are the kandic and sombric horizons, plinthite, acric properties, and base saturation. The 14 differentiae used to establish subgroups are andic-like properties, Al + Fe percentage, aquic conditions, CEC, ΔpH ($pH_{KCl} - pH_{water}$), histic epipedon, kandic horizon, lithic contact, organic C content, petroferric contact, plinthite, soil color, soil depth, and the sombric horizon.

At the family level, criteria include particle size class, mineralogy class, and soil temperature regimes. Soil mineralogy is an important factor, and eight mineralogical classes are recognized (Table 33.60). When the soil has >18% Fe_2O_3 or gibbsite in the fine earth fraction (<2 mm), the mineralogical class is determined by

TABLE 33.60 Mineralogy Classes of Oxisols

Mineralogy Class	Fe_2O_3 (%)	Gibbsite (%)	Kaolinite (%)	Halloysite (%)
Ferritic	>40			
Gibbsitic	0–40	>40		
Sesquic	18–40	18–40		
Ferruginous	18–40	0–18		
Allitic	0–18	18–40		
Kaolinitic	0–18	0–18	>50	
Halloysitic	0–18	0–18	<50	>50
Mixed	0–18	0–18	<50	<50

Source: Soil Survey Staff. 2010. Keys to soil taxonomy. 11th edn. USDA-NRCS. U.S. Government Printing Office, Washington, DC.

the amounts of each of these. If there is <18% Fe_2O_3 (12.6% Fe) or gibbsite, the composition of the clay fraction is considered.

33.13.3.4 Taxa

The class of Oxisols has 5 suborders, 22 great groups, 213 subgroups, an estimated 400 families, and perhaps as many as 1000 soil series. The great groups and possible subgroups of the Oxisols are presented in Table 33.61.

33.13.4 Geography of Oxisols

33.13.4.1 Global Extent and Geographic Distribution

The Oxisols of the world comprise about 9.8 million ha (Table 33.1), approximately equivalent to 7.5% of the global and 25% of the tropical land area. The distributions of Oxisol suborders in South America, Africa, and globally are illustrated in Figures 33.80 through 33.82. Over 95% of the Oxisols occur in the

TABLE 33.61 Listing of Suborders, Great Groups, and Subgroups in the Oxisols Order

Suborder	Great Group	Subgroups
Aquox	Acraquox	Plinthic, Aerie, Typic
	Plinthaquox	Aerie, Typic
	Eutraquox	Histic, Plinthic, Aerie, Humic, Typic
	Haplaquox	Histic, Plinthic, Aerie, Humic, Typic
Torrox	Acrotorrox	Petroferric, Lithic, Typic
	Eutrotorrox	Petroferric, Lithic, Typic
	Haplotorrox	Petroferric, Lithic, Typic
Ustox	Sombriustox	Petroferric, Lithic, Humic, Typic
	Acrustox	Aquic Petroferric, Petroferric, Aquic Lithic, Lithic, Anionic Aquic, Anionic, Plinthic, Aquic, Eutric, Humic Rhodic, Humic Xanthic, Humic, Rhodic, Xanthic, Typic
	Eutrustox	Aquic Petroferric, Petroferric, Aquic Lithic, Lithic, Plinthaquic, Plinthic, Aquic, Kandiustalfic, Humic Inceptic, Inceptic, Humic Rhodic, Humic Xanthic, Humic, Rhodic, Xanthic, Typic
	Kandiustox	Aquic Petroferric, Petroferric, Aquic Lithic, Lithic, Plinthaquic, Plinthic, Aquic, Humic Rhodic, Humic Xanthic, Humic, Rhodic, Xanthic, Typic
	Haplustox	Aquic Petroferric, Petroferric, Aquic Lithic, Lithic, Plinthaquic, Plinthic, Aqueptic, Aquic, Oxyaquic, Inceptic, Humic Rhodic, Humic Xanthic, Humic, Rhodic, Xanthic, Typic
Perox	Sombriperox	Petroferric, Lithic, Humic, Typic
	Acroperox	Aquic Petroferric, Petroferric, Aquic Lithic, Lithic, Anionic, Plinthic, Aquic, Humic Rhodic, Humic Xanthic, Humic, Rhodic, Xanthic, Typic
	Eutroperox	Aquic Petroferric, Petroferric, Aquic Lithic, Lithic, Plinthaquic, Plinthic, Aquic, Kandiudalfic, Humic Inceptic, Inceptic, Humic Rhodic, Humic Xanthic, Humic, Rhodic, Xanthic, Typic
	Kandiperox	Aquic Petroferric, Petroferric, Aquic Lithic, Lithic, Plinthaquic, Plinthic, Aquic, Andic, Humic Rhodic, Humic Xanthic, Humic, Rhodic, Xanthic, Typic
	Haploperox	Aquic Petroferric, Petroferric, Aquic Lithic, Lithic, Plinthaquic, Plinthic, Aquic, Andic, Humic Rhodic, Humic Xanthic, Humic, Rhodic, Xanthic, Typic
Udox	Sombriudox	Petroferric, Lithic, Humic, Typic
	Acrudox	Aquic Petroferric, Petroferric, Aquic Lithic, Lithic, Anion Aquic, Anionic, Plinthic, Aquic, Eutric, Humic Rhodic, Humic Xanthic, Humic, Rhodic, Xanthic, Typic
	Eutrudox	Aquic Petroferric, Petroferric, Aquic Lithic, Lithic, Plinthaquic, Plinthic, Aquic, Kandiudalfic, Humic Inceptic, Inceptic, Humic Rhodic, Humic Xanthic, Humic, Rhodic, Xanthic, Typic
	Kandiudox	Aquic Petroferric, Petroferric, Aquic Lithic, Lithic, Plinthaquic, Plinthic, Aquic, Andic, Humic Rhodic, Humic Xanthic, Humic, Rhodic, Xanthic, Typic
	Hapludox	Aquic Petroferric, Petroferric, Aquic Lithic, Lithic, Plinthaquic, Plinthic, Aquic, Inceptic, Andic, Humic Rhodic, Humic Xanthic, Humic, Rhodic, Xanthic, Typic

Source: Soil Survey Staff. 2010. Keys to soil taxonomy. 11th edn. USDA-NRCS. U.S. Government Printing Office, Washington, DC.

FIGURE 33.80 Distribution of Oxisols in South America.

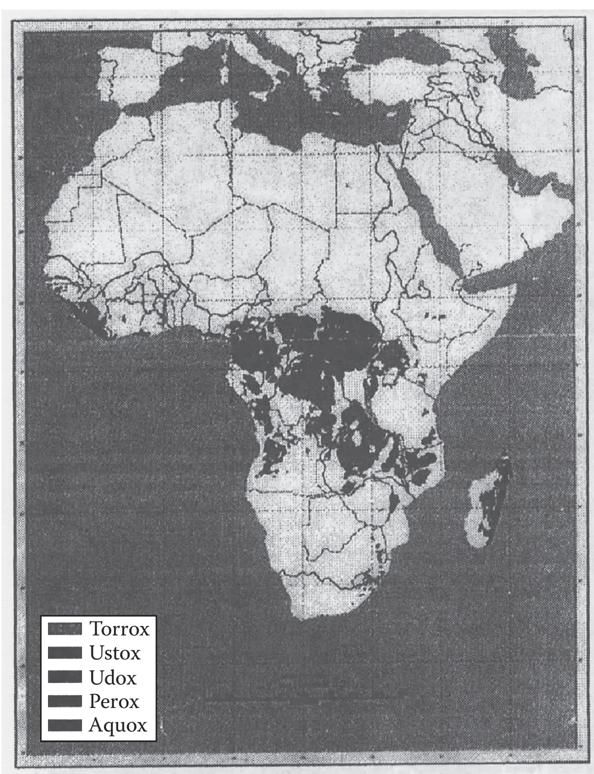

FIGURE 33.81 Distribution of Oxisols in Africa.

tropics (Table 33.2) where udox occupy 13.2% and ustox 7.9% of the land area. South America has the largest extent of Oxisols and is home to 57% of the world's Oxisols, most of them occurring in Brazil. In Africa, Zaire probably has the largest areas of Oxisols. In Southeast Asia, Oxisols only occur in small isolated areas with the largest presumed to be in Borneo (Kalimantan). In Oceania, there are small areas of Oxisols in Australia and on the islands of the Pacific and Caribbean Basins.

33.13.4.2 Extent and Geographic Distribution in the United States

Oxisols have not been officially identified in the conterminous United States and only occur on the islands of Hawaii, Guam, Micronesia, and Puerto Rico where they account for about 4.3% and 6.7% of the land area, in Hawaii and Puerto Rico, respectively (Table 33.62). Small isolated areas of Oxisols have been observed in the coastal hills of northern California formed from peridotite, but have not been established as a soil series.

33.13.5 Formation and Landscape Relationships of Oxisols

33.13.5.1 Factors of Oxisol Formation

Soil is the cumulative result of the interaction of pedogenetic processes, which are controlled by the factors of soil formation. Although the factorial approach to understanding soil formation has serious conceptual and operational limitations (Smeck et al., 1983; Wilding, 1994), it remains a unifying philosophy in pedology. The formation of Oxisols is, therefore, discussed here in the context of this paradigm.

33.13.5.1.1 Climate

With respect to rainfall, the definition of Oxisols implies two conditions. First, in the case of Oxisols with kandic horizons, there must be some time during the year when evapotranspiration exceeds precipitation as this appears to be a prerequisite for the formation of the kandic horizon (van Wambeke, 1992). Second, precipitation must exceed the capacity of the soil to retain water during some time of the year, so that water percolates through the solum (Miller, 1983). This causes the leaching of soluble weathering products and favors the residual concentration of kaolinite and sesquioxides (van Wambeke et al., 1983). The wide variety of tropical climates that meet these conditions corresponds to the udic, perudic, and ustic soil moisture regimes of *Soil Taxonomy* (Soil Survey Staff, 1998). The few Torrox that now have an aridic soil moisture regime are considered relics of a more humid climate in the past.

Regarding temperature, most Oxisol areas have mean annual air temperatures >15°C with minimal annual fluctuations and, therefore, have isothermic or isohyperthermic soil temperature regimes. A few areas in southern Brazil and South Africa may have a thermic temperature regime. The climate conducive to the formation of Oxisols is thus typically humid tropical.

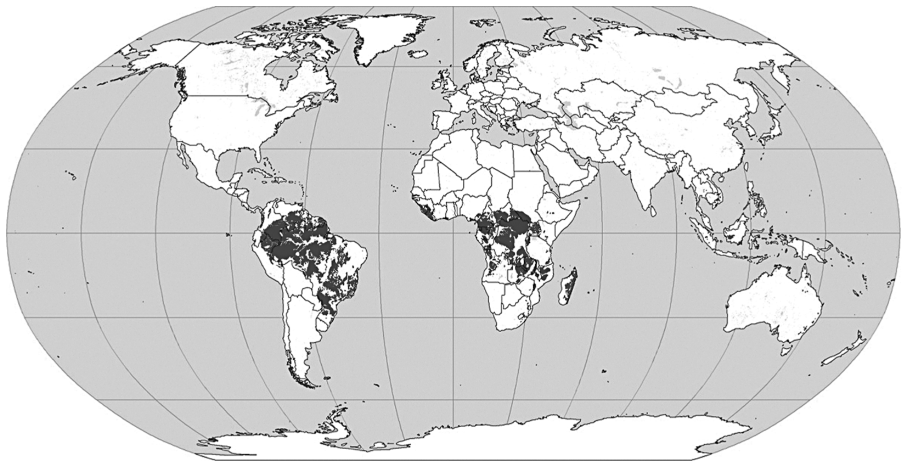

FIGURE 33.82 Global distribution of Oxisols. (Courtesy of USDA-NRCS, Soil Survey Division, World Soil Resources, Washington, DC, 2010.)

TABLE 33.62 Oxisol Series of Hawaii and Puerto Rico

Soil Series		Classification		Hawaii		Puerto Rico	
Hawaii	Puerto Rico	Subgroup	Code	Area (ha)	%	Area (ha)	%
	Delicias	Rhodic Haplustox	CCEN			280	0.03
	Moteado	Humic Haplaquox	DADD			0	0.00
Molokai		Typic Eutrotorrox	DBBC	14,143	0.84		
Molokai variant		Typic Eutrotorrox	DBBC	578	0.03		
	Matanzas	Lithic Eutrustox	DCCD			1,786	0.20
		Kandiustalfic					
Wahiawa		Rhodic Eutrustox	DCCH	8,566	0.51		
Helemano		Rhodic Eutrustox	DCCN	11,250	0.67		
Lihue		Rhodic Eutrustox	DCCN	6,300	0.38		
Niu		Rhodic Eutrustox	DCCN	1,368	0.08		
		Inceptic					
Mahana		Haplustox	DCEJ	4,918	0.29		
Makapili		Humic Haplustox	DCEM	963	0.06		
		Anionic					
Halii		Acroperox	DDBE	1,968	0.12		
Kapaa		Anionic Acrudox	DEBF	9,198	0.55		
Pooku		Anionic Acrudox	DEBF	3,154	0.19		
	Nipe	Anionic Acrudox	DEBF			896	0.10
Kunuweia		Typic Acrudox	DEBO	325	0.02		
	Cotito	Lithic Eutrudox	DECD			280	0.03
Hanamaulu		Humic Kandiudox	DEDK	3,150	0.19		
Puhi		Humic Kandiudox	DEDK	5,205	0.31		
Lawai		Hapludox	DEDN	664	0.04		
	Daguey	Typic Kandiudox	DEDN			6,393	0.72
	Zarzal	Typic Kandiudox	DEDN			0	0.00
	Rosario	Lithic Hapludox	DEED			1,158	0.13
	Los Guineos	Inceptic Hapludox	DEEH			32,553	3.67
	Limones	Humic Hapludox	DEEL			993	0.11
	Catalina	Rhodic Hapludox	DEEM			108	0.01
	Bayamon	Typic Hapludox	DEEO			9,336	1.05
	Coto	Typic Hapludox	DEEO			5,262	0.59
Total Oxisols				71,750	4.28	59,045	6.66
Total land area				1,677,308	100.00	886,216	100.00

As most Oxisols have evolved over geologic time periods, consideration of the paleoclimate is essential for understanding and analyzing their development.

33.13.5.1.2 Parent Material

A distinction must be made between transported or allochthonous and residual or autochthonous parent materials. Autochthonous parent materials have weathered in situ, whereas allochthonous materials have been transported by fluviocolluvial processes. The sediments that form the parent material of Oxisols have frequently been preweathered and may have gone through more than one weathering and transport cycle. Where Oxisol development occurs in situ, the parent material is usually saprolite, which is a highly weathered parent rock that, unless it is collapsed, still preserves most of the original rock fabrics. In terms of area, the allochthonous Oxisols are more extensive than the autochthonous (Tavernier and Eswaran, 1972). The allochthonous Oxisols on the Brazilian and Guayanian Shields (Lepsch and Buol, 1974; Lepsch et al., 1977) and the Central African Plateau (Ruhe, 1956) are formed in sediments on mid to upper Tertiary surfaces and are generally as deep as the sediments. In Hawaii, geomorphological evidence, underlying rock weathering degree, and island age and time frame of cessation of volcanic ash production suggest that these soils have formed from basaltic volcanic ash.

33.13.5.1.3 Biota

There appear to be no clear cause and effect relationships between vegetation and the geography of Oxisols as they occur under both rainforest and savannas (van Wambeke et al., 1983). Termites and ants, however, may play an important role in the formation of Oxisols. Lee and Wood (1971) studied the termite activity in soils and its effect on modifications in the solum. van Wambeke (1992) cites a French study which reports that the amount of soil that termites displace varies between 300 and 1000 kg ha^{-1} $year^{-1}$, and that their mounds typically comprise 250 m^3. Leaf cutter ants not only transport soil particles upward but also carry large amounts of plant tissue below the surface. A comprehensive account of the role of termites and the mesofauna in tropical pedogenesis has been compiled by van Wambeke (1992).

33.13.5.1.4 Relief

Oxisols occur on many landforms, including uplands, backslopes, pediments, interfluves, and river and marine terraces. The one characteristic that the loci of autochthonous Oxisols have in common is geomorphic stability over usually long periods of time. Commonly, this implies low slope gradients. Small gradients, however, are not necessarily an indication of stability or geomorphic age as a level surface may be a recent floodplain or a Tertiary peneplain.

33.13.5.1.5 Time

The time required for the formation of Oxisols obviously depends on the other pedogenetic factors, notably the weatherability of the parent rock, the composition and weatherability of transported parent materials, and the climate and its fluctuations over time. There is much scope for variability in these conditions, and consequently, in the age of Oxisols.

For Oxisols developed in transported sediments, their geologic age determines the actual maximum time available for soil formation. In Puerto Rico, for example, Oxisols have developed in preweathered materials on marine terraces of Quaternary age, but the autochthonous Oxisols of the interior commonly occur on Pliocene or Miocene surfaces that have been exposed to subaerial weathering for as long as 15 million years (Beinroth, 1981). Oxisol landscapes of comparable age have been reported in Africa, South America, and Australia (Miller, 1983). Oxisols in Hawaii can be less than 1 million years old.

33.13.5.2 Processes of Oxisol Formation

Various processes combine to produce, either concurrently or sequentially, the unique features of Oxisols. Prominent among these processes are those that (1) lead to the intensive weathering of primary minerals and the removal of soluble weathering products, the formation of 1:1 lattice clays, and the residual accumulation of sesquioxides, and (2) cause an increase of silicate clay in subsurface horizons. The first process, collectively known as laterization, is of paramount importance in the formation of most Oxisols. It involves desilication, ferrallization, ferritization, and allitization and causes the chemical migration of silica out of the solum and the relative concentration of sesquioxides in the soil. Formation of Fe coatings or aggregation at the expense of quartz grains is a common process in the tropical environment (Padmanabhon and Mermut, 1996).

Clay increase with depth that is diagnostic for the kandic horizon may result from (1) the process known as lessivage that causes clay illuviation and results in a clay maximum or bulge in the subsoil; (2) vertical downward translocation of clay without accumulation in an illuvial horizon. French pedologists refer to this process that leads to lighter textured surface horizons as appauvrissement or impoverishment (van Wambeke, 1992); (3) clay depletion in the soil surface may be caused by the selective removal of fine particles from the surface soil by erosion or mesofauna; (4) in Oxisols that are seasonally flooded, ferrolysis can cause the destruction of clay in the topsoil (Brinkman, 1970); (5) as postulated by Simonson (1949), in situ formation of clay in the B horizon may occur; and (6) lithological discontinuities may account for textural changes in the solum. In the past, the process of podzolization has been associated with Oxisols, but the evidence for it is intangible.

The formation of plinthite is often considered an extreme manifestation of laterization. The Soil Survey Staff (2010) characterizes plinthite as an Fe-rich, humus-poor mixture of sesquioxides, clay, quartz, and other diluents that commonly appear as dark red mottles in platy, polygonal, or reticulate patterns and generally forms in a horizon that is saturated with water for some time during the year. In a moist soil, plinthite is soft enough to be cut with a spade, but it changes irreversibly to ironstone or petroplinthite when exposed to repeated wetting and drying. Plinthite (Eswaran and Raghumohan, 1973) is a diagnostic feature of many soils which are hydromorphic or have gone through a hydromorphic phase during their evolution. Hardening of plinthite takes place slowly when the ground water table is lowered and the soil surface ground cover is removed. If the surface soil horizons are eroded, the underlying plinthite is exposed and hardens rapidly to form petroplinthite (Sys, 1968; Eswaran and Raghumohan, 1973). A related feature is the development of a petroferric contact, which is an abrupt boundary between soil material and an underlying layer of cemented petroplinthite gravel that is hard and impermeable to both roots and water. The definition, genesis, and kinds of plinthite and petroplinthite have been discussed in detail by van Wambeke (1992). The upland soils of Palau and Yap, where eroded and degraded, exhibit surface layers dominated by irregular jagged and vesicular ironstone gravel and gibsite pendants. The uneroded, nondegraded adjacent soils however do not exhibit plinthite (Smith, 1983).

Gleization, which is a process of importance in the formation of some Oxisols, refers to the reduction of Fe and Mn under seasonally anaerobic soil conditions and produces bluish to greenish gray matrix colors with or without yellowish brown, brown, or black mottles and ferric and manganiferous concretions. These redoximorphic features are striking characteristics of the aquic subgroups.

The transformation of raw organic material into soil organic matter (SOM) known as humification occurs in all Oxisols, but is of particular importance in the humic and histic subgroups. A related process is the illuviation of humus that results in the formation of the sombric horizon which is a dark colored subsurface horizon found in soils on old geomorphic surfaces of Central Africa and parts of South America (Eswaran and Tavernier, 1980). Although its origin and genesis are still being debated, it is used as a diagnostic horizon in *Soil Taxonomy* because it is a distinctive feature in an otherwise nondescript soil (Eswaran et al., 1986). Oxisols with sombric horizons are restricted to the cool high plateaus of the tropics at altitudes between 1400 and 3000 m above sea level that have isothermic or colder temperature regimes and a udic soil moisture regime (van Wambeke, 1992).

In summary, a broad range of environmental determinants, and pedogenetic processes and mechanisms may be involved in the formation of Oxisols. Yet, there is no single set of formative factors, processes, and mechanisms that could account for the formation of all Oxisols. The fact that the causative conditions may not have been the same or may have operated at different intensities over time, and may have occurred simultaneously or sequentially, adds complexity to Oxisol genesis.

33.13.5.3 Landscape Relationships

The geomorphic evolution of the landscape is an important factor that is more important in the formation of Oxisols than in many other kinds of soil. As pointed out by Daniels et al. (1971), the occurrence of Oxisols, as that of other soils, is controlled by the interaction of geomorphic and other formative factors and the resulting rates and degrees of expression of pedogenic processes. Beinroth et al. (1974), Lepsch and Buol (1974), and Lepsch et al. (1977) provide illustrative examples of landscape relationships of Oxisols in Hawaii and Brazil that invariably show that Oxisols occupy geomorphic positions that are older and more stable than the surfaces where Ultisols and Inceptisols occur with which they are geographically associated. However, the recent introduction of the kandic horizon, which may be diagnostic for both Ultisols and Oxisols, has blurred the geomorphic boundary between the two orders. In the case of some Hawaiian Ultisols, a truncated oxidic paleosol is overlain by oxidic depositional material of ash origin. Although no significant clay percent differences are present between the two materials, the presence of clay films on the paleosol results in the Ultisol classification but physical and chemical properties and behavior are those of an Oxisol.

33.13.6 Properties

33.13.6.1 Macromorphology

Compared with the often strikingly horizonated soils of other orders, the field morphology of most Oxisols is visually rather uniform. They nevertheless have some distinguishing attributes. Color is a prominent feature of Oxisols. The surface horizon of most lowland Oxisols is a thin, reddish to yellowish colored ochric epipedon. Oxisols at high elevations (>1000 m) frequently have a dark-colored, humus-rich surface horizon (Ruhe, 1956), which may qualify for a mollic or umbric epipedon. Many Oxisols of Central Africa also have a dark-colored layer, the sombric horizon, in the subsoil. As Table 33.63 indicates, subsurface colors range from light gray in the Aquox to various hues of red in the upland Oxisols and are generally a function of the Fe content of the original material or rock (Eswaran and Sys, 1970). Color is also related to the kind of Fe minerals, with goethite producing yellow colors and hematite red colors. Presence of colloidal organic matter darkens the soil. Although organic carbon may be 4% or more, iron oxides dominate the color leading to the typical designation of an ochric epipedon. If there is a fluctuating groundwater table, mottles or plinthite may form in the oscillation zone. If the soil remains saturated with water for long periods, reduction and removal of the Fe results in a whitish horizon.

The texture of Oxisols may vary from sandy loam to clay. A characteristic feature is that the macro structural elements are weak in the subsoil. When the soil is gently pressed between the thumb and forefinger, the material collapses or fails abruptly. This is probably a good field indicator for an oxic horizon. In surface layers with higher clay contents, macrostructure can be moderate to strong, fine to medium subangular blocky. Excessive tillage however can reduce this condition to a seemingly single grain appearance of silt and sand sized microaggregates.

Many Oxisols have stone lines with the stones being quartz or petroplinthite gravel (Ruhe, 1956). The stone line is a mark of a lithologic discontinuity indicating that the material above

TABLE 33.63 Physical Properties of Selected Pedons Representing Oxisol Suborders

Classification	Depth (cm)	Horizon	Bulk Density (g cm^{-3})	Water Retention 1/3 bar (%)	Water Retention 15 bar (%)	WRD	Particle Size (%) Sand	Particle Size (%) Silt	Clay	Soil Color
Typic Acraquox (Brazil)	0–10	A1	1.3	32.5	26.9	0.1	33.1	10.5	56.4	10YR 6/1
	10–30	Ag	1.4	21.8	17.2	0.1	34.5	11.6	53.9	10YR 7.1
	30–48	Bog1	1.3	28.7	21.1	0.1	25.1	8.8	66.1	10YR 7.1
	48–77	Bog2	1.3	29.4	23.1	0.1	44.5	13.5	42.0	10YR 8.2
	77–90	Bov	1.4	27.3	21.9	0.1	62.3	11.8	25.9	10YR 5/8
Typic Eutrotorrox (Hawaii)	0–23	Ap			21.5		17.1	40.2	42.7	2.5YR 2/4
	23–50	Bo1	1.3	29.6	21.9	0.1	10.4	4.17	47.9	2.5YR 3/4
	50–87	Bo2	1.4	28.2	22.1	0.1	24.1	30.1	45.8	2.5YR 3/4
	87–123	Bo3	1.4	28.9	21.6	0.1	18.8	34.7	46.5	2.5YR 3/4
	123–150	Bo4	1.3	30.9	20.4	0.1	11.2	39.9	48.9	5YR 3/3
Humic Rhodic Eutrustox (Brazil)	0–25	Ap			23.8		18.0	41.2	40.8	2.5YR 3/2
	25–40	AB	1.2	32.4	23.3	0.1	16.8	37.9	45.23	2.5YR 3/2
	40–64	Bo1	1.2	30.8	24.0	0.1	10.9	25.1	64.0	2.5YR 3/2
	64–110	Bo2	1.1	32.2	24.6	0.1	13.3	28.4	58.3	2.5YR 3/2
	110–210	Bo3	1.1	31.4	24.6	0.1	18.5	38.1	43.4	2.5YR 3/2
Typic Kandiperox (Indonesia)	0–10	Ap1	0.9	42.0	26.5	0.1	13.2	28.9	57.9	5YR 3/3
	10–21	Ap2	0.9	39.5	26.6	0.1	11.4	28.7	59.9	5YR 4/3
	21–51	Bo1	1.0	48.0	31.8	0.2	7.0	20.7	72.3	5YR 3/4
	51–81	Bo2	0.9	50.9	32.6	0.2	6.1	18.5	75.4	5YR 3.4
Anionic Acrudox (Puerto Rico)	0–28	A1	1.1	35.4	26.5	0.3	9.2	36.3	53.8	2.5YR 2/4
	28–46	B1	1.2	26.7	22.8	0.2	7.4	34.9	54.5	2.5YR 2/4
	46–71	Bo1	1.1	34.4	24.8	0.4	9.8	30.6	57.7	7.5YR 3/8
	71–97	Bo2	1.3	35.7	25.9	0.4	23.3	21.0	59.6	7.5YR 3/8
	97–120	Bo3	1.4	31.6	26.4	0.2	17.0	23.3	55.7	7.5YR 3/4
	120–155	Bo4	1.3	29.8	24.5	0.1	19.2	27.2	59.7	7.5YR 3/4
Humic Sombriudox (Rwanda)	0–15	A	1.0	26.5	11.3	0.2	58.2	8.6	33.2	7.5YR 3/3
	15–40	B1	1.4	15.4	11.1	0.1	56.3	9.3	34.4	7.5YR 2/4
	40–66	Bo1	1.3	17.9	12.4	0.1	52.6	8.9	38.5	7.5YR 2/4
	66–91	Bh1	1.3	21.6	15.2	0.1	42.8	8.4	48.8	7.5YR 2/4
	91–121	Bh2	1.3	22.7	16.7	0.1	43.2	7.1	49.7	7.5YR 2/4
	121–150	Bo2	1.4	19.9	15.4	0.1	42.7	7.3	50.0	7.5YR 2/4

Sources: Soil Survey Staff. 2011. National Cooperative Soil Characterization Database. Available online at http://ssldata.nrcs.usda.gov. Accessed July 8, 2011.

the line was deposited or formed at a different period than the material below. Stone lines frequently suggest that the soils are formed on transported deposits and point to the allochthonous nature of the material (van Wambeke, 1992). Some Oxisols have multiple stone lines.

33.13.6.2 Mineralogy and Micromorphology

The unique physical and chemical properties of Oxisols result mainly from the mineralogical composition of the colloidal fraction (Uehara and Keng, 1975; Herbillon, 1980). To have a CEC < 16 $cmol_c\ kg^{-1}$ clay by NH_4OAc at pH 7.0, the clay fraction must be dominated by low activity clays such as kaolinite. Iron oxyhydroxide minerals, such as goethite, hematite, and ferrihydrite, are usually associated with kaolinite, but in some Oxisols, the Fe minerals predominate (Jones et al., 1982), particularly in soils belonging to ferruginous or ferritic families. The Fe minerals can have a high positive charge (Jones and Uehara, 1973), which accounts for the special physical and chemical properties discussed later.

Gibbsite is present as a secondary mineral in many Oxisols (Eswaran et al., 1977). Weathering of primary minerals releases Si and Al; Si is lost in the soil solution while the Al crystallizes as gibbsite, mostly as nodules. The gibbsite crystals in the nodules are well crystallized as shown in the SEM micrograph taken at a magnification of ×2500 (Figure 33.83). The crystals are euhedral and twinning is common. Typically, gibbsite crystals have the size of fine silt, although dense pendants of almost pure gibbsite with the morphology of the root channels they formed in can be observed in highly degraded Oxisols of Palau (the Babelthuap Series). Oxisols usually have more gibbsite in the silt than in the clay fraction. It is for this reason that gibbsitic families are defined on the basis of the amount of gibbsite in the fine earth (<2 mm) fraction.

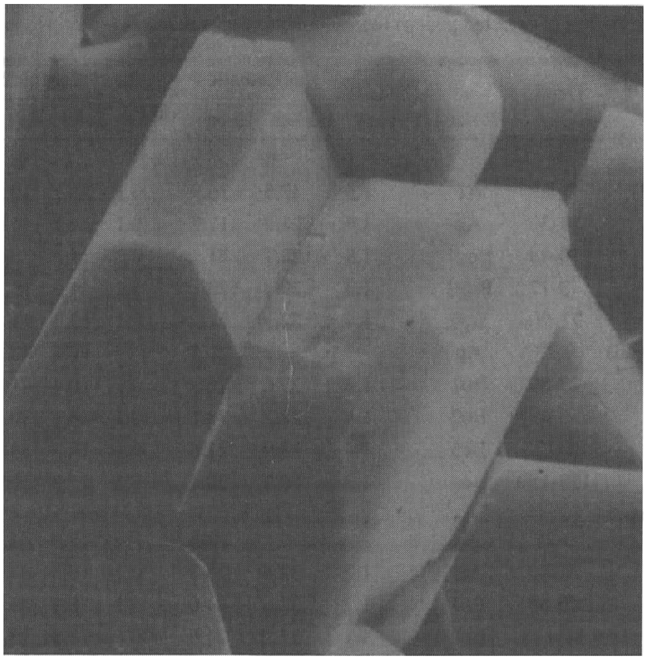

FIGURE 33.83 Scanning electron micrograph of a gibbsite nodule in a Gibbsiudox from Malaysia; the gibbsite crystals are typically euhedral and fine and medium silt size; magnification × 10,000.

The other frequent mineral in Oxisols is goethite (Eswaran et al., 1978). Plinthite, laterite, or petroplinthite frequently exhibit characteristic forms of goethite aggregates in thin sections as illustrated by the SEM micrograph (×10,000) in Figure 33.84. The goethite crystals have a typical lenticular shape and appear welded together, which gives the petroplinthic material its strength. Goethite and hematite have different habits and show different crystal forms.

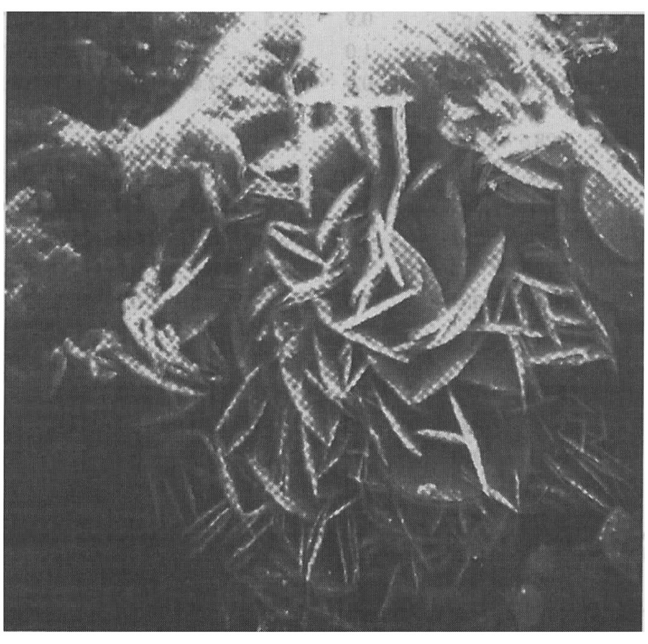

FIGURE 33.84 Scanning electron micrograph of a geothite in a laterite fragment. The crystals have a lenticular habit with split edges; laterite sample from Angadipuram, Kerala, India, which is the type locality of the Buchanan lateriate; magnification × 25,000.

In most Oxisols, the fabric is homogenous without too many specific entities like those illustrated previously (Buol and Eswaran, 1978). In some oxic horizons, a thin lining of ferriargillans (yellow coatings on the void walls) may be present. Ultrathin sections under TEM suggest that the combination of random orientation of clay particles, organic matter, and aggregates of Fe-bearing minerals accounts for the isotropic nature of the aggregates under a petrographic microscope (Santos et al., 1989). The presence of the clay skins is evidence of the transitional nature of the soil, and that clay illuviation and accumulation was an important process. The transitional nature is indicated by the kandi prefix in the soil name.

Unless eroded, the surface horizons of Oxisols have a relatively high organic matter content with high biological activity including the presence of fungal hyphae and fruiting bodies, which are generally indicators of good soil quality. Typical for the drier tropics is the presence of large termite nests, which can reach 5 m in height. Some species of termites are subsoil dwellers and their galleries may extend several meters into the soil. Bioturbation of the soil is, therefore, an important soil-forming process in tropical soils.

Oxic horizons have a friable consistency. When a large soil clod is gradually crushed in the hands, the material disaggregates and small rounded bodies become evident. These features, which are only observed in Oxisols, have been referred to as pedovites or soil eggs by Belgian pedologists and "earthy lumps" in Hawaiian pedon descriptions. The excellent and stable structure of Oxisols, and their high macroporosity resulting in moderate to moderately rapid infiltration rates, make them resistant to erosion. Where intense tillage has reduced surface structure to microaggregates and has created tillage pans, erosion rates can be dramatically increased.

33.13.6.3 Chemistry and Physics

The chemistry and physics of Oxisols are inextricably linked to the surface charge characteristics of minerals in the clay fraction. Unlike in most other soils, the surface charge of minerals in Oxisols is pH dependant and varies in magnitude and sign. Oxisols, however, are not the only soils dominated by variable charge systems. Andisols, Ultisols, and Histosols can be even better examples of soils with variable surface charge, so that the attribute that makes Oxisols unique is the low permanent negative charge of the silicate clay minerals in the clay fraction. Important physical and chemical properties of selected pedons representing Oxisol suborders are summarized in Tables 33.63 and 33.64.

The sign of the charge is readily established by the sign of the difference in pH (ΔpH) in the following equation:

$$\Delta pH = pH_{KCL} - pH_{water}$$

Oxisols with a positive ΔpH are rare, but not difficult to find if one knows where to look for them. They are rare because

TABLE 33.64 Chemical Properties of Selected Pedons Representing Oxisol Suborders

		pH				OC	Free Fe	Charge ($cmol_c$ kg^{-1})			Base Saturation (%)		Al Sat. (%)
Classification	Horizon	pH_o	H_2O	KCl	ΔpH	(%)	(%)	ECEC	CEC 7EC 8.2	CEC 8	CEC 7	CEC 8.2	ECEC
Typic Acraquox (Brazil)	Al	4.3	4.8	4.5	−0.3	2.4	0.3	1.9	6.8	13.6	3.0	1.0	89.0
	Ag	4.3	4.9	4.6	−0.3	1.6	0.2	1.2	4.8	10.1	4.0	2.0	83.0
	Bogl	5.3	5.5	5.4	−0.1	0.9	0.4	0.1	2.0	6.3	5.0	2.0	0.0
	Bog2	6.8	6.0	6.4	0.4	0.6	1.3	0.2	1.4	5.3	14.0	4.0	0.0
	Bov	6.8	6.0	6.4	0.4	0.4	1.4	0.1	1.0	4.9	2.0	1.0	0.0
Typic Eutrotorrox (Hawaii)	Ap	5.9	6.7	6.3	−0.4	1.0	12.2	8.9	9.4	16.1	95.0	55.0	0.0
	Bo1	6.1	6.5	6.3	−0.2	0.7	11.9	8.1	8.2	14.8	99.0	55.0	0.0
	Bo2	6.3	7.5	6.9	−0.6	0.2	9.9	8.9	9.4	14.9	95.0	60.0	0.0
	Bo3	6.2	6.8	6.5	−0.3	0.2	8.6	9.3	10.2	14.9	91.0	62.0	0.0
	Bo4	6.0	7.0	6.5	−0.5	0.3	8.5	7.9	8.3	13.2	95.0	60.0	0.0
Humic Rhodic Eutrustox (Brazil)Ap	Ap	5.4	6.6	6.0	−0.6	2.8	14.3	15.8	17.6	26.7	90.0	59.0	0.0
	AB	5.9	6.5	6.2	−0.3	2.2	14.7	13.7	15.4	23.6	89.0	58.0	0.0
	Bo1	5.4	6.8	6.1	−0.7	1.2	14.3	8.0	9.0	15.9	89.0	50.0	0.0
	Bo2	5.7	6.9	6.3	−0.6	0.9	14.9	6.1	6.6	13.5	92.0	45.0	0.0
	Bo3	5.9	7.1	6.5	−0.6	0.5	14.8	4.1	4.3	11.1	95.0	35.0	0.0
Typic Kandiperox (Indonesia)	Ap1	4.1	4.9	4.5	−0.4	2.0	5.6	5.9	16.9	25.5	30.0	20.0	15.0
	Ap2	3.8	4.8	4.3	−0.5	1.5	5.7	4.4	15.4	22.9	15.0	10.0	48.0
	Bo1	4.4	5.2	4.8	−0.4	1.9	5.6	5.4	15.1	23.4	34.0	22.0	6.0
	Bo2	4.9	5.3	5.1	−0.2	0.5	5.8	6.0	14.3	21.3	42.0	28.0	0.0
Anionic Acrudox (Puerto Rico)	Al	3.5	5.1	4.3	−0.8	6.0	13.0	7.9	25.4	34.8	11.0	8.0	17.7
	Bl	3.8	5.0	4.4	−0.6	2.0	12.9	1.7	12.1	21.5	1.0	0.0	52.9
	Bo1	4.4	5.0	4.7	−0.3	1.3	16.5	0.0	8.2	15.7	0.0	0.0	0.0
	Bo2	6.2	5.2	5.7	0.5	0.9	19.2	0.0	6.4	12.8	0.0	0.0	0.0
	Bo3	6.7	5.5	6.1	0.6	0.7	23.1	0.2	5.3	12.1	2.0	0.0	0.0
	Bo4	7.1	5.7	6.4	0.7	0.6	25.7	0.0	3.8	12.8	0.0	0.0	0.0
Humic Sombriudox (Rwanda)	A	3.3	4.5	3.9	−0.6	1.7	1.2	3.8	10.0	13.7	6.0	4.0	84.0
	Bl	3.5	4.5	4.0	−0.5	1.3	1.4	3.0	7.7	12.3	4.0	2.0	90.0
	Bo1	3.4	4.6	4.0	−0.6	1.2	1.8	3.8	9.6	14.6	5.0	3.0	87.0
	Bh1	3.4	4.6	4.0	−0.6	1.4	2.2	5.3	12.3	19.9	7.0	5.0	83.0
	Bh2	3.2	4.6	3.9	−0.7	1.6	2.2	5.9	15.5	25.4	5.0	3.0	88.0
	Bo2	3.2	4.6	3.9	−0.7	1.1	1.8	4.7	11.6	17.3	5.0	3.0	87.0

Sources: Soil Survey Staff. 2011. National Cooperative Soil Characterization Database. Available online at http://ssldata.nrcs.usda.gov. Accessed July 8, 2011.
ΔpH, pH in KCl minus pH in H_2O; pH_o, pH of soil at the zero point of charge where positive and negative charges are equal; OC, organic carbon content.

positive ΔpH values almost always occur in the subsoil. They rarely or almost never occur in surface horizons because negatively charged SOM masks the positive charge in the mineral–organic mixture. Organic matter, like the mineral fraction in the oxic horizon, has variable charge characteristics. The difference lies in their respective points of zero charge (p.z.c.) being <pH 3 for organic matter and >7 for oxic materials. Since soil pH values rarely fall below 3, SOM is net negatively charged in most soils.

The p.z.c. for material in the oxic horizon is highly variable ranging from pH 3–6. As a rule, it increases as organic C decreases and the silica/sesquioxide ratio of the clay fraction decreases. Hematite (Fe_2O_3), for example, has a p.z.c. of 8.5 (Parks and de Bruyn, 1962). In this sense, Oxisols are products of desilication, and the end products of desilication are the oxides and hydrous oxides of Fe and Al.

Over 82 years ago, Mattson (1928, 1932) showed that p.z.c. increased as the silica/sesquioxide ratio decreased. He also showed that the pH of the gel shifted toward the p.z.c. on leaching with distilled water, which he called isoelectric weathering. This concept is useful in explaining the chemical and physical behavior of Oxisols and related soils dominated by low activity clays. This concept is best illustrated with an example, and one of the best examples is the Nipe soil of Puerto Rico and Cuba (Anionic Acrudox in Table 33.61). (Data is for the original Type location in Puerto Rico. This site has been urbanized. The new type location (data not presented) classifies as a Typic Acrudox. The isoelectric properties of the Nipe soil are mainly determined by the p.z.c., which is measured as pH_o (pH at which positive and negative charges on variable charge surfaces are equal).

When ΔpH = 0, pH values measured in M KCl and water are identical indicating that when the material is net negatively

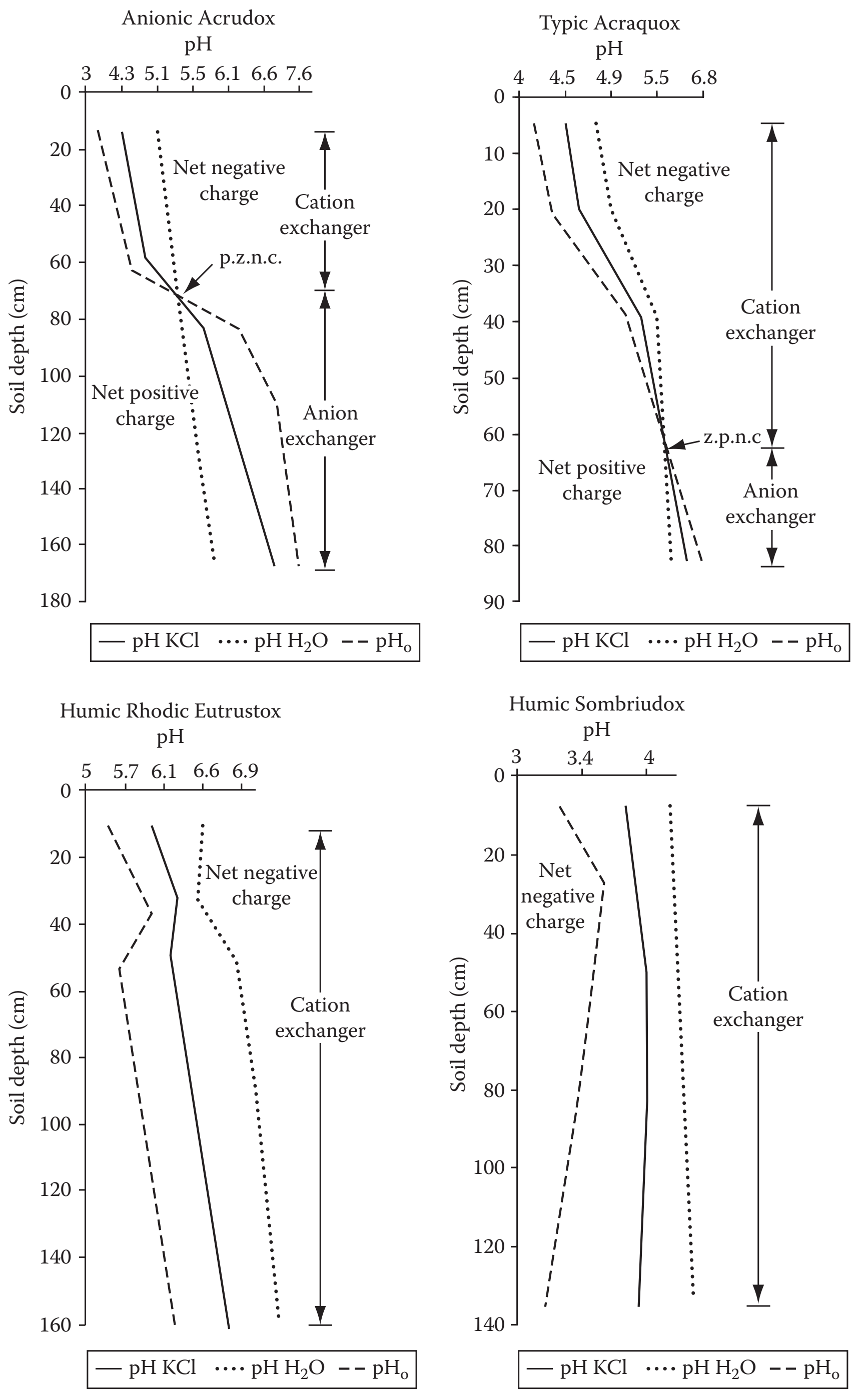

FIGURE 33.85 Surface charge characteristics of Oxisols of different suborders.

charged, pH_o is always lower than pH_{KCl} and vice versa. This does not apply to materials with significant amounts of permanent charge minerals.

The Nipe soil, like all Acric Oxisols, is a cation exchanger in the surface horizon and an anion exchanger in the subsoil. Charge characteristics of the Nipe soil and three other Oxisols are illustrated in Figure 33.85. The anion exchange capacity (AEC) of the subsoil can lead to unexpected consequences. In Hawaii, for example, 3–11 Mg NO_3–N ha^{-1} have been measured in the subsoil and deep saprolite underlying Oxisols and Ultisols (Deenik, 1997), which explains the low NO_3^- levels in the groundwater underlying these soils even after nearly a century of intensive farming. Although the NO_3^- remains trapped above the water table, pesticides banned decades ago continue to enter the groundwater. Had the soil minerals been of the permanent charge type, the NO_3–N would have reached the groundwater many years ago.

In cation impoverished Oxisols such as the Nipe soil, lime is often added as a Ca fertilizer rather than an amendment.

But because lime raises pH and increases negative charge, Ca^{2+} ions remain in the limed layer and do not move to the impoverished subsoil. To circumvent this problem, gypsum and magnesium sulfate are favored over calcite or dolomite whenever the aim is to raise subsoil Ca and Mg. The factors determining movement of surface-applied amendments in variable charge soils are discussed in detail by Sumner (1995).

The acidic organic matter near the surface and the basic oxides in the subsoil determines, in essence, the chemical properties of the Nipe soil. In most Oxisols, the desilication process has not progressed as far as in the Nipe series. In such instances, the subsoil pH will be lower than that of the Nipe soil because the p.z.c. will be lowered by the higher Si content.

An opposing process that counteracts desilication is humification of the desilicated weathering products. Humic acids, like silicic acid, have a p.z.c. below 3. At pH levels normally encountered in soils, humus and silica are net negatively charged, so that they have strong affinities for positively charged sesquioxides. In one sense, organic matter adsorption on oxide surface has nearly the same effect as resilication of the oxides (Uehara, 1995). The surface of quartz is chemically similar to silicic acid, but its low specific surface renders it virtually inert. Desilicated Oxisols can be rejuvenated by additions of soluble silicates. Large crop responses to additions of calcium silicate above those obtained from similar lime applications have been reported (Plucknett, 1971).

The moisture characteristics curves of well-aggregated Oxisols show two major desorption zones. Water in the interaggregate pores drains rapidly between 0 and −0.01 MPa. Another desorption zone occurs when the intraaggregate pores begin to drain at about −15 MPa (Sharma and Uehara, 1968a, 1968b), which results in a bimodal pore size distribution (Tsuji et al., 1975; Bui et al., 1989). Some have referred to Oxisols as behaving like aggregated sands because of their stable aggregates and high macroporosity. This description tells only half the story. The aggregates that remain nearly water saturated beyond the wilting point impart additional properties to Oxisols. The water-saturated aggregates increase volumetric heat capacity and lower thermal diffusivity. Crops such as pineapple respond to practices that raise subsoil temperature. In Hawaii, plastic sheets used to increase the effectiveness of soil fumigants also aid the crop by raising subsoil temperature (Ekern, 1967). The intraaggregate water retained at high negative pressures also affects tillage operations. This intraaggregate water is freed under the shearing action of tillage implements and causes the soil to adhere to the implement. Some farmers solve this problem by bolting a teflon sheet onto the implement's shearing surface.

In summary, the physics and chemistry of Oxisols are strongly influenced by the extent to which desilication has occurred. Basic and ultrabasic parent materials readily desilicate and produce soils that approach the central concept of Oxisols. Desilication is also aided by warm and humid conditions, so that when Oxisols occur outside the tropics, they are almost always associated with ultrabasic rocks.

33.13.7 Conclusions

Oxisols occupy about 25% of the land area of the tropics where they are the single most extensive soil type. Yet, historically they have been perceived as agriculturally unproductive and problematic for management; under low input agriculture, yields are in fact low, risk is high, and the potential for land degradation is also high (Sanchez and Salinas, 1981). The negative notion of the poor agronomic performance derives substance from the inherent chemical constraints of Oxisols, which include, to varying degrees, a low nutrient retention capacity, anion adsorption, Ca deficiency, and Mn and Al toxicity (van Wambeke, 1974; Sanchez, 1976). It is also true that the inputs required to correct these constraints may be economically prohibitive for many farmers, and in places where better endowed soils are available, the Oxisols are, therefore, at a distinct comparative disadvantage (van Wambeke, 1992). Nevertheless, with science-based management that employs the tools and techniques of modern agriculture, Oxisols can be managed to be both economically and sustainably productive. Consequently, no less an authority than Charles E. Kellogg stated that "some day the most productive agriculture of the world will be mostly in the tropics, especially in the humid parts" (Kellogg, 1967). Presumably, this assessment is based on the realization that the favorable physical attributes of Oxisols outweigh their chemical limitations. While the latter can be amended with purchased inputs, good soil structure cannot be bought. That said, excessive tillage can destroy macro aggregation.

Oxisols are the dominant soils in the humid tropical forest ecosystem, a pristine environment that constitutes an enormous reservoir of sequestered carbon and unique ecological niches of great biological diversity, in addition to being a resource for food, timber, medicine, and other products for people. Yet, the resource-poor farmers invading these areas practice shifting cultivation, and slash and burn agriculture has become the most extensive form of agriculture in the tropics. As a result, over 15 million ha of forests are being burned annually, and some plants and animals are lost permanently when their habitats are destroyed. Moreover, the resilience of these ecosystems is so low that complete regeneration may not be achieved.

Oxisols constitute a major land resource and one of the few remaining frontiers for agricultural development, particularly in Africa and South America, and also support the largest areas of tropical forests. Their conversion to agricultural land will be at the expense of the forest with the concomitant loss of biodiversity and negative impact on global climate. It is imperative, therefore, to develop viable alternatives to traditional agricultural systems. If this challenge can be met successfully, the rewards are not only to provide a means for millions of people to extricate themselves from poverty, but also to ensure the survival of the tropical forests. Landuse policies guided by an understanding of the nature, properties, and ecological functions of Oxisols are critical to sustain the integrity and productivity of these land resources.

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34

Land Evaluation for Landscape Units

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34.1 Introduction

Sustainable development has become a key goal of various international agencies, and the role of the land in achieving this is increasingly being recognized. More efficient land use is needed to, for example, feed 9 billion people in 2050, to combat climate change, and to enable sustainable biofuel production (e.g., Hartemink and McBratney, 2008). Studies have explored the potential world food supply and associated environmental quality issues, among them land degradation (Penning de Vries et al., 1995). At another scale, new farming systems are being developed to optimize fertilizer and biocide use, applying modern information technology, summarized as precision agriculture (PA; CIBA Foundation, 1997; Bouma et al., 1997a; Stoorvogel et al., 2004c). Regional planners consider alternative land use scenarios combining agricultural production with nature conservation and other functions such as water storage (Bouma et al., 2008). In all cases, land use is the core issue to be studied, while socioeconomic considerations often play a crucial role as well. Consideration of actual and potential land use as a function of land properties fits under the broad umbrella of land evaluation as advocated by FAO (1976, 1983). Land evaluation can be realized with descriptive, qualitative methods but increasingly includes quantitative simulation models for crop production, land use, and environmental impacts in combination with geographic information systems. All these methods need to be fed with adequate soil data, often derived from soil surveys (e.g., Bouma, 1994). This sometimes occurs mechanistically using existing databases with little attention to natural soil dynamics or landscape relationships, and this may lead to unrepresentative results. Soil survey has much to offer, but this expertise has to be applied and presented more effectively than at present. Soil survey interpretations in soil survey reports present broad possibilities and limitations for a variety of land uses for each mapping unit. However, these represent rather narrowly defined land uses as compared with land evaluation that enables the actual assessment of land performance. Aside from the variation in space, there is also the need to consider variation in time, be it days, growing seasons, or decades. Increasingly, land evaluation is realized in close interaction with the stakeholders, ranging from farmers to planners and politicians. Innovative developments in land evaluation, emphasizing interdisciplinary approaches at different spatial and temporal scales, will be discussed in this chapter.

34.2 Land Evaluation: The FAO Framework

34.2.1 Definitions

Land evaluation is defined as follows:

> the process of assessment of land performance when used for specified purposes, involving the execution and interpretation of surveys and studies of land use, vegetation, landforms, soils, climate and other aspects of land in order to identify and make a comparison of promising kinds of land use in terms applicable to the objectives of the evaluation.
>
> **(FAO, 1976)**

Clearly, evaluating the performance of land has been the topic of many studies in the past in different disciplines. There is, for instance, a large amount of literature on land use planning from a regulatory and social perspective, but here the focus will be on agroecology as related to sustainable development. Adopting the international FAO approach has the advantage of using a widely known procedure in which definitions are so broad that they offer many opportunities for expansion and modification. Two elements stand out when considering the definition of land evaluation: (1) Performance can only be assessed when a specific land use has been defined. In other words, judgment cannot be made as to performance in general but only for specific types of land use. Land may, for instance, function quite well as a campground but poorly when growing a wheat crop. (2) Attention is paid not only to current land use but also to potential forms of land use, which may be more or less promising depending on the objectives of the evaluation.

Land is defined as follows:

> an area of the Earth's surface, the characteristics of which embrace all reasonably stable, or predictably cyclic attributes of the biosphere, vertically above and below this area including those of the atmosphere, the soil and underlying geology, the hydrology, the plant and animal populations and the results of past and present human activity to the extent that these attributes exert a significant influence on present and future uses of the land by man.
>
> **(FAO, 1976)**

A land unit is defined as follows:

> an area of land possessing specified land characteristics and/or land qualities which can be demarcated on a map. Land is often represented as a georeferenced land unit on soil maps, which present additional data on climate in a soil survey report.
>
> **(FAO, 1983)**

The broad term "land use" is specified in terms of land utilization types (LUTs), which define a particular type of land use in varying degrees of detail. In the context of rainfed agriculture, an LUT refers to a crop, crop combination, or cropping system with a specified technical and socioeconomic setting (FAO, 1983). When combining the Land Unit with the LUT, one obtains the so-called land use system (LUS) defined as a specified LUT practiced on a given land unit and associated with inputs, outputs, and possible land improvements (FAO, 1983). Previous work in Costa Rica has defined such LUSs also in terms of the type of technology (T) being used in each particular production system. They refer, therefore, not to a LUS but to a LUST, an example of which is provided in Table 34.1.

A key element in land evaluation is the assessment of land performance. This is done, in principle, by comparing the requirements of a particular type of land use with what the land has to offer. When the two match, the land is suitable for a particular LUST. When they do not match, suitability is less to varying degrees. Land suitability is correspondingly defined as follows: *the fitness of a given type of land for a specified kind of land use* (FAO, 1976). This matching process, which is central in land evaluation, is handled by defining land qualities and land characteristics. Land qualities are *complex attributes of land that act in a manner distinct from the actions of other land qualities in its influence on the suitability of land for a specified kind of use*. The matching process is realized by expressing both land use requirements and what the land has to offer in terms of land qualities and by comparing the two expressions.

Although this is not mentioned by FAO (1976), land qualities usually cannot be measured directly. Examples are the moisture supply capacity, the workability, and the trafficability of land. They vary over the years and are determined by land behavior over extended periods of growing seasons. Modern methods will be discussed later to determine land qualities.

TABLE 34.1 A Listing of Data Documenting an LUT with LUST for Maize, Sown January 15 on a Fertile, Well-Drained Soil, Typical for the Neguev Area of the Atlantic Zone of Costa Rica

Operation	Date	Labor (h)	Equipment	Materials
Land preparation	January 31, 1991	20	Machete	
Herbicide application	January 2, 1992	10	Knapsack sprayer	2 L Gramoxome
Sowing	January 15, 1992	10	Planting stick	20 kg local variety maize seed
Fertilizer application	January 30, 1992	10		50 kg ammonium nitrate
Harvest	May 15, 1992	50		100 bags dry cobs

Source: Adapted from Jansen, D.M. and R.A. Schipper (1995). A static descriptive approach to quantify land use systems. Netherlands Journal of Agricultural Science 43:31–46.

However, in the older land evaluation work, attempts were made to find proxies for land qualities based on land characteristics, which are *attributes of land that can be measured or estimated*. In this context, one may think of soil texture, organic matter, carbonate content, etc.

The basic elements of "classical" land evaluation have now been introduced. One has land that is being used for a particular purpose in a particular way. Not only current land use should be looked at but also other possible forms of land use that are of interest. One wants to assess land performance for these different alternative forms of land use and does so by comparing land requirements for each alternative form of land use with what the land has to offer. This matching process is made possible by defining important land qualities, often defined in terms of land characteristics in different classes. The overall analysis results in statements as to relative suitabilities of a given piece of land for a series of LUSs. Much practical experience with this system is reflected in the work of Sys et al. (1991) who provide excellent case studies. Some examples to further illustrate the procedure and to discuss some underlying concepts will be examined.

34.2.2 An Example

An example given by FAO (1983) addresses land suitability for sorghum (Table 34.2) and may serve to illustrate some basic decision steps to be taken that have general validity. Suitability of the land unit being considered is expressed in terms of highly, moderately, marginally, and not suitable. Three land qualities are distinguished: rooting conditions, oxygen, and nutrient availability. This selection reflects the expert judgment of the land evaluator. Quite possibly, other evaluators would have selected other land qualities. Because land qualities, as mentioned here, cannot be measured directly, land characteristics are used as proxies, called "diagnostic factors," and are as follows: drainage class (soil survey reports), effective soil depth (to be estimated from structure descriptions in soil survey reports), and soil reaction (pH). Clearly, some selections reflect lack of good data. Soil drainage classes are very broadly defined and only remotely related to O_2 diffusion. The effective soil depth is more direct but does not consider any particular demands by sorghum; it could apply to any crop. The pH, finally, is a very general indicator for nutrient availability, and its selection reflects an apparent desperate lack of information on the soil fertility status. Finally, the evaluator has to couple classes of the diagnostic factors somehow to demands by the sorghum crop (crop requirements) to arrive at the suitability classes. In summary, there are three moments when important decisions have to be made based on expert knowledge and data from the literature: (1) selection of land qualities, (2) selection of land characteristics to describe the land qualities, and (3) selection of gradations of these characteristics to form suitability classes based on estimated crop requirements.

34.3 Beyond Classical Land Evaluation

The classical land evaluation scheme has been widely applied and has influenced many land evaluation methodologies (e.g., Sonneveld et al., 2010b). Its application has been facilitated by automated computer-driven decision support systems (e.g., Rossiter, 1990), often with a strong agroecological focus. However, five problems have become clear over the years:

1. Even though the need to define clear land evaluation objectives in close interaction with stakeholders has been stressed from the start, the development of a mechanistic, rather sterile approach has been observed in which land suitability is defined for a large number of LUSTs (Sys et al., 1991). It is not clear who is asking the questions or, worse, whether the answers being provided address the questions that are really being raised. Thus, several projects have failed because of a lack of stakeholder participation (FAO, 2007).
2. Defining land qualities in terms of land characteristics has become a rather rigid qualitative procedure, even, and particularly, in automated computer-driven decision support systems, allowing little input from modern process-driven land research focused on landscape processes and climatic change.
3. The procedure is almost exclusively driven by the properties of the land, and even though the importance of socioeconomic conditions is acknowledged, little is done to take these conditions into account. Land use and its possible changes are usually more a reflection of socioeconomic developments in society than of differences in soil suitabilities for different forms of land use. Implicitly, classic land evaluation assumes that *use* follows *land*, while in reality more often *land* follows *use*. Think of roads and shopping centers on prime agricultural land. This requires a different approach than the soil-centered one.

TABLE 34.2 Land Qualities and Land Characteristics (Diagnostic Factors) for a Sorghum LUT, Expressed in Terms of Crop Requirements

Crop Requirement			Factor Rating			
			Highly Suitable	Moderately Suitable	Marginally Suitable	Not Suitable
Land Quality	Diagnostic Factor	Unit	s1	s2	s3	n
Oxygen availability	Soil drainage class	Class	Well drained/ excessive	Moderately well drained	Imperfectly drained	Poorly/very poorly drained
Rooting conditions	Effective soil depth	cm	>120	50–120	30–50	<30
Nutrient availability	Soil reaction	pH	5.5–7.5	4.8–5.5 and 7.5–8.0	4.5–5.5 and 8.0–8.5	<4.5 and >8.5

Source: FAO. 1983. Guidelines: Land evaluation for rainfed agriculture. FAO Soils Bulletin No. 52. FAO, Rome, Italy.

Moreover, land units, as distinguished in earth science, hardly ever correspond with legal units on which decisions are made. For example, a farmer farms a field with most often different land units, not a single one. A district or county where land use decisions may be made may cut through different land units, etc.

4. The procedure was implicitly defined as being scale independent. Most applications of classical land evaluation have been at the regional level, but many land use questions are raised at farm or field level or at the continental or world level. Not only are the questions then quite different (see (1) above) but procedures to be followed should be different as well.
5. Finally, there was a lack of dealing with uncertainties. Map units, agronomic relationships, and climatic variability are characterized by high levels of uncertainty, requiring expression of land use options and their effects in terms of probabilities. This aspect was not considered in the FAO framework.

What is now needed is a better evaluation of questions being asked at different spatial scales. Next, proper procedures need to be defined to deal with these questions, realizing a wide variety of stakeholders are involved. And, finally, the proper role of the land in affecting land use decisions needs to be defined. Certainly, decisions are not made for land units but for georeferenced surfaces on the Earth that may contain many land units.

34.4 What Is the Question?

Questions regarding land use vary a great deal. Let us analyze some of them:

1. A farmer wants to know how he can obtain a high and secure yield of profitable crops at minimal cost, while operating in line with increasingly detailed environmental laws and regulations. Cutting costs to be achieved by, for example, precision application of fertilizers and biocides or minimum tillage is increasingly important. He certainly is not interested to hear that his land is moderately suitable for wheat growing. He will already know that. He wants quantitative information in terms of what to do and when, where, and how to do it within the context of given environmental regulations.
2. An environmentalist will ask how agrochemicals use can be reduced or even abolished, thus avoiding leaching into ground or surface water. To answer such questions, detailed process-based simulation models may have to be applied at both farm and regional level to estimate the adsorption and leaching of agrochemicals as a function of management. Of course, the ideal is to combine the desires under (1) with those under (2), which is the basic concept of PA.
3. A regional planner may want to formulate alternative land use options within a region all in the context of achieving sustainable development. Here, the existing land evaluation procedure may be useful, but data are not specific enough to allow quantitative trade-offs among environmental, economic, and social considerations for each of the options.
4. A policy maker may see options for land use in a region or a country, formulated by procedures under (3). His question, however, is how attractive options can indeed be realized? How can stakeholders be influenced to do the right thing: special educational efforts, taxes, bonuses, subsidies? Clearly, land evaluation does not primarily focus on such issues, but they are increasingly important for land use and should, therefore, be considered because in the end the success of any scientific activity will be measured by its effect on society. If land evaluation reports wind up on the shelf, the activity itself will turn out to be nonsustainable. This fourth question can be handled by considering the policy cycle to be discussed later.

Considering the range of questions that may be encountered, an exploratory approach is attractive, defining a number of realistic land use options for the area to be considered. The stakeholders or policy makers make a choice from options presented. The type of question to be answered is as follows: What are the options for land use? How can one optimize land use for certain objectives and what are the trade-offs between these objectives? When dealing with sustainable land use, a balance has to be found for each option between environmental, economic, and social objectives. Different groups of stakeholders or different political parties will have different affinities with different options. The primary task of land evaluation is not to develop and select the "best" option but to present a series of well reasoned and transparent and alternative land use options allowing a rational selection process. Whether or not selected options are realized in practice depends on the stakeholders, but scientists can still be involved in the selection and implementation process because—as stated above—they also have an interest that some of the options they have derived will materialize in practice. This can be achieved by following the policy cycle, including signaling, designing, deciding, implementing, and evaluation (e.g., Bouma et al., 2007). The exploratory approach does not predict but explores what have also been called "windows of opportunity," a challenging concept.

Bouma (1993) has discussed five case studies on land use at different scales, showing that each was associated with quite different questions, while knowledge used in answering the questions varied very much as well. We therefore first analyze the different types of knowledge involved in land evaluation.

34.5 Some Basic Elements in Land Evaluation

34.5.1 Considering Different Types of Knowledge: The Knowledge Chain

A diagram (Figure 34.1) has been helpful to illustrate the use of different types of knowledge when analyzing land use questions at different spatial scales (Hoosbeek and Bryant, 1992; Bouma and Hoosbeek, 1996). They considered two perpendicular axes,

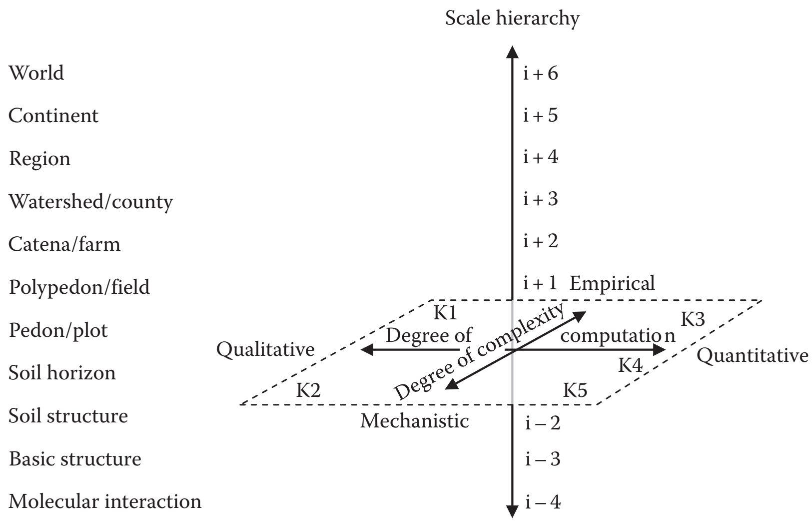


FIGURE 34.1 Scale diagram showing a series of hierarchical scales (i levels) and modeling approaches expressed in terms of four characteristics, which are summarized in terms of knowledge levels K1–K5. (Reprinted from Bouma, J., and M.R. Hoosbeek. 1996. The contribution and importance of soil scientists in interdisciplinary studies dealing with land, p. 1–15. *In* R.J. Wagenet and J. Bouma (eds.) The role of soil science in interdisciplinary research. SSSA Special Publication No. 45. SSSA, Madison, WI. Soil Science Society of America.)

one ranging from qualitative to quantitative and the other from empirical to mechanistic. Different research approaches occur within the plane thus obtained: K1 represents user knowledge; K2 represents expert knowledge; K3 represents knowledge to be obtained through semiquantitative models, in which real soil processes are not known; K4 represents knowledge through quantitative models where processes are characterized in general terms; and K5 represents the same, but processes are described in great detail, which can imply that the entire soil/crop system cannot be characterized anymore, and attention is focused on one aspect only (Bouma et al., 1996a). The vertical axis represents the scale hierarchy, where the pedon level (the individual soil) occupies the central position (i level). Higher levels are indicated as i+, while lower levels are i−. The scale in Figure 34.1 ranges from molecular interaction (i – 4) to the world level (i + 6).

One can now place the classic land evaluation in the scheme at scale hierarchy of watershed or region while the knowledge level is K2. For other questions raised above, one needs different hierarchies and knowledge levels. For example, the farmer and environmentalist would require the i + 1 field scale and a K4 knowledge level to get the necessary quantitative answers for their questions. The regional planner would operate at level i + 4 and would need at least a K3 knowledge level, because the K2 level would be too descriptive not allowing a quantitative trade-off analysis. A planner would be smart, though, to combine K2 with K3, by restricting the more detailed analyses of K3 to areas where a simpler K2 analysis could not provide answers. For example, Van Lanen et al. (1992) made a land evaluation for Europe in which potential for crop growth was established. Using soil maps, they first screened out strongly sloping land and land with shallow bedrock using a K2 approach. Then, in the remaining 40% of the land area, they ran a K3 simulation model to predict crop growth. The K2 approach for these soils (land is moderately suitable for wheat) would not have been satisfactory. This scheme introduces the possibility of combining approaches. Another example is derived from a study of De Vries et al. (1992). They did a study to determine the possible impact of acid rain on soil acidification. They dealt with nonagricultural areas without fertilization. They divided Europe (scale i + 5) (Figure 34.2) into grids, and for each grid, they determined the dominant soil type, using the soil map of Europe (K2 knowledge). Then, they selected a limited number of soil units (level i) that were considered to be representative for European soils (using K2 knowledge). In these soils, they made some detailed measurements of weathering rates (scale i – 4; knowledge level K5). Next, this knowledge was scaled back up, resulting in what is assumed to be effective K4 knowledge at the European level. It would have been impossible to make the detailed K5 measurements of weathering rates in all European soils. By using expert knowledge at different scales, measurements could be made more efficient. Ideally, measurements should always include a measure for reliability and accuracy. An overall K5 approach for all soils would have the highest reliability but would be too expensive. One must know how much is lost in terms of reliability when one goes from K5 to K4, K3, and K2. Decisions as to what to do can only be based on this type of information. The lines in Figure 34.2 represent a so-called research chain demonstrating how a given problem can be analyzed by combining knowledge at different scales. The study by De Vries et al. (1992) had a disciplinary character, covering one particular aspect of land use. The four questions raised in Section 34.4 have, however, a much broader scope focusing on sustainable land use and require therefore a more comprehensive and interdisciplinary approach than was needed in the study by De Vries et al. (1992), including consideration of the role of different disciplines and of the policy cycle. Before discussing the latter two items, it is important to discuss the role

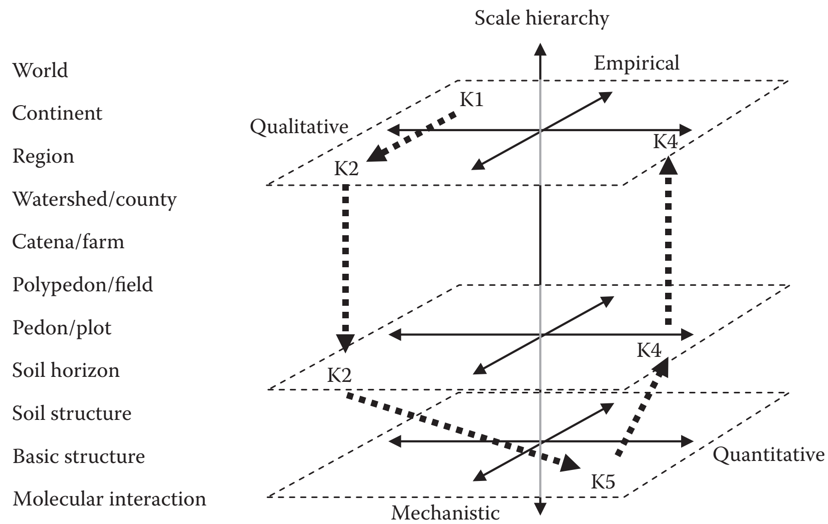


FIGURE 34.2 Scale diagram for the study by De Vries et al. (1992) on acidification of European forests. Three hierarchical scales are distinguished. At each scale, different methods corresponding to different knowledge levels are applied, forming a knowledge chain. (Reprinted from Bouma, J., and M.R. Hoosbeek. 1996. The contribution and importance of soil scientists in interdisciplinary studies dealing with land, p. 1–15. *In* R.J. Wagenet and J. Bouma (eds.) The role of soil science in interdisciplinary research. SSSA Special Publication No. 45. SSSA, Madison, WI. Soil Science Society of America.)

of soil scientists within broad, interdisciplinary land evaluation approaches involving many colleague scientists and often also stakeholders and policy makers. This role is quite different from the one described by FAO (1976, 1983, 1993) where soil scientists were in charge.

34.5.2 Applying Soil Science in the Broader Context of Land Evaluation

Soil science is back on the global agenda as evidenced by policy documents of several international agencies, listing the importance of soils in studying the Millennium Development Goals, climate change, environmental degradation, hunger alleviation, and production of biofuels (Hartemink, 2008; Hartemink and McBratney, 2008). In this context, land evaluation, as discussed in this chapter, can and should play a crucial role. There are, however, some potential difficulties (e.g., Bouma, 2009). Broad issues, as listed, require an interdisciplinary and participatory approach involving many scientists, stakeholders, and policy makers. To define a clear and visible role for soil science in such a complex context is not easy but necessary to allow input of relevant soil science expertise. Two suggestions are made here to strengthen this role. The *first* relates to three ways in which soil can be considered: (1) as an object to be studied with quantitative scientific methods ("it"); (2) as considered by society, expressed, for instance, by rules and regulations or by shear beauty ("we"), and (3) as viewed by the person studying the soil ("I"). Communication by soil scientists to larger groups often suffers when the three approaches are mixed as when personal feelings overly influence scientific study or by whatever society wishes to hear or feels. Being aware of these three views may help to make communications to the outside world more effective (Bouma, 2005). The *second* suggestion relates to the recently introduced Soil Protection Strategy of the European Union (Commission of the EC, 2006) containing, in turn, two important suggestions: first, it defines seven soil functions that are crucial for land use, and second, it defines the DPSIR approach as a means to characterize soil dynamics. The DPSIR system defines the drivers of land use change (D), the associated Pressures (P) their impact (I) and Response (R) and the subsequent state of a given landscape (S). The seven soil functions are as follows: (1) production of food and biomass; (2) storage, filtering, and transformation of compounds; (3) habitat for living creatures and gene pool; (4) physical and cultural environment; (5) source of raw materials; (6) carbon pool; and (7) archive of geological and archeological heritage. (Note that the term soil, used in the strategy, is equivalent to the term "land," as used in this chapter.) Focusing soil research on these seven functions and consideration of the DPSIR approach can help to clarify and focus the role of soil science in the teams studying the issues mentioned above, all relating to sustainable development (Bouma, 2010). The next question is how to apply this in terms of defining exploratory procedures to obtain options for land use, as discussed in Section 34.4.

34.5.3 Defining Options for Future Land Use

A scheme (Figure 34.3) was introduced by Bouma et al. (2008) allowing the systematic characterization of the state (S) of a given landscape for which a land evaluation is to be made. First it is important to broadly characterize the area in terms of land and water dynamics and ecological conditions. This can be done at K2 level. Hydropedology is a relatively new research activity within the soil science community and is very helpful in this context (Lin et al., 2005, 2006). Also, mapping locations of roads, railways, and canals, as well as settlements, helps to define the context in which the land evaluation will take place. Next, the

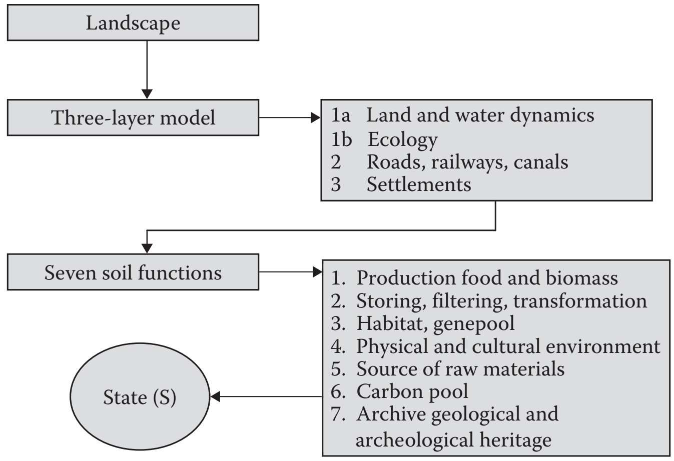


FIGURE 34.3 Flowchart indicating how the state (S) of a given landscape can be characterized by systematically considering the three-layer model and the seven soil functions.

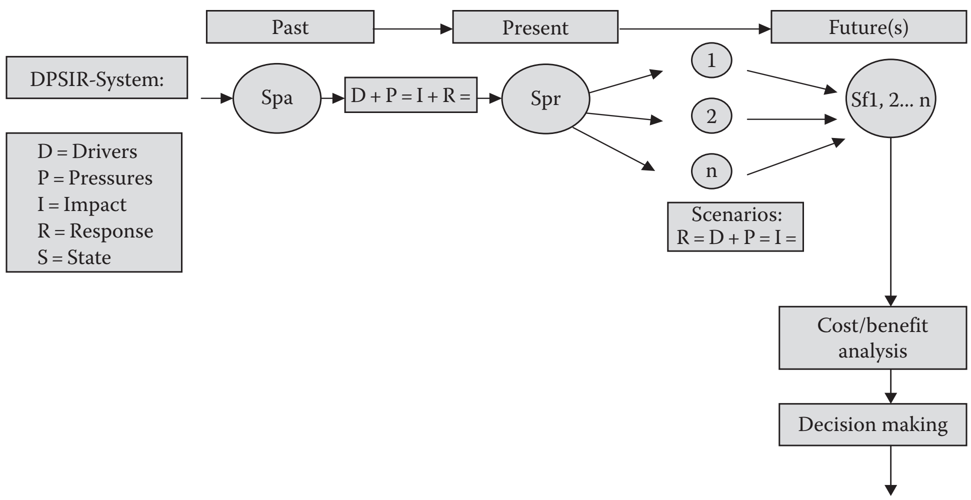


FIGURE 34.4 Illustration of the DPSIR procedure, illustrating how the past (Spa), current (Spr), and future states of landscapes (Sf1, 2, …, n) can be derived from considering D(rivers), P(ressures), R(esponses), I(mpacts), and R(esponses).

seven soil functions are systematically analyzed for the area, where, again, a selection has to be made of the level of knowledge to be applied for each function (see Section 34.5.1). The next step is to put this scheme in a time frame in terms of past, present, and future, where the latter is important for defining land use options as discussed in Section 34.4. Figure 34.4 illustrates the procedure. Drivers for land use change in the past and the associated pressures have led to certain impacts and responses. Effects of selected future options or scenarios are made visible by the responses to drivers and pressures of land use change that are associated with each option. As stated in Section 34.4, options are derived by discussions with stakeholders and policy makers.

34.5.4 Considering the Policy Cycle

Figure 34.4 could suggest that options derived are just presented to stakeholders and policy makers asking them to make a selection of the most attractive one. Even though the procedure of deriving the various options involved interaction with stakeholders and policy makers, it is important for researchers to more strongly involve both types of users. Again, as discussed above, the success of any land evaluation will in the end be measured in terms of successful acceptance of whatever was created in the procedure. This requires substantial involvement over time. Bouma et al. (2007) have illustrated this by emphasizing the need to consider the policy cycle in planning the research process. This includes various phases:

1. Signaling of a problem, preferably based on a characterization of the current state S.
2. Designing, involving development of options, as discussed above.
3. Deciding: Here, scientists should keep their distance. They design but do not decide. Still, it can be quite effective to give some advice at the right time to keep the decision process moving along.

4. Implementing: Here, researchers have a more distant role, but when they are still involved at this stage, they can answer questions, make minor modifications if certain assumptions turn out to be incorrect, etc. We see many examples where nice options for land use have been derived but where the implementation fails in ways that could have been avoided.
5. Evaluation: Very important for all involved stakeholders including scientists as a learning opportunity.

When dealing with land use issues in land evaluation, scientists often restrict themselves to the design phase (2). To make sure that land evaluations have a real impact in society, we advise the broader involvement described here. Various examples are presented and discussed by Bouma et al. (2007) and Van Latesteijn (1995). Of course, the design phase will remain the core activity and to not become overstretched as a scientist is a real challenge. This calls for a certain division of responsibilities within the scientific community, corresponding with the particular abilities of each member of the team in dealing with various aspects of the policy cycle. In this context, Bouma (2005) proposed establishment of so-called *Communities of Scientific Practice*.

34.6 Operational Procedures

34.6.1 Three Basic Steps

The role and function of soils in a given region, following the approach indicated in Figures 34.3 and 34.4, will become clear when the proper research procedure is selected. This procedure includes three basic steps:

1. Problem definition in interaction with stakeholders and policy makers, resulting in a number of alternative options for land use that are expected to solve problems being identified. The problems are also interpreted in terms of what they mean for the seven soil functions. All functions are not necessarily relevant for each land area, but it is wise to analyze them all. The level of analysis (K1–K5) depends not only on the relevance of each function but also on data availability.
2. Selection of research methods to broadly characterize the land and water dynamics and the soil functions; again, implying a selection in the range K1–K5. This selection not only includes basic data but also simulation models of varying kinds. If not enough data are available, new research can be proposed and initiated. It is attractive to define research needs on the basis of an analysis of shortcomings that would result when using available data only as this is more convincing than research needs that are defined out of context. Attention should be paid to the accuracy and reliability of the various procedures to allow an overall risk analysis in the end, which should, ideally, be an integral part of each land use option being devised (Heuvelink, 1998).
3. Effective presentation of results using modern communication technologies to be focused on prime user groups and policy makers and an evaluation of impacts, including lessons learned.

When performing land evaluation procedures, including an analysis of past drivers of land use change (Figure 34.4), one should realize that there has been a shift in focus during the last decades. Exclusive emphasis on food production and food security after the Second World War has resulted in a technology explosion: problems in food production were there to be solved. Land that was too wet was drained; land that was too poor was fertilized, even at very high rates; land that was too dry was irrigated; and land where crops were suffering from pests and diseases was treated with biocides. Initially, these measures were taken with only food production in mind, and this has led to considerable pollution of land and water locally. Later, concern for the environment played an increasingly important role, and as this process was unfolding, and as a balance had to be struck between agricultural production, on the one hand, and environmental quality of soil and water, on the other, the importance of agroecological features of the land increased dramatically. There used to be the technology-driven spirit: anything can be done anywhere. Now one realizes again that a sand will never be a clay and that the natural dynamics of soils in different agroecological zones are the basis for developing sustainable production systems that are in harmony with natural conditions and the environment.

It is, of course, important to realize that soils do not occur in random patterns in a landscape. Soil scientists know this but the general public and many of our scientific colleagues are hardly aware of this. There is too much talk about “Soils” and “Land” in general terms while, for example, there are 1500 soil types in the Netherlands alone. Much work in soil science has been done on soil classification, grouping soils that are comparable in their basic soil properties. Such groupings are used to describe soil behavior in terms of suitability for a given use (see Section 34.2). But the procedures, illustrated in Figures 34.3 and 34.4, allow extension of this limited approach toward defining characteristic: “windows of opportunity” for any given soil series (which is the lowest hierarchical unit of classification). Different forms of management will lead to different soil conditions within the same soil series, the range of conditions (the window) will be characteristically different for each series. Note that in soil classification, the effects of management have always been ignored for the simple reason that classifications should be more or less permanent and should not change as a result of normal agricultural management. But, in contrast, for land use and land evaluation, the effects of management are very important and need therefore to be associated with the classification.

34.6.2 Considering Effects of Management: The Phenoforms

Droogers and Bouma (1997) studied a prime agricultural soil in the Netherlands (a fine, mixed, mesic Typic Fluvaquent according to Soil Survey Staff, 1998). They suggested the terms genoform for the genetic soil name and phenoform for the management variant, of which three were studied, one resulting from 60 years of organic farming (BIO), one from modern intensive arable farming (CONV), and one from continuous grassland (PERM).

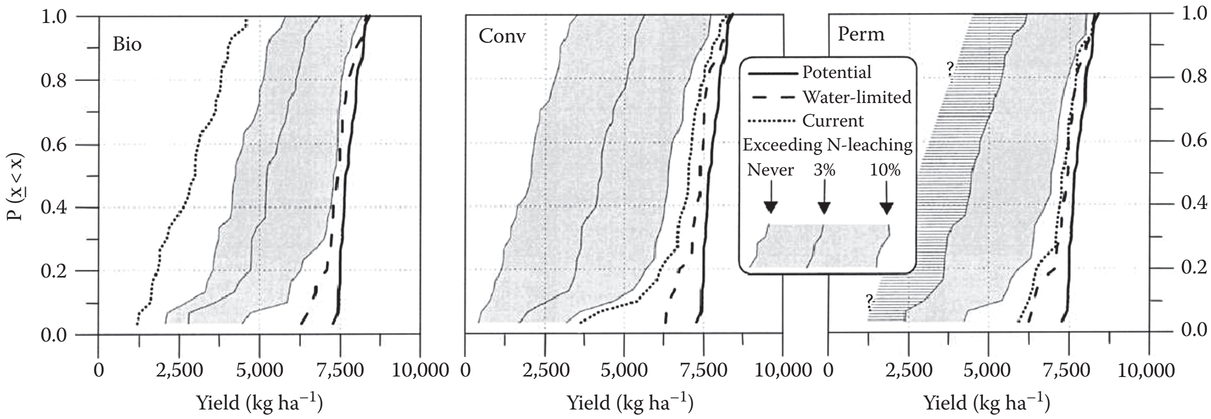


FIGURE 34.5 Cumulative probability function of simulated yields for the defined scenarios (see text). Probabilities were obtained by using 30 years of climatic data and are expressed for three probabilities that the threshold for nitrate leaching are exceeded. (Reprinted from Droogers, P., and J. Bouma. 1997. Soil survey input in exploratory modeling of sustainable soil management practices. Soil Sci. Soc. Am. J. 61:1704–1710. Soil Science Society of America.)

They used the validated WAVE model (Vanclooster et al., 1994) to calculate crop yields for wheat as a function of a wide range of N fertilization scenarios. By using yield calculations for a 30 year period with a wide range of weather conditions, they could express both yields and NO_3^- leaching to the groundwater in probabilistic terms (Figure 34.5). This, in turn, allowed the identification of a soil quality indicator for yield-expressing effects on the environment and risk. This indicator was defined by dividing actual calculated production (×100) by potential production where the latter is characteristic for each agroecological zone (Bouma and Droogers, 1998). Pulleman et al. (2000) applied the genoform/phenoform concept to the same soil series and after studying 50 fields, they derived a surprisingly robust regression equation, relating the organic matter content of surface soil (SOM) to previous land history as follows:

$$\text{SOM (\%)} = 20.7 + 29.7\ C_p + 7.5\ C_c + 7.5\ M \quad (34.1)$$

where

C_p is the past crop type (31–63 years ago)
C_c is the current crop type (1–3 years ago) with grass = 1 and arable = 0
M is the management type (3–7 years ago) with organic = 1 and conventional = 0 ($r^2 = 0.74$)

Sonneveld et al. (2002) found a similar relationship for a common type of sandy soil in the North of the Netherlands, a coarse loamy, siliceous, mesic plagganthreptic alorthod (Soil Survey Staff, 1998) where soil organic carbon (SOC) was related to land use as follows:

$$\text{SOC (\%)} = 3.40 - 1.54 \times \text{Maize} + 0.19 \times \text{Old} + 0.55 \times \text{GWC} \quad (34.2)$$

where

Maize is 1 for continuous maize, otherwise 0
Old is 1 for old grassland, otherwise 0
GWC is groundwater class in which 1 for class VB and 0 for the dryer class VI

Distinguishing well-defined phenoforms of existing soil series can be quite effective in extending the use of existing soil survey interpretations. Such effects are of major interest to many users and also define a meaningful new and innovative field procedure once the systematic soil mapping of land has finished, as is the case in many countries. Moreover, these findings illustrate the need to view soils as continuous spatial bodies that respond to land use.

This approach can be used at different scales once a database is filled with a number of phenoforms for any given soil series. Larger landscape units on small-scale maps, such as soil associations, are still defined in terms of their internal composition, for example, 30% soil series x, 50% series y, and 20% series z. Then, the effects of different types of management on yield and environmental side effects can still be expressed in a probabilistic manner, be it nongeoreferenced. Also, the fact that mapping units in the field may have quite some internal variability should be considered when making interpretations (Mausbach and Wilding, 1991; Bouma et al., 1996b; Young et al., 1997). The current development of digital soil mapping techniques, using, for example, gamma-ray spectrometry, does open up new possibilities for mapping soil properties at high spatial resolutions for relatively large areas.

34.7 The Crucial Importance of Effective Interaction with the Stakeholders

Interaction with stakeholders has been emphasized many times in the above sections during problem definition and the research process including final reporting. In the past, much research has been top down and supply oriented, which has led to the failure of several development projects (FAO, 2007). The researcher had an impression as to what the problem was that needed to be investigated, and he or she pressed forward using a favorite model, expert system, or data-gathering technique. In the end, the results of the study were presented to the stakeholders.

For sure, there are many examples of fine and effective research being executed in this way that have led to successful implementation. However, there are also too many examples of research efforts that were less successful, ending up in a desk, covered with dust. The challenge, now, is to involve stakeholders to the extent that research is being executed jointly with frequent interaction, with the desirable effect that the end result of the work is experienced as a joint product. This is easier said than done, and it may take a lot of time as illustrated by Bouma et al. (2008) for a regional study of environmental quality in the Northern Frisian Woodlands (NFW) in the Netherlands (see Section 34.8.5). Necessary interaction with stakeholders and policy makers during the research process is not only time-consuming but also potentially perilous. Any researcher has the obligation to make independent studies, whether or not potential results are liked by the stakeholders or not. If he or she gets too involved, objectivity may indeed suffer. This is a reason for many scientists to avoid interaction altogether. Though understandable, it is unwise certainly in the area of land use research that is so strongly tied to societal concerns (Bouma, 2009).

Modern information technology has an important role to play in stimulating interaction with stakeholders. Visualization of alternative land use patterns associated with different options is a very powerful tool to involve stakeholders. A picture says more than a thousand words. Interactive computer technology allows, for instance, instant generation of alternative land use scenarios with all associated input data by researchers and stakeholders. Also, the mentioned improvement of results when moving from K2 to K3, K4, and K5 approaches can be visualized as well by showing the accuracy of the land use maps obtained (Bouma, 1997b). The stakeholder can decide whether or not the costly improvement of the product is worth the cost. As Bouma (1993) has pointed out, several problems can be solved well with a relatively cheap K2 approach based on soil survey reports. This may be scientifically less challenging, but it is important from a practical point of view.

34.8 Case Studies

Six case studies will be discussed covering three spatial scales: the farm, the regional, and the world level. The three steps in the research procedure, as described in Section 34.6, will be followed for each case. Cases will be reviewed broadly, and reference will therefore be made to more detailed source publications.

34.8.1 Precision Agriculture: Arable Farm in the Netherlands

34.8.1.1 Problem Definition and Associated Soil Functions

PA is receiving increasing attention because fine-tuning of the application of fertilizers and biocides to the temporal needs of plants within spatially variable fields can be attractive from both an economic and ecological point of view (CIBA Foundation, 1997). By not applying more chemicals than are used or necessary, costs and losses to the environment are minimized: a "win-win" proposition. Prime agricultural lands in alluvial deposits all over the world are stratified. Sandy and more clayey spots alternate at small distances. Also other soils, for example, those formed in glacial deposits, may be quite heterogeneous. In this study, precision fertilization and application of biocides were tested in an operational farm setting to help fine-tune management with the objective of maximizing production while minimizing adverse environmental side effects, particularly nitrate and biocide pollution of groundwater. Research results from the past, as reflected in current fertilization advice by extension services, cannot be used for this fine-tuning as they base their advice on simple and highly generalized input-response functions derived from many different plot experiments under undefined environmental conditions. Also, the fertilizer advice implicitly assumes fields to be homogeneous. Existing exploratory or reconnaissance soil maps cannot be used either because they are not detailed enough to characterize fields. A more fundamental problem is the fact that pedological differences, even when expressed on large-scale soil maps, do not necessarily correspond with functional differences in terms of nutrient and biocide transformations, as defined by soil function 2. Functional characterization of soil is therefore a key element of PA as it is "translated" into so-called management units for the farmer, each of which requiring particular treatment in time. Their boundaries do not necessarily correspond with those of mapping units on soil maps. In all of this, aside from function 2, function 1 is crucial, while function 6 may become important as increasing the organic matter content by management in the long run will improve functions 1 and 2 (see Section 34.5.2).

34.8.1.2 Method Selection and Results

The need to define N dynamics and biocide adsorption in soil in relation to crop development requires the use of a quantitative mechanistic K4 model. More empirical and qualitative approaches are inadequate because environmental regulations are strict, and research results are only convincing when obtained at least K4 level. The WAVE model was therefore applied (Vanclooster et al., 1994). Even though a complete 1:50,000 soil map is available for the Netherlands, we lack the necessary detailed 1:5,000 maps to identify differences within individual fields. Besides, the concept of representative soil profiles is used for mapping units even on such large-scale maps and individual borings taken during mapping are not georeferenced. We therefore decided to do a new detailed soil survey for the 110 ha farm being studied (Van Alphen and Stoorvogel, 2000a). The procedure consisted of (1) detailed soil survey; (2) using geostatistics to estimate optimal boring distances; (3) making the soil borings; (4) performing simulations for each boring in the context of functional characterization, leading to the definition of management units to be used in precision management. Pedotransfer functions were used to obtain physical parameters for the simulation model, by relating them to basic soil characteristics (Van Alphen et al., 2001), and (5) perform simulations for each boring on the adsorption of various biocides being used in practice. Functional characterization (point 4 above) is based on items that are relevant for the particular arable system being studied and is site specific. In this case, four items were relevant to characterize N dynamics: (1) water stress in a dry year, (2) N stress in a wet year, (3) N leaching from the root zone in a wet year, and (4) residual N content at harvest in a wet year. Simulations of N dynamics were made for each boring, and management units were derived with fuzzy clustering and kriging interpolation techniques (Van Alphen and Stoorvogel, 2000a, 2000b). Calculations for three biocides were also made for each point with the WAVE model, focusing on simulated leaching values and associated risks. Patterns of leaching risks for each biocide were obtained by interpolation. Here, the management units, derived for N fertilization, were not used (Van Alphen and Stoorvogel, 2002). The operational farm system for PA consisted of real-time daily modeling with WAVE, using weather data. The stock of nitrogen in the root zone was tracked by calculated crop uptake and leaching as a function of time. As soon as a threshold was reached, the farmer was advised to fertilize. This moment is different in each different management unit (Figure 34.6), as the water and nutrient dynamics are different, even though the weather is identical. Unit 1 needed three fertilizations, and unit 3 needed four. The PA system reduced the amount of applied fertilizers with 25% as compared with the existing system based on generalized functions and resulted in significantly less leaching (Van Alphen and Stoorvogel, 2000b). The biocide studies showed significantly different risks being associated with using the different biocides, and the farmer was advised to use certain biocides only in certain parts of the farm, while using expensive biocides only in parts where risks were particularly high (Van Alphen and Stoorvogel, 2002).

34.8.1.3 Impact and Lessons Learned

This study on PA clearly showed the significant impact of fine-tuning the application of fertilizers and biocides to the varying needs of the plant in time. The soil science focus of this study is rather unique. Much PA work focuses on using remote or proximal sensing to establish crop conditions and to define management units. There has been much progress in developing these sensing techniques. For example, the N status of crop leaves can be well established by now. However, fertilization is already too late when applied after observing a low N status of plant leaves: growth retardation has already occurred. The soil-focused procedure, illustrated above, provides a signal before the N supply is too low, leading to N stress in the plant. It also allows estimates of possible leaching of nutrients, which is, of course, not possible with remote sensing. A second very powerful thrust in the PA research arena is the technical focus, developing global positioning systems (GPS) equipment and machinery allowing PA procedures. Unfortunately, the various approaches have not been combined and integrated well enough and that is one of the reasons that PA has not been adopted as widely as one would have expected and as one would wish. One lesson learned is that results from soil-centered studies have to be communicated more effectively to provide a counterweight to one-sided commercially motivated approaches by other disciplines. Finally, the soil studies, briefly reviewed above, were only possible thanks to the rapid development of information technology and computing power. As such, this offers unique and quite attractive new opportunities to "assess the performance of land" as required in land evaluation. Here, no abstract judgment is given as to how suitable a soil might be for a given use, but an operational procedure is presented for a farmer that can help to reach the potential of any given soil within commercial units of production, such as fields.

34.8.2 Precision Agriculture: Banana Finca's in Costa Rica

34.8.2.1 Problem Definition and Associated Soil Functions

Where European agriculture is being dictated by a plethora of legislation controlling agricultural management at the national and European level, institutional infrastructure to enforce legislation is lacking in most tropical countries. Despite the fact that national legislation is minimal, the producers of export crops are increasingly being constrained by the importing countries. A good example is the international banana sector in Costa Rica. Costa Rica provides over 25% of the bananas consumed in Europe. Banana producers are influenced in several ways by the market: (1) The European Union has supported the banana production in African, Caribbean, and Pacific (ACP) through

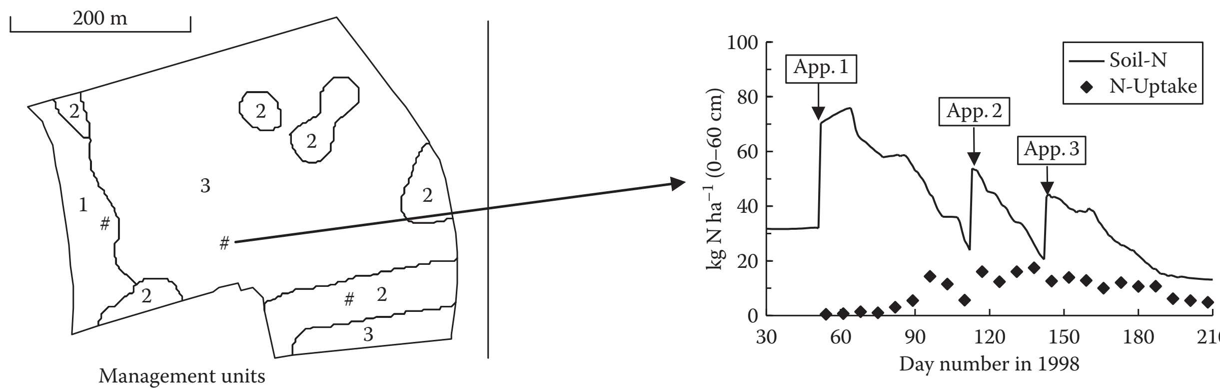
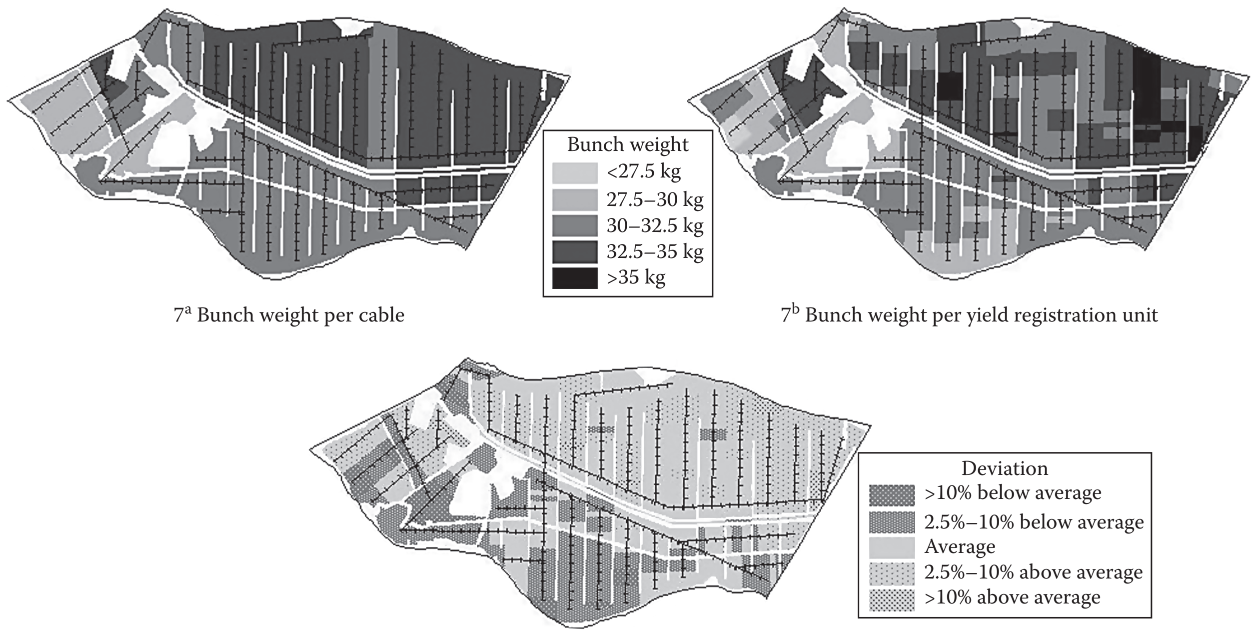


FIGURE 34.6 Management units for a Dutch arable farm with real-time simulations for soil available N contents for management unit 3 (line; kg nitrate-N ha^{-1} in the top 60 cm) and crop N uptake rate (dots; kg N ha^{-1} week^{-1}).

a complex tariff quota system restricting imports from, among others, Costa Rica; (2) the use of biocides is constrained by the regulatory agencies in the United States and Europe. They determine which pesticides are allowed and how they can be applied; (3) a plethora of different labels have been developed in Europe and the United States focusing on chemical use (organic bananas), social production conditions (fair trade), and the environment (ISO 14001, the environmental management standard); (4) in many cases, producers have to follow the guidelines of these labels to be able to export; and (5) large buyers of bananas (e.g., supermarket chains) force the banana plantations to produce under a detailed tracing and tracking system. Given the above, it is clear that the production has moved away from the small farmer. Currently, bananas are produced at large plantations that are either owned by independent producers (50%) or multinational companies (50%). All producers are organized in the Costa Rican Banana Corporation (Corbana). The banana producers have to operate in this complex set of political conditions. That does not only require management adaptations, it also requires the development of alternative management strategies and coordination among producers. Already in 1992 when the environmental lobby pushed hard on the banana sector, Costa Rican producers organized themselves in the Environmental Banana Commission aiming at the improvement of the production system in a more environmental friendly way. Before a plantation can be established, a land suitability assessment has to indicate that the land is suitable for banana production. As a result, Costa Rican banana production takes place under very favorable conditions. One of the plantations, the Rebusca plantation, has been very innovative in the past and maintains close links to the research department of Corbana. The Rebusca plantation is located in the perhumid lowlands in the northeast of Costa Rica (10.5°N, 84.0°W). The plantation measures approximately 124 ha of which 111 ha is used for the cultivation of banana. Like many plantations, the plantation exhibits significant soil variability with soils ranging from Typic Udivitrands to Andic Eutropepts and Typic Dystropepts (Soil Survey Staff, 1998). Like most Costa Rican banana plantations, the production of bananas is intensive with high use of agrochemicals in terms of fertilizers (2400 kg ha^{-1} $year^{-1}$), fungicides (48 aerial applications per year), and nematocides (3 applications per year). During the last decade, Rebusca has exported through different production channels ranging from a small exporting company owned by a few independent producers to sales through multinational companies. In the first case, production techniques were determined by plantation management, and although they had to fulfill European regulations, there was relatively little control. Exports through the multinationals had many more restrictions and controllers frequently visited the plantation. As a commercial enterprise, the first objectives of management are economical and can be described as profit maximization. Nevertheless, the owner of the plantation is also looking at the environment and tries to minimize the use and indirectly the emissions of agrochemicals to the environment. In contrast to other plantations, the owner is seeking innovations in the production system. This can be explained in part by a vision toward the future where he expects environmental and social regulations to tighten. After confronting the owner with the concepts of PA, the owner was highly interested. He saw major advantages in terms of chemical use efficiency and the registration of production activities. Just like the previous PA case, the focus is on soil functions 1 and 2 to produce efficiently with minimum losses to the environment.

34.8.2.2 Method Selection and Results

An integrated system for PA was developed for the Rebusca plantation. A detailed description of the system and the process of implementation is presented in the study by Stoorvogel et al. (2004c). In contrast to common practices in PA, we could not make use of standard yield mapping techniques. In addition, the technological focus of PA was not easy adoptable in a continuous cropping system with almost no mechanization and managed by manual labor. Banana plantations in Costa Rica are frequently (2–3 times per week) screened for bunches ready for export. Bunches are harvested and transported by a dense cable system (every 100 m) to the packing plant. For yield monitoring, groups of bunches are coded based on their origin (cable number and the location within the cable). A weighing balance is installed in the main cable just before the packing plant. Codes are registered at the balance, and bunches are weighed. A software package denominated BanMan was developed for data processing and to create and analyze the yield maps. An example is presented in Figure 34.7. In addition to the yield maps, a detailed soil survey has been carried out for the Rebusca plantation Stoorvogel, J.J., Kooistra, L., and Bouma, J. (1999). Spatial and temporal variation in nematocide leaching, management implications for a Costa Rican banana plantation. In: Corwin, D.L., Loague, K., and Ellsworth, T.R. (eds.). Assessment of non-point source pollution in the Vadose Zone. Geophysical Monograph 108. p. 281–289. Key element in an effective system for PA is the translation of site-specific soil and production data into management recommendations. There is an increasing call for the use of crop growth simulation models. However, we have to realize that for many tropical crops, these simulation models do not exist. In this case, we analyze the yield maps in relation to the soil conditions and determine the location of so-called problem areas. These areas are characterized by a relative low production compared to the average production for areas with similar soil conditions. This is an important signaling function after which more detailed studies have to pinpoint to the exact causes of the low production and management can intervene. In addition, fertilizer experiments were carried out on the three main soil types in the plantation where during a period of a year 1 ha blocks received 75%, 100%, and 150% of conventional fertilization. In these blocks, crop performance as well as nutrient leaching was monitored. Depending on the soil type, changes in fertilization resulted in changes in production and/or nutrient leaching. Results for the three main soil types are presented in Table 34.3. In soils 1 and 2, almost all of additional nitrogen is taken up by the plant if we move from 75% to 100% fertilization. However, further increasing fertilization resulted in increased leaching.



7[a] Bunch weight per cable

7[b] Bunch weight per yield registration unit

7[c] Deviation from average bunch weight per soil type

FIGURE 34.7 Yield maps and problem areas for a Costa Rican banana plantation.

TABLE 34.3 Nitrogen Concentrations in Soil Water

	Treatment (%)		
Soil	75	100	150
1	3.61	3.81	4.96
2	4.12	4.08	6.36
3	1.91	2.34	3.29

In soil 3, however, leaching already increased moving from 75% to 100% (with a stable production) indicating that fertilization could be reduced. The low production was caused by the heavy soil textures, poor internal drainage, and low pH. These results formed the basis for site-specific fertilizer recommendations. The recommendations resulted in a decrease in fertilizer use of 12%. An important new development in banana management has been the renovation of areas that experience a decline in production and for which no clear cause can be identified. Often, the deterioration of the plants has been given as the main cause. Due to the site-specific yield monitoring, we can accurately identify the areas that may improve through renovation.

34.8.2.3 Impact and Lessons Learned

The experiences with PA in the banana crop have been very favorable. Farm management has been able to improve the productivity of the plantation, and it has provided them with an improved insight into the performance of the plantation. In addition, they are equipped with an excellent system for the registration of farm management and productivity to answer questions in terms of tracking and tracing. Since the concepts of PA have been introduced in the Costa Rican banana sector, many companies and plantations have used particular elements of PA. A good example is the site-specific application of nematocides in those areas where nematode concentrations exceed the threshold of 16,000 per 100 g of roots. However, the spatial variation in nematode populations makes sampling and detection an awkward practice. The success of PA at the Rebusca plantation can be attributed to the integrated approach in which all the elements of the cropping system are included and the intensive involvement of the owner of the plantation in the development and implementation phase. Currently, the banana sector is under heavy pressure. Corbana recognizes the value of PA and starts to link their extension work to PA. The strategy of Corbana is that the sector needs to take a proactive approach. Tighter regulations with respect to the environment, the social conditions of workers and tracking and tracing are to be expected and the plantations have to be well prepared. This study presents a high-tech, interdisciplinary approach, developed in close and continuous interaction with the banana growers, to the “assessment of the productivity of the land.” In fact, this PA system, operating in a developing country, functions much better than PA systems in so-called developed countries.

34.8.3 Organic Farming in the Netherlands

34.8.3.1 Problem Definition and Associated Soil Functions

The need to create sustainable agricultural production systems that are economically feasible, ecologically acceptable, and socially attractive has led to much discussion in society as to the possible effects of introducing alternative farm management systems. Many of these discussions have an ideological character. For example, to some, organic farming serving regional markets presents a unique road to sustainable development.

To others, industrialized agriculture with a global perspective offers better perspectives. Obviously, this type of discussion is beyond the scope of this text. It is, however, of interest to see if and, if so, how soil conditions may change as different management practices are followed in a given type of soil. This way soil scientists can have independent input into broad discussions on sustainable development. This line of thinking is different from the classical one followed in soil survey interpretations where each soil series is defined in terms of a number of use limitations for a variety of land uses, implicitly assuming that soil properties within a given soil series remain the same, independent of use. To assess the effects of management on soil properties of a given soil series, representing prime agricultural land in the Netherlands, we compared soil conditions on an organic and a conventional arable farm. In contrast, also permanent grassland was added as a variant (Droogers and Bouma, 1996, 1997; Droogers et al., 1996). Several soil functions (Section 34.5.2) are relevant here: (1) biomass production, (2) transformations in the soil, (3) habitat and gene pool, and (6) acting as a carbon pool, function (4) does to a certain extent relate to organic farming as this tends to better preserve the local traditional character of the land as compared with industrial agriculture.

34.8.3.2 Method Selection and Results

Soil conditions can be well expressed in terms of land qualities (FAO, 1976; see Section 34.2.1), which are the moisture supply capacity, nutrient supply, trafficability, and workability, representing a K2 approach. This, however, is descriptive and not diagnostic. We, therefore, decided to use a deterministic, quantitative K4 model (WAVE; Vanclooster et al., 1994) that allows quantitative characterization of soil moisture regimes and associated N transformations. Common soil properties such as bulk density, porosity, moisture retention, and hydraulic conductivity reflect the effects of short-term management. Tillage or soil traffic under wet conditions or poor management may in a given year lead to compaction, puddling, and structure degradation, which may not occur in the same soil where soil traffic is avoided or where the farmer does a better job. We, therefore, focused here on soil properties that are not significantly influenced by short-term management such as the organic matter content, which is affected by long-term management in terms of decades. Moisture retention and hydraulic conductivity data needed to be measured because estimates with pedotransfer functions could not be made as they did not cover the range of organic matter contents observed. To adequately represent the effects of different contents of organic matter on N mineralization, we measured rate constants to be used in the N module of WAVE. Thus, a relatively high data demand materialized. Simulation runs were made for 30 year periods using real weather data, allowing expressions of production and leaching of NO_3^- in probabilistic terms. Predictions of moisture contents in surface soil allowed estimates to be made for trafficability and workability (Droogers et al., 1996). Runs for 1995 were compared with measured data, showing that the model performed satisfactorily. Validation was thus assured, but no thorough sensitivity or error propagation analysis was performed.

Results were presented in graphical form. The increased organic matter content originating from organic farming resulted in a 20% increase of potential productivity (Droogers and Bouma, 1996, 1997). However, trafficability and workability decreased because the higher moisture contents during the year resulted in shorter periods with adequate trafficability and workability (Droogers et al., 1996). Indeed, strong compaction was observed in fields of the organic farm where conventional plowing to 30 cm depth was used. We also addressed the question as to how production and NO_3^- leaching could be balanced, using a probabilistic graphical expression (Figure 34.5). This graph was also used to define soil quality in terms of production and leaching of NO_3^-, illustrating that the user must choose the level of risk he or she is willing to take. This risk relates to yields (probabilities that a certain yield is exceeded) and to NO_3^- leaching (probabilities among the years that the threshold value for NO_3^- leaching is exceeded).

34.8.3.3 Impact and Lessons Learned

Organic farming resulted in higher organic matter contents that, in turn, allowed higher potential productions because of a higher water supply capacity. However, risks of causing structure degradation were higher as well, and it was shown clearly that traditional plowing to 30 cm depth did indeed result in such degradation. We advised to explore minimum tillage as an alternative tillage practice, and this was followed up by the farmers. This study also led to the initiation of a broader study relating past land use successfully to organic matter content of surface soil, as reported in Section 34.6.2 by the study of Pulleman et al. (2000). The K4 land evaluation approach followed in this chapter is a far cry from the traditional K2 procedure. But it still focuses on the central issue of land evaluation, that is, "the assessment of land performance when used for specified purposes," as it addresses real questions of the land user. Computer simulations for a series of years allow expressions in terms of risks, and this is important: scientists do not only provide a verdict as to the suitability of a given form of land use but define risks associated with particular actions. It is up to the land user to decide what to do based on the risks he or she is willing to take. One lesson learned was the fact that this study defined a welcome "niche" for soil science in the very broad and often emotional societal discussion about sustainable land use and the role of organic farming. Looking back, we should have pursued this approach systematically for other major soil series offering a model for post soil mapping research. In fact, this type of work was done later only for a major soil series in sandy soil (see Section 34.6.2).

34.8.4 Alternative Land Use in Ecuador

34.8.4.1 Problem Definition and Associated Soil Functions

Although the commercial farmers of the Carchi region in Northern Ecuador are probably some of the better endowed in the rural communities of the Andes, it became increasingly apparent that the use of pesticides in intensive potato production was not only a blessing but a curse as well. In the early 1990s, researchers, farmers, and NGOs paid increasing attention to the negative effects of the intensive use of pesticides. The farmers in the Carchi region were very much dependent on these pesticides to control pernicious pests and blight. Intensive on-farm research revealed some of the major health and environmental impacts associated with these pesticides. The intensive use of highly toxic pesticides resulted in significant neurobehavioral effects on farmers, and pesticides were detected in groundwater as well as streams that serve irrigation and domestic use downstream. A broad signaling phase with large groups of stakeholders, however, also drew the attention to soil erosion. Many fields revealed the light-colored subsoil on the upper parts of the fields, and the common perception was that water erosion was the main cause. However, research showed that water erosion was not responsible for the erosion of the topsoil due to low rainfall intensities in combination with a high infiltration capacity of the volcanic ash soils. The steep fields cultivated with potatoes require intensive tillage. All tillage and harvest operations transport topsoil material down the slope. On fields tilled with tractors, the situation is even more serious. As slopes are too steep for contour plowing, tractors plow downslope. Rather than water erosion, tillage erosion was in this case the main cause for the observed erosion processes. This example clearly shows not only the importance of stakeholder input but also the impact of scientists in the signaling phase, avoiding misperceptions that could easily have led to irrelevant routine research on water erosion. After signaling and quantifying the key sustainability factors, that is, human health and environmental impacts of pesticide use and tillage erosion, the question that remained was how to intervene. Clearly, the reduction of pesticide use and pesticide handling issues were the two elements to focus on to reduce the environmental and health impacts of pesticides. To reduce the impact of tillage erosion, very few technical solutions were available. The potato crop requires intensive tillage, and any tillage practice on these steep slopes automatically results in soil erosion. The only practice that significantly reduces tillage erosion is based on the little-known pre-Columbian limited tillage/cover potato system: Wachu rozado that is still being practiced in Northern Ecuador. In this system, seed potatoes are placed on top of the pasture, and the grass mat is folded over the potatoes.

34.8.4.2 Method Selection and Results

A 2 year dynamic farm survey provided insight into the management decisions of the farmers in the Carchi region. This resulted in an economic simulation model (TOA) allowing for the ex ante evaluation of alternative policy and management scenarios (Crissman et al., 1998; Stoorvogel et al., 2004a, 2004b). The TOA methodology addresses two key elements: first, it provides an organizational structure around which to design successful interdisciplinary research that assesses the sustainability of production systems; second, it provides a successful means to communicate research findings to policy makers and the public. Farmers are the Andes' most numerous and most important soil resource managers. Agricultural technology ranges from traditional, extensive, low-input, low-output systems to modern, intensive, high-input, high-output systems. The traditional systems have to be maintained within their ecological constraints and, as a result, are generally perceived as environmentally friendly and sustainable. However, due in part to shrinking farm size, traditional systems have proven to be economically and socially nonsustainable. With a closed agricultural frontier in most parts of the Andes, the fundamental option for Andean farmers is to increase the physical and financial output from the existing farm. This inexorable pressure provides a strong incentive to shift to the higher output modern systems. The basic quest of agricultural and environmental research for sustainable farming systems is to match the environmental friendliness of traditional farming systems while reaching the higher outputs and, thus, the economic and social sustainability found in modern farming systems. Figure 34.8 shows an example of the type of answer one can expect from the TOA system. The trade-off curves between net returns and carbofuran (one of the most commonly applied highly toxic insecticides) leaching are

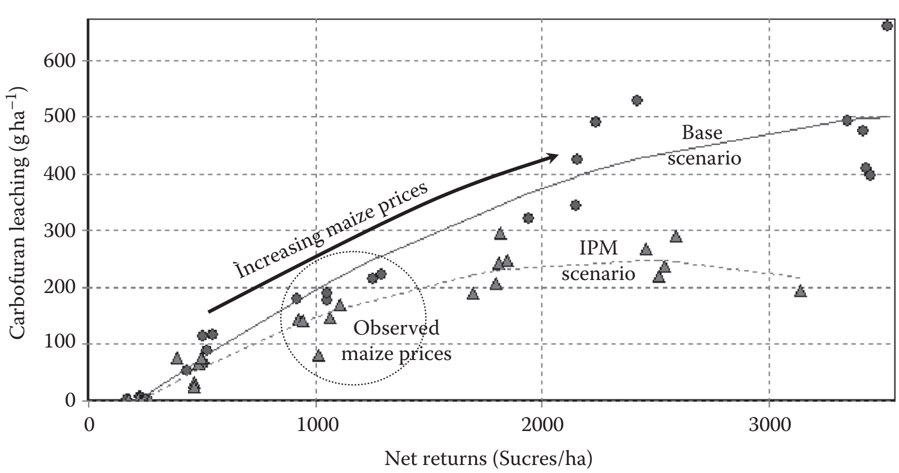


FIGURE 34.8 Trade-offs between net returns and carbofuran leaching under two different management scenarios for the potato-pasture system in Northern Ecuador.

constructed by varying the potato prices. In the base scenario, the current management system is evaluated. The trade-off curve shows that with increasing potato prices, the net returns of the systems increase coinciding with an increase in carbofuran leaching. The latter is not only due to an increase in the potato area but also due to a more intensive management of the potatoes. Two alternative scenarios have been evaluated: the effect of tillage erosion and the effect of the introduction and adoption of Integrated Pest Management (IPM). Due to tillage erosion, we see that carbofuran leaching increases. This can be explained by the removal of the organic matter-rich topsoil that fixes carbofuran. IPM shows an opposite effect where the trade-off curve is moving down. Less carbofuran is being applied, and as a result, leaching is going down. However, the reduction of IPM comes at a cost. Alternative management practices are required to control the pest requiring farm labor. As a result, net returns are slightly decreasing.

34.8.4.3 Impact and Lessons Learned

The TOA provided information on possible interventions both at the political level and at the farm level. In various follow-up projects, the policy cycle has been closed. While pesticides have not been eliminated from the Carchi communities, they are now generally used more cautiously. There is also momentum at the policy level for reducing pesticide dependence. In 1999, all stakeholders were brought together to discuss pesticides and health. This meeting resulted in the Carchi declaration demanding the elimination of the highly toxic products, the inclusion of IPM in university level agriculture training, and a wider dissemination of information on the effects of pesticides. In addition, farmer field schools have been set up in which farmers and the research community developed not only IPM technologies but also pesticide handling measures to reduce the impact of pesticides. The effects of various management changes, as part of IPM, on these farmer field schools are striking. The number of pesticide applications was reduced from 12 in conventional plots to 7 in plots with IPM. Even more important, the overall amount of pesticides applied dropped dramatically. The amount of fungicides decreased by 50%, while insecticide quantities dropped between 40% and 75%. The Carchi story is illustrative for a combined effort in which farm surveys, advanced simulation modeling, geographical information system (GIS) techniques, IPM research, stakeholder meetings, and farmer field schools led to a strong reduction of pesticide use in the Carchi study area. It illustrates the strength of the research chain rather than a single method and/or project. The project was successful in designing innovative production systems for potatoes that were environmentally friendly while protecting the health of farmers. Political decisions were made about environmental and health regulations, and they were implemented, so far only at the regional level. Educational programs for farmers were initiated during this research projects and are continued up to this day. This study not only "assessed the performance of land" but also added and integrated crucial economic and social components, which were part of the original definitions of land evaluation but were hardly incorporated in conventional land evaluation studies focusing on land suitabilities and limitations.

34.8.5 Environmental Quality in the Northern Frisian Woodlands, the Netherlands

34.8.5.1 Problem Definition and Associated Soil Functions

Environmental losses from dairy farming systems in the Netherlands are among the highest in Europe. A strong focus on increasing dry matter production though increases in fertilizer N inputs resulted in average N losses of more than 300 kg N ha^{-1} in the beginning of the 1990s. The excessive loading of groundwater and surface water in the past has stimulated the development of environmental policies at European and national level. Especially, the European Nitrates Directive (EC, 1991) and its translation through the Dutch MINAS legislation have had a profound impact on management practices at farm and field level.

The Nitrates Directive has adopted a threshold for the nitrate concentration in the upper groundwater that corresponds with the E.U. drinking water directive: 50 mg L^{-1}. In 1998, the Netherlands introduced the MINAS policy, a budgeting tool for N and P and farm level with a maximum N surplus of 180 kg N ha^{-1} to be reached in 2003 for sandy soils. Dairy farmers in the Friesian Woodlands in the North of the Netherlands were also confronted with high N surplus levels in the 1990s. Provincial groundwater monitoring networks reported nitrate concentrations for this region well above the 50 mg L^{-1} threshold (Sonneveld and Bouma, 2003). A regional nutrient project was initiated by Wageningen University in close collaboration with local farmers to develop and disseminate knowledge on possibilities for reducing N losses and hence increasing N efficiencies at farm level. A need was expressed in 2004 to provide regional policy makers and farmers with information on the quality status of groundwater in the Friesian Woodlands.

A number of land functions (Section 34.5.2) are especially relevant for the question raised: (1) biomass production; (2) transformations in the soil; and (6) acting as a carbon pool, which is specifically linked to land use on the dairy farms. The functions, physical and cultural environment (4) and geological and archeological heritage (7) became relevant in 2004. In that year, the area was designated as a National Landscape, because of the small-scale elongated fields in the area bordered by hedges and a unique high concentration of pingo-remnants from the Weichsel glacial period.

34.8.5.2 Method Selection and Results

The NFWs cover 60,000 ha and occur in the northern part of the Netherlands. Land use consists predominantly of grassland (80%) with some silage maize (5%). The area mostly consists of hydromorphic podzols developed in Pleistocene coversands. The regional nutrient project (Verhoeven et al., 2003) that aimed at lowering N surpluses adopted a strong participatory approach with study groups among farmers and contributions from scientists from different disciplines (animal science, soil science, and grassland science). Farm management practices and N surpluses were monitored and used for feedback within the study groups. Different methods were adopted to assess nitrate concentrations in the upper groundwater and N and P concentrations in surface waters in the NFW region:

1. The available STONE model was used to calculate N and P losses at regional level. This quantitative model is based on process description and can be described as a K5 model. The STONE model (Wolf et al., 2003) is used to calculate nutrient loads from the soil system to the surface water and groundwater. STONE calculates nutrient fluxes for different forms of land use like grassland, arable land, and nature and nitrate concentrations in the upper groundwater. The region was divided into 398, 250 × 250 m grid cells for which specific hydrological, soil physical, and soil chemical characteristics could be derived by downscaling existing soil survey maps. All cells are hydrologically separated but are connected to various drainage levels, like ditches, canals, and subsoil. This implies that STONE behaves spatially like a semi-3D model. All cells have a specific soil type and corresponding soil hydraulic characteristics for the unsaturated zone. The P status and mineralization capacity of the soil are also input for the model. In the STONE model, the water balance model SWAP (Belmans et al., 1983; Van Dam et al., 1997) is used to simulate water fluxes, and the ANIMO model (Groenendijk and Kroes, 1999) is used to simulate nutrient dynamics and nutrient transport. In the initial STONE simulations, the amount of fertilization on the 250 × 250 m grid was calculated based on "downscaled data 2000" of the Dutch national database on manure application and the assumption that mineral fertilizer was used according to the Dutch national fertilization recommendation. The available data represent the situation in the year 2000. In the second STONE simulation, regional data (2004) on manure application and fertilizer use were used ("regional data"), and grid cells of 25 × 25 m were used to be able to apply the regional data of manure application on the scale of the fields. Data available for farms were allocated to the small grids. Meteorological data on a daily basis are used as input.
2. A regression method was used based on a modeling study for sandy regions in the Netherlands (Roelsma et al., 2003) where nitrate concentrations had been related to autumn mineral N contents in soil (K3 level) was also applied for a selection of 29 farms on sandy soils (Sonneveld et al., 2010a). In general, agrienvironmental indicators that can be used for monitoring by stakeholders are useful in providing learning feedbacks. The following model was applied:

$$NO_3^-[\text{year}+1] = C + 0.764N_m[\text{year}] - 37.9P \qquad (34.3)$$

   where

   NO_3^- is the nitrate concentration in the upper groundwater in spring
   year is a particular year
   C is a constant depending on hydrology and soil type
   N_m is soil mineral nitrogen (kg ha^{-1}) in autumn for 0–90 cm
   P is a dummy for the occurrence of a peat layer (1 = yes and 0 = no)

   For application of this regression model, a monitoring campaign for soil nitrate contents was performed in the autumn of 2006. This method was introduced to see whether soil mineral N contents could be used as an easy indicator by farmers for assessing nitrate concentrations in the upper groundwater.
3. Finally, to validate the above mentioned K3 model, an intensive monitoring campaign was performed in 2007 on more than 300 locations to measure nitrate concentrations in the upper groundwater. In itself, this can be described as a K4 approach.

N losses at farm level were considerably reduced from 1997 to 2003. In this period, the average N surplus dropped from 327 to 168 kg N ha^{-1} (Verhoeven et al., 2003). The reductions were largely achieved through reduced inputs of fertilizer N inputs and improved efficiencies of manure N. For the years 2000 and 2004, STONE calculated relatively low average nitrate concentrations in the upper groundwater in the area with a mean value for the region of 22 mg NO_3 L^{-1} for the "downscaled national 2000 data" and 15 mg NO_3 L^{-1} for the "local 2004 data" (Figure 34.9). The estimates are far below the E.U. limit of 50 mg NO_3 L^{-1}. However, there is a large variation in nitrate concentrations for the year 2000 ranging from higher values in the drier sandy soils to lower values in the peat soils, which is related to differences in denitrification rates. The results for the year 2004 indicate overall low values.

The regression model predicted average nitrate concentrations in the upper groundwater for 2007 to be around 82 mg L^{-1}, which was much higher than the predictions for 2004 by the STONE model. Predicted 90 percentiles of nitrate concentrations in the upper groundwater using the K3 regression method were >150 mg L^{-1} for farmers on sandy soils. Based on the validation data (the K4 method), however, the estimated 90 percentiles were <50 mg L^{-1}. Several observations with high concentrations were especially found on field with arable cultivation (silage maize). Thus, Sonneveld et al. (2010a) concluded that the regression model based on national data was invalid for this particular region and lead to significantly different conclusions. The field measurements appeared to correspond with the STONE model calculations for the year 2004.

34.8.5.3 Impact and Lessons Learned

The results of the participatory approach used in this nutrient management project gave a strong impetus to involve farmers in searching for possibilities for reducing N and P losses to the environment. The strongly reduced N losses in the Northern sandy

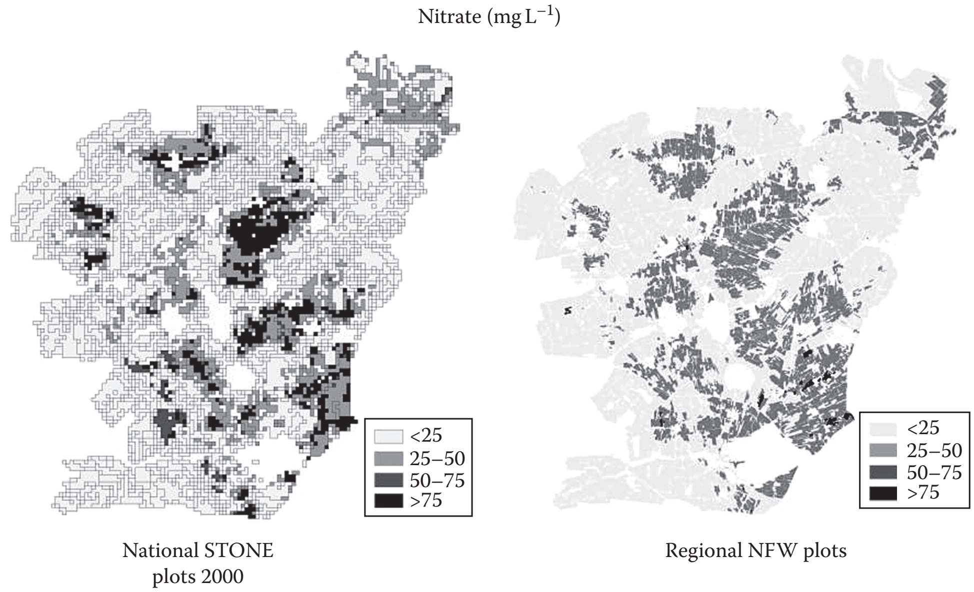


FIGURE 34.9 Mean nitrate concentrations in upper groundwater calculated for the years 2000 (a) and 2004 (b).

regions have also been documented in national inventories. The annual environmental report reported that currently, high nitrate concentrations are mainly a problem of southern sandy regions (>80 mg L^{-1}) and not of the northern regions. The STONE model simulations yielded especially relevant results when regional data on land use were introduced in the 2004 model simulations. This strongly suggests that regional calibrations remain a vital component of introducing and applying K5 models at regional scales. The K3 regression models clearly overpredicted nitrate concentrations substantially. It was also found that actual measurements on nitrate concentrations were found to be most convincing for farmers and regional policy makers. Several scale levels were involved in reaching the environmental objectives. Policy formulation at E.U. level resulted in national legislation (MINAS), which was translated to individual farms through a regional nutrient management project. Although national legislation has changed in past years, the insights that were gained have not been lost, and surpluses have continued to decrease over the years. The questions raised clearly went beyond classical land evaluation that is strongly focused on biomass production and follows a top-down strategy. Here, questions did not so much center around options for main types of land use as grassland is considered to be the most important component of dairy farming systems in the region. However, specific management strategies as to how to manage and fertilize pastures (LUTs) were being developed. Options for changing land management practices especially evolved from close collaborations between scientists and farmers and resulted in decreased environmental losses. Model simulations at the regional level as well as field measurements at local level confirmed that the reduced N surpluses resulted in nitrate concentrations <50 mg L^{-1}. Discussions have risen whether the small proportion of silage maize at some dairy farms cannot be replaced with grassland as the higher nitrate concentrations were largely found under fields with maize cultivation.

34.8.6 The World Food Crisis: Assessment of Food Security

34.8.6.1 Problem Definition and Associated Soil Functions

The world population is expected to double by the year 2030. Enough food is produced now, even though its distribution is poor, leaving 800 million people with inadequate food supplies. Many studies have been made on the potential of the Earth to feed its people. We will refer here to the still highly relevant work by Penning de Vries et al. (1995) because they used a modern land evaluation procedure and made some interesting decisions on procedures that correspond to the global level of detail. This study tried to answer questions about global food security by considering three population growth scenarios and two types of agriculture, one with high external inputs (HEI) and the other with low external inputs (LEI). Also, three diets were considered to reflect the demand side: a vegetarian, moderate, and affluent diet. The reader is referred to the above publication for more details on the interesting results of the study. Research at this scale can only be exploratory in a broad sense, and many simplifications are involved. Emphasis was placed on exploring agricultural production potentials. Whether or not these are reached is realistically considered to be beyond the scope of the work as it will depend on socioeconomic and political decisions. What is offered are characteristic windows of opportunity and, as such, the study has led to interesting insights into possible future developments. This study at world level focuses exclusively on biomass production (soil function 1) under a series of conditions relating to management, food consumption, and population growth.

34.8.6.2 Method Selection and Results

A K2 approach might appear to be realistic at first sight. However, no expert can oversee the immensely diverse production conditions prevailing in different areas of the world. The authors chose, therefore, a K3 approach using a very simple model to calculate crop yields based on available radiation, water, and nutrients. Thus, a universal approach was introduced allowing comparisons among all calculated values. They assigned all possible crops to a cereal (wheat in moderate climates and rice in the tropics) or grass equivalent to obtain a so-called grain equivalent (GE), which makes comparisons relatively easy. The NASA database was used for 1° × 1° grid cells, which occupy a 110 km × 110 km area at the equator. Confining attention here to soils for each grid cell, the authors selected values for slope, soil phase, and soil texture. Selections were based on the 1:5 million FAO World Soil Map and other available data. Clearly, choices were highly arbitrary. Cells often contain different land units, each representing major soil associations with many soil types. Soil texture of the dominant association was applied to the entire cell. Dominant soils in each cell were considered to be well drained, homogeneous without layers or cracks, to be 60 cm deep and without runoff. Degradation of soils was ignored, and so were current land use patterns. For the entire world, 15,500 land units had to be considered along with 700 climatic zones. The model was run for all cells, divided into 15 major regions, which were also distinguished for U.N. population studies, and for different scenarios representing population increase and management types. Estimates for grid cells could have been compared in some areas with estimates obtained using more reliable basic data derived from large-scale soil maps. This, however, has not been done. Therefore, the accuracy of the resulting data cannot be determined, which illustrates the broad exploratory character of the exercise. Still, the alternative is to have nothing at all, and we prefer such a rough estimate, which can be improved upon, over a lack of any estimate.

GIS was used to present maps for the different scenarios, which was effective in communicating the major results of the study. For example, a problem was perceived for Southeast Asia, where food shortages are likely in future even at low rates of population increase and with high-input agriculture.

34.8.6.3 Impact and Lessons Learned

This exploratory study has functioned well to define some major worldwide trends in food security, some of them rather surprising. Food production at the global level can increase significantly, but a strong spatial variability exists. When HEI farming is practiced, all regions can produce the required food, except some areas in Asia. Europe, the Americas, and Central Africa are well-off. Depending on the level of consumption selected, Europe, for example, can grow its food on 30%–60% of its suitable soils and in the America's, this is 20%. When choosing the LEI approach, South Asia will have a food shortage even at the minimum food demand. The impact of studies such as these is difficult to measure. Reports of U.N. agencies have, for example, recently taken a more specific approach to defining food security by focusing on particular regions, and this is supported by the type of studies discussed here. Also, showing the implications of HEI versus LEI management systems in different regions serves to focus ideological discussions dealing with the most desirable type of farming by presenting plausible consequences of following either HEI or LEI approaches. The study adds an important element to the concept of land evaluation, by including the demand side of the issue of food security. "Assessment of land performance" is not only determined here by the productive capacity as such, which is divided into the LUTs HEI and LEI, but also as a function of demand, which depends on the type of food consumption and on population growth. Thus, land evaluation is placed in a relevant societal context, while the input of soil data is rudimental, to say the least. The tendency of soil scientists is to implicitly assume that the more detailed soil data are provided, the better the land evaluation. The study shows that this is not necessarily true at world level. Unfortunately, soil scientists have not followed up this study by showing what might be gained by improving the soil input data for the model. Finally, one lesson learned is the necessity to repeatedly present studies to a large audience as this work has not received the attention it deserves. The work is not finished when the research report is finished (see also remarks about implementation in the policy cycle in Section 34.5.4).

34.9 New Thrusts for Soil Survey and Land Evaluation

34.9.1 Using New Technologies in the Proper Context

New technological developments are likely to change soil studies in the years to come (see NRC, 2009). Low-cost GPS allow accurate positioning of observers and equipment anywhere in the world. New satellites will strongly increase the opportunity to observe the surface of the Earth in more detail than ever before using an increased number of diagnostic features. Remote sensing from satellites or airplanes will allow better evaluation of crop conditions allowing improved yield predictions. In addition, on-the-go yield monitoring is becoming well established and will revolutionize the assessment of the production capacity of the land by providing a continuous record for many years. In turn, measured differences in yield will give rise to specific research on the underlying causes, which may be many.

Technological developments not only apply to measurements above the soil surface but also to measurements below the soil surface. Many new sensors are being developed to be used for the continuous registration of soil water and solute contents. Already, time domain reflectometry is widely used to measure soil water contents, replacing neutron probes. Transducer tensiometers allow accurate, instant registration of pressure heads in the soil within the range of interest for plant growth. Sensors for N contents in soil are being developed in the context of PA. Information technology allows not only rapid registration but also transmission to central computers. Indeed, some study sites resemble patients in intensive care.

Opportunities to interpret soil data have dramatically increased as well. A wide range of simulation models is available to predict soil water contents and solute fluxes, even in heterogeneous soils that swell and shrink and may show hydrophobic behavior. This must be calibrated and validated to fit the needs of specific areas where land use questions are raised. Also, expert systems have been perfected allowing a more efficient application of the vast body of knowledge residing with users of the land. GISs, finally, allow integration of different types of data, coupling of databases with models, and construction of digital terrain models with unique opportunities to visualize landscape processes. Interactive use of GISs will enhance effective contacts with end users and policy makers. How will this new technology be used to improve our land evaluation practices? The above case studies already indicated some approaches, but this is only a start. A major challenge for the future is to use this modern technology in a meaningful way keeping in mind the overall objectives of the particular land evaluation being considered. There is a clear risk of a technology push where application of fancy new technologies becomes a purpose in itself. Continued contact with stakeholders and policy makers, as advocated above, is usually quite effective in keeping track of objectives.

34.9.2 Future Land Evaluation Based on Soil Studies

In many, so-called developed countries, standard soil surveys of agricultural lands at scale 1:20,000–1:50,000 have been completed. Revisions of old maps are being made, but funds are often not available to do this. Rather, surveys for specific purposes are increasingly commissioned, or maps are generated using available data that are available in well-accessible databases in many countries. Such data, rather than classical soil maps are then used for a wide variety of applications, using pedotransfer functions and modeling techniques and other means of interpretation. The use of modern GIS software allows rapid generation of flashy reports by nonsoil scientists who are hardly aware of the underlying processes and restrictions to be imposed on data and interpretations being used. Soil scientists should be quite aware of this and make sure that they inject and generate new data as needed to keep their evaluations relevant and up to date. If we only mine existing data, soil science will die (Bouma, 2009). Recent developments in digital soil mapping illustrate innovative state-of-the-art approaches for soil characterization and land evaluation, proving that soil science is still very much alive in contributing to land evaluation in a manner that is highly relevant for society at large now and in the future (Lagacherie and Mc Bratney, 2007; Hartemink et al., 2008).

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35

Hydropedology

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35.1 Introduction

Hydropedology is an emerging interdisciplinary science that grew out of the need to address complex natural processes in the earth's critical zone (Wilding and Lin, 2006). Increasingly, scientific research is being conducted by multidisciplinary teams as demanded by the complex, dynamic, and spatial–temporal variability of natural systems that require inputs from various disciplines (Sposito and Reginato, 1992). Hydropedology seeks to bridge disciplines to address (1) knowledge gaps between pedology, soil physics, hydrology, and other related bio- and geosciences; (2) scale differences in microscopic, mesoscopic, and macroscopic studies of soil and water interactions; and (3) data translations from soil survey databases into soil hydraulic properties (Lin, 2003).

Hydropedology is a union of hydrology, pedology, and soil physics disciplines with a focus on soil–water interactions. Hydrology is defined as *the science that treats the waters of the Earth, their occurrence, circulation and distribution, their chemical and physical properties, and their reaction with their environment, including their relation to living things* (NRC, 1991). Pedology is defined as *the branch of soil science that integrates and quantifies the formation, distribution, morphology, and classification of soils as natural or anthropogenically modified entities* (Wilding, 2000; Buol et al., 2001; Lin et al., 2006). Soil physics is defined as *the study of the physical properties of the soil and the relation of the soil physical properties to the study of the state and transport of matter and energy* (Scott, 2000). Although these definitions are specific and cover the breadth of each discipline, water dynamics, response, and the interactions with the environment are common underlying themes. As a need to unite these disciplines, hydropedology has been defined as *an intertwined branch of soil science and hydrology that encompasses multiscale basic and applied research of interactive pedologic and hydrologic processes and their properties in the variably-unsaturated zone* (Lin, 2003). More specifically, hydropedology focuses on the synergistic integration of pedology and hydrology to enhance the holistic study of soil–water interactions and landscape–soil–hydrology relationships across space and time. Its aim is to understand pedologic controls on hydrologic processes and properties and hydrologic impacts on soil formation, variability, and functions (Figure 35.1; Lin et al., 2008b). Even though hydropedology has its foundation in pedology, soil physics, and hydrology, it is also linked to other bio- and geosciences such as geomorphology, geology, geography, hydrogeology, hydroclimatology, ecohydrology, biology, and other branches of natural sciences (Figure 35.2). In a broader sense, hydropedology seeks to identify feedback mechanisms that allow for a holistic approach to the study and prediction of ecosystem functions.

This chapter provides an overview of the fundamentals and applications of hydropedology, including a review of some guiding principles and several recent advances in soil architecture and preferential flow, soil hydromorphology, scaling, digital soil mapping (DSM), pedotransfer functions (PTFs), and coupling biogeochemistry with hydropedology.

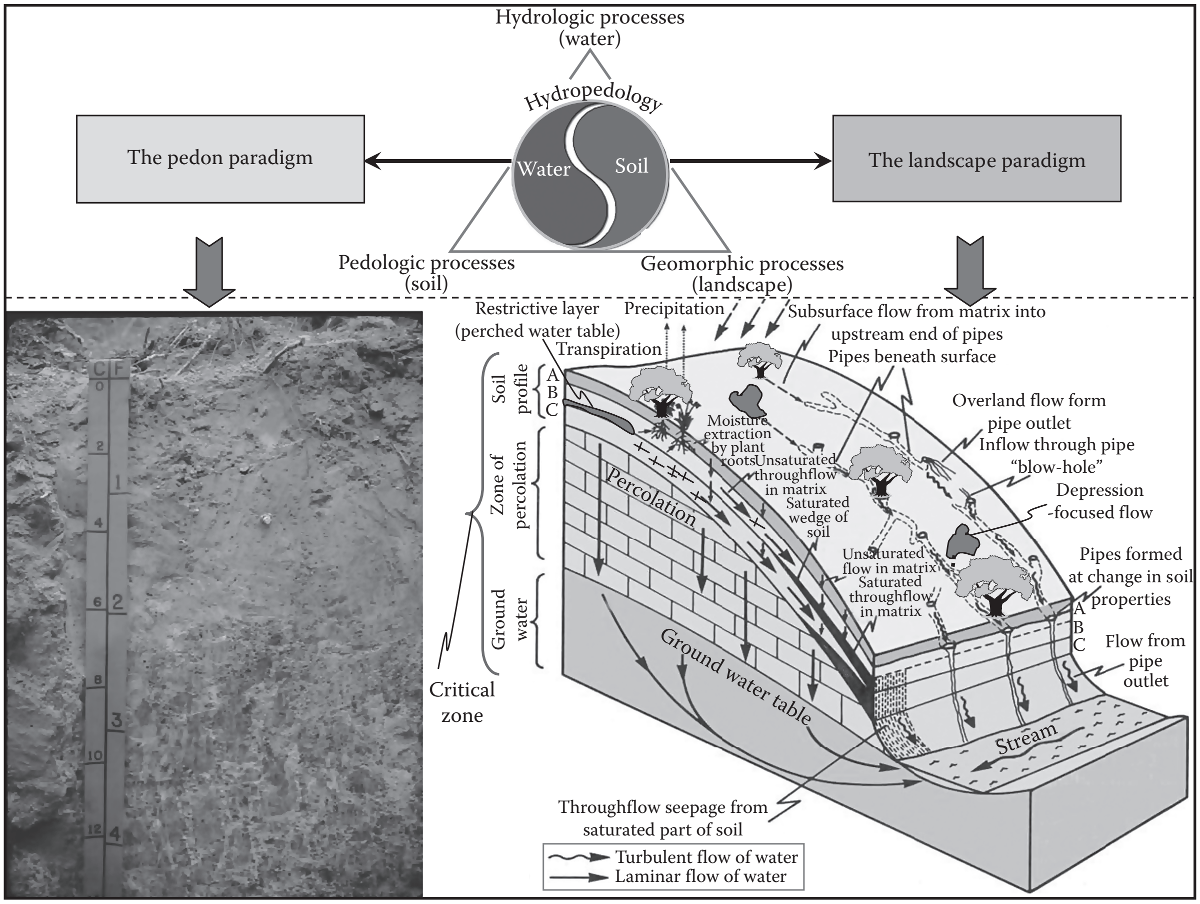

FIGURE 35.1 Hydropedology connects the pedon and landscape paradigms through linking phenomena occurring at the microscopic (e.g., pores and aggregates) to mesoscopic (e.g., pedons and catenae), macroscopic (e.g., watersheds and regional), and megascopic (e.g., continental and global) scales. (From Lin, H.S. 2010. Earth's critical zone and hydropedology: Concepts, characteristics, and advances. Hydrol. Earth Syst. Sci. 14:25–45; and Atkinson, T.C. 1978. Techniques for measuring subsurface flow on hillslopes. *In* M.J. Kirkby (ed.) Hillslope hydrology. John Wiley & Sons, Chichester, U.K.)

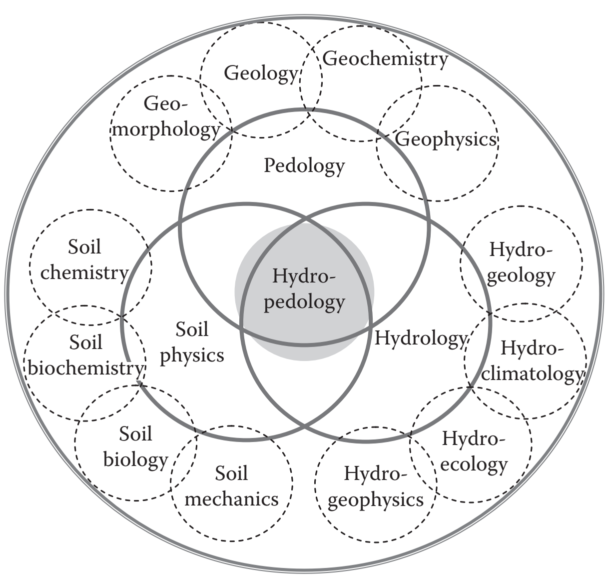

FIGURE 35.2 A conceptual diagram illustrating the relationship of hydropedology to other related disciplines. (From Lin, H.S. 2003. Hydropedology: Bridging disciplines, scales, and data. Vadose Zone J. 2:1–11.)

35.2 Fundamental Questions and Basic Characteristics of Hydropedology

Two fundamental questions of hydropedology are as follows (Lin et al., 2008b, 2008c):

1. How do soil architecture and the distribution of soils over the landscape exert a first-order control on hydrologic processes (and associated biogeochemical and ecological dynamics) across spatio-temporal scales?
2. How does water at the landscape scale (and the associated transport of energy, sediment, chemicals, and biomaterials by flowing water) influence soil genesis, evolution, variability, and functions?

Water at the landscape scale encompasses the source, storage, availability, flux, pathway, residence time, and distribution of water in the near-surface terrestrial environment. While source, storage, availability, and flux of water in the soil have been studied extensively in the past, attention to flow pathways (especially flow networks), residence time (age of water), and spatiotemporal pattern of flow dynamics (and its underlying organizing principle) has been much limited (Lin et al., 2006; McDonnell et al., 2007). Two basic characteristics of hydropedology are linked to the above two questions:

First, hydropedology emphasizes in situ soils in the landscape, where distinct pedogenic features (e.g., aggregation, horizonation, and redox features) and soil–landscape relationships (e.g., catena, soil distribution patterns, and soil map units) are essential in understanding interactive pedologic and hydrologic processes. Developing quantitative relationships between complex natural soil architecture and soil hydrologic functions across scales is an important research area of hydropedology. Three related key aspects are as follows:

- Hydropedology calls for a new era of soils research that is based on soil architecture (broadly defined as the entirety of how the soil is structured) that is beyond soil texture alone, so that the prediction of flow (and reaction) pathways, patterns, and residence times can be made realistically. This requires innovative techniques for improved quantification of soil architecture at different scales, especially in situ, noninvasively, and linking such soil architectural parameters to field-measured soil hydraulic properties.
- Hydropedology considers the soil in the real world as a "living" entity in the landscape rather than "dead" material that is manipulative mechanically. As Kubiena (1938) pointed out, a crushed or pulverized sample of the soil is related to the soil formed by nature like a pile of debris is to a demolished building. The full benefit of understanding this living system lies in studying them close to their natural setting in the real-world setting.
- Hydropedology attempts to link the form and function of soil systems across scales (Lin et al., 2006), rather than mapping soils without considering soil functions or modeling soils without incorporating soil architecture and soil–landscape patterns. Jenny (1941) clearly noted at the end of his famous book *Factors of Soil Formation*,

> The goal of soil geographer is the assemblage of soil knowledge in the form of a map. In contrast, the goal of the "functionalist" is the assemblage of soil knowledge in the form of a curve or an equation… Clearly, it is the union of the geographic and the functional method that provides the most effective means of pedological research.

- Such a union of soil maps and soil functions is what hydropedology hopes to promote in quantitative ways.

Second, hydropedology deals with the variably unsaturated zone in the terrestrial near-surface environment, including the shallow root zone, deeper vadose zone, temporally saturated soil zone, capillary fringe associated with groundwater table, wetlands, and subaqueous soils (soils formed in sediment found in shallow permanently flooded environments such as in an estuary; Demas and Rabenhorst, 2001). Three related key aspects are as follows:

- Hydrology has the potential to be an integrating factor for quantifying soil formation and evolution and for understanding soil changes upon global climate and land use changes (Lin et al., 2005). Hence, a focus on water can provide a potentially powerful means of quantifying and predicting dynamic soil functions.
- New ways of characterizing and mapping soils could or should be linked to hydrology, such as wetland boundary delineations, riparian zones, and hydropedologic functional units that are considered as soil–landscape units with similar pedologic and hydrologic functions (Lin et al., 2008b).
- The interpretation and quantification of soils as historical records of environmental changes in the past could be significantly improved if hydrologic data are considered simultaneously. This has been demonstrated in some paleosols and paleohydrology studies (Ashley and Driese, 2000).

35.3 Fundamental Scientific Issues of Hydropedology

At this stage of its development, the fundamental issues of hydropedology may be considered under the following four headings (Lin et al., 2005; Lin, 2010):

1. *Soil structure and horizonation in relation to in situ water flow and chemical transport*: Hydropedology emphasizes quantitative soil architecture of field soils across scales and their links to preferential flow in different spatial and temporal dimensions (see Section 35.4.1).
2. *Soil morphology and pedogenesis in relation to soil hydrology and soil change*: Hydropedology utilizes quantitative soil hydromorphology as a signature of soil hydrology and also uses soils as valuable records of environmental change over time (see Section 35.5).
3. *Soil catena and distribution pattern in relation to water movement over the landscape*: Hydropedology focuses on quantitative relationships between field soils and their surrounding landscape and the impacts of such relationships on hydrologic (and related biogeochemical/ecological) processes (see Section 35.6).
4. *Soil functions and maps in relation to carriers of soil quality and soil–landscape heterogeneity*: Hydropedology promotes quantitative delineations of functional soil units in the landscape as well as precision soil–landscape mapping for diverse applications (see Section 35.9).

35.4 Applications of Hydropedology

35.4.1 Soil Architecture and Preferential Flow

The natural soil "architecture" is of essence in understanding soil physical, chemical, and biological processes as well as landscape and ecosystem dynamics. It has been suggested that most that can be learned from sieved and repacked soil samples has already been done, so whatever we do with soil hydrology should be done on intact, undisturbed soils, and preferably in situ. A new era of soils research will have to rely on "structure-focused," passing the stage of "texture-focused," to achieve better ways of quantifying flow pathways, residence times, and spatiotemporal patterns of landscape water.

Soil architecture is used here broadly to mean the entirety of how the soil is structured, which encompasses at least three parts: (1) solid components, including soil matrix (represented by soil texture and soil microfabric) and soil aggregation (represented by the type, quantity, and size distribution of peds and aggregate stability); (2) pore space, including the size distribution, connectivity, tortuosity, density, and morphology of various pores; and (3) interfaces between solid components and the pore space, such as coatings on peds or pores, and the macropore–matrix, soil–root, microbe–aggregate, and horizons interfaces.

Since soil structure generally refers to a specific soil horizon, the broader term of soil architecture used here also encompasses the overall organization of a soil profile (e.g., horizonation), a soil's relationship with the landscape (e.g., catena), and the overall hierarchical levels of soil structural complexity (Lohse and Dietrich, 2005; Lin et al., 2005).

Soil horizonation or layering is ubiquitous in nature, so it must be adequately addressed when measuring, modeling, and interpreting hydrologic processes in watersheds. Various kinds and thicknesses of soil horizons and how they organize in soil profiles reflect longtime pedogenesis and the past and current landscape processes (Mausbach and Wilding, 1991). The fact that natural soils are layered has at least two significant implications for hydrology: (1) interface between soil layers of contrasting textures and/or structures would slow downward water movement, which often leads to some kind of preferential flow (e.g., fingering flow, macropore flow, or funnel flow); and (2) soil layering or discontinuity in soil hydraulic properties between layers would promote lateral flow or perched water table, especially in sloping landscapes with a water-restricting layer underneath.

A catena (also called toposequence) is a chain of related soil profiles along a hillslope, with about the same age, similar parent material, and similar climatic condition, but differs primarily in relief that leads to differences in drainage and soil thickness. Catenary soil development often occurs in response to the way water runs down the hillslope and recognizes the interrelationship between soil and geomorphic processes (Hall and Olson, 1991; Moore et al., 1993; Thompson et al., 1997; Lin et al., 2005). Catenae are often called hydrosequences of related soils, especially in depositional landscapes (Schoeneberger and Wysocki, 2005). Another important aspect of soil architecture along the hillslope is related to preferential flow network, which is further discussed later in this section.

While the importance of soil architecture across scales has long been recognized in soil science and hydrology, its quantification and incorporation into models have been notoriously lagged behind. This problem is due to many factors, including (1) inconsistent and fragmented concepts of soil structure; (2) overemphasis on ground-sieved soil materials and soil texture in the past, thus ignoring or downplaying the importance of the soil's "natural architecture." That natural soils are structured to various degrees at different scales is the rule, whereas the existence of a macroscopic homogeneity is the exception (Vogel and Roth, 2003); (3) lack of a comprehensive theory of soil structure/architecture formation, evolution, quantification, and modeling that can bridge orders of magnitude in scale and integrate physical, chemical, biological, and anthropogenic impacts. Although a hierarchical organization of soil aggregates has been well recognized (Tisdall and Oades, 1982; Vogel and Roth, 2003) and fractal characterization of soil structure has been proposed (Bartoli et al., 1998; Perrier et al., 1999), there is still a lack of means of representing field soil structure at different scales in a manner that can be coupled into models of flow, transport, and rate processes (Lin, 2003; Lin et al., 2005); and (4) lack of appropriate techniques and devices to quantify soil architecture directly, especially in situ noninvasively. Traditionally, soil structure has been evaluated by pedologists in the field using morphological descriptions or thin section observations, while soil physicists have employed wet and dry sieving, elutriation, and sedimentation to conduct aggregate analysis. In the absence of direct quantification, soil structure has been frequently evaluated by methods that correlate it to the properties or processes of interest (such as water retention, saturated hydraulic conductivity, infiltration rate, and gas diffusion rate). In recent years, noninvasive methods that permit soils to be investigated without undue disturbance of their natural architecture have become increasingly attractive, allowing 3D visualization of internal soil structure and its interactions with water. These methods include x-ray computing tomography, soft x-ray, nuclear magnetic resonance, gamma-ray tomography, ground-penetrating radar, and other methods (e.g., Anderson and Hopmans, 1994; Perret et al., 1999; Luo et al., 2008). Image analysis has brought new opportunities for analyzing soil structure, especially that of pores, their sizes, shapes, connectivity, and tortuosity (Vogel et al., 2002; Vervoort and Cattle, 2003). However, although numerous attempts have been made to find either statistical relations or deterministic links between soil structural data and hydraulic properties, a significant gap remains between in situ soil structure/architecture and field-measured soil hydraulic properties at different scales.

Quantifying soil architecture in the field across scales and at desirable spatial and temporal resolutions has been technologically limited (Tillotson and Nielsen, 1984). While landforms (e.g., digital elevation models or DEMs) and vegetation (e.g., land use/land cover) can now be mapped with high resolution (e.g., using LiDAR and IKONOS, respectively), there is a "bottleneck" phenomenon for in situ high resolution (e.g., submeter to cm) and spatially temporally continuous and noninvasive mapping or imaging of subsurface architecture including flow networks. This "technological bottleneck" has constrained our predictive capacity of many soil and hydrologic functions.

There is also a *conceptual bottleneck* that needs to be resolved for developing a new generation of hydrologic models. That is, should a continuous field or discrete objects be used to model surface and subsurface flow? Traditionally, hydrologic processes are generally conceptualized within the field domain (e.g., the Navier–Stokes equation and the Darcy's law) (Goodchild, 2007). Classical hydrology has applied findings from fluid mechanics, together with the necessary constitutive relations to develop sets of governing equations (much the same as atmospheric and ocean sciences have done). However, heterogeneities in land surface, hierarchical structures of soils, channel geometries, and preferential flow networks all make the land surface and subsurface different from the continuous field assumption (Kung, 1990; Noguchi et al., 1999) (CUAHSI, 2007; Lin, 2010). It is becoming more and more recognized that solid earth is not a continuous fluid; rather, it poses hierarchical heterogeneities with discrete flow networks embedded in both the surface and the subsurface. As McDonnell et al. (2007), Kirchner (2006), Beven (2002), and many others have noted, current models in watershed hydrology are based on well-known small-scale physics or theories such as the Darcy's law and the Richards equation built into coupled mass balance equations. It has been observed that the dominant process governing unsaturated flow in soils may change from matrix flow to preferential flow under different conditions when moving from the pore scale to the pedon scale (Blöschl and Sivapalan, 1995; Hendrickx and Flury, 2001). When moving from the pedon scale to the landscape/watershed scale, our knowledge for extrapolating the Darcy–Buckingham's law and the Richards equation to large heterogeneous area is further constrained (Weiler and Naef, 2003).

Because of heterogeneous soil architecture, variability in energy and mass inputs to soils (Addiscott, 1995; Warrick, 1998) diversity in biological activities, and nonlinear dynamics of hydrologic processes, preferential flow can occur in practically all natural soils and landscapes (Lin, 2010). As Clothier et al. (2008) summarized, preferential flow can occur spatially at the pore scale of spatial order 10^{-3} m, at the core scale (10^{-1} m), in pedons (10^{0} m), down hillslopes (10^{1}–10^{3} m) (Lehmann et al., 2007), through catchments (10^{4}–10^{5} m), and across large regions of $\geq 10^{6}$ m. Time-wise, preferential flow can operate during fluid flows at the temporal order of 10^{0}–10^{1} s, during hydrological events 10^{0}–10^{2} h, throughout seasonal changes 10^{0} year, and across interannual variations of 10^{1} years.

Based on three theoretical considerations and numerous published evidence, Lin (2010) has attempted to justify the likely universality of preferential flow in natural soils, inferring that the potential for preferential flow occurrence is everywhere in nature, although the actual occurrence of preferential flow depends on local conditions (Lin and Zhou, 2008). Lin (2010) also showed that networks are abundant in soils, such as root branching networks, mycorrhizal mycelial networks, animal borrowing networks, crack and fissure networks, artificial subsurface drainage networks, and pore networks between soil particles and aggregates. These networks provide preferential flow conduits, which in return reinforces or modifies the existing networks.

35.5 Soil Hydromorphology and Quantification

35.5.1 Water Tables and Inferred Soil Hydrology

Much of the research conducted in hydromorphology has focused on identifying the depth and duration of water tables to understand soil hydrology and the effects on pedogenesis and land use. Water tables are described as the top of the zone of saturation where water fills all or most of the pores in the soil to create saturated or satiated conditions (Daniels and Buol, 1992). The knowledge of soil hydrology has many practical applications for engineering, plant growth, reclamation, and remediation. The presence of free water in soils influences many related properties such as oxygen concentration (decreased aerobic respiration), reduction and transformation of redox couples (i.e., denitrification, metal reduction, and carbon dioxide production), change in solubility of compounds (bioremediation), change in soil pH, and decreased soil strength. Soils with high water tables generally lead to some challenges for land use such as decreased landscape stability (landslides), poor trafficability, limitations for onsite wastewater disposal, increased compaction, and poor fertility status (denitrification) in agriculture fields (Veneman et al., 1998). However, in more recent years, the benefits of wetlands and hydric soils have been recognized for their contribution to renovating contaminated waters and providing a source for groundwater recharge (Rabenhorst et al., 1998). Horizons are considered saturated when the soil–water pressure is zero or positive (Soil Survey Staff, 1999). The depth of the water table is related to multiple factors: (1) landscape position; (2) precipitation; (3) evapotranspiration; and (4) permeability of the surface and subsurface (Vepraskas, 1995). Understanding and interpreting the presence of water tables from landscape and soils information is important for understanding pedogenesis and land use. Soil morphology and redoximorphic features (formerly known as mottles) are used to imply the presence of water tables and inferred oxic–anoxic conditions.

35.5.2 Process of Redoximorphic Feature Formation

The presence of a water table results in a distinct morphology within a pedon and is the result of reduction–oxidation processes. The redox process results in the redistribution of primarily iron in natural soils, causing the soil to appear mottled. Although manganese is sometimes present and is an indication of alternating redox conditions, manganese is not as ubiquitous as iron. Additionally, the relationship between water tables and manganese is not as well correlated to water table durations and periodicity (Khan and Fenton, 1996; Fielder and Sommer, 2004). The iron segregations in areas within a soil are recorded as iron depletions and concentrations (Vepraskas, 1995; Vepraskas and Faulkner, 2001). The processes of reduction is driven by microorganisms; therefore, the soil conditions required to cause reduction are heterotrophic facultative aerobic and anaerobic microorganisms, carbon source, saturated conditions (to decrease diffusion of O_2), and a temperature above biologic zero.

Microorganisms use carbon sources to obtain energy for life. For a simplistic example, Equation 35.1 represents the use of glucose and Equation 35.2 represents oxygen as final electron acceptor (Ponnamperuma, 1972).

$$C_6H_{12}O_6 \dashrightarrow 2CH_3\text{–}CO\text{–}COO^- + 6H^+ + 4e^- \quad \longmapsto \text{ATP} \tag{35.1}$$

In aerobic soils, O_2 acts as the final electron acceptor as in Equation 35.2.

$$1.5O_2 + 6H^+ + 4e^- \rightarrow 2H_2O \tag{35.2}$$

As soils become saturated, the microorganisms continue to use oxygen until it is depleted. Once the oxygen is depleted, there are several other molecules that act as a final electron acceptor for the microorganisms. They are NO_3^-, Mn oxides, Fe oxides, sulfates, and CO_2. Given the temperature is above biologic zero, saturated conditions and a carbon source, the microorganisms through the production of electrons decrease the Eh, which creates conditions where reduced phase of the aforementioned compounds is thermodynamically stable.

The difference in solubility of iron and manganese provides the mechanisms for redistribution and formation of redoximorphic features. Iron is stable and insoluble in aerobic environments as Fe^{3+} valence state. Once iron is reduced from Fe^{3+} to Fe^{2+}, it is soluble and mobile and can be redistributed in the soil matrix. The solubility product constant for $Fe(OH)_3$ is 6.3×10^{-38}, whereas for $Fe(OH)_2$, it is 7.9×10^{-15} (Kotz et al., 1994). The reduced ferrous iron will remain as Fe^{2+} until it encounters enough higher redox potential O_2 to oxidize, commonly when the water table begins to drop allowing O_2 to enter the system and increase the Eh. The formation of iron oxide minerals often occurs in pore channels, on structure surfaces, and within the soil matrix (Vepraskas, 1995; Figure 35.3). Soils with a dominant gray matrix color are depleted of iron and display the color of minerals with minimal or no iron oxide coatings.

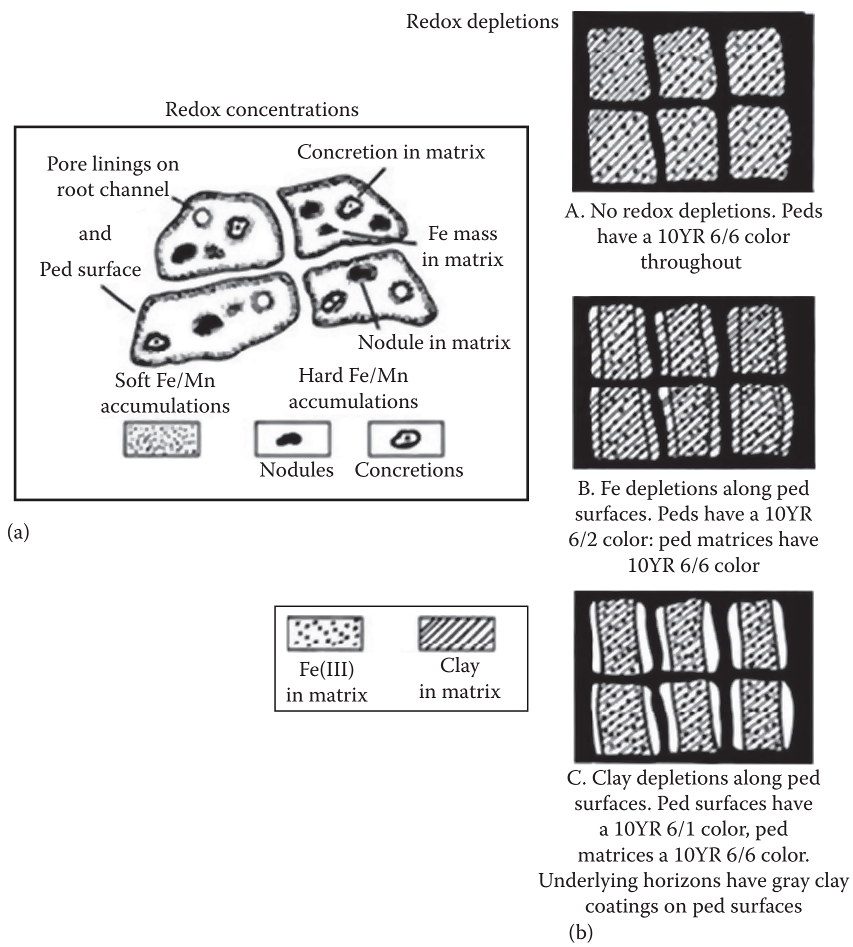

FIGURE 35.3 Schematic illustration showing different kinds of redox concentrations and their relationship to soil macropores and matrices (a). Schematic illustrations of redox depletions showing changes in color and texture as ped surfaces develop Fe depletions and clay depletions (b). (From Vepraskas, M.J. 1995. Redoximorphic features for identifying aquic conditions. Technical Bulletin No. 301. North Carolina State University, Raleigh, NC.)

35.5.3 Relationship of Redoximorphic Features and Soil Hydrology

Data on soil morphology and soil–water relationships are important because collecting water table information with wells and peizometers is time-consuming and expensive. There have been many studies relating soil morphology to the presence or absence of free water in soils (Daniels et al., 1971; Simonson and Boersma, 1972; Veneman et al., 1976; Franzmeier et al., 1983; Vepraskas and Wilding, 1983a, 1983b; Zobeck and Richie, 1984; Griffin et al., 1992; Jenkinson and Franzmeier, 2002). Researchers have focused on correlations between water table levels and morphological soil characteristics. One of the diagnostic redox features commonly used to define limitations due to the duration of water tables is the depth to chroma ≤2 depletions. Depth to gray mottles has been characterized extensively in many soils (Latshaw and Thompson, 1968; Simonson and Boersma, 1972). Yakovleva (1980) found a good general relation between depth to water table and degree of gleying in soils, and Daniels et al. (1971) found that the gray mottles were associated with different degrees of saturation in different landscape positions. More recent investigations by Evans and Franzmeier (1988) and Jenkinson et al. (2002) in Indiana and Megonigal et al. (1993) in South Carolina correlated length of soil saturation with color in alfisol soil toposequences derived from glacial till and a floodplain-terrace soil toposequence developed from tertiary bedrock, respectively. Thompson and Bell (1996) in Minnesota, observed a relationship between the surface accumulation of organic matter in mollisols and duration of saturation. Several researchers (Veneman et al., 1976; Vepraskas and Bouma, 1976; Richardson and Lietzke, 1983) have found that, in general, very brief periods of saturation are associated with high chroma peds with manganese and iron concentrations. Intermittent periods of saturation are associated with bright Fe concentrations, ped ferrans, and some Mn cutans. Saturation persisting for several months in a year is associated with <2 chroma matrix interiors, ped surface depletions, pore and root channel depletions, and virtually no Mn cutans or nodules. Since soil color patterns are strongly influenced by the chemistry and mineralogy of iron and manganese compounds, their solubility status determines the color of the soil matrix and argillans and determines the type and distribution of iron concentrations in a horizon.

Quantification of the time of saturation related to redox features has been reported in a few studies. In Indiana, Franzmeier et al. (1983) found the following soil morphology–soil–water relationships: (1) gleyed horizons with dominant colors of chroma ≤2 are saturated most of the year, (2) horizons with a dominantly brown matrix with gray depletions are saturated a few months of the year, and (3) horizons with a chroma of 5–6 that lack depletions ≤3 are seldom saturated during the growing season unless they were near a gleyed horizon. In study within the coastal plains of Georgia, Jacobs et al. (2002) found that soil horizons with chroma <2 were saturated >50% of the time. Horizons with low chroma iron depletions can be saturated as little as 18% of the time. In three representative floodplains in Maryland and Delaware, Vaughan et al. (2009) found that with lower temperatures, longer durations of saturation were needed for iron reduction with respect to ferrihydrite (FH). For temperatures greater than 90°C, less than 2 days were required for iron reduction with respect to FH; and as many as 20 days for iron reduction with soil temperatures between 1°C and 3.9°C (Vaughan et al., 2009).

However, these observations and measurements have been mostly limited to a point pedon scale and other research has shown that there are various degrees of expressions of redoximorphic features described that relate to movement and residency of water as controlled by landscape (Richardson et al., 1992; Jenkinson, 1998).

35.5.4 Soil Hydrology within Landscapes

Landscape hydrology has been the focus of research since the 1960s culminating with an intensive group of studies titled the Wet Soil Monitoring Project (WSMP). The WSMP was funded and coordinated by USDA–NRCS between 1990 and 2001. The studies were national in scope and included most climatic temperature and moisture regimes. This group of studies found that the geomorphic surface and the stratigraphy of soil horizons control water table depths in the soil, the amount of time the soil stays saturated, and the direction (vertical and or horizontal) water flows into or across the landscape. In turn, landscape hydrology governs the range of soil moisture regimes, oxidation–reduction processes, and soil color patterns (Arndt and Richardson, 1988; Griffin et al., 1992; Richardson et al., 1992; Hopkins, 1996; Thompson and Bell, 1996; Jenkinson, 1998, 2002; Feigum, 2000; Owens, 2001).

The presence and movement of groundwater through the landscape influences many of the physical and chemical processes that are involved in soil development. The water table, duration of saturation, and the flow of water across and through the landscape are controlled by topography and stratigraphy (Jenkinson, 1998). Richardson and Daniels (1992) concluded that changes in topography exert a strong influence on groundwater flowing through a landscape, and that landscapes can be subdivided into a system of groundwater recharge, flowthrough, and discharge that, under stable long-term conditions, can transfer soluble materials through the groundwater system in distances measured in kilometers. Richardson and Daniels (1992) observed these effects on landscapes in North Dakota and in the southeastern region of the United States. Thompson and Bell (1996) related the hydrological profile of soils along a toposequence in Minnesota to the geomorphology of the landscape. Daniels et al. (1971), in characterizing North Carolina landscapes, found a statistical correlation between water table depth and closeness to the dissected edge. These data indicated that shoulder positions within landforms are well drained due to lateral water movement and decreased residency time.

Additionally, the knowledge accumulated by the WSMP studies created a platform for the interpretation of hydrosequences, especially of soil drainage conditions, that are critically important when making land use decisions. The WSMP studies, gave researchers insights into what types of instrumentation worked well and what measurement processes could benefit from new technologies. Traditionally, water tables and anaerobic conditions induced by poor drainage conditions have been evaluated by wells, peizometers, tensiometers, open boreholes, alpha, alpha-dipyridyl, and platinum electrodes for Eh measurement. However, the Eh values obtained from platinum electrodes were generally considered circumstantial and occasionally suspect (Owens et al., 2008; Rabenhorst et al., 2009). Recent research by Rabenhorst et al. (2009) indicated that more accurate Eh values can be determined by using voltmeters with high input resistance (20 GΩ) and data loggers with standard configuration that maintains an open circuit during the instantaneous measurement. The need for a new method to address the issue of saturation and anaerobiosis together led to the development of a new device called indicator of reduction in soils (IRIS). IRIS was developed at Purdue University (Jenkinson, 2002) at the end of the WSMP period and further refined by Rabenhorst et al. (2008). IRIS mimics natural soil processes, visually indicates soil reduction, that can be quantified and be robust (Castenson and Rabenhorst, 2006). IRIS is a PVC tube coated with colored Fe minerals, mostly FH, that dissolves under reducing conditions (Jenkinson and Franzmeier, 2006). In practice, IRIS tubes are inserted into a soil, removed after a few weeks or longer, and evaluated by visually inspecting the coating in the field. If the coating was not dissolved, no reduction occurred, but if it was dissolved, reducing conditions must have prevailed (Figure 35.4). Another field technique using zero valent iron rods was developed to infer O_2 concentration (Owens et al., 2008). Iron rods were polished and placed in toposequences and compared with O_2 measurements, piezometer data, and Eh measurements. The coatings on iron metal rods correlated well with specific oxygen ranges. Rods in soils with O_2 concentrations below about 3 mg L^{-1} did not develop bright (7.5YR 4/4–5/8) oxide/oxyhydroxide coatings but instead formed black (10YR 2/1–2/2) coatings. Rods in soils with O_2 concentrations between about 2% and 5% developed variegated bright (7.5YR 4/4–5/8) oxide/oxyhydroxide coatings indicating microsite differences in O_2 concentrations. Rods in soils with O_2 concentrations above about 5% with adequate moisture were almost completely coated with bright (7.5YR 4/4–58) iron oxide/oxyhydroxides. This method provides a simple and inexpensive means to qualitatively estimate the ranges of O_2 status in soils.

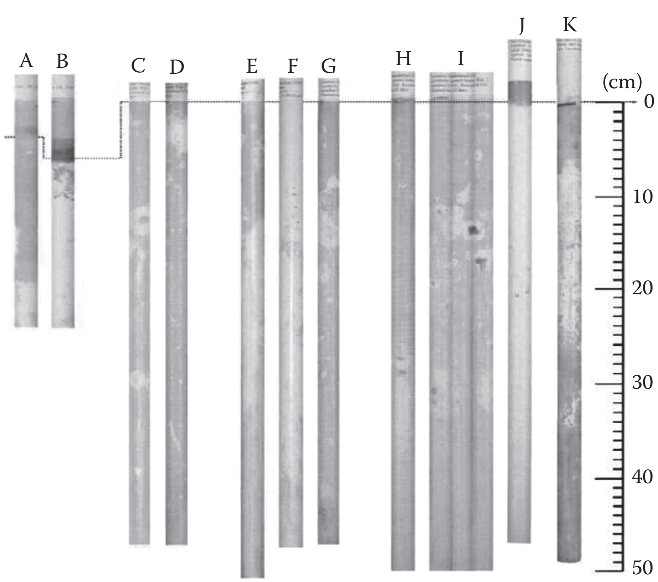

FIGURE 35.4 Photos of IRIS illustrating the degree of iron reduction on the rods. The degree of reduction and stripping of iron relates to the intensity and duration of saturated conditions in the soil.

Many soil hydrology studies were conducted at the pedon scale and along catenae. In these studies, soil–water tables were highly correlated to landscape position or topography of the landscape. The term hydrosequence is used to a series of soils with different degrees of wetness due to topography (Schoeneberger and Wysocki, 2005). Soil moisture regime and presence of water tables can be attributed to several factors; however, precipitation and topography are the two most dominant factors at field scale. Other important factors contributing to water table dynamics are the underlying stratigraphy and particle size of the parent material. Water may flow through hydraulically conductive upper horizons and "perch" on horizons with low hydraulic conductivity. In sandy parent materials with high hydraulic conductivity, the presence of a water table may not be evident.

Studies have indicated that within similar climate conditions, topography and parent material stratigraphy are the two most important controlling factors of water tables. In central Iowa, Khan and Fenton (1994) found that within an internally drained glacial till landscape, soils in depressions had longer duration of water tables with higher organic carbon when compared to higher topographic soils. Soils in the depressions also had different chemical properties. In the wettest part on the landscape, Bkg horizons formed where calcareous water was discharged. Similarly Richardson et al. (1992) found that flownet analysis based on stratigraphy and water table position was useful for describing observed differences in soils related to recharge and discharge positions in the landscape. Considering the soils in the context of the landscape, Richardson et al. (1992) identified four kinds of water movement that dominate soil development: (1) recharge or dominantly downward flow of water; (2) flowthrough or lateral groundwater movement; (3) discharge or movement from the water table either to or near the soil surface; and (4) stagnation or slow water movement creating water table mounds. This analysis indicated that leached soils and absence of carbonates were found in the recharge soils. Soils within discharge zones and flowthrough positions had accumulations of carbonates and gypsum reflecting the addition of carbonate-rich water at lower landscape positions from higher landscape positions (Figure 35.5). Rolling topography with steep hillslopes commonly have complex shallow hydrologic system where water moves laterally above a limiting layer (Daniels and Buol, 1992). Evans and Franzmeier (1988) found water flowing laterally on hillslopes remained oxygenated and did not lead to reducing conditions. Therefore, the interpretations of redoximorphic features must be placed in the context of landscapes.

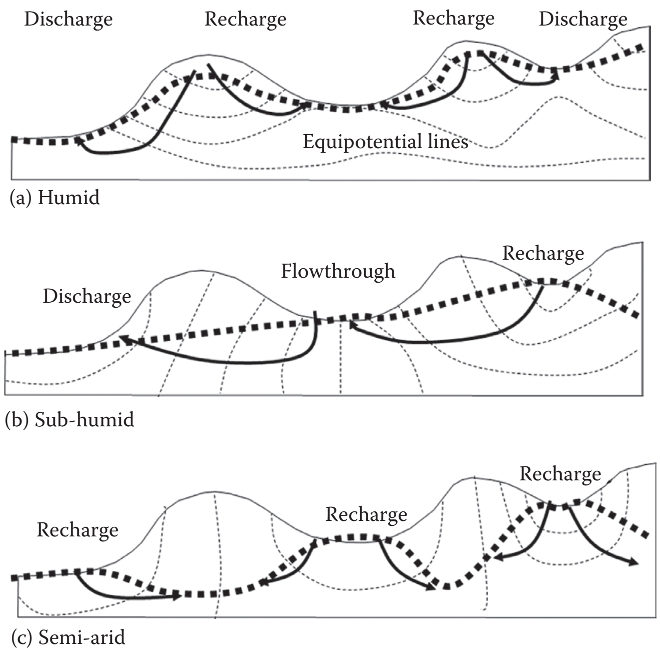

FIGURE 35.5 An examination of soils from (a) Iowa, (b) North Dakota, and (c) Saskatchewan found that the soils in the depressions varied such that the soils from Iowa indicated discharge, North Dakota had recharge in the higher areas and flowthrough and discharge in the lower areas. The Saskatchewan had a dominance of recharge wetlands with argillic horizons. (From Richardson, J.L., L.P. Wilding, and R.B. Daniels. 1992. Recharge and discharge of groundwater in aquic conditions illustrated with flownet analysis. Geoderma 53:65–78.)

Utilizing hydropedology concepts in soil morphologic studies is promising because soil properties that influence water distribution can be better understood and water–soil morphology relationships can be placed in the landscape context. The availability of digital information coupled with the ability to display spatial information will provide the potential to make significant advances for quantifying the spatial–temporal relationship between soil morphology and soil hydrology within larger landscapes. This will be discussed further in Section 35.9.

35.6 Coupling Hydropedology and Biogeochemistry

A biogeochemical cycle is a pathway by which a chemical element moves through biotic (biosphere) and abiotic (lithosphere, pedosphere, atmosphere, and hydrosphere) compartments of the earth system. Biogeochemical cycles are inseparable from the hydrologic cycle (NRC, 1991), and fluxes of elements and rates of biogeochemical cycling are often linked to hydrologic conditions (Richardson et al., 2001). Zones of enhanced fluxes and reaction rates where terrestrial and aquatic ecosystems meet have been recognized, or suspected, for decades (Lohse et al., 2009). *Biogeochemical hot spots* are patches in the landscape that show disproportionately high reaction rates relative to the surrounding matrix, whereas *hot moments* are defined as short periods of time that exhibit disproportionately high reaction rates relative to longer intervening time periods (McClain et al., 2003). Hot spot and hot moment activity is often enhanced at terrestrial–aquatic interfaces. For example, using examples from the carbon and nitrogen cycles, McClain et al. (2003) showed that hot spots occur where hydrological flow paths converge with substrates or other flow paths containing complementary or missing reactants. Hot moments occur when episodic hydrological flow paths reactivate and/or mobilize accumulated reactants.

Since hydrology often triggers hot spots and hot moments of biogeochemical reactions and ecological functions (Bundt et al., 2001; McClain et al., 2003), improved modeling and prediction of soil architecture and preferential flow will have considerable implications for determining chemical fluxes and calculating elemental budgets in soils and ecosystems. Interpretation of point measurements without knowing preferential flow paths is now often questioned (Gottlein and Manderscheid, 1998; Netto et al., 1999), because the uncertainty of whether soil solution is extracted from stagnant or high velocity flow paths makes it practically impossible to reliably determine mass flux rates. Additional complications arise in structured soils for reactive components due to locally variable chemical conditions. In addition, macropore linings and aggregate coatings restrict lateral mass transfer and reduce sorption and retardation; hence, physical and biochemical nonequilibriums are enhanced. All these suggest the need to identify and model preferential flow pathways and networks in real-world soils if we are to improve the modeling and prediction of interactive soil physical, chemical, and biological processes.

At the pedon scale, soil aggregation, macropore networks, horizonation, and the soil–bedrock interface have important impacts on preferential flow and biogeochemical reactions in soil profiles (Deurer et al., 2003; Field et al., 1984; Flury et al., 1994). At ped surfaces, for example, C and N contents and microbial activities are higher than in those in ped interiors (Tisdall, 1995; Alef and Kleiner, 1989; Kavdir and Smucker, 2005). Soil layering can trigger on or off preferential flow between soil horizons with different textures and structures, which tend to promote lateral flow in sloping landscapes. The soil–bedrock interface has also been recognized as important subsurface preferential flow pathway (Li et al., 1997; Freer et al., 1997; McGlynn et al., 2002). Flühler et al. (1996) have suggested three regimes of flow and transport within a soil profile during a preferential flow process: (1) lateral distribution flow in the *attractor zone* where preferential flow is initiated; (2) downward preferential flow in the *transmission zone* where water moves along preferential flow pathways and bypasses a considerable portion of the soil matrix (Quisenberry et al., 1993; Vervoort et al., 1999); and (3) lateral and downward dispersive flow in the *dispersion zone* where preferential flow pathways are interrupted and water flow becomes more or less uniform again.

Based on a review of studies documenting soil microbial biomass distributions with depth (Balkwill and Ghiorse, 1985; van Gestel et al., 1992; Dodds et al., 1996; Murphy et al., 1997; Richter and Markewitz, 2001; Blume et al., 2002; Taylor et al., 2002; Fierer et al., 2003), the system of Flühler et al. (1996) can be applied. Various studies have demonstrated a consistent pattern of changing microbial biomass size and activity distribution, characterized by three broad zones with distinct biogeochemical potentials that correspond to the three flow zones of Flühler et al. (1996): (1) *attractor zone*, equivalent to the A horizon and representing high-biomass surface soils under the strong influence of plant roots, moisture, and temperature; (2) *transmission zone*, equivalent to B and C horizons and representing lower more spatially heterogeneous biomass than in the *attractor zone* (Konopka and Turco, 1991; Rodríguez-Cruz et al., 2006); and (3) *dispersion zone*, representing higher-biomass, water-capture zones at the soil–bedrock interface or right above a water-restricted soil layer because of moister and more nutrient-rich conditions resulting from impeded or lateral flow (Dupuy and Dreyfus, 1992; Buss et al., 2006). These three zones are distinguished from "deep-subsurface" aquifers and geologic materials, many of which have extremely low biomass (Balkwill et al., 1985; Kieft et al., 1995). Contributions of greenhouse gas fluxes from the *attractor zone* have been studied intensively (Konopka and Turco, 1991; Krasovskaia et al., 2003; Rodríguez-Cruz et al., 2006), whereas contributions from the *transmission and dispersion zones* are essentially uncharacterized and will likely depend on their connectivity to the atmosphere or on preferential gas flux associated with preferential water distribution.

35.7 Scale Dependency of Soils and Modeling

As previously discussed, a lot of work has been conducted on measuring and understanding of preferential flow at point pedon scale and its role on soil development and hydrological process at landscape scale; however, a much broader context especially in terms of varying scales is needed. The spatial and temporal scales have been long recognized by soil science (Jenny, 1941; Soil Survey Division Staff, 1993; Wilding, 2000; Soil Survey Staff, 2003; Sposito, 1998; Tugel et al., 2005). Different schemes of soil classifications have captured soil variability at different scales from field to hillslope, landscape, watershed all the way to regional continental and world scales (Soil Survey Division Staff, 1993; Boul et al., 2003; Soil Survey Staff, 2004). Various criteria and principles have been used to classify soils and their properties across various scales represented by different modes such as state factor model (Jenny, 1941), process model (Simonson, 1959), energy model (Runge, 1973), statistical landscape models (Shovic and Montagne, 1985; Bell, 1990; Zhu et al., 1996, 1997, 2001, 2006; Havens, 1998; Zhu, 2000), and geostatistical models (Odeh et al., 1992, 1994; McBratney et al., 2003). The evolution of soil models along with that of tools and purposes of soil mapping have highlighted some of the issues related to the current soil maps and soil survey in general. The issues identified encompass a wide range related to the type and level of soil information provided and current delivery format of soil data (McBratney et al., 2003; USDA–NRCS–NGDC, 2006; Hempel et al., 2008; MacMillan, 2008).

The majority of the soil information comes from field descriptions of pedons and laboratory measurements of soil samples collected from these pedons and it is of three types: (1) quantitative, (2) qualitative, and (3) semiqualitative/quantitative. The field descriptions include qualitative statements about soil horizons, soil structure, consistency, color, and texture; quantitative statements about soil horizon thickness, redoximorphic features, degree of mottling, and nutrients; and semiqualitative statements about roots and pores among others (Schoeneberger et al., 2002). The need to use this information in hydrologic modeling has led to the development of PTFs as a way to translate qualitative and semiqualitative soil information to quantitative one (Bouma, 1989; Pachepsky et al., 1999; McBratney et al., 2002; Pachepsky and Rawls, 2003, 2004; Wagenet et al., 1991). The value of PTFs has been widely recognized because of their multiple applications in estimating and predicting soil hydraulic properties (McBratney et al., 2000; Lin et al., 2005; Pachepsky et al., 2006) and also for their use in modeling in general (McBratney et al., 2002). McBratney et al. (2002) points out that PTFs are typically limited to specific geomorphic regions or soil types and lack quantifiable uncertainty. Such efforts have led to the emerging of hydropedology as a promise to bridging pedology and hydrology (Lin et al., 2005; Bouma, 2006; Pachepsky et al., 2006, 2008). These advancements have highlighted one of the major challenges related to the use of PTF and hydropedology, that is, the transferability of the qualitative, quantitative, and semiqualitative/quantitative field measured and estimated soil information across multiple scales (Bouma, 2006). In his discussion about different conceptual frameworks and methods that have been developed to address this issue, Bouma (2006) challenges the soil scientists for the fact that soil expertise is not adequately represented in modeling. Bouma (2006) recognizes the fact that pedologists in general are not comfortable with representation of soils in terms of homogeneity and isotropy needed by hydrologists for modeling purposes. Evidence shows that certain measured hydraulic soil properties display both stochastic and deterministic behavior (Harlan and Franzmeier, 1974; King and Franzmeier, 1981; Evans and Franzmeier, 1986; Franzmeier, 1991; Haws et al., 2004; Zeleke and Cheng Si, 2005; Botros et al., 2009). Franzmeier (1991) measured K_{sat} on different soils across Indiana and found similar values by grouping soils based on the parent materials, soil horizon, and soil texture categories indicating the presence of stochastic behavior of K_{sat} within each category and deterministic behavior across categories. Haws et al. (2004) used the representative measurement area (RMA) concept to describe the behavior of measured K_{sat} and found that an RMA greater than 400 cm^2 was needed to filter out smaller-scale heterogeneities and a distance between 120 and 200 m to capture the effective or deterministic behavior of K_{sat} at landscape scale. The conclusion of the study was that in modeling water flow into and through soils, the measured K_{sat} should integrate across the range of variability at the scale of interest. Iqbal et al. (2005) using geostatistical analysis (kriging and structured variograms) of soil physical properties like soil bulk density (ρb), saturated hydraulic conductivity (K_s), and soil–water content (θ) showed that soil texture displayed spatial structure or

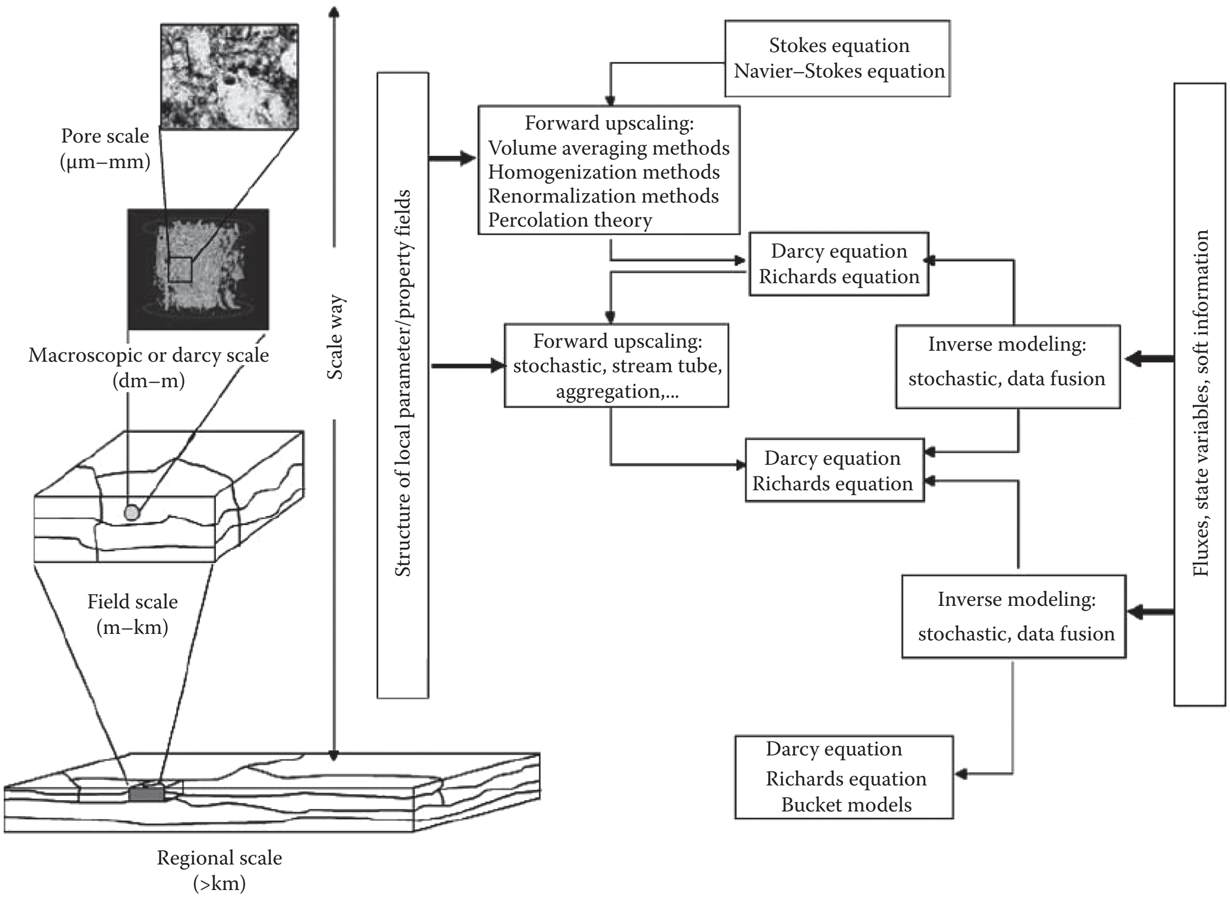

FIGURE 35.6 Forward and inverse upscaling methods to upscale soil–water flow processes from the local to the field scale. (From Vereecken et al., 2007.)

autocorrelation within a 400 m distance while hydraulic properties and bulk density within a 100 m distance beyond which the distribution of measured values displayed random behavior. Similar results have been found by other researchers (Ersahin, 2003; Kozak and Ahuja, 2005; Vereecken et al., 2007) for soil infiltration rates and other soil properties (Iqbal et al., 2005). Vereecken et al. (2007) discusses the "scaleway" approach first introduced by Vogel and Roth (2003) in upscaling soil hydraulic properties across multiple scales (Figure 35.6) and provides a summary of the upscaling methods.

Methods like fractals have also been proposed for upscaling soil hydraulic properties (Shepard, 1993; Peyton et al., 1994; Baveye and Boast, 1999; Rodriguez-Iturbe and Rinaldo, 1997; Perret et al., 2003) along with other approaches. The method is based on the concept of self-similarity of geometrical shapes displayed by disordered systems that appear similar over a range of length scales. Such structures are self-similar and are referred to as geometrical fractals (Perret et al., 2003). The characteristic of a geometrical fractal is its fractal dimension, D, which can be 1, 2, and 3 for 1D, 2D, and 3D geometric shapes. The dimensions can be determined by measuring the length, area, or volume depending on the geometric shape. The method has been used to quantify pore volumes and pore tortuous paths and relate them to soil hydraulic conductivity (Shepard, 1993). The fractal methodology approach can also be useful for understanding and upscaling soil properties using pattern recognition.

The majority of the discussions about soil spatial variability has been focused on soil properties especially hydraulic properties (Grayson and Bloschl, 2000; Ersahin, 2003; Kozak and Ahuja, 2005; Corwin et al., 2006; Vereecken et al., 2007). However, the majority of soil properties are represented spatially by soil series and soil map units. In this context, two levels of information could be distinguished to characterize the representation of spatial distribution of soil property data, that is, within and between soil map units. Research conducted on deriving certain soil hydraulic properties from soil series and soil map units (Ersahin, 2003; Kozak and Ahuja, 2005) has highlighted some issues related to the spatial representation of these properties. The majority of soil property data has been measured or estimated at point or pedon scale and transferred to polypedon, soil series, and soil map unit scales (Soil Survey Division Staff, 1993). The spatial distribution of these transferred properties comes in the form of representative values (Soil Survey Staff, 2004) or mean values with ranges with no indication about their spatial distribution within soil map units (Libohova et al., 2010). Furthermore, often ranges between soil series and soil map units overlap rendering them not statistically significant. In addition, soil map units are not always pure with only one representative soil as they could be complexes and/or associations with smaller inclusions of other soil types (Soil Survey Division Staff, 1993). Taxonomy and management related interpretations have both played a role in the current design of soil map units. Least but not last, political boundaries and time differences between soil survey conductance has introduced further variability that may not necessarily reflect the reality. How much of this variability is real and how much is overrepresented and could be simplified is a question that relates directly to the representation of spatial variability of soil properties. Efforts to address these issues have led to the development of most recent concepts like "functional homogeneity" versus "structural heterogeneity" of soil series and soil map units for parsimonious watershed modeling and predictions (Basu et al., 2009). The presence of overlapping ranges of soil properties between soil map units constitutes at the same time recognition of their continuum spatial variability from the soil scientists and surveyors. It is a reflection of the purpose of soil surveys and the limitations of the tools used at the time of soil survey campaigns. Recent developments in spatial tools and technologies offer an opportunity to significantly improve the current delivery format of spatial soil information (McBratney et al., 2003). U.S. Soil Survey Program has adapted the major land resource area (MLRA) concept (Indorante et al., 1996) and started the so-called edge-matching in an effort to create seamless soil maps for the conterminous United States. European Union has launched similar efforts despite the surmountable obstacles due to differences in soil classification systems, time, and purpose of the soil survey conductance and resistance to share data deemed often a matter of national security for some countries (European Communities, 2007). The most recent global soil map initiative aimed at creating the first world digital soil map (McMillan, 2009) has taken these efforts to another level and created an opportunity for the soil surveys around the world to come together like never before.

A plethora of methods have been developed to aid the efforts for creating spatially continuous soil maps and especially soil property maps from the existing soil information. McBratney et al. (2003) provide an extensive review of methods developed for DSM by establishing quantitative relationships between soil properties or classes and their "environment" often referred to as DEM-derived terrain attributes. Some of the methods discussed include generalized linear models, classification and regression trees (CART), neural networks, fuzzy systems, and geostatistics (McBratney et al., 2003). These and other methods have been necessary to address the existence of different levels and types of soil information and in some cases the lack of soil information. The major drives for these developments have been the great progress in geographic information systems (GIS), GPS, remote sensing, and data sources.

DEMs coupled with increasing demands for soil data and information from environmental sciences and modeling (McBratney et al., 2003). Soil science is now faced with the challenge of looking beyond its own domain and sharing paths more than in the past with other disciplines such as ecology and hydrology, just to name few (Lin et al., 2005; Wilding and Lin, 2006). This may require soil scientists to become familiar with the concepts and terminology used by other disciplines in order to establish mutually beneficial communication. A closer look at hydrological and/or ecological sciences reveals similarities with soil science for engaging in modeling different aspects of natural process being soil, water, vegetation, and wildlife (Corwin and Wagenet, 1996; Laurenson, 1974; Ibanez et al., 1995; Sayfried and Wilcox, 1995; Phillips, 2001; Austin, 2002; Burke and Lauenroth, 2002; Epstein et al., 2002; Kie et al., 2002; Shirazi et al., 2003; Hopmans and Schoups, 2005; Lu et al., 2005; Sivapalan, 2005; Logsdon et al., 2007; McDonnell et al., 2007; Wagener et al., 2007; Simmons et al., 2008). A revisit of different concepts used by these disciplines in the light of spatial variability representation is very important as "a better understanding of spatial soil variability, its development over time (pedogenesis) and its functional relationships to recent processes in soil landscapes is one of the biggest challenges in soil science" (Sommer, 2006).

In soil science, the pedon definition and its spatial limits are used to separate soils within a small-scale pattern of local variability, usually 1–10 m^2, while polypedon is presented as a unit of classification or a relatively homogenous soil body whose limits are extremely hard to find (Soil Survey Division Staff, 1993). Because of this difficulty, the polypedon concept has been ignored by most soil scientists due to its circulatory nature, and, thus, its properties and extent have been transferred to the pedon. The concept has been further expanded spatially by the use of soil series and soil map units that are viewed often as a collection of polypedons (Soil Survey Division Staff, 1993). The practical applications of these concepts have been difficult in representing the soil as true continuum. This has resulted in soils and soil map units with overlapping ranges for most of the soil properties (Lin et al., 2005), thus introducing more spatial variability than could be present. Using the analogy from Logsdon et al. (2007), the spatial variability within soil map units would be considered as random fluctuations or noise. However, the spatial variability between soil map units may or may not be random depending on the soil property and the predicted function (Libohova et al., 2010). The distinction between these two spatial variability categories is scale dependent (Logsdon et al., 2007). Hydrology and hydrologic modeling in particular utilizes similar concepts but uses terminology such as "deterministic length scale" (Sayfried and Wilcox, 1995), representative elementary area (REA) (Wood et al., 1988; Wood, 1999) closely related to the representative elementary volume (REV) (Peck, 1983), and RMA (Haws et al., 2004) used in soil science in order to emphasize scale dependence of variability. Other terms like "stochastic," "deterministic," "homogenous," and "heterogeneous" are used in the same manner in describing hydrologic processes; and the distinction among them is scale dependent (Haws et al., 2004). Sayfried and Wilcox (1995) acknowledge that spatial variability is "rarely entirely deterministic or stochastic" as noted also by other researchers (Laurenson, 1974; Vogel et al., 2000). Both trends are found in every hydrologic parameter, and the dominance of one or the other has been found to be scale dependent. In this context, the authors describe the definition of "deterministic length scale," meaning that depending on the scale, spatial variability of any parameter can be characterized as deterministic, thus the variability can be represented by a single value or function rather than a range of values with some probability distribution as in the stochastic case. Ecological sciences use terms like "numerical richness" (Kempton, 1979) and "species density" (Hurlbert, 1971). Ibanez et al. (1995) relates these terms to the ratio of the number of different soil taxa over the total number of pedons described and the number of soil taxa per sampling area, respectively. Burke and Lauenroth (2002) discuss the challenges related to representation of the regional scale variability facing the ecological sciences especially the limitations of local-scale and short-term studies in capturing regional scales. The common feature of discussions about the spatial distribution of soils or soil properties, hydrological process, or ecological species diversity is the fact that they are all bound within a certain spatial domain and the nature of their variability is scale dependent. The focus so far has been on spatial variability, but the same can be stated for temporal variability (Tugel et al., 2005; Sommer, 2006; Logsdon et al., 2008) especially when considering the human impact as Basu et al. (2009) showed on the role of tile drains on controlling hydrologic response in the Midwest. The difference in scale representation of variability can be addressed, and soil structure and hydrologic functioning at different scales can be related within the same spatial units as pointed by Pachepsky et al. (2008).

The continuous soil maps and soil property maps most likely will be in a raster format; thus, the selection of the appropriate pixel resolution is also important. The selection of pixel size is also scale related and depends on the soils and soil properties to be inferred. Soil properties manifest varying degrees of spatial dependence (McBratney and Webster, 1983) as do the terrain attributes (Gessler et al., 1995), and they need to be taken into consideration; otherwise the derived terrain attributes may have no meaning and the soil pattern or process could be lost as Moore et al. (1991, 1994) have found. McBratney et al. (2000) provides a review of resolutions and scales used for soil survey and suggests pixel sizes of >2 km, 20 m^{-2} km, and <20 m for national to global, catchment to landscape, and local extents, respectively. In a more detailed overview, McBratney et al. (2003) provide pixel resolution based on the order and scale of the soil survey and identify the 5–20 m pixel resolution as the suggested one for U.S. soil survey order 1 and 2 corresponding to scales 1:5,000–1:20,000. Winzeler et al. (2008) found that a pixel size resolution between 5 and 10 m in relatively flat glaciated landscapes in Northern Indiana can capture the local relief and the spatial distribution of selected soil properties without losing valuable information on patterns and variability at the landscape scale, avoiding at the same time unnecessary microrelief details often described as "noise."

The challenge for soil science is not only to accurately and completely represent the spatial variability of soil properties but also to relate this variability to scales. For hydrologic purposes, this means that soil morphological differences need to be viewed also from the hydrologic processes perspective, thus leading to the generation of soil functional hydrologic property maps with a focus on commonality rather than morphological differences. The approach parallels that of the catchment hydrology major challenge described by Sivapalan (2005) and McDonnell et al. (2007) as a shift from prescribing patterns of catchment heterogeneity to focusing on the search for (1) geomorphic or landform processes that generated these patterns; (2) processes that underlined this heterogeneity and complexity; and (3) the ecological, pedological, and geomorphological functions of these processes. Just as the focus on pattern, process, and function will revolutionize hydrology (Sivapalan, 2005), the same might be stated for soil science in general and soil modeling in particular. This means that mapping soils and soil map units, soil morphological and taxonomic differences as well as soil heterogeneity, need to be linked with scales, hydrological, and soil-forming processes in order to look at the soil map units from the functional in addition to compositional perspective. The variability is everywhere, and before any attempts to characterize it quantitatively, we should ask questions about "purpose" and "scale" or "boundary conditions" as there may not be a universal solution or a silver bullet to address all the variability due to the fact that the nature of variability (stochastic, deterministic, chaotic, or homogenous) is scale dependent as are the hydrological process in hydrology and other disciplines from ecology all the way to social sciences. Recent developments in DSM (McBratney et al., 2003), coupled with PTFs and the emergence of hydropedology (Pachepsky et al., 2006), provide a platform for addressing the scale dependency of soil variability related to modeling.

35.8 Pedotransfer Functions

"Data rich, information poor" (information here connotes interpretation, synthesis, and utilization of data) has been a common syndrome in many disciplines (Lin, 2003). This problem is largely due to data fragmentation, incompleteness, incompatibility, inaccessibility, or lack of interpretation and synthesis in spite of past extensive and costly data collections. In soil science and hydrology, it is recognized that gaps exist between what is available (e.g., the National Cooperative Soil Survey [NCSS] Program databases) and what is needed (e.g., soil hydraulic parameters and PTFs needed for simulation models). Improved procedures to extract useful information from the available databases and to improve and interpret soil survey data for flow and transport characteristics in different soils are needed.

With the increasing popularity of coupling GIS with vadose zone models and soil survey databases for diverse natural resource applications, the demand for soil hydraulic information has increased significantly in recent years. However, the lack of sufficient field data on soil hydraulic properties often limits the application of contaminant transport and hydrological modeling. Existing methods for direct field measurement of soil hydraulic properties remain complex, time-consuming, and costly (Mualem, 1986; Bouma, 1989), despite decades of work by soil physicists, hydrologists, and others representing different disciplines. Another limitation of direct field measurement is significant spatial and temporal variability, hence demanding a large number of measurements that are often prohibitive in terms of time and money (van Genuchten et al., 1999b). This has prompted efforts to indirectly estimate soil hydraulic properties using more readily available data often found in soil surveys (such as particle-size distribution, bulk density, organic matter content, and others). Such indirect methods, now often referred to as PTFs as suggested by Bouma and van Lanen (1987, 1990), have been attempted for estimating water retention curve parameters, saturated hydraulic conductivity, the unsaturated hydraulic conductivity function, and other soil hydraulic parameters (Vereecken et al., 1990; Batjes, 1996; Lin et al., 1999b; Tietje and Hennings, 1996; van Genuchten et al., 1999a; Wösten et al., 2001; Shaw et al., 2000). Compared to other methods of estimating soil hydraulic parameters (e.g., pore-size distribution models and inverse methods), PTFs are inexpensive, easy to derive and use, and, in many practical cases, they provide good estimators for missing hydraulic parameters (Verhagen and Bouma, 1998; van Genuchten et al., 1999b; Wösten et al., 2001). Besides conventional regression or functional analyses, new techniques such as neural networks (Schaap et al., 2001), group methods of data handling (Pachepsky and Rawls, 1999), and CART (McKenzie and Jacquier, 1997) are increasingly being explored for developing PTFs using a growing number of large soil databases such as the NCSS databases, UNSODA (Leij et al., 1996), HYPRES (Wosten et al., 1999), WISE (Batjes, 1996), SoilVision (SoilVision Systems Ltd., 2002), and many others.

While various degrees of success have been achieved with different PTFs (Pachepsky et al., 1999; Wösten et al., 2001), limitations of existing PTFs remain. For example, the vast majority of existing PTFs are completely empirical, and limited efforts have been put into systematic probing of underlying mechanisms for the existence of such functions. There is a tendency among soil scientists and others to estimate soil hydraulic functions by regression analysis in a given region, but efforts to apply PTFs derived from one area to soil hydrologic studies in other soil regions are futile (Kutílek and Nielsen, 1994). Only a few quasi-physical PTFs exist, such as those by Arya and Paris (1981), Haverkamp and Parlange (1986), and Arya et al. (1999), in which particle-size distribution is first translated into an equivalent pore-size distribution model and then further related to water retention curve. However, the flow system in the bundle of capillary tubes differs considerably from the water flow network in real-world soils (Miller and Miller, 1956; Beven and Germann, 1982). Thus, the practical application to field soils of these quasi-physical PTFs, based essentially on the Hagen–Poiseuille's law, is very limited. In addition, existing PTFs have not fully incorporated soil structure and land use information and have lacked scale and temporal considerations. As such, the accuracy, reliability, and utility of existing PTFs are constrained. In the mean time, the NCSS databases developed over the past century have been underused in addressing environmental issues. Interpretations and applications of the NCSS databases are challenges facing soil scientists in general and pedologists in particular. There are pressures on both pedologists and soil physicists/hydrologists to disseminate soil survey information and utilize it in a variety of applications (Simunek et al., 2003).

The combined efforts of pedologists, soil physicists, and hydrologists could open up new opportunities for the next generation of PTFs. For instance, five general categories of PTFs may be identified for potential improvement in estimating dynamic soil properties (Figure 35.7). PTF type I relates use-dependent

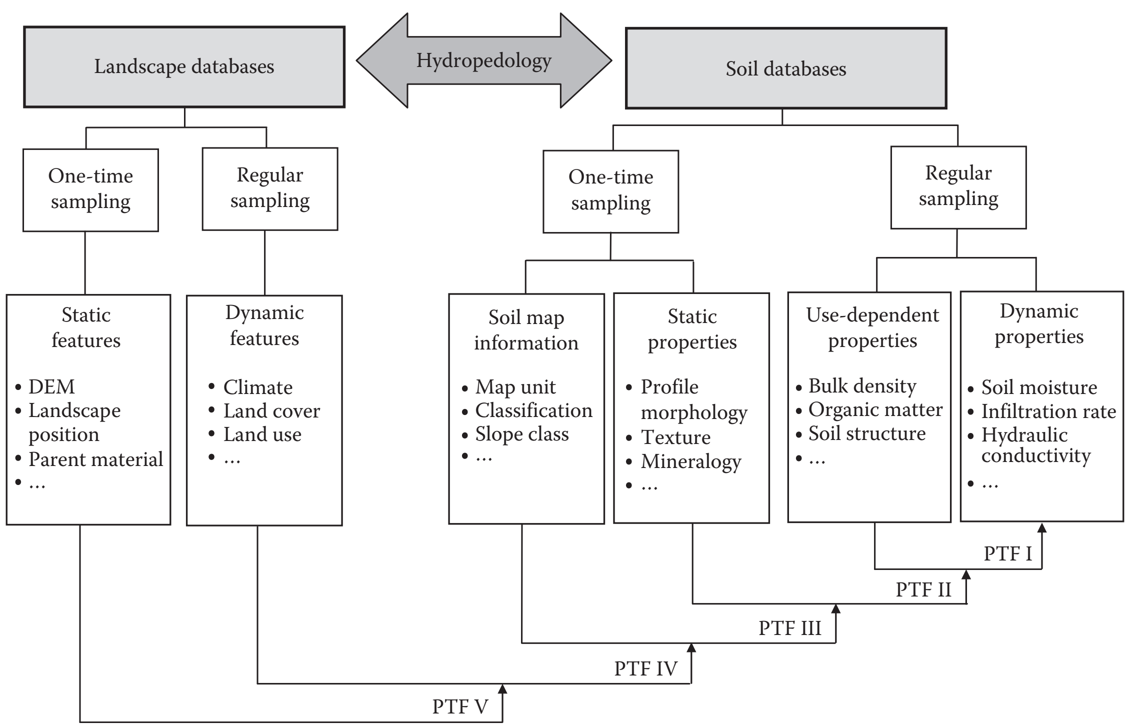

FIGURE 35.7 Five general categories of PTFs utilizing different types of input data for estimating dynamic soil properties. Landscape features such as DEMs, land use/land cover, and others serve as additional inputs to PTFs, hence connecting the pedon and the landscape scales.

soil properties to soil hydraulic information, both of dynamic nature requiring regular sampling. PTF type II includes relatively static soil properties that could be sampled only once into the prediction of dynamic soil properties. PTF type III further considers soil mapping and classification-related information to improve the prediction. Landscape features such as DEMs, land use/land cover, and others could serve as additional inputs in PTF type IV and V, hence connecting the pedon and landscape scales. It is likely that PTFs in combination with routine spatial information from soil survey, topography, and land use could improve the regional estimates of soil hydraulic parameters. In terms of land use and use-dependent soil properties, Droogers and Bouma (1997) suggested the terms "genoform" for genetically defined soil series and "phenoform" for soil types resulting from a particular form of management in a given genoform. Such distinction between major soil management types within the same soil series could potentially enhance PTFs that involve soil series and land use as carriers of soil hydraulic information. Realizing the importance of dynamic and use-dependent soil properties, the NCSS program is now considering the possible development of a dynamic soil properties database, which, once developed, would significantly facilitate the enhancement of PTFs.

Recent development in DSM coupled with earlier work on PTFs (Bouma, 1989; Bouma et al., 1999) has improved significantly the ability of soil science to transform itself from a semiquantitative to a quantitative discipline. This opportunity presents challenges at the same time, especially those related to understanding the scale dependency of hydrologic processes and their relation with spatial distribution of soil properties.

35.9 Digital Soil Mapping with Hydropedologic Concepts

In order to assess hydrologic function, soils and hydrologic processes must be fully understood within the context of the landscapes in which they occur. Soil serves as an interface between the solid and fluid envelopes, which dissipate mass and energy fluxes at the earth's surface (National Research Council, 2001). The majority of United States and World Soil Surveys are built upon soil–landscape model (Hudson, 1992), which recognizes the fact that in many landscapes, soil development occurs in response to the water movement through and over the landscape (Moore et al., 1993). This model limits the analysis of soil development at the scale where all the other soil-forming factors, namely, climate, parent material, organisms, and time, vary less compared to relief or topography. Jenny's model of five main soil-forming factors is expressed in the following equation:

$$S = f(cl, o, r, p, t\ldots) \tag{35.3}$$

where S, soil, is considered to be a function of climate (cl), organisms (o), relief (r), and parent material (p) acting through time (t).

Aside from scientific value of the model, its widespread use may have been also a result of technical and practical considerations at the start of the Soil Survey Program related to purpose of the survey, scale, and resolution of the available spatial information. The use of Jenny's conceptual model leads to the development of maps relating soil pedogenesis and properties (Gessler et al., 1995) to landscape positions in a qualitative way (Walker et al., 1968; Furley, 1976; Daniels et al., 1985; Stone et al., 1985; Kreznor et al., 1989; Carter and Ciolkosz, 1991). The conventional soil maps reflect this approach in delineated soil polygon map units that represent the spatial continuum of soil properties with single values and ranges with no clear indication about their spatial distribution within map units. Soil scientists have recognized the inability to characterize the continuum and have compromised by providing polygon values with overlapping ranges for many soil properties. The use of this information in its current format presents a challenge for hydrologic modeling that requires a spatial continuum representation of soils and their properties (Libohova et al., 2010) as they modify material and energy fluxes at the earth's surface (Gessler et al., 1995). The challenge is to use hydropedologic principles to develop numerical and spatially realistic soil–landscape models useful for a variety of purposes beyond taxonomic classification (McSweeney et al., 1994). However, soil–landscape processes and relationships express themselves over a wide range of spatial and temporal scales, which reflect past climates, truncations by overriding processes, and perhaps feedback mechanisms (Malanson et al., 1990), and often exhibit different and complex scale-dependent variations (Butler, 1964; Beckett and Webster, 1971; Burrough, 1993; Wilding et al. 1994; Wilding and Drees, 1983). The scale dependency of soil variation indicates that more soil-forming factors may be at work at a particular location offering flexibility in incorporating other factors, especially location for explaining soil variability. Soil variability often relates well with terrain attributes, thus offering a potential use as spatial predictors of soil properties at specific locations (Gessler et al., 1995). With the dawning of GIS and powerful georeferencing tools, specific georeferenced spatial location has become a key consideration in the understanding of soils and their variability. A new formalization of the factors equation offered by McBratney et al. (2003) gives the following equation:

$$S_a = f(s, c, o, r, p, a, n) \tag{35.4}$$

where S_a is a set of soil attributes at a given location understood as a function of other observed soil attributes at either the present location or other proximal locations (s), climate (c), organisms (o), relief (r), parent materials (p), age or time (a), and spatial location or position (n). The formula offers an expansion of the equation discussed by Jenny (1941) to include soil properties observed at specific locations. It implies that, by using geostatistical tools, formal relationships between soils at measured distances can be developed to strengthen soil models. This is not a new idea; Jenny himself spent a great deal of time validating spatial soil relationships (Hudson, 1992), but it is a new formulation of the importance of spatial relationships and their power to help explain soil properties. With available soil polygon data and soil point data, quantifiable terrain attributes (now often easily derived from DEMs in a GIS platform), and remotely sensed earth-surface-reflectance data, numerical analysis of soil–landscape relationships are possible to a much higher degree of accuracy and precision than ever before. The SCORPAN (S = soil, C = climate, O = organisms, R = topography, P = parent material, A = age and N = spatial position) model reflects the new emphasis on spatial location as a key attribute capable of providing predictive power in soil models. The increase recently in spatial resolution of DEMs and other spatial data coupled with the emerging of powerful terrain analysis software and geostatistical tools has provided an opportunity for soil scientists and digital soil mappers to numerically characterize soil–landscape models to develop raster-based predicted soil maps and soil property maps (Ryan et al., 2000). Soil–landscape model provides the appropriate scale for DSM to characterize soil variability as attributed to soil–water movement and distribution. This means that when climate, parent material, vegetation, and time are relatively constant, topography can explain the majority of differences in patterns of soil distribution. Topography itself is not the differentiating factor. However, topography controls water movement, which in turn affects pedogenesis. The use of terrain attributes as derived from topography captured by DEM provides the means to quantitatively represent the spatial distribution of soils and their related properties.

As previously discussed in the scale dependency of soils for modeling section, environmental modeling and management require explicit and quantitative spatial prediction of soil and landscape attributes (Gessler et al., 1995). Spatial variability affects hydrologic response over a wide range of scales (Sayfried and Wilcox, 1995; Robinson and Sivapalan, 1997) and is characterized as either stochastic and/or deterministic depending on the scale and the hydrologic process. This, in turn, means that any hydrological model has to represent soil model input parameters related to the scale of the research question, which can only be accomplished through DSM. Physically based distributed hydrologic models utilize spatially distributed inputs such as topography, soil type, and vegetation that in theory could provide a better description of hydrological processes (Bathurst and O'Connell, 1992). However, the amount of data input requirements could constitute a disadvantage for the physically based hydrologic models compared to lumped parameter models (Sayfried and Wilcox, 1995), and it may not always lead to better model predictions of hydrologic processes (Beven, 1987; Loague, 1990; Wicox et al., 1990; Anderson et al., 1992; Grayson et al., 1992).

The development of high-resolution spatial data and increase in computing power has led to demands for higher-resolution input data for hydrological modeling, especially for physically based distributed hydrology models. However, problems related to the limits of detail representation and parameterization of subgrid scale processes (within pixel) will still remain (Beven, 2001) and need to be addressed in the context of multiscale heterogeneities to identify their broad scale or general patterns (Sivapalan, 2005). The discussion on hydrological parameters, multiscale heterogeneities, and their scale dependency pattern manifestation is closely related to the stochastic versus deterministic representation of spatial variability of hydrological parameters. This definition resonates with the soil–landscape model that could be considered as the "deterministic length scale" for describing the spatial variability of soils and their related properties. Both trends are found in every parameter, and the dominance of one or the other is scale dependent as previously discussed, which means that the future efforts in characterizing spatial variability of soil properties in a raster format as input to hydrological models should consider other scales in addition to the scale of soil–landscape model. This approach utilizing DSM is not limited only to hydrology and can be applied to other areas like ecology and environmental sciences and would lead to the development of soil property maps that emphasize their functions in relation to the process being hydrological, ecological, or other applications (Forman, 1995).

New spatial technologies (including GPS and GIS) have allowed the soil survey to georeference all of their recent soil samples as part of their soil characterization database (Soil Survey Staff, 2009) developed by the Soil Survey Laboratory (SSL), National Soil Survey Center. This database provides access to the raw characterization data, including morphology, saturated hydraulic conductivity, depth of soil, and bulk density. Using the source data to produce continuous maps of hydrologically important soil parameters should more clearly link real measurements with landscape position and provide a better set of soil parameters for hydrologic modeling. Given the recent developments in digital technologies like GIS, GPS, and spatial terrain analysis software coupled with DSM, spatially distributed raster-based predicted soil property maps for modeling can now be produced (McKenzie and Gallant, 2007; McKenzie and Ryan, 1999), validated, and tested with hydrologic models.

35.10 Conclusions

In 2001, the U.S. National Research Council (NRC, 2001) released a report entitled Basic Research Opportunities in Earth Science. This report highlighted the need for understanding the earth system and identified the pedosphere as an area crucial to the critical zone function. Hydropedology can contribute significantly to the holistic study of the earth's critical zone because of the key roles that soil and water play together in the critical zone's evolution and functioning. Understanding and quantifying the interactions between water and soil is critical for understanding multiscale ecosystem functions that sustain life. The future land use planning and use of natural resources under changing climate require multidisciplinary approaches where prediction and forecast based on quantifiable metrics must be achieved. Wilding and Lin (2006) pointed out a common agenda where soil science has been identified as a major stakeholder in the growing interests in bio- and geosciences as highlighted in the NRC (2001) report. This agenda was to (1) broaden constituencies beyond traditional partners; (2) expand focus in the near-surface "critical zone" to include food security, food safety, ecosystem management, biosphere sustainability, environmental protection, and urban environment; (3) enhance fundamental knowledge of earth systems that is more systematic, interdisciplinary, dynamic, and process oriented; (4) identify early warning systems of natural hazards and resource degradation; and (5) develop joint research and education partnerships that attract coupling of public and private support. As described in this chapter, hydropedology provides one important mechanism where linking disciplines with a common theme can provide the avenue for a greater understanding of complex and dynamic interactions between the pedosphere and the hydrosphere.

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36

Subaqueous Soils

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36.1 Introduction and Historical Development of Subaqueous Soil Concepts

One of the new frontiers in soil science that has come into focus over the past two decades has been the study of subaqueous soils. Although the concept has appeared in the literature at times (Kubiëna, 1953; Muckenhausen, 1965), only recently have these substrates received recognition in the United States and as such are now accommodated under the definition of soils (Soil Survey Staff, 1999). Under the new definition, soils may occur in permanently flooded or ponded environments with water depths up to approximately 2.5 m (Soil Survey Staff, 1999).

Previous opposition to the concept of subaqueous soils was primarily focused upon the idea that subaqueous substrates are not, in fact, soils but sediments. The ruling dogma was that by definition a soil must be able to support the growth of plants (Soil Survey Staff, 1975). Thus, the essential absence of higher plants in many subaqueous environments excluded these substrates from the pedologic realm. A secondary issue was related to the defining boundaries of a soil. The first edition of *Soil Taxonomy* (1975) stated that the upper limit of soils is "…. air or shallow water." In this sense, "shallow water" was meant to exclude soils permanently under water. Thus, these materials were also excluded from being soil by their permanent inundation. Over the 25 years that spanned the development of the second edition of *Soil Taxonomy* (Soil Survey Staff, 1999), pedological thinking continued to evolve such that pedologists began distancing themselves from their agricultural roots with a loosening of the link between the definition of soils and the growth of plants. Rather, pedologists began to emphasize what had already become deeply entrenched as the foundation to the taxonomy itself, namely, the formation of horizons resulting from those generalized pedogenic processes described by Simonson (1959). For example, Bockheim (1990, 1997) and Campbell and Claridge (1987) showed that in the harsh-cold climate of Antarctica, where higher plants are not able to grow, soil horizons were still observed as a result of pedogenic processes (i.e., additions, losses, transfers, and transformations). Thus, the idea that these areas should be recognized as soils was gaining support among the pedologic community, even though they were not necessarily capable of supporting the growth of higher plants.

Much of the credit for the emergence of subaqueous soils as a field of soil science has to be given to Dr. George P. Demas. The story goes that the concept formed in George's mind as he was standing on the edge of the marsh in Maryland that he was mapping, and looking down into the shallow tidal water of Sinepuxent Bay, he posed the question "Why should we stop mapping here?" He began to consider that such submersed aquatic vegetation as eelgrass (*Zostera marina*) and widgeongrass (*Ruppia maritima*) was rooted in these substrates (Figure 36.1) and as he began to closely examine them observed what could be construed as pedogenic horizons. Soon afterward he published his paper "Submerged Soils: A New Frontier in Soil Survey" in *Soil Survey Horizons* (Demas, 1993). Over the next 6 years under the guidance of Dr. Martin Rabenhorst, Demas further developed the ideas and concepts for the characterization, formation, and mapping of subaqueous soils in his PhD dissertation (Demas, 1998).

The works of Demas led to a number of additional studies that form the basis for most of the discussion in this chapter.

FIGURE 36.1 Eelgrass meadow growing on a shoal in 1.5 m of water. The soil supporting the eelgrass is a sandy skeletal Typic Haplowassent. (Photo Jim Turenne.)

In particular, the most important findings or accomplishments included in the work and dissertation were as follows:

- Subaqueous soils form as a result of the generalized processes of additions, losses, transfers, and transformations (Demas and Rabenhorst, 1999). This led to a change in the definition of soil to include substrates that are permanently under significant water (approximately 2.5 m) and that show evidence of pedogenesis (Soil Survey Staff, 1999).
- Bathymetric maps could be constructed to use as a soil survey base map and identify subaqueous landforms in a manner analogous to the subaerial landforms that soil scientists had been studying for most of the past century (Demas and Rabenhorst, 1998).
- Similar soils occurred or formed on similar landforms. Therefore, the "soil–landscape" paradigm typically used to map subaerial environments could be applied to the mapping subaqueous soils, and, thus, specific soil–landscape relationships began to be documented for the coastal lagoons of the Mid-Atlantic United States.
- The establishment of the first official soil series for subaqueous soils: Demas,* Sinepuxent, Southpoint, Tizzard, Trappe, and Whittington.

Kubiëna (1953) was the first to use the term "subaqueous soils" to describe permanently inundated soils. Those soils composed of layers and forming in low-energy subaqueous environs were classified into four groups (Kubiëna, 1953). Most of the focus was on soils having considerable soil organic matter: dy, gyttja, and sapropel; terms often applied to substrates in limnological studies (Saarse, 1990). Horizon sequences were typically A, AC, and C regardless of the soil type. Dy soils formed below water columns that were acidic, nutrient poor, and having high concentrations of soluble organic compounds. These soil materials have a gel-like form indicative of amorphous organic matter. Gyttja forms in subaqueous soils rich in nutrients. The majority of the materials are coprogenic in origin, having a loose arrangement (Jongerius and Rutherford, 1979) typical of high n-value soil materials (low bearing capacity; Soil Survey Staff, 2010). These subaqueous soil materials have also been referred to as sedimentary peat, coprogenous earth, or limnic materials (see Fox, 1985; Soil Survey Staff, 2010). Subaqueous layers that contain various amounts of more or less unrecognizable organic debris that are enriched in sulfides were termed sapropels. Most of these sulfides are Fe-monosulfides or pyrite in the solid form or hydrogen sulfide gas as recognized by the rotten egg smell. Colors are typically black that changes to gray upon drying.

The classification of subaqueous soils by Kubiëna (1953) was cited in the national soil classification system in Germany (Muckenhausen, 1965), but there is no evidence that these soils were the focus of any serious investigation. A decade later, although he did not elaborate nor focus much on what he called "subaquatic soils," Ponnamperuma (1972) did affirm that these soils forming under water reflected "horizon differentiation distinct from sedimentation." Nevertheless, between the publication of this paper and the time when Demas began to focus on these systems more than two decades later, apparently little attention was paid to subaqueous soils.

* Originally proposed as Wallops but posthumously named Demas following the untimely death of innovator Dr. George P. Demas in December 1999.

36.2 Soil Genesis in Subaqueous Environments

In addition to the generalized model of soil genesis (additions, losses, transfers, and transformations) described by Simonson (1959), pedologists have often invoked the state factor equation of Jenny (1941) to describe and conceptually model the formation of soils. While Jenny's model has limitations, it recognizes the contributions of various soil-forming "factors." In considering the genesis of subaqueous soils, similarities to the processes and factors described by Jenny (1941) were recognized, but significant differences were also noted. The generalized model for estuarine sediments of Folger (1972) was noted where he described their origin as being derived from source geology (G), bathymetry (B), and hydrologic condition (H) (flow regime).

$$S_e = f(G, H, B) \tag{36.1}$$

The concepts of both of these previous equations were joined, with some further modifications, into a state factor equation to describe the formation of subaqueous soils (Demas and Rabenhorst, 2001).

$$S_s = f(C, O, B, F, P, T, W, E) \tag{36.2}$$

where
S_s is subaqueous soil
C is the climatic regime
O is organisms
B is bathymetry
F is flow regime
P is parent material
T is time
W is water column attributes
E is catastrophic events

Climatic regime (C) was not included in Folger's equation and does not include precipitation as in Jenny's model. The climatic component in this model primarily represents temperature. Temperature, for example, will affect the rate of organic matter decomposition (and other biogeochemical reactions).

Organisms (O) was also not included by Folger and represents the role that biota play in subaqueous pedogenesis. As an example, the burrowing of benthic organisms (essentially irrigating their burrows with oxygenated water) often contribute to the development of light-colored, surface horizons, as well as the obvious contributions of plant carbon to the upper soil horizons.

Bathymetry and flow regime (B and F) replace relief (R) in Jenny's equation. The catena concept per se is not applicable in a permanently submersed environment because relief or topograph do not play the same role as in subaerial soils. Bathymetry contributes to the effects of internal and wind-generated waves on the subaqueous soil surface. Flow regime helps to shape underwater topography and accounts for differences in the energies associated with currents and tides. Together, these two factors (B, F) essentially play the same genetic role as relief does in subaerial soil environments.

Parent material (P) was a factor in both the equations of Folger and Jenny and explains the effect of the source material on subaqueous soil profile attributes. For example, subaqueous soils that form in areas where they receive barrier island washover materials are predictably sandy textured.

Time (T) of course represents the amount of time available for the expression of subaqueous soil attributes.

Water column attributes (W) was not included in either Jenny's or Folger's equations and has been added to include variations in the chemical composition of the water column that could have an impact on subaqueous soil characteristics. Those subaqueous soil profiles developed in freshwater regions or fresh portions of estuaries will likely be significantly different than those formed in more saline or brackish environments where sulfate is available for reduction to sulfide and the potential formation of solid phase sulfide minerals. Similarly, the dissolved oxygen levels in the water column could dramatically impact the formation or the thickness of light-colored, oxidized, surface horizons.

Catastrophic events (E) is included in this equation to account for the possibility that subaqueous soil profiles may be dramatically impacted by major storm events or other uncontrollable or unknown factors. The effects of storms or modest hurricanes, however, do not seem to cause wholesale alterations to large areas of subaqueous soils.

36.3 Mapping, Research, and Agency Efforts

Subaqueous soils research and mapping projects have been completed or are in progress along the eastern seaboard of the United States from Maine to Florida and along the Gulf Coast in Florida and Texas (Table 36.1). Projects have covered a range of topics including mapping protocols, soil–landscape relationships, carbon sequestration, soil variability, pedogenesis, use and

TABLE 36.1 Summary of Current or Completed Subaqueous Soils Projects

Investigators	Affiliation	Location	Project Focus	Publications
Demas	UMD/NRCS	Sinepuxent Bay, MD	Soil survey methods and pedogenesis	Dissertation, Demas (1993); Demas et al. (1996); and Demas and Rabenhorst (1998, 1999, 2001)
Bradley	URI	Ninigret Pond, RI	Soil survey and eelgrass methods	Thesis, Bradley and Stolt (2002, 2003, 2006)
Flannagan	UME	Taunton Bay, ME	Soil survey	Thesis, Osher and Flannagan (2007)
Jespersen	UME	Taunton Bay, ME	Carbon accounting	Thesis, Jespersen and Osher (2007)
Angell	UMA	Freshmeadow Pond, MA	Soil survey	Report
Ellis	UFL	Cedar Key, FL	Soil survey	Dissertation
Fischler	UFL	Indian River Inlet, FL	SAV	Thesis
Casby-Horton/Brezina	NRCS	Padre Island, TX	Soil survey and ecological site descriptions	—
Payne	URI	Greenwich Bay, RI	Water quality methods	Thesis
Coppick	UMD	Rehoboth Bay, DE	Soil survey	Thesis in progress
Balduff	UMD	Chincoteague Bay, MD	Soil survey methods	Dissertation
Keirstead/Hundly	NRCS	Little Bay, NH	Soil survey	—
Surabian/Parizek/McVey	NRCS	Little Narragansett Bay, CT, RI	Soil survey and mooring interpretations	Report, Surabian (2007)
MapCoast	MapCoast	Rhode Island estuaries	Soil survey methods	Web available data
Salisbury	URI	Quonochontaug Pond, RI	Shellfish and dredging interpretations	Thesis in progress
Pruett	URI	Point Judith Pond, RI	Eelgrass and carbon accounting	Thesis in progress
Wong	NCSU/NRCS	Jamaica Bay, NY	Soil survey and eelgrass	Thesis

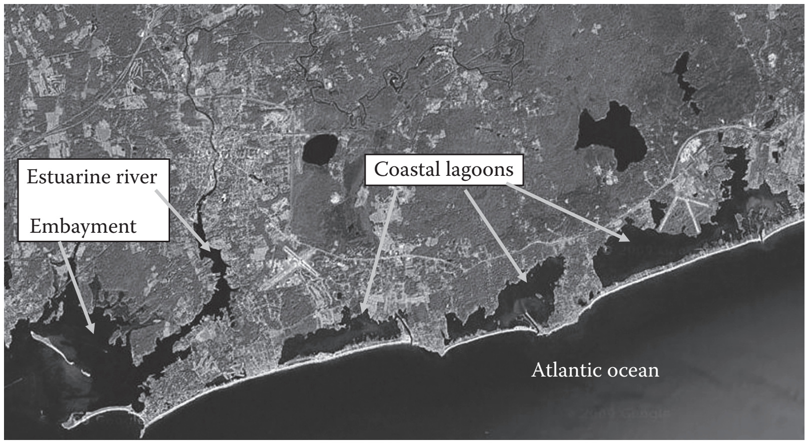


FIGURE 36.2 Examples of embayment, coastal lagoon, and estuarine river systems typically examined in subaqueous soil studies. Water depths generally average less than 2 m at low tide. Image was taken from a 2009 Google map of the southern Rhode Island and Connecticut shoreline.

management interpretations, and relationships between subaqueous soils and submerged aquatic vegetation (SAV) and water quality. Study and mapping areas have primarily concentrated on estuarine areas such as coastal lagoons and shallow water embayments (Figure 36.2). In response to mapping efforts, regional and national subcommittees within the National Cooperative Soil Survey have been established to advance national mapping standards and procedures for subaqueous soils. New parent material and landscape–landform terms have been added to the *National Soil Survey Handbook* (NCSS, 2005), and *Soil Taxonomy* has been revised to accommodate the classification of subaqueous soils. In Rhode Island, a partnership was developed (MapCoast, 2009) among Federal agencies such as US. Environmental Protection Agency (USEPA), USDA-Natural Resources Conservation Service (NRCS), National Oceanic and Atmospheric Administration (NOAA), state-level agencies, and university researchers and scientists to apply the information provided in subaqueous soil investigations to coastal resource issues and problems. The MapCoast partnership is a consortium dedicated to multidisciplinary mapping of coastal underwater resources, including bathymetry, habitat, geology, soils/sediment, and archeological resources in shallow waters (August and Costa-Pierce, 2007). NOAA is redesigning their classification system for shallow subtidal habitats (similar to the Cowardin et al. (1979) system for wetlands) and including a subaqueous soils component (Madden et al., 2008).

36.4 Methods for Characterizing Subaqueous Soils and Mapping Their Distribution

One of the big hurdles in investigations of subaqueous soils and mapping their distribution was the lack of methodologies for identifying, sampling, characterizing, and mapping subaqueous soil properties and their distribution. Discussions on topics such as collecting bathymetric data, using a vibracore, collecting soil samples under water, and handling subaqueous soil samples for subsequent laboratory analysis were essentially absent from the soil literature.

36.4.1 Creating a Subaqueous Terrain Map

Landscape units provide a first approximation of the distribution of soils on the landscape and offer an objective delineation of soil types. Numerous studies have emphasized the importance of landscape components for predicting and explaining soil distributions (Jenny, 1941; Ruhe, 1960; Huddleston and Riecken, 1973; Stolt et al., 1993; Scull et al., 2005). Subaqueous landscapes are fundamentally the same as subaerial systems and have a discernable topography from which subaqueous landforms and landscape units may be identified. However, because of the overlying water, submerged landscapes and landforms cannot be identified easily using standard methods such as stereo photography or visual assessment of the landscape. Therefore, identification and delineation of subaqueous landscape units is somewhat more complicated than that of terrestrial landscapes, and development of subaqueous topographic maps is a critical first step toward delineating subaqueous landscape units.

Bathymetric data (water depths) are used to produce a contour or bathymetric map from which subaqueous landforms can be identified and delineated to begin the soil survey. Thus, x, y, and z coordinates are necessary to create a contour map. XY locations are obtained with a differential GPS receiver (DGPS) with submeter accuracy.

Water depths can be determined in a number of ways. The quickest approach is to use a fathometer or bottom profiler. The transducer portion of the profiler is attached to the boat just below the water line. As the boat moves along the water, depths are obtained and stored in the fathometer computer or a device such as laptop. Demas and Rabenhorst (1998) reported that soundings were collected at approximately 4 $km^2\ h^{-1}$. Soundings should be collected essentially along fairly evenly spaced transects that are perpendicular to the shoreline. The depths are corrected by adding the depth between the water surface and bottom of the transducer and correcting for changes in the tide while the data are being collected.

Tide corrections are made from data collected from tide gauges operating at the same time that the bathymetric data are being collected. One to three tide gauges are generally required depending upon the size of the area of interest and the complexity of the tidal cycle within the estuary. Tide gauges should be surveyed in from United States Geological Survey (USGS) benchmarks. Tide corrections can be made in a number of ways. Most simple tidal fluctuations can be corrected using equations developed from the tidal cycles and applied via a spreadsheet. Complicated corrections may require the use of software designed for the purpose.

There has been some discussion and some attempts to use LIDAR (light detection and ranging) to obtain bathymetric data more rapidly. The SHOALS (U.S. Army Corps of Engineers) bathymetry system uses a scanning, pulsed, infrared (1064 nm), and blue–green (532 nm) laser transmitter where the depth of water is determined from the difference in return times of the two beams and knowing the speed of light in water. Optimally, this system can be used to measure water depth from 0 to 40 m with a vertical accuracy of 20 cm. While this may have promise, there are a number of limitations that can be especially problematic including the following: (1) water clarity limits the ability of light to penetrate to the bottom; (2) high surface waves and heavy fog decrease the depth penetration of the lasers; and (3) heavy vegetation and fluid mud influence the depth penetration of the lasers. Also, in many systems where the maximum water depth is only a few meters, a resolution of 20 cm may not be adequate. Hopefully, advances will continue, so that more rapid acquisition of bathymetry becomes possible.

Bathymetric data should be reviewed to remove aberrant depths and aberrant XY locations. A number of software programs are available to construct topographic maps. As an example, an ArcView TIN model was created using the bathymetric soundings and a hard break line (depth = 0) consisting of the wetline from recent orthophoto. The TIN was converted to a GRID (10 m pixel size). The land–water interface observed from the orthophoto wetline was used as a mask to set all land-based pixels to NODATA. The bathymetric GRID was smoothed by applying a 3 pixel radius averaging filter, and contours were created from the smoothed bathymetric GRID. Although the TIN model was used in our example here, other modeling approaches such as kriging have been applied to bathymetric data to create topographic maps.

Using the fathometer from a boat is limited to water depths greater than 50 cm. In areas where there is considerable tidal fluctuation (a meter or more tidal fluctuation), shallow water may be profiled at high tides. If tidal fluctuations are less, surveying of the shallow water may be necessary. This can be done with a survey grade GPS that records elevations real-time kinetic (RTK), a total station, or an all-purpose elevation rod and level. This approach can also be used to validate contour maps created from bathymetric data.

36.4.2 Landscape and Soil Delineation

Landscape unit boundaries provide the first approximation of the distribution of soils over the landscape. Landscape attributes such as slope, microrelief, surface shape, and geographic proximity to other features, and location help define landforms and landscape unit boundaries. Landscape unit boundaries are hidden beneath the water cover in the subaqueous environment and need to be deciphered from contour maps. Landforms and units such as coves, submerged beaches, shoals, and washover fan flats and slopes are some common examples found in many estuaries (Figure 36.3; NCSS, 2005). In some cases, these features can be observed in aerial photographs that penetrate the water, but in general, a contour map illustrating slope breaks is necessary to define the boundaries on each unit.

Identifying the soil types within a landscape unit is done through reconnaissance efforts and transects. The same criteria used to separate mapping units in subaerial soils can be used in subaqueous soils. Tools such as a Macaulay peat sampler, bucket auger, and tile probe are effective in providing soil samples and data for identifying soil types. Peat samplers work well in high n-value (soft, low-density) soil materials, low-energy environments. In areas where low n-value mineral soils dominate, a bucket auger can be used to sample the upper 75 cm of the soil. Some soil scientists use a sleeve with an inside diameter slightly larger than the diameter of the auger to maintain an auger boring location and to minimize slumping of sandy materials into the auger hole. Samples collected in the bucket auger are often transferred to a vinyl tray or gutter on the boat for description and possible sampling. A floating tube with the tray strapped to the tube can work in the case of wading and mapping. Tile probes can be used to find depth to bedrock or similar consolidated or semiconsolidated materials. Percentages of boulders and large stones can also be estimated with the tile probe. Side-scan sonar images, or video footage across proposed mapping units, may aid in proper placement of boundaries and provide spatial data regarding the distribution of stones and boulders extruding from the soil and into the water column.

Just as in subaerial soil mapping, map unit purity and variability need to be addressed. The most common approaches are using random points or transects to assess variability. Studies of soil variability within the landscape units (Demas, 1998; Bradley and Stolt, 2003; Osher and Flannagan, 2007) demonstrate that the concept (common to subaerial landscapes) that soil type follows landscape form (Hudson, 1992) also holds for subaqueous soils (Table 36.2). For example, Bradley and Stolt (2003) reported that 11 of the 12 map units that were used to map the subaqueous soils in a coastal lagoon had a taxonomic purity (based on the subgroup taxonomic level) above criteria used for delineation of the traditional consociation map units used in most USDA-NRCS soil surveys (Soil Survey Division Staff, 1993).

One of the criticisms of inventorying subaqueous soils is that these environments are considered "ever-changing, unstable, shifting sands." Although some areas such as flood–tidal delta landscapes are quite dynamic, Bradley and Stolt (2002) showed that a detailed 1950s NOAA bathymetric map (NGDC, 1996) of the coastal lagoon (Ninigret Pond) was essentially no different than a bathymetric map created 50 years later. With the lifespan of a soil survey on the order of 25–30 years, subaqueous soil surveys should provide descriptions and interpretations for two to three decades of most areas having subaqueous soils (Table 36.3).

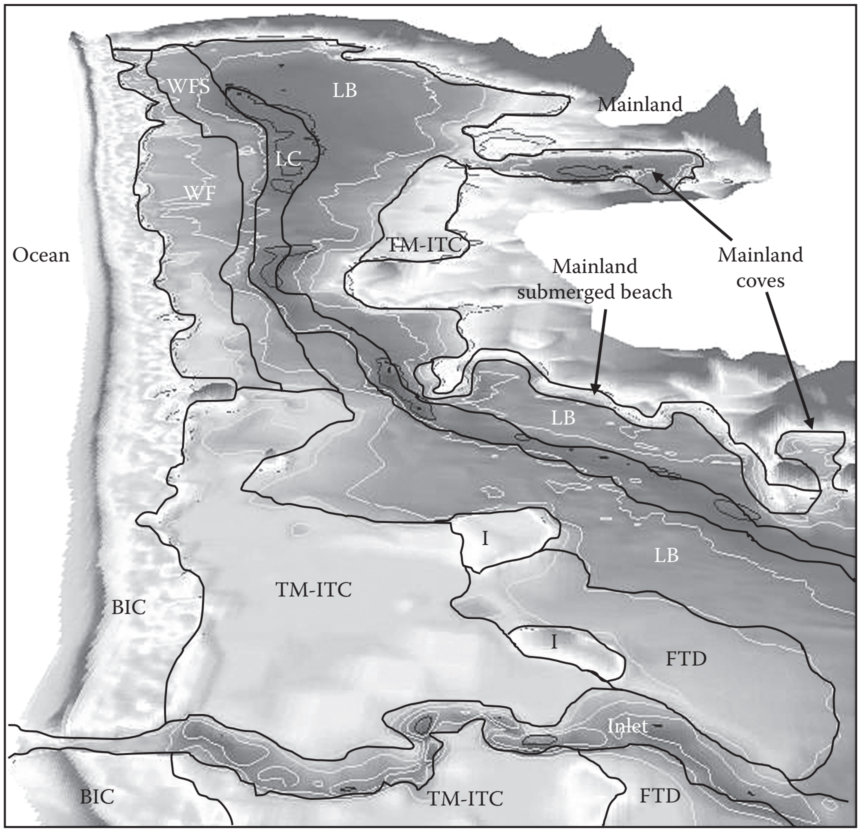


FIGURE 36.3 Examples of landscape units within a coastal lagoon. The barrier island complex (BIC), islands (I), and mainland are subaerial system. The tidal marsh-intertidal complex is sometimes subaqueous and sometimes subaerial. The flood-tidal delta (FTD), lagoon bottom (LB), washover fan (WF), washover-fan slope (WFS), lagoon channel (LC), mainland cove, and mainland submerged beach landscape units are subaqueous systems. The inlet brings in water from the open ocean on high tides and flushes the lagoon during out-going tide. Contour lines are for the subaqueous areas only have a 50 cm interval.

TABLE 36.2 Examples of Landscape Unit, Parent Material, and Soil Type Relationships from a Rhode Island Coastal Lagoon

Landscape Unit	Parent Materials	Typical Soil Subgroup Classification[a]
Lagoon bottom	Silt, fine sand, and organic material	Typic Sulfiwassents
Washover fan flat	Holocene sand	Typic Fluviwassents
Flood–tidal delta flat	Holocene sand	Typic Psammowassents
Washover fan slope	Holocene sand	Sulfic Fluviwassents
Flood–tidal delta slope	Holocene sand	Fluventic Psammowassents
Mainland submerged beach	Glacial fluvial sand and gravel	Typic Haplowassents
Barrier cove	Silt, fine sand, and organic material over glacial fluvial sand and gravel or Holocene sand	Typic Sulfiwassent
Mainland shallow cove	Holocene sand over glacial fluvial sand and gravel	Haplic Sulfiwassents
Midlagoon channel	Glacial fluvial sand and gravel	Typic Haplowassents
Barrier submerged beach	Glacial fluvial sand and gravel	Typic Haplowasssents
Shoal	Glacial fluvial sand and gravel	Typic Haplowassents
Mainland cove	Silts, fine sand, and organic material over buried organic material	Thapto-histic Sulfiwassents

Source: After Bradley, M.P., and M.H. Stolt. 2003. Subaqueous soil-landscape relationships in a Rhode Island estuary. Soil Sci. Soc. Am. J. 67:1487–1495.

[a] Classification based on Keys to Soil Taxonomy (Soil Survey Staff, 2010).

TABLE 36.3 Summary of Selected Subaqueous Soils Interpretations Identified by Federal and Regional Resource Managers for Managing Shallow Subtidal Coastal Areas in the New England (MapCoast, 2009) and Mid-Atlantic States (King, 2004)

Specific resource-based soil interpretation
Submerged aquatic vegetation restoration
Crab habitat
Shellfish stocking
Sustainable shellfish production
Mooring and dock locations
Identifying anthropogenic sites
Nutrient reduction
Pfesteria cyst residence sites
Benthic preservation site identification
Wildlife management
Waterfowl, nurseries, and spawning areas
Habitat protection for horseshoe crab
Tidal marsh protection and creation
Bathymetric maps and navigation
Dredging island creation
Effects of dredging on Benthic ecology
Dune and beach maintenance/replenishment

Sources: Mapping Partnership for Coastal Soils and Sediments. 2009. Soil survey data for Ninigret Pond. http://www.ci.uri.edu/projects/mapcoast/data/default.htm; accessed December 17, 2009; King, P. 2004. Subaqueous soil interpretations. Northeast Region National Cooperative Soil Survey Subaqueous Soils Working Group. *In* M.C. Rabenhorst, M.H. Stolt, P. King, and L. Osher (eds.) National workshop on subaqueous soils. Georgetown, DE.

36.4.3 Sample Collection for Characterization of Typifying Pedons

In high n-value materials, a relatively undisturbed half-core (5 cm diameter) can be collected using a Macaulay peat sampler, which is an excellent tool for providing samples for description and characterization. One limitation is that samples collected with a peat sampler are a bit small, which when coupled with the fact that the materials that can be sampled with the peat sampler have a low density, their dry mass is quite small. Thus, for characterization and descriptive purposes it may be necessary to take multiple adjacent samples if a Macaulay peat sampler is employed. High n-value materials can also be collected in a core barrel. A handle for pushing the core barrel in and pulling it out is attached to the barrel and weight (usually one or two persons lean on the handle) is added to push the core barrel into the soft materials. Several people are usually needed to pull the sample out.

Vibracore sampling is the most effective approach to obtain minimally disturbed samples for detailed description and sampling of typical pedons. A vibracore rig consists of an engine, a cable, and a vibracore head (Lanesky et al., 1979). The engine creates a high frequency, low amplitude vibration. The vibration is transferred through a cable to the vibracore head that is bolted to the top of core barrel or tube. This vibration essentially liquefies the soil materials in a thin zone immediately adjacent to the tube, enabling the core barrel to penetrate into the soil materials. Weight is often added to the top of the core barrel to assist in pushing the core barrel into the soils. Vibracores come in a variety of forms from small, lightweight, and portable to large heavy machines.

Core barrels are generally made out of 7.5–10 cm inside diameter aluminum pipe (irrigation pipe). Some barrels are made out of polycarbonate (these are clear and light but also six to seven times more expensive than aluminum). Core barrel lengths should be as long as the soil to be sampled, plus water depth, and an extra 50 or 60 cm.

In many sampling systems, a core catcher is attached to the cutting end of the barrel. The catcher keeps the soil from sliding out of the barrel when the core is removed from the soil. Other systems rely on a rubber plug inside the barrel that rises up as the barrel is pushed into the soil. The plug maintains a short, nearly airtight space just above the sample to minimize disturbance. This approach also minimizes the suction of the sample out the bottom of the core when the core is removed.

A 2 m core of subaqueous soil that is collected in several meters of water is generally quite heavy and difficult to pull out of the soil. Thus, in most vibracore systems a tripod is set up with a winch or chainfall on the top to extract the core from the soil. The tripod is set up over an opening in the bottom of the sampling vessel. In most cases, a pontoon boat is used for sampling with a vibracore. Sampling is done through an opening (60 cm × 60 cm) in the deck ("moon pool"). The tripod setup can also be placed over a 2 m × 2 m barge that is towed behind the boat. An opening is cut in the middle of the barge. These barges are built from floating dock Styrofoam that is encased in marine plywood.

36.4.4 Subaqueous Soil Sampling and Description

Standard descriptive terminology as outlined in the *Soil Survey Manual* (1993), and by the National Soil Survey Lab (Schoeneberger et al., 2002), and horizon designations outlined in the *Keys to Soil Taxonomy* (2010) should be used to describe subaqueous soils. Samples collected with a Macaulay or bucket auger can be described and sampled immediately.

Cores collected with a vibracore can be sampled and described on the boat or returned to the lab and kept in cold storage prior to sampling. Keeping the samples in cold storage may not be adequate enough to inhibit sulfide oxidation entirely. Cores are extracted from the barrels by cutting the barrels lengthwise on both sides and removing the half barrel. Electric shears designed to cut metal are the best option. A circular saw can also be used to cut through most of the thickness of the barrels and a razor knife used to complete the cut. Prior to using the razor knife, the shards of aluminum or plastic should be whisked away.

36.4.5 Sample Analysis

Most sample analysis can be made following standard procedures outlined in the *Soil Survey Laboratory Methods Manual* (Burt, 2004). Certain soil properties will be affected by the subaqueous environment, and laboratory procedures should be conducted with this in mind. Many subaqueous soil horizons (especially those collected from brackish or coastal environments) contain sulfides that may oxidize upon exposure to air. If samples are meant to be collected for classification purposes, treatment of the samples to avoid oxidation of the sulfides is critical. The most common approach is to immediately transfer the sample to a labeled bag, press out an air trapped in the bag, and put the sample on ice in a cooler. If the soil materials appear to be very reactive (unstable), or the amount of time between sampling and return to the lab is extended, pressurized nitrogen gas can be used to sparge the bags to remove oxygen prior to sealing. If deemed necessary, liquid nitrogen can be applied to the bag in the field to immediately freeze the sample after which it should be placed on ice.

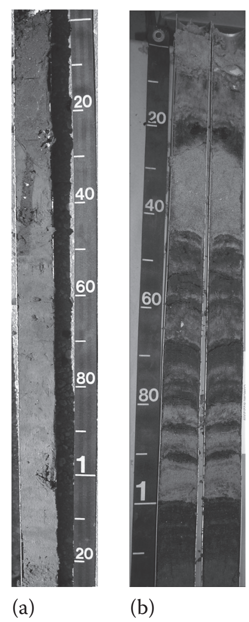
Sulfides, salts, and shell fragments are the most important to consider when analyzing the soil and are worth noting. The presence of sulfides in the soils has been noted above and should be accommodated. Measurements of sulfides are not well documented in soil survey publications. Thus, if this characteristic is to be measured, consideration should be given to the numerous methods to measure chromium reducible and acid volatile sulfides (see Bradley and Stolt, 2006; Payne, 2007). Particle-size distribution analysis may need to be altered to accommodate for the weight and flocculation capability of salts. Samples can be washed to remove salts using dialysis tubing. Carbonates in shell fragments can be an issue in measuring organic carbon and should be considered when organic and calcium carbonate carbon is being determined.

36.5 Classification of Subaqueous Soils

Soil classification is much different than traditional sediment classification, where the substrate is termed mud, silty sand, muddy sand (Fegley, 2001), or other somewhat subjective class, and the focus is often on the upper portions of the profile. The soil classification approach offered in *Soil Taxonomy* is more comprehensive and addresses the larger soil profile. For example, a sediment classification of silty sand (e.g., see Fegley, 2001) could be better described using *Soil Taxonomy* as a coarse-silty over sandy skeletal, mixed, Typic Sulfaquent soil. This soil classification conveys that the upper portion of the soil has <18% clay and >70% silt-sized particles; the lower soil materials (to a meter depth or more) are sandy with >35% gravels or larger particles; the silt and sand-sized particles are not dominated by a particular mineral; and that there are enough sulfides within 50 cm of the soil surface that when the soil materials are allowed to oxidize the pH drops to below 4. Such additional knowledge could be important for decisions regarding the use and management of a portion of the estuary.

The latest version of *Keys to Soil Taxonomy* (Soil Survey Staff, 2010) includes taxa within Entisols and Histosols ("Wassents" and "Wassists") to accommodate subaqueous soils. The formative element "wass" is from the German word for water, "wasser" (Ditzler et al., 2008). Criterion for identifying both suborders is a positive water pressure at the soil surface for at least 21 h each day in all years. The intent of the definition is to identify soils that are inundated everyday, every year, with no exceptions for periodic short- or long-term drought cycles. In certain areas with large tidal fluctuations, such as northern Maine in the United States, soils are inundated with 1–2 m of water everyday with the exception of a few hours at low tide. The 21 h minimum is proposed to allow for short daily exposure of these subaqueous soils.

Six great groups within Wassents are included keying out in the order: Frasiwassents, Psammowassents, Sulfiwassents, Hydrowassents, Fluviwassents, and Haplowassents (see Figure 36.4, e.g., profiles). Freshwater subaqueous soils key out as Frasiwassents based on an electrical conductivity of a 5:1 ratio of water to soil of <0.2 dS m^{-1}. Subaqueous soils that have sandy textures throughout the upper meter are Psammowassents. Sulfiwassents have at least 15 cm of sulfidic materials within the upper 50 cm of the soil. Soils with high n values (low bearing capacity) classify as Hydrowassents. Those soils with an irregular decrease in organic carbon with depth key out as Fluviwassents. Subgroup taxa include Sulfic, Lithic, Thapto-histic, Aeric, Psammentic, Fluventic, Grossic, Haplic, and Typic. With the exception of Grossic, all of these apply in a similar manner to previous applications in other taxa. Grossic is used to identify subaqueous soils that have very thick layers with high n values.



(a) (b)

FIGURE 36.4 (a) Vibracore profile of a Haplic Sulfiwassent from a shallow mainland cove. The tape shows 10 cm increments. Note the clam krotovina from 30 to 50 cm (C/A horizon). The white pieces in the krotovina are shell fragments. At 120 cm, there is a change in parent material from the Holocene aged estuarine sediments to the Pleistocene outwash sand and gravels. (b) Vibracore profile of a Sulfic Fluviwassent. The soil was collected from a washover-fan slope. Note the many buried A horizons that represent storm surges. The buried A horizon starting at a meter represents the upper part of a previous Typic Sulfiwassent prior to encroachment of the washover fan over the lagoon bottom landscape unit.

Subaqueous Histosols are classified as Wassists. There are three great groups: Frasi-, Sulfi-, and Haplo-wassists. The Frasiwassists have low electrical conductivity (<0.2 dS m^{-1}); Sulfiwassists have >15 cm of sulfidic materials within 50 cm; and Haplowassists are all other Wassists. Three subgroups are proposed: Fibric, Sapric, and Typic, depending on the dominant type of organic materials present. Examples of Wassent–landscape relationships are shown in Table 36.2. The *World Reference Base* (WRB) FAO (2006) has taxa similar to those proposed for *Soil Taxonomy*. WRB's Subaquatic Fluvisols correlate to Wassents, and the Subaquatic Histosols correlate to Wassists.

36.6 Applications of Subaqueous Soil Information

Shallow-water coastal habitats, including coastal lagoons, shallow bays, and estuarine areas, are highly valued and heavily used resources. Almost two-thirds of the worldwide population currently lives in coastal areas (Trenhaile, 1997), and recent demographic studies suggest that in the next 25 years 75% of the U.S. population will live in close proximity to the coast (Bush, 2004). As subaqueous soil science progresses, a wide range of use and management interpretations are expected to be developed for use with estuarine subaqueous soil maps (August and Costa-Pierce, 2007; Surabian, 2007; Payne and Turenne, 2009). These interpretations will aid in coastal, estuarine, and marine restoration, ecosystem management, and conservation efforts. For example, subaqueous soils information can be used to assist in the restoration of SAV and shellfish; identifying shellfish aquaculture sites; design and placement of shoreline protection, docks, and moorings; and identifying subaqueous soils that are of beneficial use from dredging (Demas and Rabenhorst, 1999; Bradley and Stolt, 2003; Bradley and Stolt, 2006; Surabian, 2007). Since subaqueous soil investigations are a relatively new focus in pedology, and most of the subaqueous soil efforts have concentrated on developing field and laboratory methodology, few studies have concentrated on interpretations. Thus, the breadth of information relating soil type with the use and management of these resources is quite limited.

36.6.1 Dredging and Dredge Placement

Dredging of subaqueous soils is a common practice to deepen navigable waterways and to replenish beaches. Subaqueous soils often have layers or horizons where sulfides have accumulated in subaqueous soils as a result of sulfidization. When the sulfide-bearing soils are dredged and placed in the subaerial environment, sulfides oxidize releasing sulfuric acid, lowering the pH, and creating acid sulfate soils (Fanning and Fanning, 1989). Acid sulfate soils may persist for a number of years and are uninhabitable for plants and animals. If deposited near water, these acid sulfate soils can also create runoff that is toxic to aquatic systems (Demas et al., 2004).

To test for sulfides and for taxonomic purposes, subaqueous soils are allowed to oxidize in a moist environment. In general, those soil materials that after at least 16 weeks of moist incubation reach pH values <4 are considered to have sulfidic materials (Soil Survey Staff, 2010) and are potential acid sulfate soils. Whether these soils reach the incubation pH of potential acid sulfate soils is dependent upon a number of factors such as the buffering capacity, rate and extent of acid production, weathering, and leaching due to environmental factors are not considered in the laboratory approach. For example, during a moist incubation, a small amount of sulfide would lower the pH in a sandy soil with low organic carbon because of the limited amount of buffering capacity. In such soils, the amount of acid produced would be much less than a soil with a similar incubation pH but a higher buffering capacity because of a finer texture or greater organic matter levels. In a natural setting, a small amount of acid could potentially leave a sandy, minimally buffered soil very quickly as the acidity generated by oxidation would be washed out of the system as a result of precipitation and leaching. In contrast, acid sulfate conditions may remain for decades in a fine-textured, buffered soil. Thus, understanding a number of soil parameters is critical to identifying the subaqueous soils that can be deposited in a subaerial environment as dredged materials and maintain conditions conducive for plant growth and a safe environment.

36.6.2 Water Quality

Estuarine ecosystem integrity and sustainability has received tremendous interest in recent years. These interests are being driven by concerns over the negative effects of rapid urbanization and related anthropogenic activities on the coastal environment. As the use of these natural resources has increased, the most urbanized estuaries have been ecologically compromised and common ecosystem functions and values are being lost. Obvious signs of these degraded environmental conditions are accumulations of metals and other contaminants, an increase in emerging diseases and algal blooms (Harvell et al., 1999), and anoxia related fin and shellfish kills (RIDEM, 1998, 2003). Most of these issues are related to poor water quality.

Water quality has traditionally been used in coastal areas as an indicator of the overall health of an estuary (Stevenson et al., 1993; Glasgow and Burkholder, 2000; Granger et al., 2000). Because water quality can fluctuate with tidal cycles and seasonal and yearly weather changes, water quality trends are difficult to predict or to use as a reliable indicator of extended changes in the health of an estuarine system (D'Avanzo and Kremer, 1994; Cicchetti et al., 2006). Soil properties and characteristics develop in response to the environment, making subaqueous soils a potential long-term indicator of the degree that these ecosystems have become degraded (Germano and Rhoads, 1988; Valente et al., 1992). Such an indicator would allow estuary management teams to target particular estuaries for conservation, protection, and restoration of resources based on soil survey information. Understanding the degree and spatial distribution of the degradation is critical to managing coastal estuaries for any number of functions and values, especially aquaculture and restoration of commercially important shellfish populations.

Redox conditions in a soil have important impacts on chemical processes that occur in the soil such as denitrification, changes in forms of iron, manganese, or sulfur, and the solubility of heavy metals (Tomaszek, 1995; Teasdale et al., 1998). The decomposition of organic matter by microbes fuels the redox reactions in soil. Oxygen is the strongest oxidizing agent in aqueous systems and acts as an electron acceptor during microbial decomposition. In subaqueous systems, however, oxygen can quickly be depleted and other electron acceptors are used by the microbes. These other electron acceptors include nitrate, manganese, iron, sulfate, and carbon dioxide. Each species, respectively, is reduced at a lower range of redox potentials depending on the pH (Bohn et al., 1979). These processes produce a vertical profile of decreasing redox potential with depth as each oxidizing agent is reduced until all organic matter has been decomposed (Teasdale et al., 1998).

The first of these boundaries, where all oxygen has been depleted or reduced, is generally known as the redox boundary, the redoxocline, or the redox potential discontinuity (RPD) (Knox, 1986; Teasdale et al., 1998; Hinchey and Schaffner, 2005). The depth of the RPD can be influenced by the grain size, organic matter content, temperature of the soil, as well as the movement and dissolved oxygen content of water above the soil surface (Knox, 1986). A redox potential gradient found in subaqueous soils often includes the oxidized surface layer where oxygen is still present in the interstitial water, a zone of transition where other species are being reduced, and a sulfide zone that is totally anaerobic, H_2S is prevalent, and redox potentials are very low (Knox, 1986). This zonation plays an important role in determining layers in which chemical processes involving organic carbon, nitrogen, and sulfur occur and can serve as an indication of estuary health.

36.6.3 Submerged Aquatic Vegetation

One part of the definition of soil is the ability of soil to support rooted plants in a natural environment (Soil Survey Staff, 1999). Dense beds of SAV (or seagrass) are often found in subtidal estuaries. Unlike macroalgal species, which anchor themselves to a substrate, seagrasses are rooted vascular aquatic plants in which roots serve both structural and nutrient uptake purposes (Barko et al., 1991). One of the most important interpretations from an inventory of subaqueous soils may be seagrass restoration. SAV such as eelgrass provides nursery habitat for economically important fin and shellfish and is important for sediment and nutrient filtering, nutrient cycling, and buffering wave effects. In many estuaries, aerial coverages of seagrass beds have severely declined over recent years. Therefore, seagrass restoration has become a focus of many coastal managers. Seagrass revegetation and restoration projects cost on the order of $100,000 per acre, but few of these projects have been successful. It is highly likely that the projects fail because of site selection. Seagrass revegetation sites are commonly located where past seagrass meadows had been, not at sites where present soil conditions are optimum for success. However, loss of seagrass tends to result in erosion of the subaqueous soils. Thus, soils within areas that previously supported seagrass may be significantly different following the loss of vegetation and subsequent erosion. A detailed knowledge of the relationship between subaqueous soil properties and SAV is essential for improving the success of seagrass revegetation efforts. An understanding of the seagrass–subaqueous soil system will help resource managers identify the sites where revegetation efforts can be most successful (Bradley, 2001). Few studies have looked at these relationships. Bradley and Stolt (2006) examined subaqueous soil–eelgrass relationships in a northeastern U.S. coastal lagoon. Similar studies need to be made across regions with a focus on the predominant seagrass species and the breadth of tidal ranges and gradient of temperatures.

36.6.4 Carbon Storage and Sequestration

With the concern with global warming mounting as a result of increasing greenhouse gas emissions, there is a significant interest in carbon storage and sequestration in soil systems. These interests have led to numerous studies focused on soil carbon for various land types and covers. Although forested and emergent wetlands have been well studied in regard to carbon sequestration, carbon sequestration and storage studies have largely overlooked estuary soils as important carbon sinks (Chmura et al., 2003; Thom et al., 2003). Considering that the shallow subtidal component may occupy as much as 90% of the estuary, these areas likely represent a significant and unaccounted for sink for carbon. Little is known, however, regarding the contribution of the shallow subtidal portions of the estuaries to the regional carbon stocks.

Geologic studies focused on estuarine and oceanic substrates have included organic carbon as a parameter inventory; however, most of these studies focus on surficial soil samples with the goal of understanding the origin and formation of petroleum (Hedges and Oades, 1997). Utilizing a pedologic approach, it is possible to quantify the organic carbon content of the subaqueous soil with depth, where it is actually stored, not just within the soil surface. Once soil organic carbon is determined for specific sites within an estuary, it will be possible to scale up to a regional or global scale in order to better determine the estuarine soils importance as a global carbon storage unit. Jespersen and Osher (2007) and Payne (2007) investigated the carbon storage capabilities of subaqueous soils in the Taunton Bay estuary in Maine and three embayments in Rhode Island, respectively. In both studies, a soil survey of the estuary was completed as a component of the study to relate organic carbon storage to soil–landscape unit. In addition, carbon pools to a depth of a meter in subaqueous soils were compared to their adjacent subaerial upland and wetland soils. The estuarine soil organic carbon pools were found to be equal to, and in some cases greater than, subaerial soil organic carbon pools. Payne (2007) reported higher energy, sandier soil–landscape units, such as shoals and shorefaces, had lower carbon pools than the lower energy soil–landscape units such as bay bottoms. Similar relationships were observed by Jespersen and Osher (2007).

The studies made by Jespersen and Osher (2007) and Payne (2007) were focused on northeastern tidal embayments. Little is known regarding the expansive coastal lagoons of the Atlantic coast or Gulf of Mexico estuarine subaqueous soils. Carbon pools and sequestration rates in freshwater subaqueous soils are also unknown. Future studies should be designed and implemented to investigate these subaqueous soil systems.

36.6.5 Moorings and Docks

With any body of water there are typically structures built or deployed to secure boats. The foundation for these docks or mooring (permanent anchor that boats are secured to in a harbor) are the subaqueous soils. Thus, how well the mooring or dock functions is dependent upon subaqueous soil type. The bearing capacity or n value of the surface and near surface soils is one of the most important characteristics. Surabian (2007) examined relationships between subaqueous soils and moorings and found that mushroom anchors work best in high n-value soils. These moorings sink into the low bearing capacity soils and are kept in place by surface area and suction forces. Deadweight anchors are best suited for low n-value soils or soils dominated by coarse fragments (Surabian, 2007).

36.6.6 Shellfish

Subaqueous soils are critical to the structure and function of many of the plants and animals in the estuarine ecosystem and are the foundation for commercial shellfish production and aquaculture. Worldwide the aquaculture industry continues to develop and expand. Although the economics are difficult to quantify worldwide, the value of aquaculture products per acre typically far exceed those of traditional agriculture. For example, in 2007 the average value of Rhode Island aquaculture products (oysters and clams) was $32,000 per hectare (Alves, 2007). Considering the cash value of these aquaculture products, developing an understanding of the relationships between the submerged landscape, the subaqueous soils, and the growth and productivity of aquaculture species such as clams, oysters, scallops, and mussels is essential. To date, very little information is available regarding the relationships between shellfish productivity and subaqueous soil type.

The few works that have studied aquaculture–soil type relationships have focused on clams (Pratt, 1953; Pratt and Campbell, 1956; Wells, 1957; Grizzle and Lutz, 1989; Grizzle and Morin, 1989). These studies investigated clam abundance and shell-size growth rates with environmental factors such as soil type. Soils in these studies were fairly crudely characterized (i.e., sand or mud); however, most of the studies found a relationship between growth and particle size existed. In general, increases in fines (muds, silt, and clay) were associated with retardation in growth of clams. Grizzle and Lutz (1989) concluded that substrate type is important in some instances but seston flux (the amount of suspended particulate matter including plankton and organic detritus that passes by over a given period in the water column) is more important. Since sandy substrates typically have higher energies, the seston flux is often higher relative to finer textured soils. Thus, current views on the shellfish–soil relationship are that the increased growth associated with sandier substrates in the earlier studies has been reinterpreted to be a secondary result of sandier soil being associated with higher current velocities (Rice and Pechenik, 1992). This suggests that subaqueous soil type may not directly relate to shellfish growth, but may serve as a surrogate for identifying areas of favorable seston fluxes, and could thereby be used to predict areas of the subtidal estuary with the highest potentials for shellfish growth. A better classification of the soil that would come with a subaqueous soil survey (i.e., better than sands and muds) may prove a better predictor of shellfish growth and provide delineations for the best locations for aquaculture of clams and oysters.

36.7 Future Considerations for Subaqueous Soils

Although subaqueous soils have received occasional mention in the literature for more than 50 years, only in the past decade or so have these soils been investigated with any intensity or focus. The limited number of investigations to date suggests that additional mapping, characterization, and research is needed to better understand these soils. In the United States, nearly all of the subaqueous soils projects have been conducted on the eastern seaboard in coastal waters. The same resource, habitat, and ecosystem service issues that have begun to be addressed from a pedological perspective in eastern U.S. estuaries also need to be examined in other shallow subtidal habitats as well as freshwater systems. As we have shown, the application of subaqueous soils investigations to addressing environmental and ecosystem questions related to restoration, aquaculture, carbon accounting, water quality, etc. is dependent upon an inventory of the subaqueous soil resources. The soil survey landscape-level models developed for mapping soils of embayments and lagoons need to be tested further in other Atlantic shallow subtidal habitats and then in other areas of the country and of the world. Concerted efforts should be made to conduct widespread subaqueous soil survey projects that are founded on established standards and protocols such as those used in the National Cooperative Soil Survey. These subaqueous soil resource inventories should be conducted and published at a scale that will be useful to resource managers attempting to balance both use and conservation of aquatic ecosystems that are heavily taxed and impacted as increasing populations choose to inhabit areas near the water.

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37

Digital Soil Mapping

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37.1 Introduction

The current explosion in computation and information technology comes with vast amounts of data and tools in all fields of endeavor. This has motivated numerous initiatives around the world to build spatial data infrastructures aiming to facilitate the collection, maintenance, dissemination, and use of spatial information. Soil science potentially contributes to the development of such generic spatial data infrastructure through the ongoing creation of regional, continental, and worldwide soil databases. The principal manifestation is soil resource assessment using geographic information systems (GISs), that is, the production of digital soil property and class maps with the constraint of limited relatively expensive fieldwork and subsequent laboratory analysis.

The production of digital soil maps ab initio, as opposed to digitized (existing) soil maps, is moving inexorably from the research phase (Skidmore et al., 1991; Favrot and Lagacherie, 1993; Moore et al., 1993) to production of maps for regions and catchments and whole countries. The map of the Murray–Darling basin of Australia (Bui and Moran, 2001) comprising some 19 million 250 m × 250 m pixels or cells and the digital Soil Map of Hungary (Dobos et al., 2000) are the most notable examples. The progress and development of digital soil mapping is marked by adoption of new mapping tools and techniques, data management systems, innovative delivery of soil data, and methods to analyze, integrate, and visualize soil and environmental data sets (Grunwald, 2009).

McBratney et al. (2000) reviewed pedometric methods for soil survey and suggested three resolutions of interest, namely, >2 km, 20 m–2 km, and <20 m, corresponding to national to global, catchment to landscape, and local extents. Table 37.1 provides an overview with five resolutions of interest. The third one (D3) that deals with subcatchments, catchments, and regions is the one that attracts the most attention. In the language of digital soil maps (Bishop et al., 2001), which differs from that of conventional cartography, scale is a difficult concept and is better replaced by resolution and spacing. D3 surveys, which in conventional terms have a scale of 1:20,000 down to 1:200,000, have a block or cell size from 20 to 200 m, a spacing also of 20–200 m, and a nominal spatial resolution of 40–400 m (see Table 37.1).

Unfortunately, the existing soil databases are neither exhaustive nor precise enough for promoting extensive and credible use of the soil information within the spatial data infrastructure that is being developed worldwide. The main reason is that their present capacities only allow the storage of data from conventional soil surveys, which are scarce and sporadically available.

The Netherlands has complete coverage at a nominal spatial resolution of 100 m. In France, on the other hand, a highly developed western economy, but with a large land area, only 26% of the country is covered at a nominal spatial resolution of 500 m and 13% at a nominal spatial resolution of 200 m (King et al., 1999). One third of Germany is covered with soil maps at a nominal spatial resolution of 10 m (1:10,000) but most of these are not yet digital (Behrens and Scholten, 2006). On the other hand, complete coverage of Germany at coarser resolutions (nominally 100 and 400 m) is available. It was projected that in 2010, 65% of Germany's soil maps will be digitized.

TABLE 37.1 Suggested Resolutions and Extents of Digital Soil Maps

Name	Approx. USDA Survey Order[a]	Nominal Resolution and Spacing (Pixel Size)	Resolution "loi du quart"[b]	Extent[c]	Cartographic Scale[d]
D1	0[e]	<(5 m × 5 m)	<(25 m × 25 m)	<(50 km × 50 km)	>1:5,000
D2	1, 2	(5 m × 5 m)–(20 m × 20 m)	(25 m × 25 m)–(100 m × 100 m)	(500 m × 500 m)–(200 km × 200 km)	1:5,000–1:20,000
D3	3, 4	(20 m × 20 m)–(200 m × 200 m)	(100 m × 100 m)–(1 km × 1 km)	(2 km × 2 km)–(2,000 km × 2,000 km)	1:20,000–1:20,0000
D4	5	(200 m × 200 m)–(2 km × 2 km)	(1 km × 1 km)–(10 km × 10 km)	(20 km × 20 km)–(20,000 km × 20,000 km)	1:200,000–1:2,000,000
D5	5	>(2 km × 2 km)	>(10 km × 10 km)	>(200 km × 200 km)	<1:2,000,000

[a] Soil Survey Staff (1993), Table 2.1.

[b] According to Boulaine (1980), the smallest area discernible on a map is 0.5 cm × 0.5 cm or one quarter of a square centimeter, hence the term "loi du quart." The USDA *Soil Survey Field Handbook* (Table 2.2) (Soil Survey Staff, 1993) quotes 0.6 cm × 0.6 cm. Both of these really refer to conventional map delineations, and resolution estimates based on these minimum areas should be regarded as very conservative.

[c] Calculated as minimum resolution times 100 (pixels) up to maximum resolution times 10,000 pixels.

[d] Digital soil maps are defined by their resolution and spacing—which here we equate with pixel size—the cartographic scale is calculated as 1 m/(side length of 1,000 pixels), assumes that the smallest area discernible is 1 mm × 1 mm. Conversely, the resolution (ρ) of a 1:100,000 conventional map can be calculated as, $\rho = 1/\chi \times \lambda = 100{,}000 \times 0.001 = 100$ m if we consider the smallest area resolvable on a map (λ), with representative fraction χ, to be 1 mm × 1 mm. It could be argued that the minimum resolution is the size of 2 × 2 pixels.

[e] This resolution was suggested by Dr. Pierre C. Robert, University of Minnesota, for applications in precision agriculture.

The situation in larger countries such as Australia and Brazil is much worse. In Australia, for example, prior to Moran and Bui's (2002) work, the Murray–Darling Basin, Australia's most important agricultural region comprising some 14% of the land area, had 50% coverage at 500 m and 3% at 200 m. In Brazil, the country is uniformly covered by the Soil Map of Brazil and the Agricultural Suitability Map of Brazil at a nominal spatial resolution of 10 km, exploratory soil maps by the RADAM/EMBRAPA Solos project (1:1,000,000 or nominally 2 km) and Agroecological Zoning (diagnosis of environmental and agro-socioeconomic features, nominally 4 km or 1:2,000,000).

The main reason for this lack of soil spatial data is simply that conventional soil survey methods are relatively slow and expensive. Furthermore, there is a worldwide crisis in collecting new field data in general that leads some to be very pessimistic about future developments in conventional soil surveying. To address this situation, current spatial soil information systems have to extend their functionalities from the storage and the use of digitized (existing) soil maps to the production of soil maps ab initio. This is the aim of digital soil mapping, which can be defined as the creation, and population of spatial soil information systems by the use of field and laboratory observational methods coupled with spatial and nonspatial soil inference systems (Lagacherie and McBratney, 2006).

Thus, digital soil mapping refers to the production of spatial soil databases, based on soil observations combined with environmental data through quantitative statistical relationships. In that sense, digital soil mapping is not simply digitizing existing soil maps and is more than just producing paper maps. There are other terminologies such as "predictive soil mapping" (Scull et al., 2003) that refers to the production of digital soil maps, and "environmental correlation" (McKenzie and Ryan, 1999) that is an aspect of the spatial soil prediction function.

Digital soil mapping uses a range of technologies allowing for more accurate and efficient prediction of soil properties through optimal sampling strategies, rapid analysis of soil properties, and rapid acquisition of environmental variables over a large extent. Combined with pedometric techniques, it can provide the best prediction of soil properties at the required resolution with associated uncertainties. Digital soil mapping can be thought as a means for modernizing and systematizing traditional soil survey.

In this chapter, we review various approaches with numerous examples from the literature, which are largely seen as special cases of the approach suggested here. First, we trace the development of the quantitative ideas and methods over the last 60 or so years.

37.2 Brief Review of Approaches to Soil Spatial Prediction

Hudson (1992) contended that soil survey is a scientific strategy based on the concepts of factors of soil formation coupled with soil–landscape relationships. Hewitt (1993) pleaded for the need for explicitly stated, but not necessarily quantitative, models for soil survey. Such models may be knowledge based (Bui, 2004). In this view, soil maps are representations of soil surveyors' knowledge about soil objects. In this chapter, we argue in favor of quantitative predictive models.

37.2.1 Jenny

Recalling Jenny's famous equation (Jenny, 1941), which he intended as a mechanistic model for soil development,

$$S = f(c, o, r, p, t, \ldots) \tag{37.1}$$

implicitly

S stands for soil
c (sometimes cl) represents climate
o represents organisms including humans
r represents relief
p represents parent material
t represents time

some have asserted that it cannot be solved; nonetheless, since Jenny published his formulation, it has been used by innumerable surveyors all over the world as a qualitative list for understanding the factors that may be important for producing the soil pattern within a region. Numerous researchers have taken the quantitative path and have tried to formalize this equation largely through studies of cases where one factor varies and the rest are held constant. So quantitative climofunctions, topofunctions, etc. have been developed. Much of this work was done before sophisticated numerically intensive statistical methods became available. Here are some brief examples.

c (sometimes *cl*): Climofunctions were the ones most developed by Jenny in his 1941 book. Jones (1973) found relationships between carbon, nitrogen, and clay and annual rainfall and altitude in West African savanna using linear and multiple linear regression. Simonett (1960) found a power–function relationship between mineral composition of soil developed on basalt in Queensland and annual rainfall.

o: There seems less development of organofunctions, many believing that the principal organofunction or biofunction, that of vegetation, is dependent on soil rather than the converse. Noy-Meir (1974) found relationships between vegetation and soil type in S.E. Australia. The other principal organofunction, the anthropofunction, has only been recently quantified—much of the work on soil degradation and soil quality is evidence of the effect of humanity on soil. The classic work of Nye and Greenland (1960) is an early quantitative example.

r: The relationship between soil and topographic factors has been evident at least since Milne's paper (Milne, 1935). Quantitative topofunctions are manifold. For example, Furley (1968) and Anderson and Furley (1975) found a piecewise linear relationship between organic carbon, nitrogen, and pH of surface horizons and slope angle for profiles developed on calcareous parent materials around Oxford in England.

p: Quantitative lithofunctions have not been developed often, perhaps due to a difficulty in recognizing and quantifying the dependent and independent variables. Barshad (1958) quantified mean clay content as a function of rock type.

t: Some consider this the only truly independent variable (but if that is the case why is space not also included?). Chronofunctions are often theoretical or hypothetical rather than observed. Hay (1960), however, found an exponential relationship between clay formation and time for soil developed in volcanic ash on the island of St. Vincent, as would be expected from first-order kinetics.

A lot of this early quantitative work was very well summarized by Yaalon (1975). Many of the relationships found are not linear. It should be remembered that the aim of these investigations was to understand soil formation and not necessarily to predict soil from the other factors.

Recognition of interactions between the soil-forming factors is potentially important because it is one possible source of detailed soil pattern. It is difficult to find work that considers such interactions explicitly. Webster (1977) perhaps came closest with his canonical correlation studies of sets of soil properties and environmental factors. From this work, he suggested, for example, that soil will reflect a strong interaction between topography and lithology particularly on upper slope positions, but this will be time dependent. Odeh et al. (1991), using closely related methods, made similar findings.

37.2.2 Geographic and Neighborhood (or Purely Spatial) Approaches

Since the late 1960s, there has been an emphasis on what might be called geographic or purely spatial approaches, that is, soil attributes* can be predicted from spatial position largely by interpolating between locations of soil observation. Another way of thinking about this is as a *neighborhood law* expounded first perhaps by Lagacherie (1992), but is the basis underlying the soil combinations of Fridland (1972) and also of soil geostatistics (Giltrap, 1977), etc. Generally, we can consider the soil at some location (x, y) to depend on the geographic coordinates x, y and on the soil at neighboring locations $(x + u, y + v)$, that is,

$$s(x, y) = f((x, y), s(x + u, y + v)), \tag{37.2}$$

the dependence usually being some decreasing function of the magnitude of u and/or v.

This approach arose originally out of the need for spatial prediction to make soil maps, and because of a failure to obtain prediction from the soil-forming factors largely because the quantitative variables describing these factors were not readily available to do such predictions. These purely spatial approaches are almost entirely based on geostatistics and its precursor trend-surface analysis, although thin-plate smoothing splines have been suggested and used occasionally (Laslett et al., 1987; Hutchinson and Gessler, 1994). Exact-fitting splines do not perform well (Laslett et al., 1987; Voltz and Webster, 1990).

37.2.2.1 Geostatistics and Related Methods Trend Surfaces: $s(x, y) = f(x, y)$

Trend surfaces are low-order polynomials of spatial coordinates. Several applications have been reported in the literature. Davies and Gamm (1969) applied this technique to soil pH values from the county of Kent in England. Edmonds and Campbell (1984) described the average annual soil temperatures at locations within a network of stations from Virginia and neighboring states with a third-degree polynomial that explained 71% of the observed variation. On the other hand, Kiss et al. (1988) found

* Soil attributes is a general term to mean that which can be attributed to the soil by measurement or inference, for example, soil properties like pH, or classes like a soil profile class, or the presence or the absence of a soil horizon class.

that the spatial pattern of ^{137}Cs activity in well-drained, native noneroded soil in the agricultural portion of Saskatchewan was complex and could not be adequately described by a second-order trend surface. There appears to be no literature on trend surfaces for soil classes; nevertheless, Wrigley (1978) has made an attempt to map the probability of the occurrence of soil classes. Spatially, trend surfaces are rather simplified "unnatural" representations, and more complex spatial patterns often need to be modeled.

$$\text{Kriging—}s(x, y) = f(s(x+u, y+v)). \quad (37.3)$$

It was recognized that more complex spatial patterns could be accommodated by treating soil variables as regionalized variables using the methods of geostatistics, particularly various forms of kriging. The papers by Burgess and Webster (1980a, 1980b; Webster and Burgess, 1980) are probably regarded as the most seminal. These kriging methods, reviewed by Burrough (1993), Goovaerts (1999), and Heuvelink and Webster (2001) could deal with continuous soil properties and classes, could give estimates for blocks or pixels of varying size, and moreover could estimate uncertainty.

$$\text{Co-kriging—}s(x, y) = f(s(x+u, y+v), \{c, o, r, p, t\}(x, y)). \quad (37.4)$$

It was recognized early in the development of soil geostatistics that soil could be better predicted if denser sets of secondary variables (spatially cross-) correlated with the primary variable were available; this technique is called co-kriging. In the early co-kriging studies (McBratney and Webster, 1983; Vauclin et al., 1983; Goulard and Voltz, 1992), these ancillary variables were other soil variables, indicating that other soil variables are themselves useful predictors of soil. Later, with the advent of GIS and improved technology, co-kriging was performed with detailed secondary data sets of environmental variables derived from digital elevation models and satellite images (Odeh et al., 1994, 1995).

37.2.2.2 Jenny and Geography—Corpt or Clorpt: $s(x, y) = f(\{c, o, r, p, t\}(x, y))$

An alternative spatial prediction strategy to the purely geographic approaches was developed in the early 1990s, although there were precursors. In these studies, the state-factor equation was put explicitly into a spatial framework, and the factors were also observed in the same spatial domain. This approach probably resulted from the advent of the first GISs, and also possibly as a pedological response to geostatistics. It seems to be based on a much earlier 1D example of using environmental (terrain, representing *r*) attributes for soil prediction, namely, that of Troeh (1964) and Walker et al. (1968). Probably the first of its kind, Aandahl (1948) quantitatively relates landscape attributes to soil properties. He derived the distribution of N based on slope length. Troeh (1964) analyzed the elevation data from two catenas and derived slope and profile curvature. He then plotted the slope gradient and profile curvature and found that the soil drainage classes could be distinguished by paraboloid of revolution equations. Walker et al. (1968) used slope, curvature, aspect, and distance from the local summit, in combination with multiple linear regression to predict soil morphological properties such as A horizon depth, depth to mottling, and carbonates along a transect. An early, perhaps the first, 2D example is Legros and Bonneric (1979), based on earlier work by Legros (1975). They described a soil–environment relationship using various factors (altitude, slope, exposure, and parent material) that were observed on a 500 m grid-cell basis to predict the degree of podsolization in Massif du Pilat of France, and mapped it digitally at a resolution of 500 m. The prediction was achieved by a kind of taxonomic distance relative to reference sites. This was done well before the advent of formal GIS.

The GIS-based studies started at the beginning of the 1990s. Terrain analysis had improved and secondary rasterized layers, providing a kind of complete enumeration of the area could be put in GIS. The soil observation points were intersected with the layers of secondary data, a model fitted by various means, and then the model was used to predict all other locations on the raster. Moore et al. (1993) gave the first 2D example using a set of terrain attributes derived from a digital elevation model on a 15 m grid to predict continuous soil properties such as A horizon thickness and pH for a small catchment in Colorado. Odeh et al. (1994) did a similar study in South Australia. Skidmore et al. (1991) predicted forest soil classes in New South Wales from layers of natural vegetation data (representing *o*) and terrain attributes on a 30 m grid. Bell et al. (1992, 1994) predicted soil drainage class from terrain data, and Lagacherie and Holmes (1997) predicted soil classes in the Languedoc using layers of lithological and terrain data. Favrot and Lagacherie (1993) foreshadowed this as a general approach for making soil class maps.

For quantitative prediction purposes, this has been called the "*clorpt*" or "*corpt*" equation (McBratney et al., 2000). Some people have termed the approach "environmental correlation" (McKenzie and Austin, 1993). McKenzie and Ryan (1999) used environmental correlation associated with stratigraphy, digital terrain models and gamma-radiometric survey, respectively, to predict soil properties in Australia. Ryan et al. (2000) reviewed the concepts and applications of spatial modeling using the "environmental correlation" approach and used it to predict forest soil properties at the landscape level.

For predicting soil classes, S_c, or soil properties, S_a, often only a subset of the five soil-forming factors has been used, for example, when information from a digital elevation model is available,

$$\begin{aligned} S_c &= f(r), \quad \text{e.g., Bell et al. (1992)} \quad \text{or} \\ S_a &= f(r), \quad \text{e.g., Moore et al. (1993),} \end{aligned} \quad (37.5)$$

or relief and a lithology map,

$$S_c = f(r, p), \quad \text{e.g., Lagacherieand Holmes (1997),} \quad (37.6)$$

or relief and vegetation,

$$S_c = f(r, o), \quad \text{e.g., Skidmore et al. (1991).} \quad (37.7)$$

37.2.2.3 Combinations: Clorpt (or Corpt) and Kriging

Alert readers will have noted that there has been a certain similarity and convergence between the co-kriging and the environmental correlation approach. Some workers recognized this in the mid-1990s and combined the two in what is generically known as regression-kriging (Knotters et al., 1995; Odeh et al., 1995). In this approach, "*clorpt*" is used to predict the soil property of interest from environmental variables, and kriging is used on the residuals. Bourennane et al. (1996) used kriging with external drift, which is related to regression-kriging but only allows for linear relationships between the variable of interest and the environmental variables (the external drifts).

37.2.3 Predicting Soil Attributes from Other Soil Attributes: $s_1 = f(s_2)$

As noted above, some of the co-kriging studies (McBratney and Webster, 1983; Vauclin et al., 1983) showed that soil could be predicted from other soil attributes. This observation in itself is not very useful unless there are much denser secondary variables available. Remote (e.g., gamma radiometrics) and proximal sensing (e.g., electromagnetic induction) offer this possibility. This becomes increasingly important because Phillips (2001) gives several examples where "*clorpt*" apparently does not work, particularly at fine resolutions. This suggests that for predictive purposes, *s* (for soil) should be added to the "*corpt*" list.

37.2.4 Some Brief Conclusions

From this brief review, we see that

1. Quantitative relationships have generally been most easily found between soil and topography but there is evidence of quantitative relationships with the other four soil-forming or soil-altering factors.
2. In general, the relationships cannot be assumed to be linear.
3. Little work has been done on interactions between factors.
4. Soil can be spatially predicted from geographic position using a variety of techniques.
5. Soil can be predicted from other soil attributes at the same location.
6. Soil can be predicted from itself, other soil attributes, and environmental attributes at neighboring locations.

37.3 The Scorpan Model

McBratney et al. (2003) generalized and formalized the digital soil mapping approach using a Jenny-like formulation not for explanation but for empirical quantitative descriptions of relationships between soil and other spatially referenced factors with a view to using these as soil spatial prediction functions. This is called the "*scorpan*" model, which can be written as follows:

$$S_c = f(s, c, o, r, p, a, n) \quad \text{or} \quad S_a = f(s, c, o, r, p, a, n), \quad (37.8)$$

where
S_c is soil classes
S_a is soil attributes

There are seven factors:

1. *s*—soil, other properties of the soil at a point
2. *c*—climate, climatic properties of the environment at a point
3. *o*—organisms, vegetation or fauna or human activity
4. *r*—topography, landscape attributes
5. *p*—parent material, lithology
6. *a*—age, the time factor
7. *n*—space, spatial position

Soil is included as a factor because soil can be predicted from its properties, or soil properties from its class or other properties. The *s* refers to soil information either from a prior map, or from remote or proximal sensing or expert knowledge. Implicit in this are the spatial coordinates *x*, *y* (and probably not *z*) and an approximate or vague time coordinate $\sim t$. This time coordinate can be expressed as "at about some time *t*." So explicitly, for example,

$$S_c[x, y, \sim t] = f(s[x, y, \sim t], c[x, y, \sim t], o[x, y, \sim t], r[x, y, \sim t], p[x, y, \sim t], a[x, y], [x, y]). \quad (37.9)$$

Each factor will be represented by a set of one or more continuous or categorical variables, for example, *c* by average annual rainfall and average annual temperature or climate class.

We shall not consider the direction of causality. For example, many reckon vegetation to be dependent on soil and we could write $o = g(S)$, where *o* is set of vegetation classes or percentage cover of a species, *g* is some arbitrary function, and *S* is a set of soil classes or attributes. For our purpose, we could write $S = g^{-1}(o)$, where g^{-1} is the inverse function of *g*, $S = g^{-1}(o) = f(o)$. We stress that the approach in general is not theoretical, and it is largely empirical—where there is evidence of a relationship we use it. Clearly, although we do not require causality, we should be mindful of potential problems of nonuniqueness if *g* is not a monotonic function, as well it might be if *S* is say topsoil pH and *o* is the number of plants of a particular species per unit area.

A general soil prediction model would be

$$S(x, y, z, t) = f(Q), \quad (37.10)$$

where *Q* is predictor variable(*s*). Here we will consider some restrictions, in cases where *S* stands for $S(x, y, (z), t)$, that is, the soil class or attribute at some spatial location *x*, *y*, (*z*) and at some time *t*.

37.3.1 What Is S? Soil Classes S_c or Individual Soil Attributes S_a

The model must be able to predict the probability of a set of classes, for example, for the case of five classes, say, A, …, E, the model would predict the probability vector (p[A], p[B], p[C], p[D], p[E]), for example, $S_c[x, y] = (0.01, 0.72, 0.01, 0.02, 0.25)$ along with some measure of uncertainty. The problem will generally consist of a preexisting soil class label (from some soil classification system) at each soil observation location and a set of colocated environmental variables. These are called the training data. This represents a supervised classification or supervised learning problem. (More rarely, unsupervised learning also known as numerical classification may be used on observed soil attributes to first generate the class labels.) The supervised learning rules are fitted using the training data and then used at other locations where only environmental variables are observed. Most prediction methods treat soil classes as "labels," and their prediction only considers the minimization of the misclassification error. Soil classes at any taxonomic level have taxonomic relationships between each other, and in some instances, the errors in prediction of certain classes are more serious than the others. Minasny and McBratney (2007a, 2007b) proposed the incorporation of taxonomic distance between the soil classes in the prediction.

The model should also be able to predict individual soil attributes S_a along with a measure of uncertainty. The S_a might be the value of a given soil attribute at a certain depth, for example, the clay content at 60 cm, that is, $S_a[x, y]$ = 310 g/kg, along with an uncertainty measure. Similarly to the class problem, this will generally consist of a measured soil attribute at each soil observation location and a set of colocated environmental variables. These are the training or calibration data. This represents a generic (multiple) regression problem. The generic regression equations or rules are fitted using the calibration data and are then used at other locations where only environmental variables are observed.

Heuvelink and Webster (2001) have discussed the merger of discrete and continuous models of spatial variation. Heuvelink (1996) suggested the mixed model of spatial variation, in which the soil property is treated as the sum of a global mean, a class-dependent deviation from the mean, and a spatially correlated residual. Prediction with this model boils down to kriging with an external drift (Delhomme, 1978), which in this case is a classification. Its main advantage is that it performs well over the whole range of spatial variation, from exclusively discrete realities. A more general interpretation of this kind of idea, and the one we use here, is that the external drift represents $f()$ and can be any kind of function. The discreteness or continuity of S will depend on the magnitude and form of $f()$. In the Heuvelink (1996) case, the $f()$ is a one-way analysis of variance model, a special case of a generalized linear model (GLM) (McCullagh and Nelder, 1983; Lane, 2002).

37.3.2 The General Approach

If we write the equation as $S = f(Q) + e$, then the general approach we shall use is to take some observations of S in the field at known locations $[x, y]$ and fit some kind of function to a set of pedologically meaningful predictor variables Q, which will generally be raster data layers of size M in a GIS. Once the model is fitted at the m observation points, the prediction can be extended to the M points or cells in the raster thereby giving a digital map. The efficiency of the method relies on the fact that hopefully $m \ll M$, and because S is much more difficult and expensive to measure than the Q. The success will depend on the following:

1. Having sufficient predictor variables observed everywhere or at least with a relatively high data density
2. Having enough soil observations (data points) to fit a relationship
3. Having functions $f()$ flexible enough to fit a nonlinear relationship
4. Having a good relationship between the soil and its environment

Followed by a discussion of quantitative procedures for fitting $f()$ in Section 37.3.3, we present some considerations concerning e in Section 37.3.4, and a review of previous studies in Section 37.3.5.

37.3.3 Form of $f()$

We will now discuss some forms of $f()$, most, but by no means all, of which have been or can be used for this kind of problem. For the sake of brevity, we shall not delve deeply into the mathematics of the methods. The advance in statistical learning techniques, enhanced by ongoing developments in data mining, has aided the use of different forms of $f()$ in soil science. Recent developments and technical details of the statistical modeling have been recently and extensively reviewed by Hastie et al. (2009). When predicting soil classes, some kind of *supervised classification* will be used, and for soil attributes, some kind of *generic regression* will be used. These are now discussed.

37.3.3.1 Linear Models

Linear models include regression for predicting soil attributes and classification for predicting soil classes. Linear regression included in this section is linear models using ordinary or generalized least squares (GLS). Linear methods for classification include discriminant analysis.

37.3.3.1.1 Ordinary Least Squares

For multiple linear regression, the model is written as follows:

$$\mathbf{s} = \mathbf{Qb} + \mathbf{e}, \quad (37.11)$$

where

s is the vector of response (predicted soil attribute)
Q is the matrix of predictor variables
b is parameter vector of the linear function
The error component, **e**, represents the deviations of the model to the observed value

The parameter is usually solved using ordinary least squares (OLS), with assumptions that **e**

1. Are independently and identically distributed (independence assumption)
2. Have zero mean and finite variance (homoscedasticity assumption)
3. Are normally distributed (normality assumption)

OLS has been used widely in prediction of soil attributes, because of the ease of use and wide availability. The predictors are usually continuous variables; however, qualitative factors or discrete variables can also be integrated.

37.3.3.1.2 Principal Component Regression and Partial Least Squares

When a large number of correlated predictor variables is present (such as electromagnetic spectra), usually principal component analysis is used to produce linear combinations of the original inputs. Selected principal components are then used in place of the original predictors. Alternatively, partial least squares (PLS) (Martens and Naes, 1989) are developed, which constructs a new set of components as regressor variables that are linear combination of the original variables. Unlike principal component regression, which only uses a combination of the predictors, the components in PLS are determined by both the response variable(*s*) and the predictor variables.

Principal component regression and PLS have been used quite extensively in predicting soil attributes from the electromagnetic spectrum, especially in the near and mid-infrared ranges (e.g., Chang et al., 2001). This method may be necessary if the environmental covariates consist of hyperspectral imagery.

37.3.3.1.3 Linear Discriminant Analysis

Discriminant analysis (Fisher, 1936) is the seminal supervised learning technique. It has been applied in soil science for more than 70 years. Webster and Burrough (1974) used the method to allocate soil observations into existing classes. Henderson and Ragg (1980) employed a multivariate logistic method to assess the usefulness of soil properties for distinguishing between taxonomic units. The method was perhaps first used for digital soil mapping by Bell et al. (1992, 1994) who related soil drainage classes to landscape parameters and used the resulting discriminant functions for spatial predictions.

The theory is readily accessible in Webster and Oliver (1990) and Hastie et al. (2009). Triantafilis et al. (2003) generalized the theory to a fuzzy linear discriminant analysis. This considers the a priori membership of each individual to each of the classes.

37.3.3.2 Generalized Linear Models

GLMs extend the linear regression models to accommodate nonnormal response distributions (Hastie and Pregibon, 1992). The theory and applications in soil science have been reviewed by Lane (2002). Usually, to accommodate for nonlinearity, transformation of variable is introduced, GLM attempt to modify the model rather than transforming the data (Lane, 2002). GLMs have the assumption of independence between the response and predictor variables.

37.3.3.2.1 Prediction of Continuous Soil Attributes S_a

McKenzie and Austin (1993) used GLMs to predict soil attributes (clay content, CEC, EC, pH, bulk density, and COLE) using environmental variables (geomorphic unit, local relief, etc.) as predictors. Other examples include Odeh et al. (1995, 1997) and McKenzie and Ryan (1999) who used GLMs to predict nonnormally distributed continuous variables. In both of these cases, the GLM was used in preference to standard linear regression because of the nonnormal distribution of the response variable. Park and Vlek (2002) found that GLMs performed better than neural networks and regression trees in predicting soil attributes from environmental variables.

37.3.3.2.2 Prediction of Soil Classes S_c

Gessler et al. (1995) used GLMs to predict the presence or absence of a bleached A2 horizon using digital terrain information. In this case, a logit link function was used due to the binomial distribution. Campling et al. (2002) used logistic regression to model soil drainage classes from terrain attributes and vegetation indices as calculated from a Landsat TM image. Kempen et al. (2009) developed a logistic regression model for predicting soil groups in the Netherlands from DEM, groundwater maps, land cover, paleogeography, geomorphology, and soil maps. The model-building process was guided by pedological expert knowledge to ensure that the final regression model is not only statistically sound but also pedologically plausible.

37.3.3.3 Generalized Additive Models

Generalized additive models (GAMs) attempt to characterize the nonlinear effect, which is not considered in GLMs. GAMs have the form:

$$S = \alpha + \sum_{j=1}^{p} f_j(q_j) + e, \tag{37.12}$$

where
α is the intercept
f_j is a nonparametric "smoothing" functions for predictor q_j
p is the number of covariates
e is the error

The smoothing functions can be splines, loess, kernel, and other, smoothers (Venables and Ripley, 1994). Smoothers fit the data locally, and crucial to the fit is the size of the local neighborhood, which is generally controlled by a smoothing parameter. The smoothing parameter controls the variance–bias trade-off. A large neighborhood produces estimates with low variance and potentially high bias, a small neighborhood produces the reverse effect (Hastie and Tibshirani, 1990).

The reported use of GAMs in the soil science literature has been minimal. One of the few studies has been Odeh et al. (1997) who compared GAMs with GLMs and linear regression for the prediction of organic carbon with digital terrain information as secondary information. GAMs were found to be superior. Bishop and McBratney (2001) used GAMs for mapping of soil cation exchange capacity from environmental factors (terrain attributes, bare soil color aerial photograph, bare soil LANDSAT imagery, crop yield data, and soil apparent electrical conductivity).

37.3.3.4 Tree Models: Classification and Regression

Various implementations of decision trees (DT) or classification trees (CART) have been used to extract and apply knowledge about soil spatial patterns. Tree-based models are fitted by successively splitting a data set into increasingly homogenous subsets. The response variable can be either a factor (classification trees) or a continuous variable (regression trees), and the explanatory variables can be of either type (McKenzie and Ryan, 1999). Once the partitioning has ceased, the subsets are called terminal nodes. Each terminal node is typically assigned the label of the majority class. Splits, or rules defining how to partition the data, are selected based on information statistics that measure how well the split decreases impurity (heterogeneity or variance) within the resulting subsets (Scull et al., 2005). All possible organizations of explanatory variables into two groups are examined recursively to evaluate the effectiveness of each possible split.

Decision trees have found favor in digital soil mapping (DSM) because they can handle missing values, can use continuous and categorical predictors, are robust to predictor specification, and make very limited assumptions about the form of the regression model (Henderson et al., 2005).

Lark et al. (2007) expressed some reservations about overfitting with data mining methods such as regression trees unless the methods were evaluated against truly independent test data.

Henderson et al. (2005) concluded that regression trees had been widely adopted because they were robust, could handle continuous and categorical data, and were reasonably easy to interpret. The final trees are essentially binary keys with yes/no decisions based on a single class or parameter value at each node. This makes it easy to express the splitting rules in semantic terms that are easily understood and appreciated by human interpreters.

37.3.3.4.1 Prediction of Continuous Soil Attributes S_a

This is called a regression tree. Pachepsky et al. (2001) used regression tree to predict sand and silt contents and water retention from terrain attributes (slope, curvature). One of the limitations in regression trees is the discrete predictions from each terminal node, which result in a lack of smoothness of the prediction surface. This can result in unrealistic representations of soil variability if the tree has a small number of terminal nodes (McKenzie and Ryan, 1999). An improvement to regression trees involves building multivariate linear models in each node (leaf). This type of model, which is analogous to using piecewise linear functions, has been implemented in the program Cubist (RuleQuest Research, 2000). This is used by Henderson et al. (2005) for mapping soil properties throughout Australia.

37.3.3.4.2 Prediction of Soil Classes S_c

This is called a decision tree or a classification tree. The binary decision tree algorithm uses a binary split that has exactly two branches at each internal node. There are different decision tree methods. The most commonly used is CART (Breiman et al., 1984). Lagacherie and Holmes (1997) discussed the application of CART for soil classification and its sensitivity to error. Another popular algorithm is C4.5 (Quinlan, 1992) and its later version See5 (RuleQuest Research, 2000). Bui and Moran (2003) utilized this program for mapping soil classes across the Murray–Darling Basin in eastern Australia. Moran and Bui (2002) refined the analysis of Bui et al. (1999) by using a "boosted" tree to reduce the classification error.

37.3.3.5 Neural Networks

Neural networks attempt to build a mathematical model that supposedly works in an analogous way to the human brain. Neural networks have a system of many elements or "neurons" interconnected by communication channels or "connectors," which usually carry numeric data, encoded by a variety of means, and organized into layers. Neural networks can perform a particular function when certain values are assigned to the connections or "weights" between elements. To describe a system, there is no assumed structure of the model, instead the networks are adjusted or "trained" so that a particular input leads to a specific target output. The mathematical model of a neural network comprises of a set of simple functions linked together by weights. The network consists of a set of input units, output units, and hidden units, which link the inputs to outputs. The hidden units extract useful information from inputs and use them to predict the outputs (Hastie et al., 2009).

Neural networks are now widely described in the soil science literature, mainly for predicting soil attributes. The application of neural networks as pedotransfer functions for predicting soil hydraulic properties is the most common.

37.3.3.5.1 Prediction of Continuous Soil Attributes S_a

The practice of predicting soil hydraulic properties in the form of pedotransfer functions can be found in many studies such as Minasny and McBratney (2002). Chang and Islam (2000) predict soil texture from multitemporal remotely sensed brightness temperature and soil moisture maps. Minasny et al. (2006) used a modified form of neural networks for mapping the parameters of a negative exponential depth function.

37.3.3.5.2 Prediction of Soil Classes S_c

Neural networks can be used to predict the probability of classes using multilogit transformation of the output. Another type of network is called self-organising maps (SOM, in this case, *not* soil organic matter) (Kohonen, 1982). Kohonen's network is an unsupervised classification splitting input space into patches with corresponding classes. It has the additional feature that the centers are arranged in a low-dimensional structure (usually a string, or a square grid), such that nearby points in the topological structure (the string or grid) map to nearby points in the attribute space.

Zhu (2000) used neural networks to predict the probability of soil classes from soil environmental factors. Fidêncio et al. (2001) applied two types of neural networks (radial basis function networks and

SOMs) to classify soil samples from different geographical regions in Sao Paulo, Brazil by means of their near-infrared (diffuse reflectance) spectra. Behrens et al. (2005) used neural networks to predict soil classes based on existing soil maps. The predictors are terrain attributes, geologic–petrographic units, and land use.

37.3.3.6 Fuzzy Systems

Zadeh (1965) defined a fuzzy set as "a class of objects with a continuum of grades of membership." Such a set is characterized by a membership (characteristic) function that assigns to each object a grade of membership ranging between 0 and 1. In classical Boolean logic, each individual, or site, either belongs to a class (membership = 1) or does not belong to a class (membership = 0). In Fuzzy logic, each individual, or location, is assigned a value between 0 and 1 that expresses the relative degree to which it belongs to a class. The degree of belonging to a given class (range 0–1) is assessed by comparing the properties that define the central concept for a class to the properties exhibited by any individual, or location, that one wishes to classify.

In the context of digital soil mapping, fuzzy systems represent one approach to assigning an estimate of the likelihood of a particular soil class or soil property value occurring at any given location under a given set of environmental conditions. Typically, the environmental conditions are represented by a series of covariates or predictor variables that exhaustively cover the spatial extent of an area for which predictions are required. The fuzzy likelihood that a given soil class or soil property value will occur at any point, and under any particular set of conditions, has been estimated using several quite different approaches. These approaches can be broadly characterized as unsupervised, supervised, and knowledge based (Hengl and MacMillan, 2009).

Unsupervised approaches typically do not have an a priori set of target classes defined and do not use spatially located reference locations to establish relationships between observed classes and values of the covariate predictor data sets. For example, Irvin et al. (1997) used a fuzzy *k*-means unsupervised approach to derive groupings of fuzzy soil–landform classes based solely on consideration of the input data values without reference to any existing classification. This approach used iterative processes to determine class means by minimizing distances in multidimensional attribute space. The resulting continuous classification created "partial" class memberships for each data point with membership values ranging from 0 to 1.

In situations in which the optimum number of classes to be identified, and the conditions or criteria that define the central concepts of each class, are not known a priori, statistical ordination techniques can be applied to extract class numbers and definitions. Fuzzy *k*-means or fuzzy *c*-means has been used extensively in soil science and DSM to identify optimum class numbers and central concepts and to then determine the fuzzy degree of membership of unclassified individuals or locations to the central concepts of each of the statistically defined classes (McBratney and De Gruijter, 1992; Odeh et al., 1992; De Gruijter et al., 1997; McBratney and Odeh, 1997; Burrough et al., 2000; Carré and Girard, 2002; Zhu, 2006). Detailed descriptions of the equations used in Fuzzy *k*-means, and the logic behind them, can be found in Odeh et al. (1992), McBratney and Odeh (1997), and Zhu (2006).

In digital soil mapping, most applications of fuzzy *k*-means classification have adopted an approach in which environmental covariates (e.g., slope, aspect, curvatures, climate variables, or vegetation cover) were used to define the central concepts of classes of soil of interest.

Supervised approaches all start out with a defined list of a priori classes that a user wants to predict and they all make use of spatially located reference data to establish the fuzzy membership functions that relate values of predictor variables to the corresponding likelihood that a particular soil class or property value will occur. These approaches all represent a form of data mining in which instances of known classes at known locations are used to identify values of predictor variables that are associated with the presence of a given class or property value. The training data can be drawn from a variety of different sources, including preexisting soil maps, preexisting (legacy) point observations or sample sites, newly collected, geolocated field observations or sample data or even points or areas identified interactively, on-screen, by a knowledgeable expert as being likely to belong to a particular class or to have a particular property value. A key requirement for successful use of supervised approaches is that there must be instances in the training data of all classes that exist in an area and that a user wants to predict. Additionally, the training data should cover, or represent, the full range of values exhibited by each and everyone of the covariates used to predict the fuzzy class memberships. A Latin hypercube sampling (LHS) procedure (Minasny and McBratney, 2006) can be used to analyze the feature space of the covariates to establish whether the instances in the training data represent the full range of covariate values in the feature space. If the training data do not fully cover the feature space, then there will be confusion as to which classes or property values are most likely to occur at locations in the feature space where there are no instances of training data for a particular range of covariate values.

Several different approaches have been described for collecting and analyzing training data to establish fuzzy membership functions for soil classes or properties. Some involve identifying multiple instances of each class of interest within an area of interest. Each instance, or location, is treated as a representative profile for a soil class of interest, and each unclassified location is compared to each instance of each class to identify the instance to which it is most similar (Zhu et al., 1996; Zhu, 2000). In a variation of this approach, Shi et al. (2004, 2005, 2008) treated individual locations are cases and assigned different weights to cases such that cases closer to an unclassified location carried a greater weight than cases farther away. This case-based reasoning (CBR) permitted instances of a class to influence the classification of unclassified locations that were located near to them while not exerting much influence on classification of locations that were further away.

37.3.3.7 Other Methods

Ballabio (2009) used support vector machine (SVM) algorithm for mapping soil organic carbon in mountainous regions in Italy.

SVM is a set of supervised learning methods used for classification and regression, and its algorithm aims to match model complexity to data complexity. SVM performs classification by constructing a nonlinear n-dimensional hyperplane that optimally separates the data into two categories. This is done by constructing a linear boundary in a large, transformed space of the feature space. Support vector regression is a form of SVM that is applied to regression problems.

Genetic algorithms (GAs) (Goldberg, 1989) are randomized search and optimization techniques guided by the principles of biological evolution and natural genetics. They have been used mainly in optimization of large multidimensional problems. Pal et al. (1998) developed a GA-classifier and applied it to satellite imagery (Pal et al., 2001). It attempts to approximate the class boundaries of a data set with a fixed number of hyperplanes in such a manner that the associated misclassification of data points is minimized. Nelson and Odeh (2009) explored the Genetic Algorithm for Rule-set Production (GARP) (Stockwell and Noble (1992). GARP was developed to model the habitat distribution of plant and animal species using locations of known species presence and environmental variables. The algorithm has stochastic elements that produce a population of rules prior to iterative use of the best rules in a given generation to develop the next generation until some convergence criteria are met and a solution given. GARP is noteworthy in that it was developed specifically to utilize legacy data. However, Nelson and Odeh's results showed that GARP did not perform as well as decision tree algorithm, implying the need to improve the algorithm for its full potential to be realized for digitally mapping soil classes.

37.3.3.8 Strengthening Models: Bagging, Boosting

There has recently been empirical evidence that the accuracy of $f()$ prediction can be enhanced by generating multiple models and aggregating them to produce an estimate. There are two widely used approaches for producing and using several models that are applicable to a wide variety of statistical learning methods. Bootstrap aggregating or bagging (Breiman, 1996) and boosting (Freund and Schapire, 1997) manipulate the training data in order to generate different models. These methods arise more naturally in the supervised classification problem, but they can be extended to generic regression.

Bootstrap methods (Efron and Tibshirani, 1993) assess the accuracy of a prediction by sampling the training data with replacement. Suppose the training data are composed of predictors Q and response S of size N, we draw B data sets each of size N of the training data by sampling with replacement. For each of the bootstrap data set Z^b, $b = 1, 2, \ldots, B$, we fit model $\hat{f}^b(q)$. The bagging estimate is calculated as

$$\hat{f}(q) = \frac{1}{B}\sum_{b=1}^{B}\hat{f}^b(q). \tag{37.13}$$

The bagging method is the basis of the random forests algorithm.

Boosting combines the outputs of many "weak" models to produce a powerful "committee." Boosting uses all the data at each repetition, but maintains a weight for each instance in the training set that reflects its importance. Adjusting the weights causes the model to focus on different data and, hence, leads to different models. The multiple models are then aggregated by voting to form a composite model. In bagging, each component model has the same vote, while boosting assigns different voting strengths to component classifiers on the basis of their accuracy. Moran and Bui (2002) used boosting to improve their digital soil map of the Murray–Darling basin. Henderson et al. (2005) described how fitting multiple trees through boosting and bagging permitted tree averaging, which could lead to potential improvements in the predictive power of decision trees.

Grimm et al. (2008) described random forest (RF) as an implementation of randomized classification and regression trees in which numerous trees are generated within the algorithm and finally aggregated to give one single prediction (Hastie et al., 2009). Random forests combine the tree predictors such that each tree depends on the values of a random vector sampled independently and with the same distribution for all trees in the forest. It has been used successfully for mapping soil carbon (Grimm et al., 2008). The model can handle lots of inputs, has a high accuracy, and apparently does not overfit the data. This is achieved by using different subsets of the training data (with bootstrap) and using different subsets of the predictors for training the tree (determined randomly). Thus, only patterns that are present in the data would be detected consistently by a majority of the trees. However, this claim is not always true. Hastie et al. (2009) showed that when the number of variables is large but the fraction of relevant variables for prediction is small, random forests are likely to perform poorly and random forests can certainly overfit the data; the average of fully grown trees can result in a model that is too rich and incur unnecessary variance.

37.3.3.9 Expert (Knowledge-Based) Systems

Expert systems (Dale et al., 1989) are ways of harvesting and engineering knowledge. The terms expert system and knowledge-based system have been used both interchangeably and for quite fundamentally different approaches. Zhu (2008) observed that knowledge-based approaches used in digital soil mapping have extracted and codified knowledge in three main ways, these being as follows:

- Extraction of knowledge from existing documents, legends, and keys
- Extraction of tacit empirical knowledge from human domain experts
- Extraction of knowledge from spatial data sets via data mining

Knowledge already formalized and codified in documents, legends, and keys may simply require implementation as either Boolean or Fuzzy rules and not require any specific extraction process. Empirical knowledge about likely relationships between classes of interest and formative environmental conditions can be extracted by asking local experts to describe and quantify their heuristic, and mostly tacit, domain knowledge through structured interview and evaluation sessions (Zhu et al., 1996; Zhu, 2008). Extraction of knowledge via application of statistically based data mining techniques works by comparing the spatial co-occurrence of known examples of a particular class or entity with the spatial pattern exhibited by sets of covariates that are considered likely to possess some predictive relationship with the class or entity of interest. Interestingly, the widely used SoLIM (Zhu, 1997, 2008) and SIE (Shi et al., 2004, 2008) fuzzy inference systems have both utilized all three approaches to extract expert knowledge.

Let us first consider those expert systems that make use of some form of data mining technique. Data mining methods require two main inputs. First, they require evidence in the form of known observations of the classes or values of interest at known locations. These provide the training data. Training data may exist in either point or map form or both. Second, they require data sets of explanatory variables that are spatially referenced and cover entire areas of interest exhaustively. These explanatory data sets provide the means of developing and then extrapolating predictive relationships from point (or localized map) observations to entire areas of interest. Data mining can be considered as a form of supervised classification as described elsewhere in this chapter.

Decision trees, or classification trees (see Section 37.3.3.4), have emerged as a favored data mining approach in digital soil mapping. Another data mining method that has found some use in digital soil mapping is Bayesian Analysis of Evidence (Skidmore et al., 1991). Cook et al. (1996a) and Corner et al. (1997) developed and applied the custom software program *Expector* to capture and apply human heuristic beliefs in a structured belief system. Mayr et al. (2008) used commercial Bayesian analysis of evidence software, *Netica* by Norsys, to extract rules for classifying soil types by extracting knowledge from analysis of spatially referenced auger bore data and existing soil maps.

Bui (2004) observed that knowledge engineering is the study of expert systems and how experts build mental models. She argued that soil maps and their legends are representations of structured knowledge, namely, the soil surveyor's mental soil–landscape model. Soil map unit descriptions describe, in words, the soil–landscape relationships. For example, in a narrative legend, a soil toposequence can be described as a function of relative elevation/slope position, slope length and slope gradient, and curvature. Sometimes, the soil–landscape model for each map unit is represented as a 2D cross section or 3D block diagram. Soil map unit descriptions and block diagrams are usually the only way the soil surveyors' mental models are transmitted to others. This soil–landscape model has been described by Hudson (1992) as the operative paradigm for soil survey. Bui (2004) and Wielemaker et al. (2001) suggested methodological frameworks to formalize the landscape knowledge of the soil surveyor by structuring terrain objects in a nested hierarchy followed by inference and formalization of knowledge rules.

Knowledge-based expert systems have been characterized as consisting of two main components, these being a "knowledge base" and an "inference engine" (Skidmore, 1989). Zhu et al. (1996) described knowledge-based systems as being organized on three levels of data, knowledge base, and inference engine. Data refer to information on environmental conditions of an area such as elevation, slope, and aspect. The knowledge base contains the declarative knowledge about a particular problem being solved. In the case of predictive soil mapping, the knowledge base contains information about the relationships between the spatial distribution of soils in an area and the environmental factors that are believed, or known, to influence this distribution. The inference engine applies the knowledge to the data to produce predictions or results. It controls when and how the specific problem solution knowledge is used. A particular concern of an inference engine is often the order, or sequence, in which a particular piece of information (or a rule) is processed (Skidmore, 1989).

Shi et al. (2008) recognized several useful distinctions regarding knowledge used in devising and applying fuzzy logic rules. First, they distinguished between *global knowledge* and *local knowledge*. Global knowledge applies with equal validity across an entire area of interest whereas local knowledge is applicable only within certain restricted extents. Second, they distinguished between knowledge based on understanding of relationships between classes of interest and values of environmental variables in *parameter space* and knowledge based on the existence of known classes or conditions in particular locations in *geographical space*. They differentiated global rule-based reasoning (RBR), capable of handling global knowledge in parameter space, from global and local CBR, capable of handling global versus local knowledge in geographical space. CBR is used when local experts can identify locations at which specific classes of interest are known to occur and can state whether these classes always occur in these same settings throughout an area or only occur in these settings within a restricted geographical extent. RBR is used when local experts can identify the environmental conditions, as defined by available covariates, under which conceptual classes of interest are known to occur in parameter space.

Shi et al. (2008) provide an excellent description of eight main steps that are involved in devising, applying, evaluating, refining, and then finalizing knowledge-based rules to predict the spatial pattern of soil classes using the soil inference engine (SIE). They noted that MacMillan et al. (2007) independently, but not surprisingly, employed a procedure that was almost identical to theirs. The eight steps described by Shi et al. (2008) are as follows:

1. The soil scientist provides the global knowledge, including names of the soils he/she expects to see in the mapping area and descriptions of the environmental conditions of these soils. The environmental conditions depicted by environmental values are formalized into Fuzzy rules and saved into rule bases; those represented by geographical locations are formalized into cases and saved into global case bases.
2. The soil scientist or a GIS specialist prepares data layers for characterizing the environmental conditions. The data layers may cover terrain attributes, geology, vegetation, climate, and other features. These data layers are stored in a GIS database.

3. SIE performs RBR or global CBR, using the global knowledge and the GIS database, to generate maps of the general pattern of soil distribution in the mapping area.
4. The soil scientist verifies the draft maps from (3). If he/she is satisfied with the maps, the mapping is done. Otherwise, he/she may go back to (1) to adjust the rules or global cases, or go to (5) to fine-tune the draft maps.
5. The soil scientist provides the local knowledge, in the form of cases, to address local exceptions. The cases are saved in local case bases.
6. SIE performs local CBR using the local knowledge and the GIS database.
7. The soil scientist verifies the maps from (6). He/she can adjust the cases and run local CBR again. He/she repeats this process until he/she is satisfied with the result.
8. The soil scientist uses SIE and other GIS tools to integrate the results from (3) and (6) to generate the final maps.

The development of the SoLIM inference model (Zhu and Band, 1994; Zhu, 2006, 2008) through time provides a revealing example of trends in how fuzzy knowledge of soil–landscape relationships has been captured and applied. Early implementations of SoLIM were mostly based on using interviews to capture the tacit knowledge of local experts. However, they often did not take into account the consideration of the cognitive aspects of knowledge formulation and required users to provide either exact forms for membership functions for each class to be defined or a very large set of typical cases or instances that could be used as exemplars in CBR (Qi et al., 2006). In a second stage of development, many SoLIM applications relied on acquisition of large numbers of user-selected spatially referenced instances or cases to supply the information required to define central concepts, and local exceptions, for representative or typical soil classes (Zhu, 2000; Shi et al., 2004, 2005). As SoLIM subsequently evolved and was increasingly applied for operational mapping of larger areas, the use of large volumes of geolocated field observations to serve as training data sets became increasingly more costly and less viable. More recent applications of SoLIM have notably adopted approaches that were based on rapidly identifying prototypes (Qi et al., 2006) or semantically defined central concepts (Liu and Zhu, 2009) by using expert tacit knowledge to define conceptual or mental models of central concepts of soil classes to be predicted. These semantically defined representative concepts not only require far less time and effort to produce than geolocated reference sites but they have also been shown (Qi et al., 2006; Liu and Zhu, 2009) to be more effective and to produce more accurate predictions of the spatial distribution of soil classes than earlier case-based approaches.

The equation for computing an estimate of any given soil property of interest is given by Zhu et al. (1997) as follows:

$$V_{i,j} = \frac{\sum_{k=1}^{n} \mathbf{S}_{i,j}^{k} V^{k}}{\sum_{k=1}^{n} \mathbf{S}_{i,j}^{k}}, \tag{37.14}$$

where

$V_{i,j}$ is the continuous soil property to be predicted
$S_{i,j}$ is the soil similarity value for the soil k at location i,j
V^k is the mean or representative value for the soil property of interest for soil k

While this approach produces estimates that are not quite as accurate or reliable as those produced by other, more data intensive methods such as regression-kriging or regression trees, it is an attractive option for areas that lack extensive databases of geolocated point information on soil properties. In these circumstances, all that is required is a single representative value for each soil property of interest, for each depth or horizon of each soil class of interest and an ability to develop fuzzy knowledge-based rules about which environmental factors influence the development and occurrence of the soil classes of interest.

37.3.3.10 Unsupervised Classification

In the previous sections, when we were discussing classes we were considering "supervised classification." This is also known as allocation or identification. This is where we wish to produce prediction equations for placing soil existing soil classes, such as a particular categorical level in a national or international classification system. However, we may first wish to make new classes from the observed soil properties. This is known as unsupervised classification. Much of the early work on pedometrics, in the 1960s, focussed on this topic. The numerical classification methods that have been used quite extensively in soil science more recently are k-means and fuzzy k-means (Odeh et al., 1992; De Bruin and Stein, 1998; Triantafilis et al., 2001). See the section on fuzzy systems on the application of fuzzy k-means. There is also a semi-supervised classification considering classification in the presence of some labeled data (Pedrycz and Waletzky, 1997).

Unsupervised classification is an option for making digital soil class maps particularly where the national or international scheme does not project well onto the soil–landscape. However, once the new classes have been established at the soil observation locations, then one of the previous methods, inter alia discriminant analysis, multiple logistic regression, regression trees, needs to be applied to fit equations and then make predictions from environmental covariates at the other locations where no soil properties have been observed.

Carré and Girard (2002) used a continuous method for horizon and profile classification called OSACA. This was based mainly on field soil morphological attributes. Carré and Jacobson (2009) further developed a program for quantitative grouping of soil layer descriptions into profile classes. The program calculates the taxonomic distances between observed profiles based on layer (horizon) characteristics. Characteristics can be either observed soil properties or layer class memberships. OSACA can allocate observed soil profiles to existing classes or create a new classification of the profiles. Their method is unique in that it models the taxonomic distance to each of the class centroids at each observation site. Because these distances are continuous variables, multiple linear regression on environmental variables was used as the "supervised classification" step. One regression equation was developed for each class and the distances predicted at each site on their prediction raster.

37.3.4 Spatial Considerations

The older *corpt* approach has no intrinsic or formal spatial component other than the functions are predicted in a spatial context, that is, spatial position is not taken into consideration. This seems unwise for a mapping application. Spatialization can be introduced by considering spatial components of the environmental and soil variables (Section 37.3.4.1) and by perpending the spatial correlation structure of the residuals (Section 37.3.4.1), as was briefly discussed in Section 37.3.3.1.

37.3.4.1 Decomposition of *Q* Factors into Spatial Components

DSM studies have increasingly recognized the need to identify which scales of variation, which are operative and discernable in the predictor data layers (*Q*), are most strongly related to observed variation in a property or class of interest (Bui et al., 1999, 2006; Fisher et al., 2004, 2005; Smith et al., 2006; Arrell et al., 2007; Deng, 2007; Wu et al., 2008; Zhu, 2008). Several approaches have been used to investigate the spatial structure that may be operative in predictor data sets and to identify the ranges of distance over which this structure operates.

Bui et al. (2006) observed that a spatial hierarchy emerged from their decision tree analysis of predictive relationships between environmental covariates and soil properties over the entire extent of the continent of Australia. They undertook a very logical, but frequently neglected, interpretation of their results in terms of physical and biological processes that influence pedogenesis. They sought to interpret their results in terms of the degree to which different environmental covariates, operative at different scales, could be related to soil-forming processes that influenced the values of the soil property variables they were modeling. Their results indicated that the state factors of soil formation formed a hierarchy of interacting variables, with climate being the most important factor at a continental scale, but with different climatic variables dominate in different regions. Their results also showed that lithology was almost equally important in defining broadscale spatial patterns of soil properties and that shorter-range variability in soil properties appeared driven more by terrain variables. The message to be taken from this thoughtful and rigorous analysis is that different environmental covariates are likely to drive the significant processes of soil formation at different scales and over different distances. Individuals seeking to model the spatial distribution of soil properties ought to first systematically examine the variation in those properties at different scales and seek to identify the environmental covariates (and their scales) that are most closely associated with causing variation in soil property values at specific scales.

In this same context, there has been an increasing number of studies that investigated the effects of different DEM resolutions (Zhang and Montgomery, 1994; Thompson et al., 1997, 2001; Fisher et al., 2004, 2005; Kienzle, 2004; Arrell et al., 2007; Erskine et al., 2007; Seibert et al., 2007; Wu et al., 2008) and neighborhood (window) sizes (Schmidt et al., 2005; Smith et al., 2006; Behrens et al., 2007a, 2007b; Zhu, 2008) on the performance of models designed to predict soil properties or soil classes. Typically, such studies have been applied to smaller areas over which the regional climate and lithology could be assumed to hold constant, or to vary only slightly, relative to more significant variation associated with topography. When predicting individual soil properties, the approach has often been to fit regression models in which the soil property of interest is predicted using environmental covariates computed at different neighborhood sizes and/or grid resolutions. For example, Schmidt and Hewitt (2004) showed that the predictive value of profile curvature for estimating soil properties in a regression model varied with the scale (window size) for which profile curvature was calculated. Similarly, Smith et al. (2006) showed that there was a range of window sizes, which varied from landscape to landscape, which produced the most accurate maps of soil classes, and that finer resolution DEMs did not necessarily create more accurate maps. They observed that there was no apparent physical process-based significance behind using a fixed 3×3 neighborhood for computing terrain attributes. Instead, they concluded that it is much more important to match the terrain characteristics computed from the DEM using a specified neighborhood size, with the characteristics of the real world landscape. In a subsequent publication, Zhu (2008) observed that soil-forming processes operated at specific scales, and that spatial analysis should employ a neighborhood size comparable to the spatial scale of the process under consideration. He argued for use of joint distributions of processes and parameters with recognition of effective neighborhoods, defined in terms of the extent over which a process of interest operates. This spatial extent is comparable to the concept of grain size and, for many soil–landscape studies, is equivalent to the extent defined by a hillslope from crest to channel.

Other approaches that have been used to investigate the spatial structure of environmental covariates include factorial kriging analysis (Bourgault, 1994; Wen and Sinding-Larsen, 1997; Oliver et al., 2000) and wavelets (Zhu and Yang, 1998; Carvalho et al., 2001; Mendonça-Santos et al., 2006). Both of these methods decompose the separate variables into separate hierarchical spatial components of decreasing spatial resolution. The factorial kriging method assumes stationarity while the wavelet method does not. The factorial kriging method finds the scale of the components from the observations, whereas in wavelets the various scales are dictated by the size of the image, that is, the scales are increasing powers of 2 pixels.

The various components of *Q* could conceivably all be derived at different spatial scales and used as separate layers in the fitting of *s*. It is more than likely that the short spatial range components (e.g., the nugget component) might not relate to soil and can be removed (see Smith et al., 2006). In a quite a different application Oliver et al. (2000) found that land use was related to a long-range spatial component in SPOT imagery and not to two shorter range components.

37.3.4.2 Structure in *e*: Generalized Least Squares and Geostatistics

It would be naïve to imagine that there is no spatial structure in *e*. If *e* has a spatial structure, then GLS or geostatistics can be applied. Why would *e* have a spatial structure? The answers could be as follows:

- *Scorpan* is incorrect.
- Attributes used to describe *scorpan* are inadequate.
- Interactions are misspecified.
- Form of *f*() is misspecified.
- Something intrinsic—such as spatial diffusion, interaction, or inhibition processes.

Variograms of the fitted parts of the soil spatial prediction functions for the various factors will be instructive in elaborating these possibilities.

37.3.4.2.1 Generalized Least Squares

In GLS (Cressie, 1993): $\mathbf{s} = \mathbf{Qb} + \mathbf{e}$; errors $\mathbf{e}$ belong to multivariate normal distribution with mean 0 and covariance matrix $\mathbf{V}$: $\mathbf{N}(0, \mathbf{V})$. For spatial data, it can be further simplified assuming the error is homogenous with variance σ^2; thus, $\mathbf{V}$ can be replaced by $\sigma^2\mathbf{C}$, where $\mathbf{C}$ is the correlation matrix of the errors (Lark, 2000).

GLS has been used in soil science literature, for example, Samra et al. (1991) predicted tree growth from soil sodicity parameters with spatially correlated errors. Other examples include Aiken et al. (1991), Opsomer et al. (1999), and Vold et al. (1999). Lark (2000) provided the theory and example of using GLS for mapping soil organic matter content. Hengl et al. (2003) utilized GLS as a regression-kriging procedure for spatial prediction of soil properties in Croatia.

The drawback with this method is the heavy computation time when a large volume of data is involved, as the calculation time of semivariance and inverse of the correlation matrix $\mathbf{C}$ will increase (approximately cubically with sample size); nevertheless, Opsomer et al. (1999) showed some mathematical manipulation to avoid the inversion of the whole matrix. Pace and Barry (1997) developed a spatial autoregressive model, which utilized the sparse matrix technique to allow for quick computation for large spatial data. Another approach has been to choose classes of covariance functions for which kriging can be done exactly, even though the data set is massive (Cressie and Johannesson, 2008). In this approach, a multiresolution spatial process is constructed explicitly so that (simple) kriging can be computed extremely rapidly, with computational complexity linear in the size of the data.

37.3.4.2.2 Geostatistics: Scorpan Kriging

Here, we recognize that the spatial "trend" can be described by *f*(*s*, *c*, *o*, *r*, *p*, *a*, *n*), and the residuals *e* modeled separately by variograms and kriging. The final prediction is the sum of *f*() and *e*.

37.3.4.2.2.1 Scorpan: Universal Kriging Universal kriging allows incorporation of both deterministic and stochastic components in kriging:

$$S(\mathbf{x}) = \sum_{j=0}^{p} b_j q_j(\mathbf{x}) + e(\mathbf{x}) + \varepsilon, \tag{37.15}$$

where

the first term represents the nonstationary trend, which is modeled as a set of linear functions of the environmental variables $\mathbf{Q}$ with parameter vector $\mathbf{b}$

the second term is the stochastic component modeled by variogram

the third component ε is the error term that occurs below the level of spatial resolution or support of the covariates

Universal kriging can be solved by modifying the kriging system. However, the trend function is only limited to linear functions, and when the number of variables *p* is large, the matrix inversion to solve the system can consume heavy computation time.

Alternatively, the trend function can be modeled separately, where kriging is combined with regression (Ahmed and DeMarsily, 1987; Knotters et al., 1995). This method involves regression of the soil attributes as a function of predictor variables. This is followed by kriging of the regressed values, where the variance of the predicted (from the regression model) is used as the uncertainty of the modified kriging system. This is also known as kriging with uncertain data (Ahmed and DeMarsily, 1987). Odeh et al. (1994, 1995) defined regression-kriging where model *f*() is used to describe the relationship between predictors and soil attributes:

$$S(\mathbf{x}) = f(\mathbf{Q}, \mathbf{x}) + e'(\mathbf{x}) + \varepsilon, \tag{37.16}$$

where

$f(\mathbf{Q}, \mathbf{x})$ is a function describing the structural component of S as a function of $\mathbf{Q}$ at $\mathbf{x}$

$e'(\mathbf{x})$ is the locally varying, spatially dependent residuals from $f(\mathbf{Q}, \mathbf{x})$

In regression-kriging, the soil property *S* at an unvisited site is first predicted by *f*() and then followed by kriging of the residuals of the model.

As discussed at the end of Section 37.3.3.10, Carré and Girard (2002) used a continuous method for horizon and profile classification followed by multiple linear regression on environmental variables. One regression equation was developed for the taxonomic distance to each class centroid, and the taxonomic distances were predicted at each site of their prediction raster. The residual taxonomic distances to each class centroid were then spatially predicted onto the raster using ordinary kriging and added to the taxonomic distances from regression analysis. The summed taxonomic distances were then displayed and manipulated to make class maps. In this way, regression-kriging can be used to make continuous or discrete soil class maps taking into account the spatial correlation structure of the residuals from the fitted classes at each data point.

37.3.4.2.2.2 Scorpan: Simple Kriging Because $f()$ is modeled under the assumption that e has zero mean, simple kriging can be applied to the residuals of the model. Simple kriging allows prediction of the spatially correlated residuals with known mean where the weights of the kriging equation do not need to sum to unity (Webster and Oliver, 1990).

37.3.4.2.2.3 Scorpan: Compositional Kriging So far, kriging has been used mainly to predict soil attributes, for prediction of soil classes incorporating predictor variables **Q**, a form of compositional kriging with external trend is proposed:

$$\Pr[S_c(\mathbf{x})] = f(\mathbf{Q},\mathbf{x}) + e'_c(\mathbf{x}) + \varepsilon, \tag{37.17}$$

where $\Pr[S_c(\mathbf{x})]$ is the probability of the soil at $\mathbf{x}$ that belongs to soil class c. The probability of the soil classes $c = 1, \ldots, K$ must sum to 1 and the residuals of the probability must sum to 0:

$$\sum_{c=1}^{K} \Pr[S_c] = 1, \tag{37.18}$$

$$\sum_{c=1}^{K} e'_c = 0. \tag{37.19}$$

Solution of this method will involve prediction of probability of occurrence of soil classes using a form of $f()$, such as logistic regression, and compositional kriging of their residuals (Walvoort and de Gruijter, 2001).

37.3.4.2.3 Geostatistics: Co-Kriging and Coregionalization Analysis

Another method is co-kriging. Any of the q layers can be a covariate in co-kriging. Indeed, many people have used this. The major problems with co-kriging have been twofold. First, the parameters of all the $[(q+1)q]/2$ variograms and cross-variograms have to be estimated and the parameters have to obey a strict inequality (Wackernagel, 1987). Secondly, and more importantly, the co-kriging model really assumes linear relationships between the predictor and predicted variables. The use of categorical predictors and predicted variables is also difficult. Odeh et al.'s experience (Odeh et al., 1995) was that co-kriging did not perform as well as $S = f(r) + e$ (regression-kriging) and was more cumbersome to use. We prefer the *scorpan* model, but co-kriging should not be dismissed and the difficulties and restrictions will be overcome. It can be argued that co-kriging is a kind of GLM. Co-kriging can be used for soil attributes, and compositional (co-)kriging (De Gruijter et al., 1997; Walvoort and De Gruijter, 2001) can be used for indicators or probabilities of discrete soil classes or memberships of continuous ones.

37.3.4.2.3.1 Coregionalization Analysis Even if co-kriging is not done, coregionalization analysis (e.g., Lark and Papritz, 2003) is a very instructive way of studying the linear spatial relationships between soil and the predictor variables Q. This will indicate the spatial scales over which we might expect linear relationships to hold.

37.3.4.3 Other Spatial Methods

Bayesian maximum entropy (BME) was introduced by Christakos (1990, 2000). This approach allows the incorporation of a wide variety of hard and soft data in a spatial estimation context. The data sources may come in various forms, such as intervals of values, probability density functions (pdf), or physical laws (Christakos, 2000). Bogaert and D'Or (2002) used BME algorithm and a Monte Carlo procedure (BME/MC) to generate a map of particle-size distributions from a limited number of accurate measurements and a spatially exhaustive soil map. Compared with ordinary kriging (OK), this approach has the advantage of using soft information on a sound theoretical basis.

Brus et al. (2008) used BME to estimate the probabilities of occurrence of soil categories in the Netherlands. They used "hard" observations from 8,369 soil profile descriptions, and the soil map of the Netherlands 1:50,000 as "soft" information. They concluded that BME is a valuable method for spatial prediction and simulation of soil categories when the number of categories is rather small (say <10). For larger numbers of categories, the computational burden becomes prohibitive, and large samples are needed for calibration of the probability model.

37.3.5 Recent Studies

Although the *scorpan* model was formalized in 2003, various authors have fitted parts of it. McBratney et al. (2003) summarized the work of a large number of studies. Their analysis showed that soil attributes have been estimated more often (70% of studies) than soil classes (30% of studies). The key predictor factors are r (80% of studies) followed by s (35%), o and p (both 25%), n (20%), and c (5%) whereas a does not seem to have been used as a factor. The most common combination was r and s. Most studies used a DEM as the main source of ancillary data, followed by remotely sensed imagery and preexisting soil coverage.

Grunwald (2009) reviewed more recent studies in digital soil mapping during the period of 2007–2008. Of the 90 reviewed studies, 40% focused on predictions of basic soil properties such as texture, bulk density, structure, 31% on soil carbon, 24% on environmental quality assessment, 17% on hydrologic properties, 9% on soil degradation (salinity, acidity, and erosion), and 16% on mapping of soil taxonomic/ecological classes. In particular, mapping of soil organic carbon was prominently represented. In these recent studies, about 40% utilized proximal or remote sensors, and 23% of the studies used sensor data to complement analytical soil data, which are more costly and labor-intensive to derive. The sensors include visual and near-infrared, mid-infrared, and apparent electrical conductivity.

37.4 Sources of Data: The Seven Scorpan Factors

There are seven factors or sets of variables in the *scorpan* model, which makes it different from Jenny's model. The aim is to obtain information on all of these. It will be a matter of convenience (access to data sources) and scientific contention, which variables are used to represent the factors. Indeed, this is an area that has not been well enough studied. The creation of these digital maps of the input environmental variables representing the six factors in the *scorpan* model is seen as an integral part of the digital soil resource assessment approach and a very valuable, environmentally useful, by-product of the new approach. The layers can be used for other modeling purposes. Much of the earth science and ecological research of the last 20 years has been contributing toward the creation of these layers.

These digital surfaces themselves will be created using surface modeling procedures; regression-kriging or Laplacian smoothing splines or TINS. For D3 surveys, they should probably be produced on a 100 m raster with a block size of say 100 m × 100 m.

Figure 37.1 highlights the useful parts of the electromagnetic spectrum for obtaining information on soil and environmental variables through remote and proximal sensing. Matter emits electromagnetic radiation in different parts of the spectrum, and this radiation may be measured by different types of spectroscopy depending on the wavelength. This provides a basis for remote sensing of the properties of matter. A sensing system might measure the radiation emitted by an object after the object has itself been irradiated. Two examples of this are the optical remote sensing systems that measure the solar radiation reflected by an object and the synthetic aperture radar systems (SAR) that measure deliberately long-wave radiation backscattered by an object. Alternatively, it may be possible to measure radiation emitted by an object because of its temperature (emitted in thermal infrared frequencies) or because of radioactive decay (decay of uranium, thorium, and potassium isotopes is widely measured by "passive" gamma radiometry in geophysics). The electromagnetic radiation emitted from an object will therefore depend on its physicochemical properties, some of which are of direct interest in the study of soil (the temperature, mineralogy, organic content, physical structure, or the chlorophyll content of overlying vegetation).

37.4.1 "*s*" Factor

Remote and proximal active and passive sensing gives detailed information on the soil itself—these reflections or emissions or transmissions are intrinsic properties of the soil material and profile they may indicate other soil attributes like texture or mineralogy. This factor is likely to becoming increasingly important as technology advances.

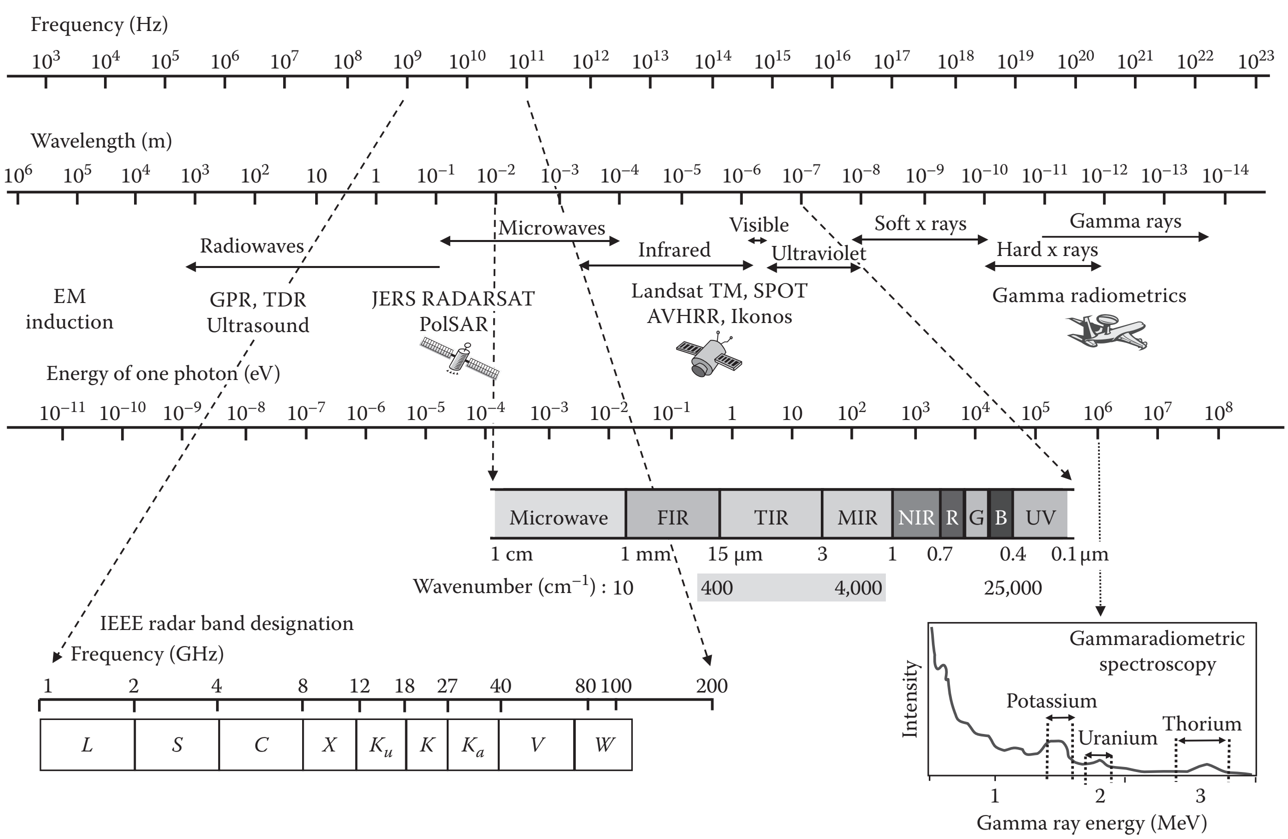

FIGURE 37.1 Electromagnetic spectrum showing various parts important for soil prediction.

37.4.1.1 Surface Multi- and Hyperreflectance

Hyperspectral sensors are those that measure a large amount of bands with spectral resolutions less than 20 nm (Palacios-Orueta and Ustin, 1998). In terms of mapping soil information, hyperspectral sensors have been found to be useful in mapping mineralogical features such as iron oxides (King et al., 1995), carbonates, and sulfates (Crowley, 1993). The amount of information generated by hyperspectral sensors poses computational problems in terms of extracting useful information in an efficient manner (see Section 37.3.3.1.2).

In a general way, the use of digital remotely sensing imagery for mapping soil has been problematic because vegetation cover obscures much of the soil response making it necessary to search for indirect evidence that may be visible at the surface (Campbell, 1987). Thus, remote sensing cannot be applied alone to soil studies (Lee et al., 1988). The use of proxies (such as topography, vegetation, drainage patterns) and field observations are important approaches for inferences about soil. In fact, mapping forest soil directly from remotely sensed data is difficult because of the complexity of environmental factors contributing to the spectral reflectance measured by a sensor. Nevertheless, forest soil was correctly mapped only where it was correlated with species or when vegetation is sparse or absent as a result of cultivation or drought (Post et al., 1994). The principal problems to soil delineation from such imagery, in addition to vegetation cover, are as follows:

- Soil moisture content can interfere with the spectral reflectance (specially in the infrared, thermal, and microwave regions (Obukhov and Orlov, 1964)
- Atmospheric effects (Cipra et al., 1980)
- Physical soil characteristics or disturbance patterns (Huette, 1988) and observation conditions (e.g., intensity and direction of illumination)

However, some research has been done in adjusting some of these drawbacks, by creating some indices with the main purpose of removing the effect of soil spectral influence (Huette, 1988) or even adjusting the images for vegetation interference, in combination with other information, and developing prediction models for improving soil mapping (Odeh and McBratney, 2000). In this way, Dobos et al. (2001) used satellite data complemented with DEM data, in order to correct the distortions caused by topographic variations of the landscape and provide additional data for soil–landscape modeling.

The presence of vegetation cover attenuates electromagnetic radiation at most wavelengths (Skidmore et al., 1997), and the reflectance at the soil surface does not always reflect soil variation at depth (Agbu et al., 1990). While infrared and visible sensors only measure surface characteristics, radar and gamma radiometry can provide spectral information beyond the vegetative cover and the soil surface.

37.4.1.2 Radar Attenuation

Information can be obtained by radar especially if there is a light-textured soil (low dielectric constant) over a heavier textured horizon (high dielectric) or a water table (very high dielectric), and radar sensors can penetrate through soil to a depth that is equal to 10%–25% of their wavelength (Lascano et al., 1998). The longest wavelength radar sensor available from space platforms is 23.5 cm (L-band) on the JERS-1 satellite.

So far, we have only considered empirical methods where the multivariate prediction methods have been used to incorporate the remotely sensed imagery into prediction models. In the case of SAR, the backscattering signal is largely dependent on the dielectric properties of the media that is reflecting the signal, in the case of soil this is the volumetric moisture content (Moran et al., 1997). Therefore, physical models based on the theory of the diffraction of electromagnetic waves have been developed, an example being the integral equation model (IEM) (Fung et al., 1992). The model calculates a backscattering coefficient that is based on the following:

- Radar sensor configuration, for example, observation frequency, polarization, and incidence angle
- Surface characteristics, for example, roughness and dielectric properties

The model was adapted successfully to predicting soil moisture from bare soil surfaces by Altese et al. (1996). The presence of vegetation complicates matters, and it is currently too difficult to create models describing the interaction between soil–vegetation layers and microwaves in real world applications. Therefore, in the presence of vegetation, empirical approaches are required (e.g., Dobson and Ulaby, 1986; Wood et al., 1993).

37.4.1.3 Electrical Conductivity

Soil bulk electrical conductivity (or its reciprocal soil electric resistivity) reflects a combination of soil mineralogy, salts, moisture, and texture; hence, it is a good compound measure of soil. Two commonly used kinds of instruments are electromagnetic induction (EMI) and electrical conductivity/resistivity based on rolling electrodes (ECRE). The most widely used instruments for soil studies are the EMI devices from Geonics in Canada and two types of ECRE devices, a U.S. design (Lund et al., 1999) and a French one (Tabbagh et al., 2000). These instruments have been used extensively in precision agriculture for mapping soil types and properties (such as Bishop and McBratney, 2001; Sudduth et al., 2001; Anderson-Cook et al., 2002). Such proximal sensing offers the possibility of producing high-resolution maps of soil properties (D1 surveys of Table 37.1). Regression equations have been developed to predict moisture content, topsoil thickness, and clay content. EMI instruments can be placed in airborne platforms for catchment and regional mapping.

37.4.1.4 Gamma Radiometrics

Gamma-ray spectrometry (GRS) provides a direct measurement of natural gamma radiation from the top 30–45 cm of the soil (Bierwirth, 1996). A gamma-ray spectrometer is designed to detect the gamma rays associated with radioactive elements and to accurately sort the detected gamma rays by the respective energies (Grasty et al., 1991). Airborne radiometrics survey measures the radiation naturally emitted from the earth surface, using gamma emitters like ^{40}K and daughter radionuclides of ^{238}U and ^{232}Th. K is a major constituent of most rocks and is the predominant alteration element in most mineral deposits. Uranium and thorium are present in trace amounts, as mobile and immobile elements, respectively. As the concentration of these radioelements varies between different rock types, we can use the information provided by a gamma-ray spectrometer to map rocks. Airborne methods (air gamma-ray spectrometer [AGRS]) provide valuable, systematic coverage of large areas, by providing information about the distribution of K, U, and Th that is directly interpretable in terms of surface geology. Nevertheless, AGRS is a surface technique only—interpretation requires an understanding of the nature of the surficial materials and their relationship to bedrock geology.

Although this technique has been employed for geological and mineral resource mapping for over 20 years, it has just become an interesting tool in soil science for detecting spatial variation of soil-forming materials (the *p* factor). It can also be considered as a direct, albeit compound, measure of the mineralogical and textural composition of the soil itself (*s*). Gamma-radiation data are usually provided in three channels—corresponding to spectral windows for K, U, and Th radiation. The apparent K concentration is likely to be most easily interpreted by pedologists. The value of gamma-radiometric data is increasing with the knowledge of their relation with soil-forming materials and when considered jointly with other information such as terrain models or aerial photography (Cook et al., 1996b) has become an important source of data for digital soil mapping. This technique has also been applied to estimate variation in surface soil moisture content (Carrol, 1981).

Rawlins et al. (2009) used airborne radiometric data and digital elevation data as covariates to map SOC distribution from an intensive survey in Northern Ireland. Radiometric data (K-band) and, to a lesser extent, altitude are shown to increase the precision of SOC predictions when they are included in linear mixed models of SOC variation. Division of the soil in Northern Ireland into three classes (mineral, organo-mineral, and peat) leads to a further increase in the precision of SOC predictions.

37.4.1.5 Existing Soil Class or Property Maps or Expert Knowledge

An existing soil map for parts of an area can be used to build a prediction model, or an experienced surveyor's expertise can be used to make simple rules that can be applied to a DEM, etc. These soil layers should be used as part of the information to build the new model—they should not really be the model itself. This is arguable, and there is further discussion in Section 37.5.12.

37.4.2 "*c*" Factor

Climate could be represented by mean annual temperature (*T*) and mean annual rainfall (*P*) and perhaps some measure of potential evapotranspiration (*E*). In the old literature, *P/E* was used to separate pedalfers from pedocals. A $P/E \ll 1$ would imply semiarid conditions and precipitation of carbonates at depth.

Climate surfaces can be produced from meteorological stations interpolated by Laplacian smoothing splines (Hutchinson, 1998a, 1998b). This has been implemented in a program called ANUCLIM. The climate variables used are monthly mean values for minimum temperature, maximum temperature, precipitation, solar radiation, evaporation, and others. The climate surfaces can be used to generate secondary information, for example, bioclimatic parameters such as mean temperature of warmest period, precipitation of driest quarter, etc., which are useful in determining the climatic envelope for plant and animal species.

Published work suggests that remote sensing analysis can be used for estimating and mapping air temperature, soil moisture, and atmospheric humidity at regional to global scales. Air temperature can be inferred from normalized difference vegetation index (NDVI) data from the NOAA advanced very high-resolution radiometer (AVHRR). Several studies have shown that the surface albedo can be estimated using remote sensing data (i.e., Brest and Goward, 1987), and that net radiation can be calculated with sufficient accuracy (Kustas and Norman, 1996; Boegh et al., 2002).

37.4.2.1 Temperature

Surface temperature can be derived from remote sensing such as AVHRR, geostationary orbiting earth satellite (GOES) (Diak et al., 1998), and TIROS operational vertical sounder (TOVS) (Susskind et al., 1997). The TOVS has two sensors: the high-resolution infrared sounder and the brightness temperatures of the microwave sounding unit. These data can provide estimates of daily air temperature, humidity profiles, and surface temperature (Susskind et al., 1997).

Goetz et al. (1995) compared surface temperature derived from a multispectral radiometer (MMR) mounted on a helicopter (resolution ~5 m pixel), a C-130-mounted thematic mapper simulator (TMS) (~20 m pixel), and the Landsat 5 thematic mapper (120 m pixel). Differences between atmospherically corrected radiative temperatures and near-surface measurements ranged from less than 1°C to more than 8°C. Corrected temperatures from helicopter-MMR and TMS were in general agreement with near-surface infrared radiative, thermometer measurements collected from automated meteorological stations while the Landsat 5 TM systematically overestimated surface temperature.

37.4.2.2 Precipitation

Spatially distributed precipitation estimates can be derived from rainfall gauge measurements (interpolated using splines or other techniques) or by remote sensing. Records of gauge measurements of monthly precipitation are available throughout the entire twentieth century, while satellite estimates can provide monthly to hourly resolution since 1974. A review has been presented by New et al. (2001).

37.4.2.3 Evapotranspiration

Li and Lyons (2002) estimated regional evapotranspiration in central Australia, using limited routine meteorological data and the AVHRR data. Their model attempts to minimize the difference between model-predicted surface temperature and satellite-derived temperature to adjust the estimated soil moisture. They suggested that radiometric surface temperature can be used to adjust simple water balance estimates of soil moisture providing a simple and effective means of estimating large-scale evapotranspiration in remote arid regions.

Boegh et al. (2002) used Landsat-TM data to estimate a composite evaluation of atmospheric resistance, surface resistance, and evapotranspiration. The input parameters were surface temperature, net radiation, soil heat flux, air temperature, and air humidity. The application of the technique in a remote sensing monitoring context was demonstrated for a Danish agricultural landscape containing crops at different stages of development.

37.4.2.4 Water Balance Components

The familiar water balance formulation is

$$P + I = \Delta S + E + T + R + D, \tag{37.20}$$

where
- P represents precipitation
- I represents irrigation
- ΔS represents change in soil moisture
- E represents evaporation
- T represents transpiration
- R represents surface runoff
- D represents deep drainage

Precipitation and evapotranspiration can be estimated from remote sensing data. The most important component relating to soil itself is soil moisture, and much research has been focused on estimating spatially distributed soil moisture (see also Section 37.4.1.2). Jackson et al. (1996) gave an overview of remote sensing techniques for estimating soil moisture. Many studies have successfully demonstrated the use of infrared, passive, and active microwave sensors to estimate soil moisture (Hoeben and Troch, 2000). Microwave remote sensing of soil moisture is based on the soil's dielectric properties. The large difference between the dielectric properties of dry soil and moisture enables good calibration. The analysis is based on a model that simulates radar backscattering given known surface characteristics such as moisture and roughness. Passive microwave sensors have the advantage of less dependence on soil surface roughness. The main disadvantage with spaceborne sensors is that they produce coarse-resolution images. This problem is overcome in active microwave sensing through the use of SAR sensors (10–100 m). This has been used to monitor spatial and temporal soil moisture at catchment scale (10–1000 km^2) both in vegetated and nonvegetated areas (Lin et al., 1994; Su et al., 1997). Mancini et al. (1999) evaluated the use of multifrequency radar observations in the laboratory for estimating soil moisture.

Murphy et al. (2009) used a fine resolution digital elevation model to model soil moisture conditions in Alberta, Canada. They compared the compound topographic (wetness) index (CTI) with a new algorithm that produces a cartographic depth-to-water (DTW) index based on distance to surface water and slope. They found that their DTW model was closer to field-mapped conditions. All major wet areas and flow connectivity were reproduced and a threshold value of 1.5 m DTW accounted for 71% of the observed wet areas.

The soil moisture and ocean salinity (SMOS) satellite mission was launched in November 2009. It operates under the L-band (1.4 Ghz) and was designed to observe soil moisture over the Earth's landmasses and salinity over the oceans. A global soil moisture map with accuracy better than 0.04 m^3/m^3 can be produced every 3 days; however, the spatial resolution is very coarse (50 km or better).

37.4.3 *"o"* Factor

The "organisms" factor is a basic term encompassing biota composition or the flora and the fauna of specific region. Flora stands for vegetation description whereas the fauna means the animals, microorganisms, and even human activities. In pristine or newly developed environments, the "natural" vegetation class should represent some kind of equilibrium relation with soil type. In that case, the vegetation is an indicator of the underlying soil type. However, "natural" vegetation is quite uncommon to find since it has frequently been disturbed by human activities (generally called "land uses"). Vegetation and land uses, generally provided by land cover analysis, can then be either an indicator of the underlying soil type or a determinant of the soil type. After defining land use and vegetation, their source of data, their relationships with soil, as determinant or indicator, can be explained.

37.4.3.1 Land Use Information and Relationships with Soil

Three kinds of land uses are generally distinguished: settlements (cities, mines, industrial areas, etc.), agricultural activities (arable lands, pastures, plantation, etc.), and natural or seminatural areas (natural forests, natural grasslands, bare rocks, and water). This information is generally provided by remote sensing images providing access to land cover. While settlements are usually easy to distinguish from above, it is often difficult to punctually differentiate agricultural pastures from natural grasslands, bare soils of arable lands from bare rocks, and plantations, or harvested areas, from natural forests. To this end, historical ground inventories, long-term imageries, or time-series analysis of remotely sensed images should be included in the mapping process of land uses. In that sense, land use information can be crucial for assessing the "*a*" (or time/age) factors. Schulp and Veldkamp (2008) made an inventory of historical land use. They showed that coarse resolution (500 m) historical patterns can explain up to 75% of the soil organic matter variability, whereas actual land use only explains 2% of the variability. Moreover, the IPCC report (2006) provided increase or decrease coefficients of soil organic carbon stocks according to land use or land cover conversions. This information is then crucial at the global level.

37.4.3.2 Vegetation Definitions, Data, and Relationships with Soil

Vegetation information is one of the most used "scorpan" factors along with landform. It is usually obtained from intensive ground survey and/or from visible and infrared reflectance by remote sensing being enhanced more recently by microwave imagery (Clevers and van Leeuwen, 1996). It comprises biomass deriving from agricultural activities and from natural or semi-natural areas. Three kinds of components should be described: the type or species of vegetation, the intensity of vegetation or biomass quantity, and the area covered by vegetation usually called "vegetation coverage."

37.4.3.2.1 Vegetation Type or Vegetation Land Cover

The vegetation type is usually assessed by doing a classification of remote sensing multispectral imagery on the spectral signatures of the image pixels (Figure 37.2). Vegetation is mainly discriminated by wavelengths at 0.55 μm in the green region, at 0.67 μm in the red region, and at 0.87 μm in the NIR region (Bunnik, 1978). Classification can be supervised using ground measurements or unsupervised when applied directly on the pixels. For tree species, and for very fine resolution, when there is a priori knowledge of the species, it is possible to use airborne laser scanning (Lidar) for differentiating the trees according to their morphological attributes such as tree height, stem diameter, and crown radius (Ørka et al., 2009). Carré and Girard (2002) used information on spectral bands for assessing the spatial distribution of soil types using linear regression. Hansen et al. (2009) used multispectral data and topography for assessing soil–landscape units.

37.4.3.2.2 Vegetation Intensity or Biomass Quantity

According to vegetation species and its state of development, chlorophyll absorbs more or less solar energy. The absorbed or transformed energy that represents biomass production is usually modeled by combining different visible and near-infrared bands. The produced indices are then not only indicating vegetation types but also primary production. These indices are NDVI (Rouse et al., 1973), EVI (Waring et al., 2006), and FAPAR (Gobron et al., 2000).

Furthermore, concerning crops, the use of yield monitors on harvesting machines also provides a source of spatial biomass information (Stafford et al., 1996). Bishop and McBratney (2001) used yield-monitored wheat yield to aid in the prediction of soil clay content. Yield-monitored data are currently useful for D1 mapping (Table 37.1).

37.4.3.2.3 Vegetation Coverage or Area Covered by Vegetation

The area covered by vegetation, whatever the vegetation type, can be assessed using the indices explained before. This allows for usually discriminating bare soil from lands covered by vegetation and thus assess potential risks related to soil vulnerability like wind erosion (Skidmore, 2006). Many publications describe the role of vegetation on different soil processes, on the spatial distribution of soil at different scale, and for different time periods (months, year, and decade). It is then important to take this factor into account for digital soil mapping.

37.4.4 "*r*" Factor

Topography has long been recognized as one of the main soil-forming factors (Jenny, 1941). Milne (1935) introduced the concept of a catena to describe a predictable and recurring pattern of soils that occupied a topographic sequence along a hillslope from divide to channel to divide. Aandahl (1948) was the first to quantitatively relate landscape attributes to soil properties. Troeh (1964) fit a cylindrical parabola to contour lines to derive

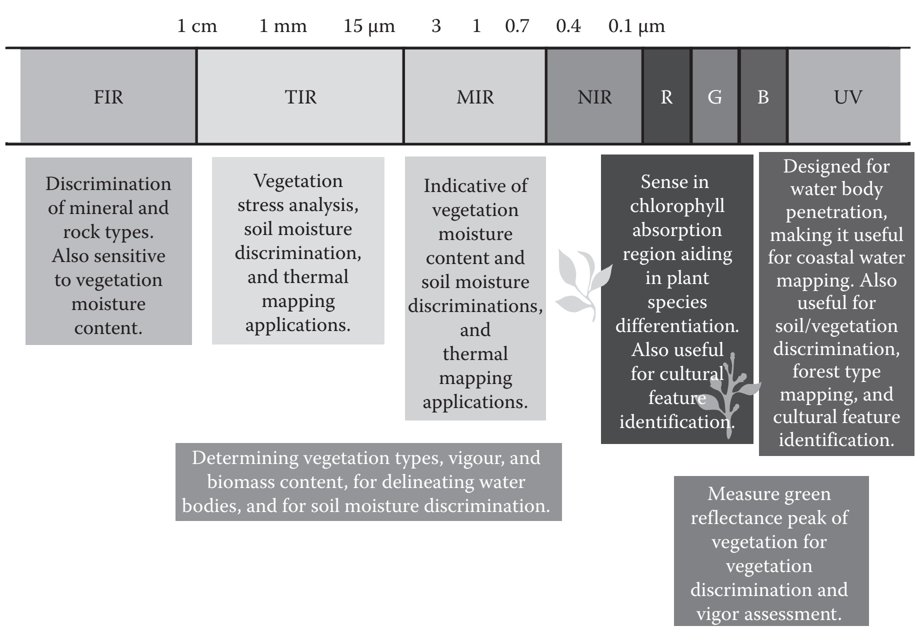

FIGURE 37.2 Ultraviolet-infrared spectrum showing important parts for sensing vegetation indices.

slope and curvatures. These landform parameters were derived to correlate to soil drainage classes.

Almost all DSM studies reflect a recognition that changes in the value of soil properties or the classification of soils occur in response to environmental gradients and that many environmental gradients of interest are strongly related to elevation, moisture, and energy gradients that are defined and controlled by local and regional topography (Milne, 1935; Burrough et al., 2000; McKenzie et al., 2000).

The principal assumption underlying the soil–landform paradigm is that the distribution of soils in the landscape is predictable and that, at least at a local scale, this distribution is largely controlled by how topography influences the distribution and redistribution of moisture, energy, and materials in the landscape. Terrain derivatives computed from digital elevation data can provide useful measures for predicting and quantifying how topographic controls influence this distribution and redistribution. They therefore represent a primary source of explanatory variables for most efforts to implement automated predictive soil mapping. Many would argue that digital terrain modeling is the most useful and quantitatively developed factor for predicting soil attributes and soil classes, at least at a local (hillslope) scale (McKenzie et al., 2000).

Topography has increasingly come to be described and quantified by means of digital elevation models (DEMs). Initially, DEMs were produced by digitizing existing contour maps or by extracting spot elevations and stream-line data from traditional land surveys and then interpolating these contour or spot elevations to a regular grid (or using them to create a triangular irregular network or TIN). More recently, DEMs have been produced more rapidly, cheaply, and at a finer spatial resolution by means of automated stereo autocorrelation of stereo image pairs obtained from airborne or satellite sensors, by interpolation of dense ground point observations obtained from vehicle-mounted high-resolution GPS receivers, or from very fine resolution clouds of point elevation data obtained using airborne lasers (LiDAR) or synthetic aperture radar (IfSAR or InSAR). See Nelson et al. (2009) for a thorough review of DEM production methods and sources.

The use of elevation data in digital form involves quantitative parameterization of the surface model or the numerical description of continuous surface form (Pike, 1988; Wood, 1996). Pike et al. (2009) observed that geomorphometry is fundamentally concerned with the quantitative measurement and extraction of both land surface parameters (LSPs) and land surface objects (LSOs). An LSP is a descriptive measure of a surface attribute (e.g., slope, aspect, and wetness index), usually arrayed as a continuous field of values and usually organized as a raster image or map. An LSO is a discrete surface feature (e.g., a watershed line, cirque, alluvial fan, and drainage network) that is usually represented as a vector object defined by points, lines, or polygons (see also Deng, 2007).

Parameterization means to quantitatively measure continuous properties of a landscape that can be used to describe form (Wood, 1996). Geomorphometry involves quantification of form but can also include consideration of different measures of pattern, texture, and hydrological, gravitational, or spatial context, as well as the extraction of geomorphologic objects. A wide range of LSPs can be extracted from a DEM, such as altitude, slope, aspect, different curvatures, upslope contributing area, and compound topographic index (CTI). Similarly, DEMs can be processed to identify and delineate a wide range of discrete LSOs.

For discussion purposes, LSPs have frequently been grouped into primary and secondary topographic attributes (Moore et al., 1993; Wilson and Gallant, 2000). Primary attributes are measures of a single aspect of the land surface whose calculation does not require combination of two or more separately computed values. Primary attributes are often further differentiated into local attributes, such as slope, which are computed within a neighborhood window of fixed (but now increasingly of variable) dimensions and regional attributes, such as contributing area, whose calculation extent is not predetermined or fixed but varies according to conditions imposed by a particular set of land surface data. Secondary terrain attributes are any attributes (such as wetness index) whose values are computed using a combination of two or more primary attributes (Wilson and Gallant, 2000).

37.4.4.1 Primary Terrain Attributes

Primary attributes have been used successfully in numerous studies to predict different soil attributes and classes. Maintaining the above distinction of local versus regional primary attributes, we observe that local attributes mostly describe the height, orientation, and shape of the land surface at a point or, more correctly, within a region defined by a rectangular (or sometimes circular) window surrounding a point. The main primary local attributes are elevation, slope, aspect, and various measures of curvature.

Elevation in its raw form mainly influences regional and local climate through its effect on surface temperature and precipitation (Deng, 2007). Elevation has also been used in numerous DSM studies as a local indictor of relative slope position within restricted areas (Shi et al., 2004). Slope and aspect describe the orientation of the local land surface within a neighborhood window. Curvatures primarily describe the shape of the local land surface within the window.

Shary et al. (2002) highlighted the importance of recognizing that many terrain attributes are field specific, insofar as they are computed with reference to a particular vector field, such as the gravitational field. Slope, aspect, profile and plan curvature are all computed relative to the orientation of the normal to the gravity vector. When interpreting these measures, the implicit assumption is that the current orientation of the land surface reflects the conditions that existed over the period of time that a soil was formed. For example, a closed depression (nondraining) can change to an open depression (draining) if the land surface tilts at some point in time after the formation of the closed depression.

Evans (1972, 1998) provided an overview of primary terrain attributes in relation to their geomorphologic meaning. Slope is frequently interpreted in terms of the amount of energy associated with a particular slope gradient that is available to drive soil-forming processes such as runoff, erosion, and deposition. Aspect, in combination with slope, is often interpreted in terms of exposure to incoming solar radiation, drying prevailing winds, or accumulation of snow or sediments in sheltered lee slopes. Curvature is generally interpreted in terms of how local shape affects surface runoff, erosion, and deposition. Areas that are convex in profile or plan are considered likely to shed water, to cause divergence of surface flow (in plan) and acceleration of surface flow (in profile), and to consequently be drier than normal sites. Areas that are concave in plan (across slope) are considered to favor convergence of surface flow and development of moister conditions. Areas that are concave in profile (downslope) are considered to favor deceleration of surface flow and deposition or accumulation of both moisture and sediments. Curvature also has often been interpreted as an indication of relative position on a hillslope (see Pennock et al., 1987). Areas that are convex in profile were interpreted as indicative of crests and shoulders while areas concave in profile were interpreted as indicating decelerating positions in toe slopes. Areas that are planar in profile have been associated with back slopes.

The majority of DSM studies appear to have preferred to make use of the most commonly computed local terrain attributes slope, aspect, profile curvature, and plan curvature computed using the most commonly available algorithms. Shary et al. (2002) described instabilities in the calculation of plan curvature and suggested that it not be used and instead replaced with its close equivalent of horizontal curvature. Many subsequent studies adopted the use of either horizontal or tangential curvature in preference to plan curvature to characterize curvature in the across slope direction (Schmidt et al., 2003; Schmidt and Hewitt, 2004).

Local terrain attributes are typically taken to provide an indication of the shape or orientation of the land surface only, that is, its form. However, the increasing trend to compute these variables over a range of grid resolutions and window dimensions produces information that can be interpreted to quantify characteristics of the land surface that describe texture, pattern, scale, and context. The magnitude and frequency of variation in these local terrain attributes within expanding neighborhood windows can provide useful measures of pattern and texture in the landscape. It can also help to establish the characteristic distances over which variation in these properties occurs. This can be interpreted as a measure of the size and scale of the landscape. Finally, multiple measures of slope, aspect, and particularly curvatures within expanding calculation windows can be interpreted in terms of the position or context of a grid cell in the landscape (e.g., a ridge in a pit located on a larger ridge) (Wood, 1996).

Many primary regional topographic attributes provide information about some aspect of topographic or hydrological position or context. The most commonly used primary regional attribute is upslope contributing area or drainage area also rescaled to compute specific catchment area (SCA). Calculation of this attribute involves tracing down flow directions and accumulating a sum for the area upslope of every cell that contributes drainage to each downslope cell. This calculation can involve using a single steepest flow direction algorithm (the so-called D8 algorithm) to assign all flow from each cell into its steepest downslope neighbor, or it can utilize one of several different algorithms that partition flow from each cell according to some estimate of the proportion of flow that might logically be expected to pass from a given cell into each of its neighbors of equal or lower elevation. This attribute provides a measure of relative position in the landscape within the context of simulated hydrological flow of surface water.

Many primary regional attributes represent efforts to capture the relative position in the landscape of a point or the amount of relief (and therefore potential energy) associated with the area upslope (or downslope) of a point with respect to some other significant reference point, such as a channel or a ridge. Measures of relative landform position have been reported that are based on calculations of the variation in elevation (Pike, 1988) or mean slope (Fels and Matson, 1996) within fixed windows of various dimensions. Wilson and Gallant (2000) identified a number of measures of variation in elevation within circular windows centered at a grid cell.

Skidmore (1990) introduced one of the first algorithms for computing a measure of relative slope position. It involved first classifying cells as either ridges (convex) or channels (concave), based on their local surface shape. An expanding search window centered on each grid cell was then used to determine the Euclidian distance from each cell to the closest ridge and channel cells. This approach has the potential disadvantage that the closest channel or ridge cell may not be the one to which a given cell is connected by a path of surface water flow (see Rennó et al., 2008). MacMillan et al. (2000a, 2000b) implemented an algorithm for computing a variety of different measures of absolute and relative landform position and relief that was based on tracing both upslope and downslope along simulated paths of surface flow to locate and measure the horizontal, flow length, and vertical distances to critical cells designated as belonging to pits, peaks, channels, or divides. This approach has been proposed or implemented by others as well (Bell et al., 1992; McSweeny et al., 1994; Bui and Moran, 2001). Interesting and useful alternate measures of relative position in the landscape have been described in terms of valley bottom index (Gallant and Dowling, 2002) and openness (Yokoyama et al., 2002). Relative landform position tends to be a key conceptual consideration in many human-devised systems of soil or landform mapping, and measures of landform position and context have increasingly come to be used as explanatory variables in digital soil mapping (Bell et al., 1992; Bui and Moran, 2001, 2003; Schmidt and Hewitt, 2004; Schmidt and Andrew, 2005; Schmidt et al., 2005).

See Wilson and Gallant (2000) for a review of primary and secondary topographic attributes including various measures of slope and area computed for cells upslope of a central cell that contribute flow into it (contributing area) and cells downslope of a given cell into which it contributes flow (dispersal area).

37.4.4.2 Secondary Terrain Attributes

Secondary terrain attributes are computed from the primary attributes. These have been described in detail by Wilson and Gallant (2000). These attributes usually combine two or more primary attributes to characterize the spatial variability of specific processes in the landscape. The most widely used is CTI or also called wetness index:

$$\mathrm{CTI} = \ln\left(\frac{A_s}{\tan\beta}\right), \tag{37.21}$$

where
A_s is the upslope area
β is the slope

Wilson and Gallant (2000) also provide routines for the calculation of erosion, solar radiation, and dynamic wetness indices.

Wilson and Gallant (2000) list and document a number of secondary topographic indices including three different versions of the wetness index or CTI, a stream power index (SPI), a sediment transport capacity (STC) index, a channel initiation (CI) index, and several radiation and temperature indices. All of these can prove useful in efforts to automatically classify landform-based terrain elements. The secondary topographic attribute that is most widely reported as being used in efforts to automatically classify landform entities is the wetness index or CTI (Beven and Kirkby, 1979; Moore et al., 1991a, 1991b, 1993; Wilson and Gallant, 2000).

37.4.4.3 Terrain or Landscape Classification

In contrast to the preceding discussion of continuous LSPs, landform classification locates and extracts identifiable LSOs from digital elevation data. These objects occur across a range of scales, and their recognition can involve varying degrees of subjectivity.

Within the context of digital soil mapping, there have been three main areas of application of landform classification. The first has been the identification of critical points, or surface-specific points (Peucker and Douglas, 1975), based largely on local surface shape. The second has been the identification and classification of areas of relatively homogeneous terrain conditions, referred to as landform elements, which are usually identified as components of a hillslope along a sequence from crest to channel. These landform elements are usually described as being occupied by a single soil series or a narrow range of similar soil series. The third application has involved recognition of regional landform patterns, which are typically described as being occupied by repeating combinations of different soils, such as soil associations or complexes. Each of these three major approaches is reviewed in brief as follows.

Recognition of critical points, or surface-specific points (Peucker and Douglas, 1975), based primarily on local surface shape has been a key component of automated landform classification since its inception. Pike et al. (2009) cite Cayley (1859) as having been the first to identify and describe the conceptual importance of these points in describing landforms. Critical points and lines identify local maxima and minima in the landscape and help to establish horizontal and vertical scale and context for landform classification. Their identification is relatively objective and crisp but some ambiguity occurs when local surface shape is not pronounced enough to identify pits, peaks, channels, or ridges definitively. Ambiguity also exists with respect to the scale at which these objects are identified. Wood (1996) showed how extraction of these critical objects within multiple windows of expanding dimensions could result in classification of the same point as, for example, a pit within a peak within a pit. This multiresolution analysis approach has been increasingly recognized as a valuable way to assess context and form across multiple scales (Deng, 2007).

The surface-specific points and lines described above delineate a mesh, or framework, of hillslope entities bounded by ridge lines and course lines. The intersection of ridge lines and course lines delineates areas that represent individual hillslopes that run from local divides down to local channels. Pike et al. (2009) cite Maxwell (1870) as having been the first to observe that the area enclosed by these lines was part of both a *hill*, whose lines of slope run down from the same summit and a *dale*, whose slope lines run down to the same pit. These semantics emphasize the relationship of these points and lines to patterns of local surface water flow. Not surprisingly then, essentially the same points and lines can be extracted by computing networks of simulated surface water flow and extracting pits, peaks, channels and divides from the flow networks. This has been done by Band (1989a, 1989b, 1989c) who used pixel to pixel hydrological connectivity to define fundamental terrain units based on the intersection of complementary divide and channel networks. Hillslopes were defined as the areas between a channel segment and its associated drainage divide.

Identification of these pits, peaks, channels, and ridges plays a significant role in other aspects of landform analysis and classification. First, several useful LSPs, such as horizontal distance or change in elevation to a ridge or channel, need to first identify these significant points and lines in order to be computed. Second, these critical points and lines establish the context and boundaries within which subsequent efforts to identify and classify landform elements operate. Third, many of the reported approaches for classifying landform patterns make use of information on the scale and spacing of ridges and channels to establish measures of context and pattern used in classifying landform patterns.

A landform element is a subcomponent of a landform pattern. Landform elements may be conceptualized as consisting of portions of a landform pattern that are relatively homogeneous with respect to shape (profile and tangential curvature), steepness (gradient), orientation or exposure (aspect or solar radiation), moisture regime, and relative landform position (e.g., upper, mid, or lower). Speight (1990) recognized more than 70 types of landform elements and gave as examples cliffs, foot slopes, and valley flats. Dikau (1990) recognized simpler form units, or form facets, nested within form patterns, that were described by inherent shape-based attributes such as curvature, gradient, and aspect. Dikau (1989) differentiated form elements with homogeneous plan and profile curvature from even more homogeneous form facets that had uniform gradient, aspect, and curvature.

The majority of applications of landform classification in digital soil mapping have represented efforts to identify and extract landform elements arranged in a topographic sequence along hillslopes from crest to channel (Park et al., 2001). These classifications were usually intended to build upon conceptual classifications of hillslopes into summits, shoulders, back slopes, foot slopes, and toe slopes as proposed by Ruhe (1960) or the 9 unit classification of hillslopes proposed by Dalrymple et al. (1968) and Conacher and Dalrymple (1977) or the 10 types of topographic landform positions described by Speight (1990).

Classification of landform patterns is generally more challenging than classification of landform elements along a single hillslope. Not surprisingly then, there have been fewer approaches described for automatically classifying landform patterns and fewer applications of landform pattern classification in digital soil mapping. Classification of landform patterns has mainly been used to establish domains, or regions, over which a particular recurring pattern of soils and terrain (e.g., a soil association) is observed to occur and within which a particular set of classification rules for a limited number of soil classes is thought likely to apply (Dobos et al., 2000, 2001; Bui and Moran, 2001, 2003; Henderson et al., 2005; Bui et al., 2006; MacMillan et al., 2007).

The de facto standard for classifying landform patterns is Dikau et al.'s (1991, 1995) computerized implementation of the manual system of landform classification proposed by Hammond. The method of Dikau et al. (1991, 1995) follows Hammond (1954, 1964) in recognizing, within a window of fixed dimensions, 4 classes of proportion of gentle slopes, 6 classes of relative relief, and 4 classes of profile type, which, when combined, lead to 22 possible landform subclasses (Brabyn, 1998). These are commonly regrouped into 24 landform classes and 5 major landform types of Plains, Tablelands, Plains with Hills or Mountains, Open Hills and Mountains, and Hills and Mountains (Bayramin, 2000).

Brabyn (1997, 1998) noted two problems with implementation of Dikau's methods. The first was a pattern of progressive zonation that developed in transition zones between areas of high and low relief. The second was that the methods classified areas with quite different macro landform into the same class. Guzzetti and Reichenbach (1994) also identified concerns with imprecise boundary locations and mixed classification and tried to address them by developing and applying a different approach that combined automated classification of DEM data with manual interpretation and digitizing of clear, sharp final boundaries.

There have been relatively few other approaches reported for classifying landform patterns from digital elevation data. Bui and Moran (2003) used the computed values of total distance from ridge to channel to produce a map of relief classes (<9, 9–30, 30–90, 90–300, and >300 m) to define physiographic regions that exhibited signature soil patterns. Similarly, MacMillan et al. (2007) used the total relief computed for individual hillslopes to define classification domains of different relief within which fuzzy rules were used to classify ecological site types along toposequences from crest to channel.

Iwahashi and Pike (2007) proposed an alternative approach for classifying landform types at multiple scales that is very simple, efficient, and effective. This approach uses only three inputs of slope gradient, local convexity, and surface texture (feature frequency and spacing). The method is locally adaptive as it does not use fixed threshold values to distinguish, for example, steep from gentle slopes or convex from concave areas. Instead, the method uses a nested-means approach in which the mean value within a window of 10 × 10 cells is used as a threshold to differentiate steep from gentle or convex from concave. The method produces 8, 12, or 16 classes depending upon whether the gentle slope class is further subdivided once (into two subclasses) or twice (into four subclasses). This method delineates undefined classes whose local meaning and description must be assigned by a postclassification review and interpretation. Examples presented for DEM grids of 1 km, 270 m, and 55 m illustrated that the method produced meaningful and interpretable results across a range of scales.

In landform classification, as in conventional image analysis, there has been a recent trend toward investigating and adopting methods that involve image segmentation, or object recognition, that classify the object rather than classifying an individual pixel. A multiscale image segmentation approach is advocated by Burnett and Blaschke (2003) and Drăgut and Blaschke (2006) because "a multivariate analysis of pixels does not include topological relationships of neighborhood, embeddedness, or shape information."

In contrast to the bottom-up agglomerative approach of per-pixel classifications, such top-down divisive approaches identify significant discontinuities or boundaries at locations of maximum change in attributes. These boundaries define the spatial extent of objects whose attributes can then be determined and used to classify the entire object. The object's attributes can include context, shape, and topological information, such as adjacency or connectivity in addition to typical measures such as slope, aspect, or curvatures.

In conclusion, it is useful to reflect upon the observation of Dehn et al. (2001) that landforms are described mainly in two different ways (i) based solely on their geometry or (ii) based on semantics used to express and capture subjective conceptual mental models. Deng (2007) also differentiated five classes of landform entities according to the degree of vagueness or subjectivity involved in their definition and recognition. Landform entities that were discrete, clearly identifiable, and objective were classified as bona fide objects or prototypical objects. Landform entities whose definition and recognition involved a degree of vagueness and subjectivity were classified as semantic or fiat objects, landform classes, or multiscale objects.

37.4.5 "*p*" Factor

Acquisition of suitable information about parent material for use in digital soil mapping often presents a challenge. The attributes of parent material that are most often of interest are depth or thickness (deep or shallow), texture (e.g., coarse, medium, or fine), lithology or mineralogy (e.g., base status), age, and mode of deposition.

At present, the most common source of information about parent material for use in soil mapping studies has come from digitized versions of published geological or soil maps. Among geological maps, those that focus on lithology, rather than age or stratigraphy, tend to be more useful for soil prediction. Reinterpretation or recompilation of existing geological maps is often necessary to emphasize lithology in place of the original stratigraphic focus (Lawley and Smith, 2007).

Given that existing geological maps are often unavailable or inadequate, many studies have sought to identify alternative approaches for obtaining or inferring the main parent material attributes of interest. Of these, interpretation of airborne gamma-radiometric data in terms of regolith properties seems to have generated the greatest interest and success (Cook et al., 1996; Wilford et al., 1997, 2001; Pickup and Marks, 2000, 2001; Hardy, 2004; Rawlins et al., 2009). Radiometric data highlight differences in the parent material that can be interpreted in terms of depth to bedrock, mineralogy and lithology, age and mode of deposition. Clear distinctions have been made between transported and in situ regolith and between rocks with varying degrees of weathering. Radiometrics can provide an indication of the degree of erosional and depositional activity, which allows actively eroding surfaces with shallow regolith to be separated from stable surfaces with deeper regolith. Differences in mineralogy show up as different colors and patterns in ternary radiometric images, and these patterns can identify the spatial extent of areas of different lithologies (e.g., base rich versus acidic) and mineralogies (potassium rich granite versus depleted quartz sands). Degree of weathering, related to geomorphic age and amount of transportation, can often be interpreted from radiometric data, with highly weathered materials emitting low responses and recently exposed or weathered materials emitting high levels of gamma radiation. Such distinctions can be used to separate, for example, recent eolian deposits from old, well-sorted and intensely weathered fluvial deposits (Bierwirth and Brodie, 2005). Typically, radiometric data are most effective when interpreted in combination with additional data sources, particularly terrain data as extracted from a DEM (Cook et al., 1996b; Wilford et al., 2001).

Radiometric data are not readily available for many areas. In such cases, researchers have investigated a number of different options for inferring specific attributes of the parent material by interpretation of airborne and satellite imagery of various types and resolutions. The use of imagery for inferring attributes of parent material typically focuses on identifying specific parent material characteristics that are associated with distinct and highly contrasting patterns in the imagery. For example, areas of salinity, affected by gypsic or calcic surface materials, often contrast sharply with surrounding unaffected areas and can be isolated using imagery alone or using a combination of imagery and digital elevation data (Fernández-Buces et al., 2006; Nield et al., 2007; Rodriguez et al., 2007; Boettinger et al., 2008). Similarly, areas of organic soils (Connolly et al., 2007), wetlands (Peng et al., 2003; Landmann et al., 2006; Wright and Gallant, 2007), and African dambos (Hansen et al., 2009) often contrast sharply with their surroundings and can be isolated using multidate, multispectral digital imagery or multidate SAR (Townsend and Walsh, 1998; Townsend, 2001; Lang et al., 2008; Liu et al., 2008a).

Ground-based measurements of soil's apparent electrical conductivity (ECa) have been widely used to infer properties of the parent material for relatively small areas of field to regional extent. Here the assumption has been that electrical conductivity reflects the combined influence of soil texture, mineralogy and moisture content, and areas of different texture will exhibit interpretable differences in electrical conductivity (Liu et al., 2008b). Additionally, other natural fields of the earth, gravitational, magnetic (Galdeano et al., 2001), and electromagnetic (Beard, 2000), have also been used to provide information on underlying geological structure.

Recent developments have seen increasing interest in modeling spatial variation in parent material classes or parent material properties in a manner that is analogous to *scorpan* modeling of soil spatial patterns (Hardy, 2004; Lawley and Smith, 2007). Pain et al. (2001) and Clarke and Pain (2004) described regolith–landform studies in Australia as being concerned with understanding the 3D distribution of regolith materials, and its representation in map form, rather than vertical profiles or cross sections. They note that regolith is very closely associated with landforms and once the interrelationships between regolith and landforms are understood, landforms can largely be used to predict regolith patterns. Thus, landforms are used as a proxy for the largely hidden regolith. They report that landforms can be identified and delineated in a number of different ways, including conventional air photo interpretation, manual or automated interpretation of remotely sensed imagery, automated analysis of airborne radiometric data, and 3D mapping of the regolith using such techniques as air- and ground-based electromagnetic, ground-penetrating radar, and shallow seismic magnetic surveys. They inferred regolith properties such as texture, mineralogy, depth, and geochemistry by interpretation of the mapped landform classes. It is anticipated that an increasing number of digital soil mapping studies will utilize predictive modeling methods to produce separate predictions of parent material attributes such as depth, texture, and lithology for subsequent use in digital soil mapping.

37.4.6 "*a*" Factor

a represents age or elapsed time. This will give limits on how long pedogenesis has been occurring and should differentiate soil classes and properties. One useful estimate of *a* is the age of the ground surface, which may be very old indeed (Twidale, 1985). Alternatively, *a* can be represented by the age of the material in which soil has developed, suggesting that the *scorpan* approach will not deal well with polycyclic soils. It is theoretically likely that soil development will follow some logarithmic or square-root time relationship, suggesting more need to differentiate between younger materials than older ones. Schaetzl et al. (1994) discuss the form of soil chronofunctions.

Geomorphologists and stratigraphers can presumably draw maps of *a* independent of soil maps. In fact, such "gues(s)timated" maps along with an estimate of uncertainty could be used to represent this factor. There are methods for soil and material dating of course, for example, ^{14}C, $\delta^{18}O$, thermoluminescence (inter alia, Matt and Johnson, 1996), and $^{40}Ar/^{39}Ar$ (inter alia, Van Niekerk et al., 1999). None of these are, as far as we are aware, capable of scanning and producing full coverage in a true remote sensing fashion. Ground electromagnetic methods have been used for stratigraphic mapping, for example, Sinha (1990). *a* remains difficult to characterize well. Indeed, it seems that more than any factor expert knowledge is still needed to derive *a*. Considerable advances in technology and knowledge are needed.

Scull et al. (2005) used decision tree in the southwestern Great Basin area for predicting soil class. Age of landform is inferred from remote sensing data, which can indicate surface age by detecting desert varnish. They found that the age of landform covariates is an important predictor in the decision tree model. Noller (2010) presented an explicit use of the age factor in digital soil mapping. He estimated geochronological information for Malheur County, Oregon based on independently compiled Quaternary geological maps, age point data, and remotely sensed data. These data were incorporated in decision tree analysis. Expert soil survey maps are used as reference in making predictions with and without implicit or explicit age information. Addition of geochronological data produces digital soil maps that are most closely aligned with expert maps. Improvements in the map are greater in areas where the age stratification of the landscape is not obvious to the soil surveyor.

37.4.7 "*n*" Factor

As was discussed in Section 37.2, soil can be predicted from spatial coordinates alone. Obtaining these is now much easier due to the advent of GPS with 5 m accuracy receivers costing less than $1000. This may indeed reflect some other environmental variable such as climate, and because of this, it can be argued that *n* is not really a factor, but simply organizes the coordinates is a simple way to ensure that spatial trends not included in the other environmental variables are not missed. Therefore, *n* could also be described by some linear or nonlinear (nonaffine) transformation of the original spatial coordinates, for example, a new coordinate could be the closest distance of each location to the coast (Webster, 1977, p. 201), or contextual representation of space, for example, distance uphill from the nearest discharge area (Bui and Moran, 2000), distance from a watershed, distance from roads, distance from a point source, etc. This factor is potentially a valuable yet cheap source of environmental information and should never be disregarded.

37.5 Applications

Having presented a review of what has been done by others and having suggested and reviewed a generic predictive model and potentially useful environmental data layers, we now put this into a framework for soil mapping based on *scorpan* and Soil Spatial Prediction Functions (SSPFs) and spatially autocorrelated errors (**e**). The *scorpan*-SSPFe approach is now outlined with some brief discussion of each of the steps. Its uses, problems, and other implications of the *scorpan*-SSPFe approach are discussed subsequently.

37.5.1 Summary of the Scorpan-SSPFe Method

The *scorpan*-SSPFe method essentially involves the following steps.

i. *Define soil attribute(s) of interest and decide resolution* ρ *and block size* β
These are the design specifications for the survey. Define soil attribute(s) of interest, that is, a soil property or set of soil properties and/or a set of soil classes, usually from some predefined classification system. The resolution may be defined by the resolution of the environmental variables, for example, 30 m Landsat pixels, but should be a design specification from the intended use of the information. Referring back to Table 37.1, we believe that the methodology discussed here is most appropriate for D3 surveys, therefore pixel or block size ρ, is equal to pixel spacing β and will be in the range 20–200 m. (The linking of ρ and β is a simplification and is a point that requires further study—see Bishop et al. (2001) for further discussion.) At this stage, the uncertainty limits that can be tolerated may be also be defined.

ii. *Assemble data layers to represent Q*
Assemble the data layers with consideration of the number of layers describing each factor and any prior evidence as to the importance of each factor. This was discussed in detail in Section 37.4. At this stage, we do not know the relative importance of the data layers. Balance is probably important. At this phase of development, because of the relative availability of DEMs, it would be easy to obtain 15 or 50 terrain attributes (*r*), for example, as described in Shary et al. (2002), and rather difficult to represent *a* or *c*. An attempt should be made to represent all the factors however.

iii. *Spatial decomposition or lagging of data layers*
This is suggested as a step because it is felt that predictions might be scale dependent, and it is important to find the appropriate spatial associations. This can be achieved either by a wavelet decomposition (e.g., Mendonça-Santos et al., 2006), or geostatistically. The geostatistical approach involves modeling the correlation structure in the imagery by decomposing the variogram into independent spatial components, and then taking each component in turn and kriging it, thereby separating it from the others. Oliver et al. (2000) used this approach on SPOT imagery. Both of these

methods will allow the removal of short-range uncorrelated noise components from subsequent sampling, modeling, and prediction stages. The spatial decomposition of the environmental variables begs the question of whether the target attribute should also be spatially decomposed. In most cases, there would probably be insufficient observations to do this effectively, but it could be done where this is not the case.

An alternative approach to spatial decomposition is that of spatial lagging, that is, to fit a model such as,

$$S(x,\ y) = f(s(x+u,\ y+v),\ c(x+u, y+v), \ldots) + g(x+u, y+v), \tag{37.22}$$

where the soil attribute of interest is "regressed" on the layers representing the *scorpan* attributes and on spatially lagged ($+u$, $+v$) copies of them, with u and v variable. This approach seems somewhat more cumbersome than the spatial decomposition approach.

iv. *Sampling of assembled data (Q) to obtain sampling sites*
In most cases, soil sampling will be required to set up the model (except perhaps when the aim is map updating). We have a lot of prior information on the environmental variables, which we can use to guide the sampling. The aim is to construct predictive equations for the soil attributes of interest in terms of the environmental variables. This is a kind of calibration exercise, and, therefore, it would seem wise to span the range of values of each variable, so that the prediction model will not be required to extrapolate beyond its bounds. Hengl et al. (2003) proposed sampling along the principal components of the *scorpan* factors; the number of samples taken from each of the components is the proportion of the total variance described by each of the principal components.

Minasny and McBratney (2006) argued that, for the purpose of spatial regression calibration, it would be beneficial to cover the range of values of each of the covariates using LHS (McKay et al., 1979). LHS is a procedure that ensures a full coverage of the range of each variable by maximally stratifying the marginal distribution. LHS involves sampling n values from the prescribed distribution of each of k variables $x_1, x_2, \ldots, x_k$. The cumulative distribution for each variable is divided into n equiprobable intervals. A value is selected randomly from each interval. The n values obtained for each variable are matched randomly with those of the other variables. This method does not require more samples for more dimensions (variables). A conditioned Latin hypercube sampling (cLHS) was proposed by Minasny and McBratney (2006) for sampling of existing covariates or *scorpan* factors. We cannot directly apply conventional LHS to the multivariate distribution of covariates. Sample points selected by conventional LHS may represent combinations of the variables that do not exist in the real world. Randomization is used in this case as the distribution among the variables is not even, and some parts of the variable space might not correspond with existing soil. The conditioned LHS algorithm attempts to select n observations (sites) from the covariates, which can form a Latin hypercube in the feature space. The algorithm solves an optimization problem: given M sites with covariates (X), select m sample sites ($m \ll M$) so that the sampled sites x form a Latin hypercube. The method is a search algorithm based on heuristic rules combined with an annealing schedule. The objective function to be minimized is an error criterion that counts the occupancy of the hypercube. This kind of sampling should produce a reasonably efficient way of sampling the soil and its environment, so that the range of conditions are encountered, ensuring a good chance of fitting relationships *if they exist*.

An alternative procedure is suggested by the work of Lagacherie et al. (2001) that define a reference area (Favrot, 1989), which through the spatial data layers of the environmental variables extrapolates well to a larger region. Sample the reference area purposively or systematically (fit the model and extrapolate to the rest of the area). This might give a better chance of fitting local relationships with a given sampling effort and should be more efficient in field time—the advantage hinges however on how well the extrapolation can be done.

v. *GPS-field sampling and laboratory analysis to obtain soil class or property data*
Step (iii) yields a set of m spatial coordinates at which the observations of the soil attribute(s) are to be made in the field. These can be located with a GPS receiver, and samples taken for subsequent laboratory analyses, in the usual manner. At a subset of these locations, say 5%, a duplicate observation should be made at a distance (say) half the resolution (ρ) to get an estimate of the short-range field variability in the soil attribute(s). This will help in subsequent spatial modeling. (If a specific purpose or numerical classification or allocation to a conventional classification system is required then the observed soil data can be processed to obtain discrete or continuous class labels for each observation site.)

vi. *Fit quantitative relationships (observing Ockham's razor)*
We can now assemble the soil data for the left side of the *scorpan* equation and the environmental data for the right side. We can now fit the model $f()$ representing the m locations using any of the techniques described in Section 37.3.3. Ockham's razor (the principle that states that "Entities should not be multiplied unnecessarily") should be applied to find the model (with the least number of parameters) that fits best. The residuals (e) of the soil property or class probability or membership at each of m sites should be estimated also (and kriged using either *scorpan*-simple or *scorpan*-compositional kriging as mentioned in Section 37.3.4.2.2).

A single model may not be adequate especially if there are strong pedogeomorphic or geological contrasts within the study area—each subregion may show quite different relationships between soil and environmental variables. In that case, it may be necessary to develop a *scorpan*-SSPFe model for each pedogeomorphic or geological subregion. Care needs to be taken that the prediction surfaces produced are realistically continuous. This approach demands a higher data burden than a single model for a whole geographic region. Once again the total number of parameters in the final model needs to be considered.

If we have a lot of local information for both sides of the equation (like an existing soil map, or proximally sensed soil data), we can fit a local model,

$$S[x,\ y] = f_l([x,\ y]) + e_l(x,\ y), \tag{37.23}$$

where the l subscript on $f_l([x, y])$ and $e_l(x, y)$ refers to models fitted to a local (moving) neighborhood centered on $[x, y]$ rather than to the whole area to be mapped. It will be rare that m will be large enough to fit this model. This kind of model was used geostatistically, that is, $S[x, y] = e_l(x, y)$ for soil salinity mapping by Walter et al. (2001).

vii. *Predict digital map*

We now have a model $f()$ fitted to the m locations, which we can now apply to the $(M - m)$ locations where we have no soil observations, but have environmental observations. Additionally, kriging of the residuals (e) at is also done at the $(M - m)$ locations and the results added together. Additionally, the uncertainties of the predictions of $f()$ and e at the M locations should be evaluated. Raster maps can then be made of the soil attribute(s) and their associated uncertainties.

viii. *Uncertainty analysis*

A principal feature of digital soil mapping is the ability to present an accompanying map of uncertainty to the prediction map. However, uncertainty is seldom quantified in digital soil maps. There are two important sources of uncertainty: the uncertainty in the input data and the uncertainty of the spatial prediction techniques.

Uncertainty in the input data can be from the soil data and *scorpan* factors. The uncertainty in soil data is usually much larger than in the covariates. Usually, digital soil mapping uses legacy soil data. Legacy soil data arise from traditional survey, which is generally empirical and based on the mental development of the surveyor. There are no statistical criteria for traditional soil sampling, which may lead to bias in the areas being sampled. Carré et al. (2007) discusses some techniques for evaluating the quality of the legacy soil data based on the *scorpan* factors.

Conventional polygon-based maps often have no uncertainty information. In the Australian Soil Resource Information system, the uncertainty of a soil property is derived from the polygon-based map (McKenzie et al., 2005). This is calculated from a standard uncertainty, which is based on scientific judgment using all the relevant information available, which may include previous measurements on related soils, experience of lab or field or prediction techniques, and assumed from spatial variation.

The major contributing factors to uncertainty in spatial prediction are the neighborhood points used for interpolation: the number and proximity of the samples, clustering of samples, and continuity of the variables. In geostatistics, the usual approach for calculating the uncertainty is by computing a kriging estimate and the associated error variance, which are then combined to derive a Gaussian-type confidence interval. For the regression-kriging approach, the uncertainty has two parts, from the regression model and the kriging of the residuals. See Minasny and McBratney (2007a, 2007b) for mathematical details on this topic.

The uncertainty for prediction of soil classes is more difficult to quantify. Kempen et al. (2009) assessed the uncertainty of the multinomial logistic regression model prediction of soil classes by using Shannon's information entropy. It is calculated based on the probability of the predicted soil classes, the maximum entropy is 1, which occurs when all outcomes have equal probability. The minimum value for the entropy is 0, which occurs when there is no uncertainty and one of the outcomes has probability 1. It should be noted that the entropy indicates whether the predicted soil group has a large probability, it does not indicate that the prediction itself is correct. What is really required is a prediction interval on the prediction probability statement.

Bui and Moran (2003) used a heuristic approach for uncertainty of soil classes. The training area was randomly sampled 10 times. Each time a decision tree was built, and soil classes were predicted into unmapped areas. Where the same class was consistently predicted by several trees, the likelihood is greater that the prediction is consistent (not necessarily correct). The modal class prediction from the 10 trees was used in making the final map. The proportion of predictions resulting in the modal class was used to create a map of uncertainty of the predictions. The map of uncertainty of predictions indicates where the trees have difficulty in capturing the spatial pattern of soils in the landscape.

ix. *Field sampling and laboratory analysis for corroboration and quality testing*

It must not be assumed that the digital information is perfect with minimal information on the quality of the information. Indeed, we cannot expect this kind of map to be more accurate than conventional ones. There are two reasons for this: (i) local variation (at whatever resolution) is a limiting factor, there is a lot of soil variation within 10 m, or 100 m, or 1 km; and (ii) there is uncertainty in environmental layers and this can propagate errors (Lagacherie and Holmes, 1997; Florinsky, 1998; Heuvelink and Burrough, 2002).

Like conventional soil maps, digital ones should also be evaluated for quality. This is the same as ground truthing in remote sensing where many measures have been advocated (see Story and Congalton, 1986; Liu et al., 2006). The main quality criteria for soil class maps are purity (user's accuracy) and producer's accuracy, and root mean square error for continuous soil property maps. These can be evaluated by a number of strategies.

One approach is by cross-validation. Some proportion of the available observations are put to one side, a common proportion is one-third. Prediction modeling is done with the remaining two-thirds of the data and the map produced. The one-third put aside are now considered test sites and compared with model predictions This can give an indication of producer's and user's accuracy for categorical maps and average bias and root mean square error for maps of continuous soil properties. The results are only indicative because the sample was probably not designed for the express purpose of quality assessment. The results might be somewhat biased for legacy data (Carré et al., 2007), and less so for observations from LHS (Minasny and McBratney, 2006), but still not optimal.

It is better to construct a new sample. Cluster sampling and stratified random sampling have been used in the past for testing the accuracy of conventional maps (De Gruijter and Marsman, 1985; Marsman and De Gruijter, 1986). Kempen et al. (2009) carried out an accuracy assessment of their digitally updated soil map of part of the Netherlands. They observed 150 new points based on a stratified random sampling design with strata based on the original map and found a small improvement in purity from 52% to 58%. They also estimated the purity and producer's accuracy (which they called "sensitivity") for all the units of the new digital soil map.

One of the principal advantages of digital soil maps is the promise of spatialized uncertainty estimators. The quality of the combination of the estimate and its associated uncertainty needs to be tested to properly evaluate digital soil maps. Thus far, this has not been attempted.

x. *If necessary simplify legend, or decrease resolution by returning to (i) or, improve map by returning to (v)*
If we find the map does not meet design specifications, that is, class purity is less than x%, or the confidence intervals for a soil property are too wide over parts of the map, then we can simplify the legend or decrease spatial resolution (i)—but these should be design specifications, or more sensibly, target further sampling (v) in areas where predictability appears to be poor, and recalculate the maps.

37.5.2 Uses

There are various potential uses of the digital soil mapping approach. Among them are production of digital soil maps as a replacement for the paper-based choropleth soil map of the past, updating existing soil maps, extrapolating existing soil maps into unmapped areas, construction of dynamic soil maps, digital soil assessment, and global soil map.

37.5.2.1 Digital Soil Maps

The main use of the *scorpan*-SSPF approach is to replace the polygon-based soil maps of the past with digital maps of soil properties and classes and their associated uncertainties for areas previously mapped, or for new areas. These maps will be stored and manipulated in digital form in a GIS creating the possibility of vast arrays of data for analysis and interpretation.

The first digital soil maps were simply representations of the observations without interpolation or relation to the environment (e.g., Webster et al., 1979). Some authors have worked on better ways to present digital spatial soil information, chronologically, De Gruijter and Bie (1975), De Gruijter et al. (1997), and Grunwald et al. (2001). This is an area requiring considerable research. One goal must be fully operational multiresolution digital soil maps.

While we cannot necessarily expect the maps to more accurate than conventional ones, we can expect to have a quantitative estimate of the uncertainty (Section 37.5.1); sampling effort should be expended to achieve this. Laba et al. (2002) compared conventional (Congalton and Green, 1999) and fuzzy (Gopal and Woodcock, 1995) methods to assess the accuracy and uncertainty of land cover maps produced at high taxonomic resolution. These methods could be applied to digital soil maps.

Survey commissioners, decision makers and users in general would perhaps be more comfortable with a concept of certainty rather uncertainty. This answers the question, "how well do we know the value at some location?" rather than concentrating on "how badly we know it." A potentially adequate standardized ($0 \rightarrow 1$ or $0\% \rightarrow 100\%$) measure of certainty is $f = 1 - \min(2s/V, 1)$, where s is the standard deviation of the estimate. For example, if we have an estimate V of clay content of 50% and an s of 5% then $f = 0.8$ or $f\% = 80\%$. More sophisticated measures may be required, such as a certainty characteristic—the probability that a statement C is true within a distance d or an increasing neighborhood A. Clearly, more work is needed on standards for digital soil maps.

A recent advancement in digital soil mapping is the capability to map the distribution of soil properties over the whole soil profile. This created a pseudo 3D distribution of soil properties. Minasny et al. (2006) fitted an exponential decay depth function for soil organic C distribution data in the Edgeroi area in NSW, Australia. Parameters of the exponential function were then mapped using neural networks with some *scorpan* factors as covariates. Carbon storage for any point in the landscape then can be obtained by integrating the exponential function to the desired depth. A more general depth function can be obtained using an equal-area spline (Bishop et al., 1999). Parameters of the spline function can be mapped creating a whole soil profile distribution (Malone et al., 2009).

37.5.2.2 Updating Soil Maps

Digital soil mapping can be used to update soil maps from previous surveys. Models predicting soil classes or series can be built based on legacy soil data (existing soil maps or soil profile descriptions) with *scorpan* factors as covariates. The spatial prediction models can be applied to the entire data. Kempen et al. (2009) used digital soil mapping principles to update the national 1:50,000 soil map for the province of Drenthe (2680 km^2) using legacy soil data. Multinomial logistic regression was used to quantify the relationship between ancillary variables and soil groups. The model-building process was guided by pedological expert knowledge to ensure that the final regression model was not only statistically sound but also pedologically plausible.

They estimated actual purity of the updated map, as assessed by the validation sample, was 58%.

37.5.2.3 Interpolation or Extrapolation of Existing Soil Maps

If *s* for a previously mapped region is put on the right of *scorpan* equation, the legend is retained largely, and new samples are collected, this might be considered by some to be map updating. There is another possibility; this is where the previous *s* is put on the left rather than the right side of the *scorpan* equation (an example is given in Bui et al., 1999). The advantage of this is that no new sampling is required for fitting—although corroboration sampling should be done. This would allow quantitative elaboration of the existing (but unknown) models, for classes this is what Girard (1983) referred to as (the usually unknown) "chorological rules," and their subsequent extrapolation to new areas. Possible problems include repeating old models that may be wrong, or extrapolation outside the range of associated environmental data sets—Lagacherie and Holmes (1997) and Lagacherie and Voltz (2000) discuss this for landscapes, while McBratney et al. (2002) were concerned about this for pedotransfer functions. It is also important to establish for which purpose existing data have been collected. Some soil surveyors use auger observations to confirm their mental models, while others use them to find inclusions or boundaries. Therefore, this approach should be used with a deal of caution.

Extrapolation of soil types from one small area to a larger extent can be done once the digital soil mapper knows the initial area is representative of the larger extent (that is why it is usually called "reference area"). This idea, initially proposed by Favrot (1981), has been applied by Lagacherie et al. (2001) for extrapolating French Mediterranean soilscapes. Soilscapes of reference defined as a combination of soil-forming factors in a buffer neighbor can be expressed as a vector composition of elementary landscape classes of different sizes. Since the landscape classes are known to be representative of the overall area, Manhattan distances between soilscapes of reference and composition of landscape classes in running neighboring windows can be calculated. The final allocated soilscape is the one that minimizes the Manhattan distance. Two outputs are then produced in that case:

- The final soilscape map
- The corresponding allocation distance giving access to the accuracy of the final map

This technique allows for minimizing field sampling. However, the main scientific question is where to stop the boundaries of the bigger extent to map?

Bui and Moran (2001, 2003) built decision tree rules in training areas where detailed soil maps were available, and the rules were extrapolated to larger areas where detailed maps were not available. Grinand et al. (2008) tested this approach and observed marked differences in accuracy between the training area and the extrapolated area. They found that sampling intensity did not appear to influence the accuracy of prediction, and spatial context integration by the use of a mean filtering algorithm on the covariates increased the accuracy of the prediction on the extrapolated area.

37.5.2.4 Environmental Change: Partially Dynamic "Scenario" Soil Maps

One major criticism of conventional soil maps is that they are essentially static statements. Digital soil maps created with the *scorpan*-SSPF methodology offer new and necessary possibilities. It is becoming increasingly important for environmental reasons to know not just $S[x, y]$ but $S[x, y, t]$. If we know any of the partial differentials, $\delta s/\delta t$, $\delta c/\delta t$, $\delta o/\delta t$, etc., the first two perhaps being the more important—we can project the existing soil map forward by some time u by calculating most simply say $c + u(\delta c/\delta t)$ for all points and running the new c layer(s) through the prediction function. "Change-detection analysis" (Mücher et al., 2000) is well developed for land use and vegetation change (components of $\delta o/\delta t$) using remotely sensed imagery and/or aerial photographs average, and localized values of o can be estimated from rasterized images taken at two or more times (Munyati, 2000). Other derivatives may be obtained from models (e.g., temperature and rainfall changes) or from a few monitoring stations (e.g., soil changes) within the area of interest (Mendonça-Santos et al., 1997).

This potential approach has limitations compared with a fully fledged dynamic simulation model, such as lack of feedback and possible extrapolation problems, where, for example, $c + u(\delta c/\delta t)$ takes us (well) outside the range of the original training data. Nevertheless, we still have a relatively quick and easy way to produce first-cut "scenario" soil maps of both properties and classes.

The European Parliament is currently implementing a new Directive on the use of renewable energy in Europe. Ten percent share of renewable energy in transport should be achieved by each Member State by 2020. The main source of renewable energy for transport is planned to be biofuel. One scientific aim addressing the sustainability question is then to assess the impacts of such a Directive on global land (soil and above-ground vegetation) carbon stock changes due to indirect land use conversions. After quantifying the biocrops needed to achieve the 10% biofuel share target, land use conversion due to biocrops and transfer of food/feed/fiber crops will be spatially estimated at the global scale. Overlaying the current stock of land carbon with the current land use and targeted land use conversion, climate and land management practices information, it is possible to spatially assess the carbon stock changes within the next 20 years by following the Tier 1 approach of volume 4 of the *IPCC Guidelines for National Greenhouse Gas Inventories* (IPCC, 2006). The main difficulty is not only to estimate uncertainties coming from the modeling (this is provided by the IPCC report) but also to quantify the uncertainties of the different sources of inputs, usually not specified.

37.5.2.5 Digital Soil Assessment

Digital soil assessment is the step following digital soil mapping (Carré et al., 2007). It is the quantitative modeling of difficult to measure soil attributes, necessary for assessing threats to soil (e.g., erosion, decline of organic matter, and compaction) and soil functions (biomass production and environmental interactions) using outputs from digital soil maps, that is, the soil properties map and the accuracy of the predictions. The threats to soil and the soil functions are generally aspatial models. They require not only data on soil (as provided through digital soil mapping) but also external data that can be also considered as soil-forming factors, and already used as covariates. These can be climate, land cover, or land management practices. DSA requires running the aspatial models, and as for DSM, the prediction of uncertainties coming from input data and the models.

Uncertainty prediction is usually performed using global sensitivity analysis of the models and simulating the distribution of the uncertainties with Monte Carlo simulation (Heuvelink, 1998). Three exercises have been tested for digital assessment problems. The first one deals with soil pollution issue (Romić et al., 2007) using regression-kriging approach and continuous limitation scores for assessing global soil pollution. The second one deals with wind erosion (Reuter et al., not published) combining regression-kriging for the soil properties prediction (used as output for wind erosion). The third one deals with estimation of biomass production in England & Wales (Mayr et al., private communication).

37.5.2.6 Global Soil Map

A Digital World Soil Properties Map consortium (www.globalsoilmap.net) has been formed and comprises representatives from universities, research centers, development organizations, and private enterprises around the world. This is a response to the urgent need for accurate, up-to-date, and spatially referenced soil information, at a global scale, as expressed by the modeling community, farmers and land managers, and policy and decision makers. The objective of this consortium is to create a digital map of the world's soil properties at a resolution of 90 m × 90 m to a depth of 1 m based on legacy and newly collected soil data.

Making a digital global soil map is a challenging task, methods for mapping soil properties at a continental or global scale are not straightforward. Moreover, different parts of the world have varying soil information of varying qualities. The soil information can be from legacy soil profile data, existing soil maps, but also data from reflectance spectra. The soil–landscape model will vary from place to place and the knowledge and techniques for regional soil mapping may not be applicable at a global scale.

Minasny and McBratney (2010) summarized a methodology for global digital soil mapping based on legacy soil data. For an area of interest, we assemble all the *scorpan* or environmental covariates and existing soil data. The second step is to check how the soil data cover the covariate space and to select possible training areas.

Methods used for digital soil mapping depend on the availability of soil data, and the possibilities in the order from the richest to the poorest soil information are as follows:

1. *Detailed soil maps with legends and soil point data*
   This is the richest information that can give the best prediction of soil properties. Soil properties can be derived from both soil maps and soil point data. The available methods are extracting soil properties from soil map using a spatially weighted measure of central tendency, for example, the mean, spatial disaggregation of soil maps, *scorpan* kriging, and combinations of these. An example of such an application is given in Henderson et al. (2001, 2003).
2. *Soil point data*
   When soil point data are available, soil properties can be interpolated and extrapolated to the whole area by using a combination of empirical deterministic modeling and a stochastic spatial component. We called this the *scorpan* kriging approach.
3. *Detailed soil maps with legends*
   When only soil maps are available, soil properties need to be extracted from the maps using some central and distributional concepts of soil mapping units.
4. *No data*
   When no data or soil maps exist in area, we will use an approach we called *Homosoil*, which means that we need to estimate the likely soil properties under the observed soil-forming factors or *scorpan* factors.

37.5.3 General Discussion

We now discuss some general points relating to the *scorpan*-SSPF approach to making digital soil maps.

37.5.3.1 Is This the Right Approach?

As stated at the beginning of Section 37.5.1, this is a proposal, not a fait accompli, albeit based on a significant amount of work and experience worldwide. Whether or not this turns out to be the right approach hinges on a number of factors, not all of them scientific, but we shall deal with those first. Scientifically one could ask a number of questions. Are there other soil-forming factors? Are we missing key variables? For example, have we successfully incorporated hydrological effects? Is the underlying idea of a soil somehow in equilibrium with its environment reasonable enough to be predictive? Or, is the soil too chaotic for prediction by other factors? Will the proposed methodology give similar answers as traditional approaches, and do we want it to? Further experiment and experience will undoubtedly answer these questions.

There are other socio-economico-political factors that will have a bearing also. The sociopolitical factors demand recognition of, and solutions to, environmental problems. We believe that the approach has the right kind of economics; it is potentially cheaper than traditional approaches and gives the desired kinds of information. Most of the hardware and software tools are in place to put this approach into practice. Clearly, integrated systems have to be devised. Research into aspects is always needed, principally efficient sampling designs and useful certainty estimation. The biggest stumbling block is identifying and assembling the teams of personnel with the skills required to complete the task. Education of skilled and knowledgeable personnel for those teams is a key priority.

37.5.3.2 Dangers: Let the User Beware!

There are real dangers with this, or any new approach, if it is misused or abused. Here, we outline briefly some obvious pitfalls.

1. *Data quantity and quality*. The first danger is not using enough real soil observations to fit the models, or with using poor quality (missing or noisy) predictor variables. This can, to a degree, be handled by uncertainty analysis—a large topic (Heuvelink, 1998), which has not been discussed formally in this paper. There is a lower limit below which any fitted models will be meaningless.
2. *Overfitting the data*. It is easy to overfit models; this could be because of lack of observations but is more often due to a lack of parsimony, especially a problem for tree-based methods. Overfitted models predict poorly. It is imperative to apply Ockham's razor—this will help with evaluating poorly fitting or overfitted models. The use of cross-validation, pruning, and boosting methods (Hastie et al., 2009) might also help.
3. *Circularity*. A third hazard comes from the possible circularity of the model, for example, a DEM producing climate surfaces producing soil variables as an input to soil class prediction. Once again uncertainty analysis will help.
4. *Databases and data mining*. During the past decade, soil scientists have created regional, national, continental, and worldwide databases. Data mining is a phrase for a class of database applications that look for hidden patterns in such groups of data (Hastie et al., 2009). (Unfortunately, the term is sometimes misused to describe software that presents data in new ways.) Proper data mining software attempts to discover previously unknown relationships among the data. Data mining is a broad concept from supervised learning (prediction) to unsupervised learning and includes all the methods described in Section 37.3.3—neural networks, classification trees with boosting. There are a large number of commercial software products available to do this. They incorporate one, or often several, of the methods described in Section 37.3.3. This will make evaluation difficult as different soil science groups use different software products for fitting *f*(); therefore, comparative studies will be important to evaluate the best approaches. In addition, large national or international databases of legacy soil data will be available (e.g., Bui and Moran, 2003); they also have potential problems because of their unknown site selection probabilities—which are not usually equal—some of the data from Britain where a 5 km grid survey has been completed, are an exception!

37.5.3.3 A New Paradigm?

Hudson (1992) described soil survey as a paradigm-based science. Paradigm is a much overused and hackneyed word these days but it has a precise philosophical meaning. (Much of the following two paragraphs is paraphrased from Rosenberg, 2000.) Paradigm is a term employed by Kuhn (1996) to characterize a scientific tradition, including its theory, apparatus, methodology, and scientific philosophy. The soil scientist's task is to apply the paradigm to the solution of problems. Failure to solve problems is the fault of the scientist not the paradigm. Persistent failure makes a problem an anomaly and threatens a revolution, which may end the paradigm's hegemony.

What is the difference between the *scorpan*-SSPFe approach and the conventional Jenny-landscape model? Both are models—they are simplified descriptions of regularities governing a natural process, usually mathematical and sometimes derived from a more general or simplified theory. Ontologically, they are similar—they both require soil objects and attributes, which are a function of their environment. The conventional paradigm is a qualitative theory. The approach outlined here is a quantitative, partially deterministic, partially probabilistic, empirical theory. So methodologically, they are quite different. The apparatus is also different; here, we require digital information, computers, GIS, etc. The Jenny-landscape model may fall under the *deductive-nomological* model of scientific explanation but, because of its somewhat probabilistic nature, the *scorpan*-SSPFe approach may fall under the *inductive-statistical* model of explanation (Rosenberg, 2000). So the *scorpan*-SSPFe approach to soil mapping probably represents an emerging paradigm eventually leading to a complete paradigm shift.

This begs the question, does *f*() have to be empirical? The Vienna school of logical empiricists would be generally happy with *scorpan*-SSPFe approach, although perhaps they would have difficulties with its partially probabilistic nature. The lack of a mechanistic theory for predicting soil tugs at the soil scientist's cloak of explanation. Perhaps this is an unnecessary concern; philosophical empiricists believe that there is nothing to causation beyond a regular sequence. Any testing of the mechanistic theory will require empirical observation of the real world. The first attempts at a mechanistic approach have begun but it will be a long time before the mechanistic theoretical approach will be competitive in the predictive sense.

37.5.4 Working Group on Digital Soil Mapping

The Working Group on digital soil mapping (www.digitalsoil-mapping.org) was formed in 2005. It operates under the auspices of the Commissions on Soil Geography (C1.2) and Pedometrics (C1.5) of the International Union of Soil Sciences (IUSS). The workgroup has biennial workshops on Digital Soil Mapping. The first workshop was held in Montpellier, France in 2004. The outcome of the workshop is published as a book: Digital Soil Mapping: An Introductory Perspective (Lagacherie et al., 2006). The second workshop was in Rio de Janeiro, Brazil in 2006. The outcome of the workshop is published as a book: Digital Soil Mapping with Limited Data (Hartemink et al., 2008). The third workshop was held in Logan, the United States, in 2008 with an outcome: Digital Soil Mapping: Bridging Research, Environmental, Application, and Operation (Boettinger et al., 2010). The fourth workshop was in Rome in 2010.

37.6 Conclusions

We have reviewed various approaches to predictive modeling and data acquisition and proposed a framework and a methodology for producing digital soil maps.

37.6.1 Summary of the Method

The *scorpan*-SSPFe method essentially involves the following steps:

i. Define soil attribute(s) of interest and decide resolution ρ and block size β
ii. Assemble data layers to represent *Q*
iii. Spatial decomposition or lagging of data layers
iv. Sampling of assembled data (*Q*) to obtain sampling sites
v. GPS-field sampling and laboratory analysis to obtain soil class or property data
vi. Fit quantitative relationships (observing Ockham's razor) including spatially autocorrelated residual errors
vii. Predict digital map
viii. Perform uncertainty analysis
ix. Field sampling and laboratory analysis for corroboration and quality testing
x. If necessary simplify legend, or decrease resolution by returning to (i) or, improve map by returning to (v)

All of the hardware and software tools, technologies and knowledge, are in place to make this approach operational. This is clearly an exciting time for soil resource assessment.

37.6.2 Future Work: Open Questions

Clearly, we need to try out the methodology outlined above and by experience we shall discover the useful forms of *f*() and the serviceable *Q* layers. These are the key open questions. In summary, topics to be further researched include the following:

1. Environmental covariates for digital soil mapping
2. Spatial decomposition and/or lagging of soil and environmental data layers
3. Sampling methods for creating digital soil maps
4. Quantitative modeling for predicting soil classes and attributes (including generalized linear and additive models, classification and regression trees, neural networks, fuzzy systems, expert knowledge and geostatistics)
5. Quality assessment of digital soil maps
6. (Re)presentation of digital soil maps
7. Economics of digital soil mapping

Nevertheless, we believe the methodology can be used now for real-world applications.

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38

Soil Change in the Anthropocene: Bridging Pedology, Land Use and Soil Management

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38.1 Overview and Objectives

Pedology is the science of soil change. Historically, this meant that pedology was mainly a basic science of natural soil processes and soil formations. Today, in the new geologic epoch that Crutzen (2002) has named the Anthropocene, pedology is fundamentally challenged to bring humanity entirely within the soil continuum (Dudal et al., 2002). During this period in which humanity exerts growing and predominant influence over Earth's systems, and especially Earth's soil (Yaalen 2000), pedology is becoming a much more interdisciplinary science with basic and applied objectives, one that requires the natural and social sciences but also the humanities (Richter, 2007).

In this chapter, discussions of "soil change" focus on the domestication of Earth's soil over historic time, the transformation of soil into a cultural-historic-natural system. This chapter describes human-altered soil processes and the forcing factors that are fundamentally changing soils across the planet, for like it or not, soils now serve as parent materials for humanity's accelerating influence. Humanity's impact on Earth's soils can now be called "global soil change" (Arnold et al., 1990).

The objectives of this chapter are to review the history of pedology as it responds to the challenges of the Anthropocene, consider soil change as it impacts soil function and ecosystem goods and services, evaluate the approaches to the study of soil change, enumerate on the factors and mechanisms affecting soil change that will improve prediction and management, and speculate on the future science, inventory, and management of soil change, specifically in the next few decades. The evolution of pedology is briefly described from its time as a basic, natural science that was focused narrowly on soil as a relatively slowly changing natural system, to its present and future as a broadly interdisciplinary environmental science that quantifies and predicts human-affected soil and soil-environment change on human timescales (years, decades, and centuries). More than anything, this chapter means to encourage the pedological community to redouble and expand the work of its pioneers such as Hilgard (1860), Darwin (1882), Dokuchaev (1883), Jenny (1941, 1961), Bidwell and Hole (1965), and Yaalon and Yaron (1966) to further integrate humanity as agents of soil formation, evaluate humanity's effects on soil function, and quantify and predict soil responses to human influence with smaller margins of error. We see this historical development as being entirely in parallel with the conceptual development of Earth's critical zone (National Research Council, 2001; Wilding and Lin, 2006), an integrative concept of the near-surface environment, first articulated by the landmark National Research Council (2001) report on the future of the earth sciences.

38.2 A Brief History of Human Influence on Soil

Soils have been used for agricultural and engineering purposes for at least 10,000 years. Human influence on soil was probably first affected by fire, but over the Holocene effects included a wide range of agricultural impacts; construction of villages, cities, and roads; leveling and terracing; irrigation and drainage; mining; and compaction and compaction and erosion. Human action is often goal oriented and is strongly affected by culture.

Human use, impact, and reliance on the soil are nothing short of staggering, and today with more than 6 billion persons on Earth, we look to the prospect in only a few decades of 10 billion persons (FAO-STAT, 2009). Over 2 billion Mg (metric tons) of cereals are harvested from soils per year, and over 1 billion Mg of vegetables plus roots and tubers; and from animals, over half billion Mg of milk, a quarter billion Mg of meat, and 50 million Mg of eggs; and from forests, more than 3 billion Mg of wood for fuel wood and industrial products. Currently, over 100 million Mg of N, P, and K are applied each year (increasing by three- to eightfold since 1960); about 250 million ha are irrigated, an area that has about doubled since 1970; and on the order of a million hectares of land is severely disturbed each year by mining. The growth of humanity's numbers and affluence means that these products of the soil will be doubled in the coming few decades and century. Both the extent and intensity of soil management are thus substantially increasing, and by any measure, we are pushing the Earth's soil very hard indeed.

More than half of the Earth's 13 billion ha of soil are today plowed, pastured, fertilized, limed, irrigated, drained, fumigated, bulldozed, puddled, compacted, eroded, leached, mined, reconstructed, or converted to new uses. Like it or not, humanity has become a major factor in soil formation and can prompt change through single events such as tillage or prolonged phenomena such as global climate change. A number of soil properties are responsive to common agricultural or forestry practices, regional air pollution, and alterations in hydrology and landforms (Table 38.1). Two general patterns are evident: Most soil properties are *dynamic*, that is, significant changes can occur over timescales of centuries, decades or less, and only a few are relatively static or *persistent,* that is, properties that are little affected by human forcings. Table 38.1 is most striking perhaps for how it illustrates that so many important soil properties are susceptible to change on timescales of decades or less.

TABLE 38.1 Soil Properties Grouped Qualitatively According to Rate of Change in Response to Human Forcings Such as Common Agricultural and Forestry Practices, Regional Air Pollution, Alterations in Hydrology, or Climate Change

Dynamic		Persistent
10 Years	10–10^2 Years	>10^2 Years
Acidity and salinity	Fragipans	Texture
pH-dependent charge	Fe/Al oxides	Rock volume
Bulk density and porosity	Occluded fractions of C, N, P	
Bioavailability of macro- and micronutrients and contaminants	Eluvial clay Non-pH dependent charge Fragipans, duripans, and plinthites	
Infiltration and hydraulic conductivity	Stabilized humic substances	
Rooting depth and volume		
Aggregates and structure		
Redoximorphic features		
Labile fractions of organic carbon		

38.3 Pedology, Anthropedology, and Earth's Critical Zone

Of the widely recognized five soil-forming factors, pedology has historically treated organisms nearly always without humanity, except as an actor that disturbs, interrupts, manipulates, truncates, arrests, or subverts the natural process of soil formation. Yet, like it or not, human influence on soils worldwide has increased so extensively in the past few decades that there is now no alternative but to embrace humanity as part of the soil system, (Amund & Jenny 1991) even to recognize human influence as a sixth soil-forming factor (Amundson and Jenny, 1991; Dudal et al., 2002).

Viewed from the perspective of the Anthropocene (Crutzen, 2002), pedogenesis and morphology have for too long overemphasized the natural environment to the neglect of humanity as a soil-forming factor (Dudal et al., 2002; Ibáñez and Boixadera, 2002). Most soil surveys provide little information about the response of dynamic soil properties, such as organic matter, bulk density, and aggregate stability, to management and the effects of these changes on soil functioning. Despite this critique however, the historic pedological emphasis on the natural environment is understandable given how daunting the tasks for traditional pedology have proven (e.g., Richter and Babbar, 1991; Lal and Sanchez, 1992). Soil, the subject of pedology, has been documented to possess extreme spatial diversity from local to global scales, a diversity derived from high-order interactions of soil-forming factors. Soil is described as the most unparsimonious of all Earth's natural entities according to Johnson (2005), and the most complex of Earth's biomaterials according to Young and Crawford (2004).

One of pedology's most challenging tasks, however, is to understand how soils function and develop over human timescales. While soils can be destroyed and reinitiated in a moment's passing due to floods, mudflows, wind storms, volcanic ejecta, earthquakes, or a plow or a backhoe, soil formation also plays out over incredibly long sweeps of time. In the natural environment, stream terraces stable for even 1000s of years have soils that are youthful, whereas ancient soils are distinguished if they develop and survive for millions of years on biogeomorphically stable landforms. With soils ranging so widely across time, it is understandable why pedology has traditionally considered humanity to be more an interruption than an integral part of soil formation. Yet in response to humanity's growing influence on Earth's systems, Bidwell and Hole (1965) and Yaalon and Yaron (1966) forcefully argued that pedology needs to embrace the vast activities of humanity. Bidwell and Hole (1965) considered not only hunting, gathering, and cultivation as integral to soil formation, but also watershed management and planning. And when Yaalon and Yaron (1966) used the term "metapedogenesis" that described human alteration of soil, pedology became a science of human-soil interactions.

Pedologists in the Anthropocene thus conceive of soil differently than did Hilgard (1860) and Dokuchaev (1883), whose pedological frontiers were focused explicitly on the natural formation of "virgin soil" according to Hilgard (1860). In the Anthropocene, pedology must increasingly focus on the science, inventory, and management of domesticated soils for purposes related to food production, human safety, and environmental quality. Because of the significance of this shift of focus, Yaalon and Yaron's (1966) ideas about anthropedogenesis can be recognized to be as fundamental to the development of pedology as the original model of soil as a natural body, attributed to Dokuchaev and Hilgard over a century ago.

No longer can soil be portrayed as it was in the past: as a relatively static component of ecosystems or as one formed by processes that do not include humanity. Recent pedological studies show how soil in the Anthropocene is a dynamic natural, historic, and cultural system subject to fundamental changes on human timescales. Boxell and Drohan (2009) found changes in soil morphology and physical properties with potential for increased runoff and erosion after *Bromus tectorum* L. (cheatgrass) invasion of a native sagebrush ecosystem in Western North America. After an evaluation of mature *Pinus palustris* Miller (longleaf pine) stands and intensively cultivated land use systems in the southeastern United States, Levi et al. (2010) discovered that near-surface soil properties were more similar by land use than by taxonomic-based soil map units.

Soil is an integral and interactive part of the rapidly changing ecosystem and wider environment and an important component of Earth's critical zone (National Research Council, 2001; Wilding and Lin, 2006), the solid surface of the planet where life is sustained by the interactions of physical, chemical, and biological processes in well integrated above- and belowground environments. Information about human impacts on these dynamic systems is needed to evaluate and predict the effects of land use and management on soil and its capacity to function. Recently, a new USDA-NRCS soil survey program was initiated to gather information about soil change in order to define potential and achievable values of dynamic soil properties for specific soils under a variety of management systems (Tugel et al., 2008).

38.4 Soil Change and Soil Function

Of great interest to anthropedologists are changes in soil functions in both natural and managed systems. Soil functions are described by what soil does and by how soil sustains, regulates, and controls the many biotic and abiotic processes in Earth's pedosphere. In the Anthropocene, soil functions and ecosystem services are often parallel concepts for the evaluation and management of human impacts on ecosystems. Soil functions and ecosystem services have developed into critical discussions not only in the soil sciences literature but also in the literature of environmental change, ecosystem ecology, biogeochemistry, and land management. The general similarity of these concepts across disciplines is striking.

Soil functions relating to plant production, human and animal health, and the environment are widely discussed in the soil quality literature (Larson and Pierce, 1991; Parr et al., 1992; Blum and Santelises, 1994; Doran and Parkin, 1994; Karlen and Stott, 1994; Acton and Gregorich, 1995; Warkentin, 1995; Harris et al., 1996; Seybold et al., 1997; Carter et al., 2003; Andrews et al., 2004) and include (1) sustaining biological activity, diversity, and productivity; (2) regulating water and gaseous flows; (3) retaining and processing nutrients, organic compounds, and pollutants; (4) resisting structural and physical degradation; (5) partitioning and processing energy; and (6) supporting aesthetic and cultural attributes. For soil survey inventories, the emphasis is on fundamental functions such as nutrient and elemental cycling, hydrologic functions, and providing a medium for plant growth (Tugel et al., 2008).

The Millennium Ecosystem Assessment (Hassan et al., 2005) provides a broad definition that does not explicitly identify soil functions but alludes to them as supporting services. The report defines ecosystem services, some of which are soil functions, as follows: "Ecosystem services are the benefits people obtain from ecosystems." Soil functions are ecosystem services that include (1) provisioning services such as food and water (products); (2) the regulating of floods, drought, land degradation, and disease; (3) cultural services such as recreational, spiritual, religious, and other nonmaterial benefits; and (4) supporting services such as soil formation and nutrient cycling, which are essential for the production of provisioning, regulating, and cultural services. In a discussion of the economic valuation of services provided by natural ecosystems, Daily (1997) specifically describes services provided by soil, such as the physical support of plants, and places a value on that service using costs of a replacement practice, such as hydroponics.

In landscape assessments, Tongway and Ludwig (1997) use the term "functional" to describe how landscapes capture, retain, and use water and nutrients. In the procedures described in *Interpreting Indicators of Rangeland Health*, soil functional status is evaluated for biotic integrity, hydrologic function, and soil and site stability (Pyke et al., 2002; Pellant et al., 2005).

Across these definitions of soil function, the consequences of change are paramount. Arnold et al. (1990) suggest that the significance of soil change depends mainly on its reversibility. Irreparable change can compromise the ability of land managers to enhance or maintain soil functional capacity, produce goods and services, and minimize adverse effects on the environment. Minimizing the negative consequences of change so that future management goals and soil functions are not compromised is an overarching goal for long-term soil conservation and ecosystem management and a primary concern with respect to soil degradation (Oldeman and van Lynden, 1997, Gerasimova et al., 2000). Inventorying and disseminating information about soil change and its consequences is a new challenge for soil survey.

38.5 Scientific Approaches to the Study of Soil Change

Because so many soil changes play out over years and decades (Table 38.1), special approaches are needed to quantify these dynamics. Although an understanding of soil change is vital for sustained land management, we cannot readily predict many soil changes with our current knowledge. Moreover, since soil changes result from high-order interactions of biology, chemistry, and physics, quantification of soil change requires direct observation of specific soils and their responses to management, at research sites over periods of decades. However, direct observation of changes over time in particular soils has been feasible for only a few soils. Five alternative and complimentary approaches are discussed: (1) long-term soil–ecosystem experiments (LTSEs), (2) short-term soil experiments, (3) space-for-time substitutions (SFTSs), (4) long-term resource monitoring (LTRM), and (5) computer models.

LTSEs are studies used to observe whole soil systems as they respond to controlled management regimes. To more fully understand the Earth's soil and its relations to the larger environment, there is no substitute for long-term observation, and repeated sampling, archiving, and analysis, in studies that pass from one generation of scientists to the next. Such foresighted research is challenging to establish and sustain. The first global inventory of such studies is being accumulated with descriptive meta-data on a real-time Web site (www.ltse.env.duke.edu).

Long-term experiments provide the base lines for understanding how and why soils change through time, for estimating rates, lags, and thresholds of change, and for documenting soil's resilience. However, long-running experiments are difficult to initiate and sustain; they require organization, data management, and unusual collaborations among scientists, even across generations. Long-term studies are susceptible to neglect or abandonment, and even productive LTSEs such as those described by Farina et al. (2000a, 2000b) in southern Africa can be summarily terminated simply due to an absence of interested scientists. Although LTSEs can be initiated with enthusiasm and the best of intentions, they may be terminated by lack of funding, shifts in research priorities, or societal instability.

Compared with complex experiments that require much labor and expense to maintain, long-term experiments that are labor efficient and straightforward in design are most likely to survive the tests of time. Even still, scientific returns from LTSEs may be few during their early years, and though challenging, the cycles of field work, samplings, archiving, and analyses must be balanced by scientific achievements. Long-term soil experiments face technical challenges as well (Steiner, 1995), as, for example, sampling soil with precision and accuracy is rarely easy and the statistical designs of many LTSEs can be questioned (Loughin, 2006). Some experimental treatments such as cultivation may eventually undermine the integrity of the experiment itself (Sibbesen, 1986). The representativeness of LTSEs is also not a trivial issue; for example, Debreczeni and Körschens (2003) estimated that >70% of the world's long-term field experiments are in Europe, and a recent web-based inventory of LTSEs (Richter et al., 2009) indicates that >80% of LTSEs test agricultural objectives and that >50% of these studies are located on relatively level Alfisols and Mollisols, many of which have remarkably high native fertility. Steiner (1995) suggested that the reason why so few long-term experiments test marginal soils was due to the unsustainability of such studies.

While LTSEs may be instrumental and necessary to the study of soil change they are by no means sufficient. Long-term studies need to be complemented by short-term soil–ecosystem experiments (STSEs) in the laboratory and field; by broadly focused SFTSs; by LTRM programs; and by computer modeling. Each is summarized in Tables 38.2 and 38.3 and discussed in detail in the following paragraphs.

STSEs in the lab and field include most studies of soils that are conducted up to a few years in duration. Nearly all of the sciences of soil physics, chemistry, and biology have been built from data arising from STSEs. Yet ironically, while STSEs often explore soil processes such as aggregation, adsorption, complexation, carbon dynamics, weathering, microbial activity, redox reactions, and soil fertility itself, extrapolating soil change over decadal timescales becomes a major challenge for STSEs. Small errors scaled across many years can readily bias long-term projections. Although STSEs can greatly enrich soil concepts and models, most are reductionist (isolating individual components

TABLE 38.2 Five Major Approaches in the Science and Management of Soil Change

Approach	Time Scale (Years)	Strengths	Limitations
STSEs	<1–10	Field or lab based, experimental control, versatile, short-term process	Extrapolation to larger scales of space and time, reductionist
LTSEs	>10	Field based, direct observation of whole soil system, experimental control, sample archive	Duration before useful data, vulnerable to loss, extrapolation to larger scales
SFTSs	>10–1000	Field based, highly time efficient	Temporal responses confounded entirely with spatial variation
LTRM	>10	Field based, direct observation, regional perspectives, sample archive	Complex planning and operational details, expense, no experimental control, duration before useful data
Computer models	<1 to >1000	Versatile, heuristic, and predictive, can support all other approaches	Dependent on quality of observational data

TABLE 38.3 Suitability of Measurement Approaches for Documenting Soil Change

	Measurement Approach			
Soil Change Attribute	STSE	LTSE	SFTS	LTRM
Variable at steady state (dynamic soil property)	R	S	M, R	S
Rate				
Short-term	F, R	F		F
Long-term		S	N	S
Fluctuation				
Short-term	F, R	F		F
Long-term		S		S
Trend, long-term		S		S
Pathway of change, feedback, thresholds, and hysteresis				
Short-term	F, R, S	F, S		F
Long-term		S	N	S
Resistance				
Short-term	R, S	S	M, R	S
Long-term		S	M, N	S
Resilience				
Short-term	R, S	S	M, R	S
Long-term		S	M, N	S

The three techniques, STSEs, LTSEs, and SFTSs, include experimental designs to facilitate comparisons among different treatments or kinds of land uses. LTRMs are not commonly designed to make comparisons among monitored sites or conditions but rather to simply track trends over time. Quantifying rates, pathways of change, and thresholds generally requires long-term studies or monitoring.

F, suitability depends on measurement frequency; M, suitability depends on management history at location sampled; N, suitability depends on the number of age classes sampled; R, suitability depends on characteristic response rate of the variable measured; S, suitability depends on time span over which repeat measurements are made. No entry indicates poorly suited.

and reactions), and do not examine the whole soil, complete with its high-order interactions and lag times that become apparent only with time in LTSEs. Even still, if STSEs are performed in conjunction with LTSEs, they can provide critical short-term process data important to the interpretation of longer-term soil change. Examples of short-term process data describe the seasonal dependence of nutrient mineralization, the depth-dependence of soil water chemistry, the rate at which annually produced plant-organic matter is harvested or returned to the soil as detritus, or management susceptibility of biologically mediated aggregate formation.

SFTSs, also called *chronosequence studies* or soil comparison studies, are used to efficiently examine temporal change in soil, ecosystems, and landscapes (Pickett, 1989; Hotchkiss et al., 2000). For questions of how soils change with time, SFTSs sidestep the great burden of LTSEs, the need for time itself to pass to directly acquire the information about soil change. One of the most well-known examples is that of Jenny's Mendocino Staircase in California, which is a set of marine terraces tectonically raised over the duration of the Pleistocene, with all soil-forming factors except time assumed to be similar (Jenny, 1980). The SFTSs are especially well suited for studies of change over geologic time and as such they are widely used to describe soil and ecosystem change over multimillennia (Jenny, 1980). SFTSs are, however, indirect in their inquiry, purposefully confound space and time, but thereby leave open the possibility for entirely faulty interpretation if misused. As an example, in a study of soil formation on mine reclaimed land, a poorly designed SFTS with improper site selection might over- or underestimate rates of soil formation on acidic mine spoils if recent soils are composed of parent materials with greater or lesser acidity than those used for older soils (Pickett, 1989). Many scientists are skeptical about the use of SFTSs (Gleason, 1927; Hotchkiss et al., 2000; Buol et al., 2003). On the other hand, only SFTSs can describe many soil changes that operate over many centuries and, if carefully designed to control for physical environment, are useful for documenting change in situations where repeating observations is not feasible (Ballantine and Schneider, 2009).

Carefully initiated comparison studies are being used in the soil survey programs of the Natural Resources Conservation Service to document land use and management effects on dynamic soil properties (Tugel et al., 2008). These studies apply the SFTS technique to make statements about change over time. They manage the limitations owing to variable historical conditions and inherent spatial variability of soil properties by restricting study sites to those with similar management history on closely similar soil and by replicating data collection across multiple spatial and temporal scales. They use state and transition models or other simple conceptual models of cause and effect to derive testable hypothesis and organize results. Selecting extensive soil map units for soil survey comparison studies as well as other SFTSs, LTSEs, and STSEs facilitates the extension of results across broad areas. LTRM programs are designed to document regional or national trends in soil resource condition over time. National monitoring programs generally use a random sampling procedure to select permanent monitoring sites where repeated observations of land use and resource condition are made at multiyear intervals. The tremendous cost of field sampling at all monitoring points is a major barrier to national scale soil monitoring programs. The Forest Inventory and Assessment reduced field sampling after initial efforts (Bechtold and Patterson, 2005) and the Natural Resources Inventory conducted field data collection for a few selected regions on an exploratory basis (Brejda et al., 2001). Vital Signs monitoring programs of the National Park Service were initiated in the early 2000s and to date have limited data to report (Fancy et al., 2009). Long-term monitoring is not limited to national scale monitoring programs. Some land managers monitor specific fields, pastures, timber stands, and other ecosystems as a part of their overall operations. Typical monitoring objectives will vary with the situation and include (1) evaluation and documentation of the progress toward management goals, (2) detection of changes that may be an early warning of future degradation, and (3) determination of trends for areas in desired condition, at risk, or with potential for recovery (Elzinga et al., 1998; Herrick et al., 2005; Madson et al., 2006).

Monitoring data produced by individual land managers are generally site specific in terms of objectives and properties measured and can be difficult to extrapolate. Compared with LTSEs, national resource monitoring programs quantify changes in soils not under experimental control but at randomly selected points across the landscape. Such LTRM programs thus have the potential to quantify soil change regionally as affected by shifts in regional land uses or other environmental conditions. But because management practices are not controlled, LTRM programs are challenged to interpret causes of observed dynamics. Causes of regional changes in soil carbon in two notable LTRM programs in England and Wales (Bellamy et al., 2005) and across Belgium (van Wesemael et al., 2004) have been controversial (Smith et al., 2007).

Computer models offer an approach to understanding soil change from the instantaneous to the multimillennial (Parton et al., 1987; Bouma and Hack-ten Broek, 1993; Pulleman et al., 2000; McBratney et al., 2002). The role and rationale of models are as much heuristic as they are predictive, as they represent refined hypotheses and depend on linkages with empirical studies. Models are instrumental for making progress in understanding soil change whatever the timescale, and are most convincing when simulation results are comparable with observational data. Whether observations are from LTSEs, STSEs, SFTSs, or LTMP's, empirical data are critical for gauging model competence and performance.

38.6 Forcing Factors, Process, Resistance, and Resilience

Understanding the causes of soil change and their effects on function requires interdisciplinary analysis at multiple spatial and temporal scales (Dent et al., 1996) as well as reductionistic basic research (Bouma, 1997). The challenge of studying open, dynamic systems is that there are few simple cause and effect relationships. Yet, motivating the study of soil change is the understanding that humanity is an integral part of the soil system, an agent affecting the flow through soils of energy, water, gases, and solid materials (Bidwell and Hole, 1965; Yaalon and Yaron, 1966), not simply an external player disturbing or perturbing the system (Goudie, 2005). Human activities impose a variety of forces on soil to affect the processes essential to the production of goods and services. This section provides a brief discussion on how forcing factors and processes involved in soil change are described by various disciplines. Also included are perspectives on the soil's ability to resist and recover from such forcings.

38.6.1 Forcing Factors

Forces driving ecosystem change are commonly described as forcing factors, stressors, disturbances, and perturbations. Although forcings often have negative connotations, their impacts can of course be positive or negative. Stressors and disturbances refer to factors that cause significant changes in ecological processes as well as the plant, soil, and animal components in ecological systems (Barrett et al., 1976; White, 1979). Stressors are forces that can be continuous, cyclic, or intermittent and include climate change, prolonged drought, agricultural impacts, grazing, fire suppression, and proliferation of invasive species. We use "disturbance" to represent relatively discrete events in time. Each occurrence of the following is an example: plowing, fertilization, irrigation, brush removal, fire, flooding, short-term drought, high-intensity storms, and wind storms. Stressors have a number of features in common. They can be either natural or anthropogenic and an integral feature of natural systems or a necessary operation in managed systems. Forcing factors can also produce direct impacts on small to large areas, which can lead to impacts on other areas. For example, climate change, the cumulative result of many forcing factors, will impact the soil system on large and small areas.

Disturbance regimes, such as recurring fire or flooding, or repeated tillage of agricultural fields are characteristic of many systems. A change in frequency or intensity of a forcing factor can place new stresses on a system. Five attributes are used to define disturbance regimes: disturbance type, spatial scale, intensity, frequency, and predictability (Herrick et al., 1999). A disturbance matrix can be constructed to help compare and understand the effect of disturbances on process (Table 38.4). In this example, fire, logging, and grazing remove aboveground vegetation that affects the supply of organic matter to the system. Logging, grazing, and vehicle traffic may compact the soil, thereby affecting runoff, water availability, soil aeration, and nutrient return to the system. In the Anthropocene, disturbance can also include an absence of fire, as fire has been removed from the landscape over many 100s of millions of hectares with major effects on soils and ecosystems.

The length of time required for a soil to change in response to outside forces is widely variable. It may take decades or longer for the cumulative effects of stressors to cause a functionally important change in soil. Discrete disturbance events, however, can precipitate almost immediate change such as the decrease in soil bulk density that occurs with the first tillage operation of the growing season (Pikul et al., 2006). Episodic events, such as hurricanes and drought, are events that may trigger a shift in systems that have experienced gradual change resulting from long-term management impacts (Scheffer et al., 2001). The combined effects of discrete events and longer-term stress can sometimes be devastating, such as those experienced in the Dust Bowl when continuous cultivation and prolonged drought were followed by intense wind storms.

38.6.2 Soil Processes

Predicting and managing the effects of human actions on soil function and ecosystem services require a detailed understanding of processes that operate within the pedosphere. However, there is tremendous complexity in the processes involved in soil change. Ross (1989) describes four increasingly complex and interlinked soil processes: (1) processes that form soil materials,

TABLE 38.4 Disturbance Matrix Illustrating a Classification of Disturbances and Their Effects on Soil Properties and Processes

		Effect			
				Nutrient Return	
Event	Pattern	Biomass Removal	Soil Compaction	Form	Distribution
Wildfire (hot, long-duration burn)	Continuous	All	Diffuse[a]	Mineral, black carbon	Concentrated and diffuse
Selective logging	Patchy	Woody	Linear and patches	Unprocessed and processed, organic	Depends on management of logging slash
Grazing (moderate)	Patchy	Herbaceous	Linear and patches	Mineral and processed organic	Discrete, concentrated (dung), diffuse (trampled vegetation)
Vehicle traffic (off-road)	Corridors	All within wheel tracks	Linear	Negligible within track	Negligible

Source: Modified from Herrick, J.E., and W.G. Whitford. 1999. Integrating soil processes into management: From microaggregates to macrocatchments, p. 91–95. *In* D. Eldridge and D. Freudenberger (eds.) People and rangelands: Building the future. Proc. Int. Rangeland Congr. Vol. 1. Townsville, Australia. July 19–23, 1999. International Rangeland Congress, Inc., Townsville, Australia.

[a] The effect here is indirect and maybe delayed: Fire may increase soil hydrophobicity, crusting, and compaction by removing protective vegetation and litter layers.

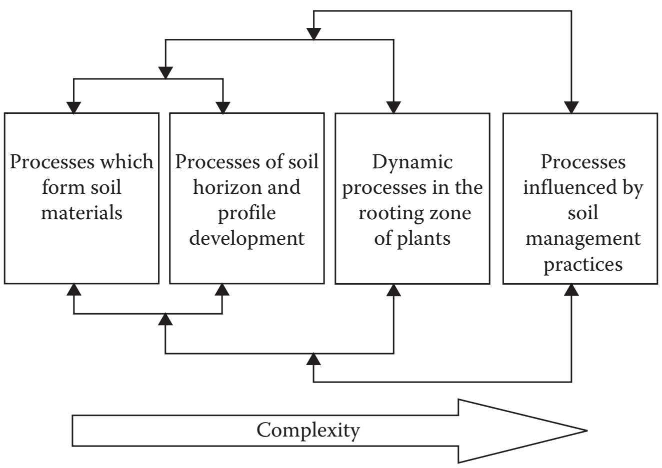

FIGURE 38.1 Processes involved in soil change. Soil systems function through many interacting processes. (Modified from Ross, S. 1989. Soil processes: A systematic approach. Routledge, London, U.K.)

(2) processes of soil horizon and profile development, (3) dynamic processes in the rooting zone of plants, and (4) processes influenced by management practices (Figure 38.1). Acknowledging that probably no soil process is uniquely biogeochemical, pedogenic, ecological, or geomorphic, it is important to draw upon many disciplines in our efforts to understand soil change. In the Anthropocene, it is important to join these traditions to enrich and broaden perspectives of soil and what has recently been called Earth's critical zone (National Research Council, 2001; Richter and Mobley, 2009). Pedology may focus on processes that alter soil material, geo- and biogeochemistry on chemical reactions, and ecology on system interactions, but the system is the same and the needs for understanding changes in the system justify a more integrated critical zone science (Wilding and Lin, 2006). The four primary processes of additions, losses, transformations, and translocations, originally formulated by the pedologist, Simonson (1959), are biogeochemical and ecological processes discussed in detail by Chadwick and Graham (2000) and Richter and Markewitz (2001). Distinctions between disciplines become blurred when considering biologically mediated processes such as the physical-displacement forces applied to soil by growing roots (Brimhall et al., 1991; Richter et al., 2007) or porosity-dependent, acid-induced transformations and weathering of soil minerals affected by soil microbes and plant respiration (Richter and Markewitz, 1995).

Processes related to forcing factors and plant growth are a primary focus in disciplines such as agronomy and silviculture, as well as the sciences of desertification and land degradation. Land use conversions and agricultural practices can quickly alter existing soil processes and even introduce new processes to the soil system (Ross, 1989). The alteration of soil physical and chemical properties and processes through plowing, drainage, and additions of new materials alter soil physical, chemical, and biological processes and primarily affect (1) soil structure, (2) porosity, (3) soil moisture and temperature dynamics, (4) heat flux, (5) soil trafficability and workability, and (6) nutrient storage and release. Tillage may control weeds, but it may also break down soil aggregates and promote soil compaction, processes that have generally negative consequences to hydrologic soil function and nutrient cycling.

Studies of soil and land degradation and desertification are concerned with processes that substantially decrease an area's biological productivity or usefulness to humanity. Soil degradation itself is defined as "a process that describes human-induced phenomena, which lower the current and/or future capacity of the soil to support human life" (Oldeman and van Lynden, 1997). Water erosion, wind erosion, salinization, and chemical, physical, and biological deterioration are among the many soil degradation processes enumerated in the literature and of grave concern with respect to the production of goods and services. The set of processes mapped in the Global Assessment of the Status of Human-induced Soil Degradation, while commonly cited, is not comprehensive and is limited to those processes occurring across extensive areas of the world (Oldeman et al., 1991; Chen et al., 2002; Lal et al., 2004).

Synthesis of process information from a wide range of disciplines is essential to understand and predict human impacts on soil systems over time. Palm et al. (2007) advocate identifying relationships between land management, degradation processes, and impacts of those processes. Positive or negative feedbacks among soils, plants, and animals, the atmosphere and climate, and management can together lead to soil change (Lawrence et al., 2007) and impact the functional capacity of soil at a number of spatial scales (Herrick and Whitford, 1999). Furthermore, a specific land management practice can trigger more than one type of soil degradation process and one type of impact (Stocking, 1995).

38.6.3 Resistance and Resilience

The response of soils to forces and stressors is controlled by soil's ability to absorb, resist, and recover from the forces. The concepts resistance, resilience, and buffering help describe soils' response. Ecological discussion over concepts of resistance and resilience is extensive and vigorous (Holling, 1973; Pimm, 1984; Gunderson, 2000; Briske et al., 2008; Bestelmeyer et al., 2009a), a discussion that extends to engineering and pedology (Greenland and Szablocs, 1994; Holling, 1996). From perspective of the soil, resistance is the capacity to maintain functional capacity through disturbances, and resilience is the capacity to recover the functional and structural integrity lost after a disturbance or prolonged stress (Blum, 1997; Seybold et al., 1999). Some concepts of resilience incorporate resistance to change as well as the capacity to recover (Holling, 1973; Szabolcs, 1994; Gunderson, 2000; Tenywa et al., 2006). Quantitative estimates of soil resilience are based on pre- and postconditions and computed using process rates (Szabolcs, 1994), dynamic soil properties (Seybold et al., 1999), or a combination of properties and processes (Lal, 1994). Soil resilience differs somewhat from ecological resilience, as ecological resilience involves multiple subsystems of the ecosystem (e.g., plant community, soil fauna) (Walker et al., 2004). Soil resilience is therefore a component of ecological resilience.

One of the interests in soil resilience is to estimate the possibility of reversing change in soil and soil function. Arnold et al. (1990) offer an approach for estimating reversibility that uses a ratio of rates of change that gives insight to the time frame required for reversal. From a land management perspective, the length of time required to reverse a change can have a significant influence on the perception of the ease of reversibility and its costs. Reversibility is characterized using rate or velocity of change and is classified on the basis of the ratio of the primary change process to the reverse process. If both the primary change and reversal processes are either rapid or slow, the system is considered reversible. This approach also illustrates that the reversal process is often different from the primary change process thus exhibiting pronounced hysteresis.

While few soil changes may be truly irreversible, reverse changes that may prove intractable include severe erosion, salinization without a suitable outlet for drainage water or salts, drainage and oxidation of sulfidic soil materials (iron pyrite, FeS_2), and heavy metal contamination.

38.7 Spatial and Temporal Patterns of Soil Change: Attributes for Prediction

The ability to predict how soils change over human timescales is critical to the success of sustainable soil management. Dynamic soil properties can be used to document change as well as quantify processes and soil functional capacity. Predicting change and its effect on processes and function requires observational data and models to describe and simulate temporal patterns exhibited by dynamic soil properties. Measuring temporal variation across space, however, as is required in LTSEs, STSEs, and SFTSs, is complicated by inherent and human-induced spatial variability of soil properties. Spatial and temporal variability of dynamic soil properties within systems are treated in this section.

38.7.1 Dynamic Soil Properties

When studying soil change within the Anthropocene, the variables of particular interest are dynamic soil properties (Tugel et al., 2005). Dynamic soil properties include use-dependent properties (Grossman et al., 2001). Many dynamic properties are indicative of soil function (see Chapter 26 of *Handbook of Soil Sciences: Resource Management and Environmental Impacts*) and recently soil survey document attainable levels and change in properties and function (Tugel et al., 2008).

Guidelines for selecting a minimum data set of dynamic soil properties for monitoring and soil survey comparison studies focus on functions of interest as well as ease of measurement, reproducibility, and cost (Larson and Pierce, 1991). Minimum soil-data sets typically will include bulk density, aggregate stability, infiltration, organic matter, pH, EC, exchangeable cations, extractable P, and potentially mineralizable N, which help characterize key soil functions and processes. Biological measures such as microbial biomass-C are also included where time of sampling can be controlled or seasonal fluctuations can be modeled.

In addition to land use and management, dynamic soil properties are affected by changes that are gradual, for example, a slowly warming climate or a prolonged drought, and that are oscillatory, for example, diurnal and seasonal regimes of temperature or moisture (Table 38.5). The dynamic soil properties' sensitivity to diurnal and seasonal environmental fluctuations may vary greatly and should be considered when selecting properties for inclusion in a study, especially SFTSs and LTRMs where seasonal measurements are not likely to be made. Repeated observations of properties sensitive to environmental fluctuations can be misinterpreted as management-related differences unless they are measured at the same time of day (diurnal sensitivity) or year (seasonal sensitivity). Environmental fluctuations

TABLE 38.5 Sensitivity of Soil Properties to Environmental Fluctuation in Temperature and Moisture

Sensitivity	
Diurnal	Seasonal
Soil temperature	Soil temperature
Composition of soil air	Moisture content
Redox potential	Bulk density, total porosity
	Aggregate stability
	Infiltration rate, hydraulic conductivity
	Bioavailability of macro- and micronutrients
	Microbial activity
	Electrical conductivity
	pH
	Redoximorphic features

Many soil properties respond to environmentally induced fluctuations in temperature and moisture. The environmental sensitivity should be considered when designing a study to ensure selection of suitable properties and the proper timing and frequency of measurement.

and anticipated changes in trends should be considered when designing a study to ensure proper selection of suitable properties, timing and frequency of measurement, and overall duration of study.

38.7.2 Spatial Patterns

The spatial variability of dynamic soil properties depends upon the type of soil, the plant community, the scale of disturbance, and the management history as well as the scale of measurement (Wilding et al., 1994). Legacies from past land use and management can persist in modern systems, such as tillage-related decreases in soil organic matter, contributing to spatial variability of dynamic soil properties (Foster et al., 2003; McLauchlan, 2006; Li et al., 2010). Understanding legacies and historic ranges of anthropogenic and nonanthropogenic variability is essential in interpreting modern changes in soil, although this knowledge does not include all of the information necessary for predicting future change (Millar and Woolfenden, 1999; Parsons et al., 1999).

The spatial variability of dynamic soil properties is often associated with spatial patterns of vegetation or local terrain, such as coppiced shrubs and concave intershrub spaces. Information about these patterns is important for the prediction of soil change and its effects on function. An example includes rows and furrows within a field and their effect on (1) salt redistribution within the soil after irrigation and (2) toxic salt damage to the crop (Wadleigh and Fireman, 1949; Miyamoto and Cruz, 1987; Ashraf and Saeed, 2006).

38.7.3 Temporal Patterns

Timescales over which dynamic soil properties change in response to land use and management vary from minutes to hundreds of years or more (Sparling, 2006; Richter, 2007). Fluctuation, trend, rate, pathway of change, feedback, hysteresis, thresholds, and fluxes are used to describe temporal patterns of soil change. Study objectives and the suitability of the various measurement techniques for documenting these attributes should be considered when initiating a project to document change. Detailed descriptions of the following attributes of change are located in Arnold et al. (1990) and Tugel et al. (2008).

38.7.3.1 Fluctuations and Trends

Fluctuation is temporal variation in soil properties and includes nonsystematic, random variation and regular periodic, cyclic variation.

Trend is the general direction of change and can be increasing, decreasing, or steady-state equilibrium. The time it takes to approach or a dynamic equilibrium varies with the property, the kind of soil, the type of management, and any continuing forcing factors.

38.7.3.2 Rate, Pathways of Change, and Feedbacks

Rates of soil change are rarely constant. Temporal variation in dynamic soil properties can follow pathways that are logistic or exponential and are uncommonly linear, owing to fluctuations and feedbacks. Feedback mechanisms are involved when the rate changes. Positive feedbacks intensify process rates, and negative feedbacks diminish or limit rates. Where feedbacks accelerate process rates, such as increasing rates of surface runoff and erosion over time, the feedback is positive, though the result may have a negative effect on soil function. Plant–soil feedbacks have been suggested to be strongest for plants growing in extreme environments (Ehrenfeld et al., 2005).

38.7.3.3 Thresholds

Thresholds in soils are tipping points and have application to many processes (Chadwick and Chorover, 2001). In ecosystem management, an ecological threshold represents the conditions at a point in time after which future management options become limited and corresponds to the shift from one alternative state to another (Hobbs and Harris, 2001; Bestelmeyer, 2006; Bestelmeyer et al., 2009b). Recognition of a threshold implies that a functionally important change in process rate has occurred.

38.7.3.4 Fluxes

Fluxes in soils can be energetic or material and occur via the gas, liquid, or solid phases. The importance of fluxes is underscored by Simonson's (1959) classical view of the processes of pedogenesis in which three of four of Simonson's general processes involved material and energy fluxes: inputs, removals, and translocations. Fluxes are measured as the rate of movement of chemical elements, or materials such as organic matter, clay minerals, or colloids. Fluxes are key to the functioning of the internal soil system and to how and why soil interacts with the wider environment. Much about soil management, in the past and in the future, involves the management of energetic and material fluxes.

Examples of how inputs, removals, and translocations have altered soils are far too few. The best examples of research and management sites that link changes in soil systems with fluxes of materials into, through, and out of the soil over decadal timescales are long-term soil experiments (Richter et al., 2007).

38.7.4 Reorganization in Space and Time

Changes in process rates, feedbacks, and thresholds are involved in pattern reorganization within systems. Because dynamic soil properties reflect processes, property–process–pattern relationships can be established. Furthermore, spatial patterns of soil properties and plant communities can shift and reorganize through time in response to stressors, disturbances, and vegetation dynamics (Bestelmeyer et al., 2006; Ravi et al., 2010).

Awareness of property–process–pattern reorganization is essential for prediction of future changes. In one example (Figure 38.2), heavy continuous grazing followed by drought may produce positive feedbacks between vegetation and soil properties that intensify physical, chemical, and biological degradation. The feedbacks lead to the following: (a) a decrease in soil organic matter and an increase in size of bare spaces, (b) a decrease in soil aggregate stability and reduced resistance to erosion, (c) a loss of topsoil through erosion and a decrease in infiltration, and (d) an additional loss of grass and increase in shrubs, which cause the feedback loop to continue (Bird et al., 2001). Eventually, a system will reach and cross a threshold to a new state with altered properties, processes, and patterns.

Of concern to sustained land and environmental management are postthreshold conditions such as reduced acid buffering capacity, impaired hydrologic processes, displaced plant communities, or different crop suitabilities when compared to the prethreshold states (Groffman et al., 2006). The management actions used to restore a postthreshold state may be different and more costly than those used to keep a system from crossing a threshold (Stringham et al., 2003). Once a threshold has been crossed, the limited reversibility of the resulting conditions may restrict future management options.

38.8 The Science and Management of Soil Change: Status and Future

About half of the approximately 13 billion ha of Earth's soil are now managed for human use: cultivated for crops; managed for pastures and hayfields; logged for wood; disturbed by mining; developed for urban, suburban, transportation, industrial, and recreational projects; and used to process burgeoning streams of human and animal wastes (FAO-STAT, 2009). Important areas are contaminated by chemical compounds and large areas lie in wait of conversion for use or reuse in the coming decades. All of these soils are also being affected by changing climates and increasing concentrations of atmospheric CO_2 and other greenhouse gases. The age of pedogenesis has given way to the age of anthropedogenesis.

Sustainably managing this change so that future management and soil function are not compromised is a goal well-worth bequesting to future generations. Creating this legacy is entirely contingent upon our ability to predict and improve management of soil change on the human timescale. Land managers and policy analysts need new soil survey products that provide information about soil's resistance to stress and disturbance, soil's ability to recover readily (resilience), and any prethreshold

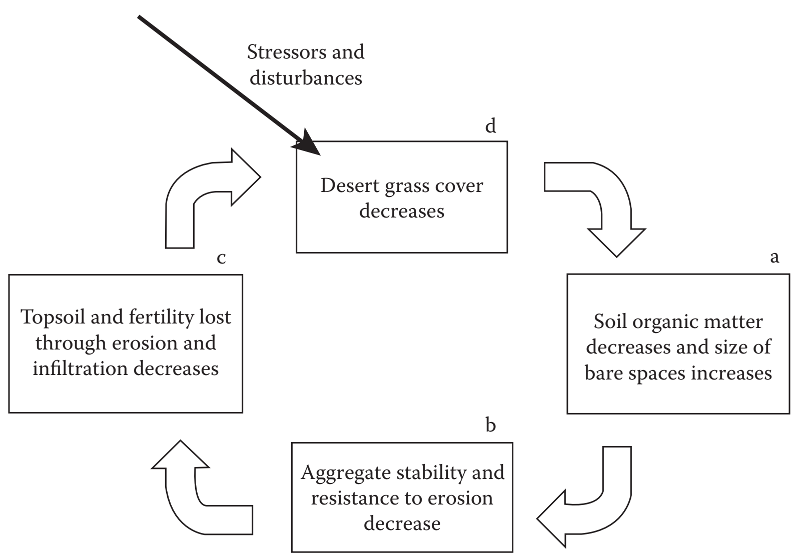

FIGURE 38.2 Effects of management on reorganization of soil property–process–pattern. Interactions between plant and soils contribute to the reorganization. Decreases in organic matter, aggregate stability, and resistance to erosion occur are associated with bare spaces that can continue to expand as degradation proceeds. (Modified from Bird, S.B., J.E. Herrick, and M.M. Wander. 2001. Exploiting heterogeneity of soil organic matter in rangelands: Benefits for carbon sequestration, p. 121–138. *In* R.J. Follett, J.M. Kimble, and R. Lal (eds.) The potential of U.S. grazing lands to sequester carbon and mitigate the greenhouse effect. CRC Press LLC, Boca Raton, FL.)

indices that warn about the loss of soil function. At present, our understanding of how soils are changing over decades' time and the potential prolonged effects of these changes on soil function can only be described as elementary (Richter and Markewitz, 2001; Tugel et al., 2005). Pedology and anthropedology must advance an interdisciplinary science of Earth's soil, that is, Earth's critical zone (National Research Council, 2001) in order to provide the foundation for long-term sustained soil and landscape management.

Pedologists in the Anthropocene (i.e., anthropedologists) use a science that is both basic and applied and has ever increasing interdisciplinary requirements to address sustainability and management challenges. Anthropedology must focus on management resilience of soils to help increase the production of goods and services without forcing systems across thresholds to undesired or difficult-to-reverse states (Bestelmeyer et al., 2009a). We suggest that improved management requires future soils research in five areas:

1. Pedogenic processes in human-altered systems
2. Soil changes and effects on functions and potentials for use
3. Mechanisms involved in soil change, degradation, and recovery
4. Prethreshold indices to predict difficult-to-reverse soil conditions
5. Restoration of highly degraded soil

The battery of approaches described in this chapter (LTSEs, STSEs, SFTSs, LTRM's, and models) can quantify soil change and improve sustainability of soil management. Many disciplines from the sciences and humanities will be needed for successful quantification and interpretation of humanity's effect on soil function.

Contemporary pedologists need to focus on the human element as a factor driving soil formation and transformation in the Anthropocene. It is humanity's transformation of Earth's soil that challenges scientists to develop a pedology with broad purview and decades' to centuries' timescales, and a pedology that supports the science, inventory, and management of the environment, ecosystems, and global change.

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39

Noninvasive Geophysical Methods Used in Soil Science

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39.1 Introduction

Geophysical methods measure changes in some physical property (i.e., density, seismic velocity, electrical conductivity, electrical resistivity [ER], magnetic susceptibility, dielectric permittivity) of the subsurface without direct access to the sampled volume (Daniels et al., 2003; Allred et al., 2008). Over the last three decades, the use of geophysical methods has significantly increased in soil investigations. The geophysical methods most commonly used in these pursuits include electromagnetic induction (EMI), ground-penetrating radar (GPR), transient EMI, galvanic resistivity, and magnetics. These methods are used to infer and better understand the spatial variability of soils and soil properties, and to guide observations and sampling. Selection of the most suitable geophysical method often requires an understanding of the soil properties that influence the method's response, and whether, and to what extent, a selected soil property affects the measured physical property (Allred et al., 2008). Whereas surface geophysical methods allow more continuous coverage than traditional approaches, they are limited in their capacity to resolve and characterize many pedologic features. Recent improvements in instrumentation and the integration of geophysical methods with other technologies (global positioning systems [GPS], data processing software, and surface mapping programs; e.g., geographic information systems [GIS], Geosoft, and Surfer) have fostered the expanded use of geophysical methods in soil investigations. The impetus has been the needs to improve quality control, provide more comprehensive coverage, and increase the efficiency of field operations. In agriculture, the three most commonly used geophysical methods are EMI, GPR, and Electrical resistivity (Allred et al., 2008). This chapter will focus on two of these geophysical methods: GPR and EMI. The initial uses, expansion, and present-day applications of GPR and EMI in agriculture have been summarized by Collins (2008) and Corwin (2008), respectively. As numbers of applications have become diverse and numerous, with a considerable amount of literature written on the uses of these geophysical methods in soil science, it is impossible for this brief discussion to cover all of these applications and papers.

39.2 Ground-Penetrating Radar

39.2.1 System Overview

39.2.1.1 Principles of Operation

Ground-penetrating radar is an impulse radar system designed for shallow (0–30 m) subsurface investigations. GPR systems transmit short pulses of very-high- and ultrahigh-frequency (from about 30 MHz to 1.2 GHz) electromagnetic energy into the ground from an antenna. Each pulse consists of a spectrum of frequencies distributed around the center frequency of the transmitting antenna. The transmitted pulses of energy are propagated downward into the soil. Whenever this wave of energy contacts an interface that separates layers with different permittivity (e.g., boundaries separating major soil, stratigraphic, and lithologic layers or features), a portion of this energy is reflected back to a receiving antenna. The more abrupt and contrasting the permittivity on opposing sides of an interface, the greater the amount of energy that is reflected back to the antenna. The receiving unit amplifies and samples the reflected energy and converts it into a similarly shaped waveform in a lower frequency range. The processed reflected waveforms are displayed on a video screen and can be stored on a hard disk for future playback and/or processing.

Ground-penetrating radar is a time scaled system. This system measures the time that it takes electromagnetic energy to travel from an antenna to a subsurface interface and back. To convert the travel time into a depth scale, the velocity of pulse propagation or the depth to a reflector must be known. The relationships among depth (D), two-way pulse travel time (T), and velocity of propagation (v) are described in the following equation (after Daniels, 2004):

$$v = 2\frac{D}{T}. \tag{39.1}$$

The velocity of propagation is principally affected by the relative permittivity (E_r) of the profiled material(s) according to the equation (after Daniels, 2004):

$$E_r = \left(\frac{C}{v}\right)^2. \tag{39.2}$$

In Equation 39.2, C is the velocity of light in a vacuum (0.3 m ns^{-1}). Velocity is often expressed in meters per nanosecond (m ns^{-1}).

Relative permittivity is a dimensionless constant that describes a material's capacity to store and release electromagnetic energy in the form of an electrical charge (Cassidy, 2009). Relative permittivity is expressed as (after Cassidy, 2009)

$$E_r = \frac{(E)}{(E_0)}. \tag{39.3}$$

where

E is the permittivity of the material

E_0 is the permittivity of free space (8.8542×10^{-12})

The relative permittivity varies with the amount of free and bound water (Cassidy, 2009). Relative permittivity ranges from 1 for air to 78–88 for water (Cassidy, 2009). Small increments in soil moisture can result in substantial increases in the relative permittivity of soils (Daniels, 2004). Using a 100 MHz antenna, Daniels (2004) observed that the relative permittivity of most dry mineral soil materials is between 2 and 10, while for most wet mineral soil materials, it is between 10 and 30. The relative permittivity of saturated peat deposits can range from 43 to 69 (Ulriksen, 1982). The relative dielectric permittivity of permafrost ranges from 2 to 8 (Cassidy, 2009). It should be noted that these reported values are simply approximations that were derived largely from laboratory experiments rather than real-world experiences (Cassidy, 2009). There is a tendency for permittivity to decrease with increasing frequency and conductivity to increase with increasing frequency (Annan, 2001).

39.2.1.2 Equipment

A typical GPR system consists of a radar control unit with transmitting and receiving antennas. The control unit consists of a screen, microprocessor, and mass storage medium. A microcomputer is used to control measurement processes, store data, and serve as a user interface. Modern GPR systems are lightweight and highly portable (Figure 39.1). Principal manufacturers of GPR include Geophysical Survey Systems Inc. (GSSI; Salem, New Hampshire); MALÅ (Stockholm, Sweden); and Software & Sensors (Mississauga, Ontario).

The most common mode of GPR data acquisition is the reflection profiling mode (Figure 39.1) in which radar waves are transmitted, received, and recorded as the antenna is moved along the soil surface. Transillumination methods are less commonly applied in soil investigations and require boreholes. In this method, the transmitter and receiver are placed and moved through two boreholes placed on opposite sides of the measured medium.

Based on GPR systems design and user preference, data are collected in either a continuous or step mode. The continuous mode involves the unidirectional movement of both transmitting and

(a)

(b)

FIGURE 39.1 Radar traverses being conducted in the continuous mode with a GSSI SIR-3000 system and a 200 MHz antenna (a) and a Software & Sensors Noggin system and a 250 MHz antenna (b).

receiving antennas (usually housed in one container or bound together) along a traverse or grid line (Figure 39.1). This mode is well suited to the rapid collection of data over relatively large areas. In the step mode, transmitting and receiving antennas are placed on the soil surface at a set distance apart (constant offset), a scan or multiple scans are obtained, and the antennas are then moved a short measured distance along the traverse line to repeat the process. This method is rather slow and cumbersome. However, averaging multiple scans collected at each measurement point results in a reduction of background noise.

Multi-offset methods are used with some GPR systems. These methods involve varying the distance between the transmitting and receiving antennas. Multi-offset methods are used to obtain estimates of the radar's velocity of propagation versus depth in the ground (Annan, 2009). Two commonly used multi-offset methods are common midpoint (CMP) and wide-angle reflection and refraction (WARR). In CMP, the distance between the antennas is increased stepwise, while the same midpoint position is maintained between the two antennas. In WARR acquisition, the distance between the two antennas is increased stepwise while the transmitter is maintained at a fixed location.

39.2.1.3 Antennas

Antennas are used to transmit energy, receive energy, or both. The resolution and penetration depth of GPR are determined by the antenna frequency and the electrical properties of earthen materials (Olhoeft, 1998; Daniels, 2004). A major constraint of GPR is the relationships between antenna frequency, penetration depth and resolution; resolution will decrease with decreasing antenna center frequency and with increasing depth of penetration (Buynevich et al., 2009). Higher-frequency (>400 MHz) antennas provide higher resolution, but have less penetration depth than lower frequency antennas (<300 MHz). The antennas most commonly used in soil investigations have center frequencies between 100 and 500 MHz. Lower frequency antennas (100–300 MHz) are commonly used for water table, lithologic, and stratigraphic studies, and for investigations in areas of more conductive and attenuating soil materials. Higher-frequency antennas (400–500 MHz) provide better results in relatively dry, electrically resistive soils. Antennas with frequencies of 900 MHz–1.5 GHz have been used for shallow investigations in some coarser-textured soils. For organic soils, where greater depths of penetration are often desired to profile the organic/mineral soil interface, lower frequency (70–200 MHz) antennas are commonly used.

39.2.1.4 Field Methods

For most GPR soil investigations, a traverse line or small grid is established across the study site. Generally, more information is obtained by using a network of traverse lines that capture the extent and variability of GPR reflections (Buynevich et al., 2009). Typically, reference points are located at uniform intervals along traverse or grid lines. The spacing of these reference points varies with the anticipated complexity of soils and soil properties, and can range from several centimeters to several tens of meters. The number of reference points depends on the feature of interest and the length of the GPR traverse line.

Typically, for soil survey investigations, an appropriate antenna is moved along the ground surface in the continuous mode (Figure 39.1a). On relatively smooth surfaces that are free of debris or other hindrances, a survey wheel can be attached or the antenna placed in a cart with an odometer wheel (Figure 39.1b). Trigger mechanisms on the survey wheel or cart control the start and stop of data recording and provide reliable distance marks (Buynevich et al., 2009). At uniform distances along the traverse line, the operator or the trigger mechanism impresses a mark on the radar record indicating a reference or distance point.

After reviewing the radar record in the field, soils are cored, described, and sampled at few selected reference points along the traverse line to verify GPR depth measurements and interpretations. Based on these observations, subsurface horizons, contrasting layers, and/or features are identified on radar records. For soil survey investigations, the presence and depth to subsurface horizons or features are used to identify soils on the radar record. Later, during the post-surveying analysis of the radar records, the depths to a soil interface of interest are interpreted and manually or semiautomatically entered into a spreadsheet. Basic statistics and descriptive summaries for each traverse line are developed and presented as survey results.

39.2.1.5 GPS Option

During the 1990s, it was realized that, in order to be more useful, GPR data needed to be integrated with available digital soil data and maps. This has led to the integration of GPR with GPS. As noted by Rial et al. (2005), the accurate positioning of radar data with GPS and its importation into GIS is a goal, which will shortly become a perquisite in GPR investigations. Modules have been developed in newer GPR systems and processing software that permits the georeferencing of GPR data. These modules not only visually georeference the GPR data (GSSI, 2008), but have widened the scope of GPR surveys (Gustafsson, 2007).

With GPS, as the radar is moved across a soil polygon, its position is continuously tracked. The number of potential reference points is determined by the speed of advance and the sampling rate set on the radar unit. During post-surveying analysis of radar data, the position of each radar scan is proportionally adjusted according to the time stamp of the two nearest positions recorded with the GPS receiver.

39.2.1.6 Soil Properties That Influence the Effectiveness and Penetration Depth of GPR

Compared with other geophysical techniques, GPR provides the highest resolution of subsurface features. However, GPR is not appropriate for use on all soils (Doolittle, 1987). There are a number of factors that may increase dissipation of radar energy passing through soils including electrical conductivity, water content, soluble salts, carbonate minerals, and gypsum (Campbell, 1990; Olhoeft, 1998; Daniels, 2004). Soils having high electrical conductivity rapidly attenuate radar energy, restrict penetration depths, and severely limit the effectiveness of GPR.

Radar energy dissipation increases with increasing soil water content (SWC) (Campbell, 1990; Daniels, 2004). When an alternating electrical field is applied, water molecules align their permanent dipole moments parallel to the direction of the applied electrical field (Daniels, 2004). The displacement of bound water molecules results in the loss of energy as heat (Neal, 2004) and storage of electrical field energy. At frequencies above 500 MHz, the absorption of energy by water is the principal mechanism for radar energy loss in soils (Daniels, 2004). Even under very dry conditions, the amount of bound water is sufficient to affect radar energy loss in soils (Daniels, 2004).

Electrical conductivity and energy loss are also affected by the concentration of salts in the soil solution (Curtis, 2001). All soil solutions contain some salts, which act as electrolytes and increase ionic conduction. In semiarid and arid regions, however, accumulations of soluble and sparingly soluble salts of K, Na, Ca, and Mg in soils increase the electrical conductivity of the soil solution and the attenuation of electromagnetic energy (Doolittle and Collins, 1995). Because of their high electrical conductivity, saline (electrical conductivity >4 dS m^{-1}) and sodic (sodium absorption ratio ≥13) soils are considered unsuited to most GPR applications. In saline soils, depending on moisture contents, penetration depths typically range from a few to less than 25 cm (Daniels, 2004; Ben-Dor et al., 2009) and GPR is considered an inappropriate tool.

Mukherjee et al. (2010) observed that the dielectric permittivity and electrical conductivity of karstified limestone increased with increasing porosity. Because of increased signal attenuation in saturated carbonate terrains, GPR penetration depths are restricted (Mukherjee et al., 2010). At similar moisture and clay contents, soils with high $CaCO_3$ equivalent have been reported to reduce the depth of GPR penetration in soils (Grant and Schultz, 1994; Lebron et al., 2004).

Because of their high adsorptive capacity for water and exchangeable cations, the penetration depth of GPR is inversely related to clay content. Olhoeft (1986), using a 100 MHz antenna, observed a penetration depth of about 30 m in clay-free sands. However, the addition of only 5% (by weight) smectitic clays reduced the penetration depth by a factor of 20 (Olhoeft, 1986). Doolittle and Collins (1998) noted that, depending on antenna frequency and the specific conductance of the soil solution, GPR penetration depths range from 5 to 30 m in dry, sandy soils, but average less than 50 cm in wet, clayey soils.

Electrical conductivity and energy loss increase with increasing cation exchange capacity (CEC) of the clay fraction (Saarenketo, 1998). Cations adsorbed to the clay particles provide an alternative pathway for electrical conduction and, therefore, contribute to electromagnetic energy losses (Saarenketo, 1998). Soils with clay fractions dominated by high-CEC clays (e.g., smectite and vermiculite) are more attenuating to GPR than soils with an equivalent percentage of low-CEC clays (e.g., kaolinite, gibbsite, and halloysite). For soils with comparable clay and moisture contents, greater depths of penetration can be achieved in highly weathered soils of tropical and subtropical regions than in soils of temperate regions. Additionally, the contribution of clay and surface conduction to electrical conductivity and electromagnetic energy loss is more noticeable in soils that have low concentrations of soluble salts (Klein and Santamarina, 2003).

39.3 GPR Soil Suitability Maps

Knowledge of anticipated depths of penetration is important for the effective use of GPR. Most radar users have limited knowledge of soils and are unable to predict the likely penetration depths or the relative suitability of soils for GPR within study sites. In many soils, high rates of signal attenuation severely restrict penetration depths and limit the suitability of GPR for a large number of applications. In saline and alkali (or sodic) soils, where penetration depths are typically less than 25 cm (Daniels, 2004), GPR is unsuited for most applications. In wet clays, where penetration depths are typically less than 50–100 cm (Doolittle et al., 2003), GPR has very low potential for most applications. However, GPR is highly suited to most applications in dry sands, where penetration depths can exceed 50 m with low frequency antennas (Smith and Jol, 1995).

A map of effective ground conductivity (Fine, 1954) provided broad guidance on anticipated rates of signal attenuation, GPR penetration depths, and the relative suitability of soils to GPR applications within the United States. Because this map was prepared at a small scale (1:2,500,000) and from a limited sample population, resolution was poor and data were generalized, which resulted in many discrepancies between predicted and actual usefulness of GPR at specific sites.

Collins (1992) developed a GPR suitability rating and created GPR suitability maps based on soil classification and soil properties within the upper 2 m. Doolittle et al. (1998) used the USDA–Natural Resources Conservation Services (NRCS) State Soil Geographic (STATSGO) database to create maps that summarized the distribution of several soil properties affecting GPR suitability. Subsequently, these general soil spatial and attribute data were used to develop and later revise the GPR Soil Suitability Map of the Conterminous United States (GSSM-USA) (Figure 39.2; Doolittle et al., 2002b, 2003, 2007), which was based on a larger sample population and offered greater detail than the effective ground conductivity map (Fine, 1954).

Soil attributes used to determine the GPR suitability indices of soil polygons shown on the GSSM-USA include taxonomic criteria, clay content and mineralogy, electrical conductivity, sodium absorption ratio, and $CaCO_3$ and $CaSO_4$ contents. Each soil attribute for each horizon is rated and assigned an index value ranging from 1 to 6. Lower index values are associated with lower rates of signal attenuation and greater penetration depths. For each soil attribute, the most limiting index value within 2 m for each soil in the map unit is selected, and these limiting indices are summed for each soil (the component index value). For each soil map unit, the relative area of soils with the same index values is summed, and the GPR suitability index (GSI) for the soil map unit is the most aerially extensive index value in the map unit.

FIGURE 39.2 **(See color insert.)** The GSSM-USA.

Soil attribute index values and relative soil suitability indices are based on observed responses from antennas with center frequencies between 100 and 200 MHz. For mineral soils, the inferred suitability indices are based on unsaturated conditions and the absence of unpredicted, contrasting soil properties within 2 m. Within any map delineation, the actual performance of GPR will depend on variations in soil properties, the type of application, and the characteristics of the subsurface target. Penetration depths and the relative suitability of soils will be less under saturated conditions, and if shallow groundwater is present, GPR penetration will be reduced and contingent on the ions present in solution.

According to GSSM-USA, only 22% of the soils in the conterminous United States are considered well suited to GPR. Areas of moderate and low potential soils for GPR applications are more extensive and occupy an estimated 33% and 36% of the acreage, respectively. Because of saline and alkaline conditions, 7% of the soils in the conterminous United States are considered unsuited to GPR. As evident from this data, a majority of the soils in the conterminous United States are not well suited for GPR soil investigations, and most successful applications of GPR in the United States has been restricted to regions that have extensive areas of soils that are well suited to GPR.

Because of the small compilation scale (1:250,000) of GSSM-USA, the minimum polygon size is about 625 ha. Larger scale GPR soil suitability maps have been prepared on a state basis using the USDA–NRCS's Soil Survey Geographic (SSURGO) database (Doolittle et al., 2006b), which is directly derived from soil surveys completed at scales ranging from 1:12,000 to 1:63,360 (minimum delineation size ranging from about 0.6 to 16.2 ha, respectively) (Soil Survey Division Staff, 1993). The same soil properties, attribute index values, and aggregation methods used to prepare GSSM-USA were used to derive these larger scale maps.

An example of these large scale maps, the GPR Soil Suitability Map of Nebraska (GPRSSM-NE), is shown in Figure 39.3. Compared with GSSM-USA (Figure 39.2), soil information contained on GPRSSM-NE (Figure 39.3) is less generalized, soil patterns are more intricate, and polygons are shown with greater resolution. Broad spatial patterns, which correspond to major soil and physiographic units within Nebraska, are evident on both maps (Figures 39.2 and 39.3). However, GPRSSM-NE provides a more detailed overview of the spatial distribution of soil properties that influence the depth of penetration and effectiveness of GPR.

The spatial information contained on GPR soil suitability maps can aid evaluations of the relative appropriateness of using GPR, selection of the most suitable antennas and survey procedures, and the need and level of data processing. GPR soil suitability maps are available for most states and can be accessed at http://soils.usda.gov/survey/geography/maps/GPR/index.html. These maps are periodically updated as additional soil information is collected and certified.

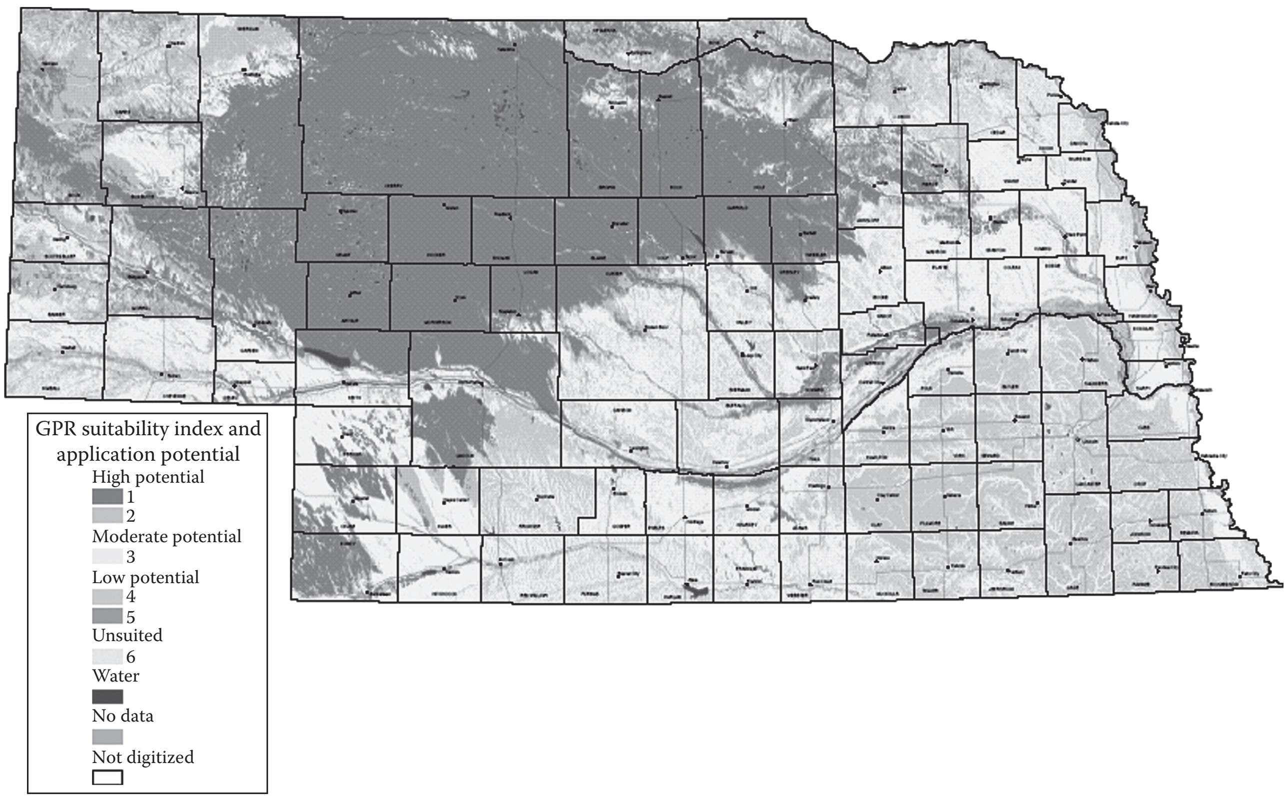


FIGURE 39.3 **(See color insert.)** The State GPRSSM-NE.

39.3.1 Factors Influencing Feature Detection

39.3.1.1 Reflection Coefficient

The amount of energy reflected back to an antenna is a function of the dielectric gradient that exists across a soil interface or boundary. The greater and more abrupt the contrast in the dielectric properties of adjoining soil materials, the greater the amount of energy reflected back to the antenna, and the more intense and conspicuous the amplitude of the reflected signal appearing on radar records. Soil horizons, layers, and features that have similar relative permittivity are poor reflectors of electromagnetic energy and are difficult to identify on radar records. The reflection coefficient, R, is a measure used to express the difference in relative permittivity that exists between two adjoining materials. The reflection coefficient is proportional to reflection strength and is expressed as (after Neal, 2004)

$$R = \frac{\sqrt{E_{r2}} - \sqrt{E_{r1}}}{\sqrt{E_{r2}} + \sqrt{E_{r1}}}, \tag{39.4}$$

where E_{r1} and E_{r2} are the relative permittivity of adjoining materials 1 and 2. As evident in Equation 39.4, R is dependent on the difference in the relative permittivity between the two adjoining materials.

The E_r values of most dry and wet, mineral soil materials range from 2 to 10 and from 10 to 30, respectively (Daniels, 2004). The E_r of soil materials is strongly dependent upon moisture content, and the amount of energy reflected back to the radar's antenna is greatly influenced by the abruptness and difference in moisture content between soil horizons, layers, or features.

Typically, strong radar reflections (high-amplitude reflections) are produced by abrupt soil interfaces that separate contrasting soil materials and often correspond to boundaries between soil horizons. Differences between horizons that result in high-amplitude reflections commonly are associated with differences in moisture contents, physical (grain size, texture, bulk density), and/or chemical (organic carbon, calcium carbonate, sesquioxides) properties of the horizons. GPR has been used to determine the depth to bedrock, contrasting master (E, B, C, and R) horizons, buried genetic horizons, frozen soil layers, illuvial accumulations of clays or organic matter, and cemented or indurated horizons. GPR does not detect subtle changes in soil properties (e.g., color, mottles, structure, porosity, slight changes in texture), transitional horizons (e.g., AB, AC, BC), or most vertical divisions in master horizons.

GPR does not directly measure the water table depth, but responds to near-saturated conditions within or near the top of the capillary fringe associated with a water table (Smith et al., 1992; Bentley and Trenholm, 2002). As the width of the capillary fringe above a water table increases, reflections from the water table have increasingly lower amplitudes, more dispersed characteristics and are, therefore, less distinguishable on radar records (Annan et al., 1991). In coarse-textured soils, the capillary fringe is narrow, the difference in permittivity between the unsaturated and saturated zones is abrupt and contrasting, and the water table is often distinguishable on radar records. The more gradual change in water content over greater distance in clayey soils results in less distinct radar reflections.

39.3.1.2 Resolution

Resolution is the ability to detect two closely spaced interfaces or distinguish a subsurface feature. In order to be resolved, the thickness of a layer must be at least the same dimension as the wavelength, or the travel time through the layer must be equal to or greater than the pulse duration (Annan, 2001). Horizontal or lateral resolution depends on the velocity of propagation, the antenna's bandwidth, and the distance from the antenna. As radar waves propagate through the soil, they expand to form a conical "footprint" area. Consequently, the more distant a layer or feature is from the antenna, the wider the footprint area and the lower the horizontal resolution (Annan, 2001). Higher-frequency antennas have larger bandwidths and therefore provide greater horizontal resolution than lower frequency antennas (Annan, 2009).

The vertical resolution is dependent on the propagated wavelength (λ), which is determined by dividing the propagation velocity (v) by the antenna frequency (f) (after Daniels, 2004):

$$\lambda = \frac{v}{f}. \tag{39.5}$$

In general, vertical resolution is considered to be about one-fourth the wavelength of the radar signal. If two features are separated in time by less than this amount, they will be indistinguishable on radar records and interpreted as one. However, as noted by Moorman and Michel (1997), the detection limit is considerably lower than the theoretical vertical resolution. Annan (2009) observed that two features must be separated in time by at least one-half of the wavelength to be identified. As can be seen in Equation 39.5, lower frequency antennas have longer wavelengths and therefore afford lower vertical resolution than higher-frequency antennas.

39.3.1.3 Scattering losses

Scattering losses are frequency dependent and become significant at high radar frequencies (Annan, 2009). Scattering is necessary for the detection of major interfaces or features. However, excessive scattering, caused by small-scale heterogeneities or scattering bodies (e.g., rock fragments, tree roots, animal burrows, and cultural features or debris) in the soil, results in unwanted background noise that impairs interpretations. Large numbers of scattering bodies confound interpretations by producing numerous, undesired, subsurface reflections, which clutter radar records and mask or obscure the signal from soil features of interest. Additionally, scattering losses may be the most significant source of signal attenuation restricting penetration depths (Annan, 2001).

39.3.2 Display and Interpretation

39.3.2.1 Two-Dimensional Display

Radar records are similar in appearance to sonar or seismic-reflection profiles. By moving an antenna along the soil surface, a large number of radar scans (or traces) are recorded at a fixed rate or intervals (time or distance) producing a 2D radar record. On 2D radar records, the horizontal scale is distance, which is based on the rate of antenna advance or the distance traveled. The vertical scale is based on the round-trip travel time of the electromagnetic wave. Radar records are displayed as either line-scan or wiggle-trace displays (Figure 39.4b and a, respectively). Wiggle traces are preferred by most engineers and geologists who are familiar with oscilloscope displays. Line-scan displays assign colors or color intensity to the amplitude range on the traces. Line-scan displays are most frequently used for GPR data because of the high data volume (number of traces) (Daniels et al., 2008).

Figure 39.4 contains two displays of the same 2D radar record, which was collected in an area of Bridgehampton soil (coarse-silty, mixed, active, mesic Typic Dystrudepts) in Rhode Island. Bridgehampton soils formed in thick, silty aeolian deposits that are underlain by stratified sands and gravels. In Figure 39.4, the horizontal scale is expressed in meters and is based on measured distances along the traverse line. The vertical scale is based on a velocity of propagation that was determined by comparing the measured depth of a feature with the two-way travel time to the subsurface interface (e.g., the contact of the silty aeolian mantle with the underlying coarse-textured outwash) and Equation 39.1. In the renditions shown in Figure 39.4, the aeolian mantle is virtually free of reflectors. The uppermost, high-amplitude subsurface reflector represents the aeolian mantle/outwash interface. The underlying outwash is characterized by linear reflectors of varying amplitudes. In the two renditions of this 2D radar record, the aeolian mantle/outwash interface is punctuated by an incised feature (between the 6 and 8 m distance marks) whose form and expression suggests an ice-wedge cast.

39.3.2.2 Three-Dimensional Display

The effective visualization of radar data is the key to modern GPR interpretations. An emerging approach in GPR is the analysis of subsurface structures and geometries from a 3D perspective. Three-dimensional GPR allows the visualization of data volumes from different perspectives and cross sections (Beres et al., 1999), which can assist identification, characterize subsurface soil, stratigraphic and lithologic structures and geometries, and improve interpretations of subsurface features. In areas of electrically resistive materials, 3D GPR can provide unrivaled resolution and detail of subsurface features as compared with 2D GPR (Grasmueck and Green, 1996). The use of 3D GPR has allowed improved definition of subsurface features and resulted in more complete and less ambiguous interpretations than traditional 2D GPR (Beres et al., 1999).

The acquisition of data for 3D GPR requires greater expenditures of time and labor than 2D GPR. However, the additional expenditures of resources needed to collect, process, and visualize

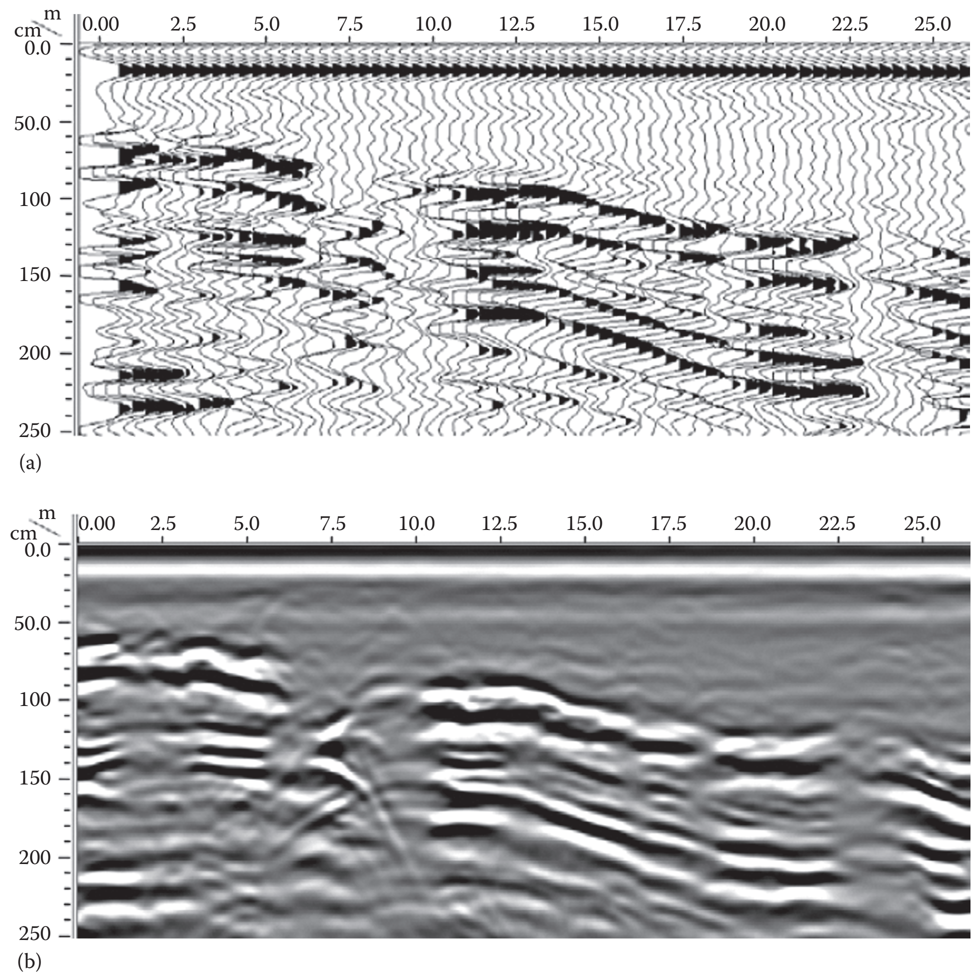


FIGURE 39.4 Two examples of the same 2D radar record, which was collected with a 200 MHz antenna in an area of Bridgehampton soils. The radar record is displayed as a (a) wiggle-trace and (b) a line-scan display.

3D GPR data are often compensated by the more comprehensive spatial coverage and greater resolution of subsurface features (Grasmueck and Green, 1996). Important considerations for collecting and interpreting 3D GPR surveys are discussed by Lehmann and Green (1999). The basic process to construct a 3D GPR pseudo-images of the subsurface is that a relatively small area (typically, 1–2500 m^2) is intensively surveyed with multiple, closely spaced (typically, 0.1–1.0 m), parallel GPR traverse lines. This relatively dense set of grid lines is necessary to resolve the geometries and sizes of different subsurface features and to minimize spatially aliasing the data (Grasmueck and Green, 1996). Appropriate computer software allows the 3D GPR pseudo-image to be viewed from nearly any perspective (Junck and Jol, 2000) and arbitrary cross sections, insets, and time slices can be extracted from the 3D data set. In addition, animated imaging capability allows users to interactively examine the entire data volume (Grasmueck, 1996).

Three-dimensional GPR has been used to identify the presence and map the geometries of subsurface features in both consolidated and unconsolidated materials. Recently, 3D GPR was used to investigate the subsurface configuration of ice-wedge polygons near Barrow, Alaska (Munroe et al., 2007), and sediment filled wedges and buried polygonal ground in mid-latitude United States (Doolittle and Nelson, 2009).

Figure 39.5 contains a 2D radar record and two 3D GPR pseudo-images from a small (30 × 30 m) grid site located in an area of Delton soils (loamy, mixed, active, mesic Arenic Hapludalfs) in Wisconsin. Delton soils formed in a thin (50–100 cm) mantle of aeolian sands overlying clayey lacustrine deposits. On the 2D radar record (Figure 39.5a), the topography of the contact between the aeolian sands and underlying lacustrine deposits (highlighted with a segmented, line) appears wavy with two conspicuous concavities. On the two 3D GPR pseudo-images, a solid cube (Figure 39.5b) and a cube with a 23 × 29 m inset removed to a depth of 100 cm (Figure 39.5c) are shown. Along the sidewalls of the solid 3D GPR pseudo-image, the interface separating the sandy aeolian mantle from the underlying clayey lacustrine deposits is clearly expressed. The 3D GPR pseudo-image with the inset cube removed shows a series of linear and intersecting, high-amplitude (colored white) reflections on the base of the cutout cube. The geometry of these reflectors suggests buried polygonal ground created in a former periglacial environment. Closed segments in the lineations resulted from finer

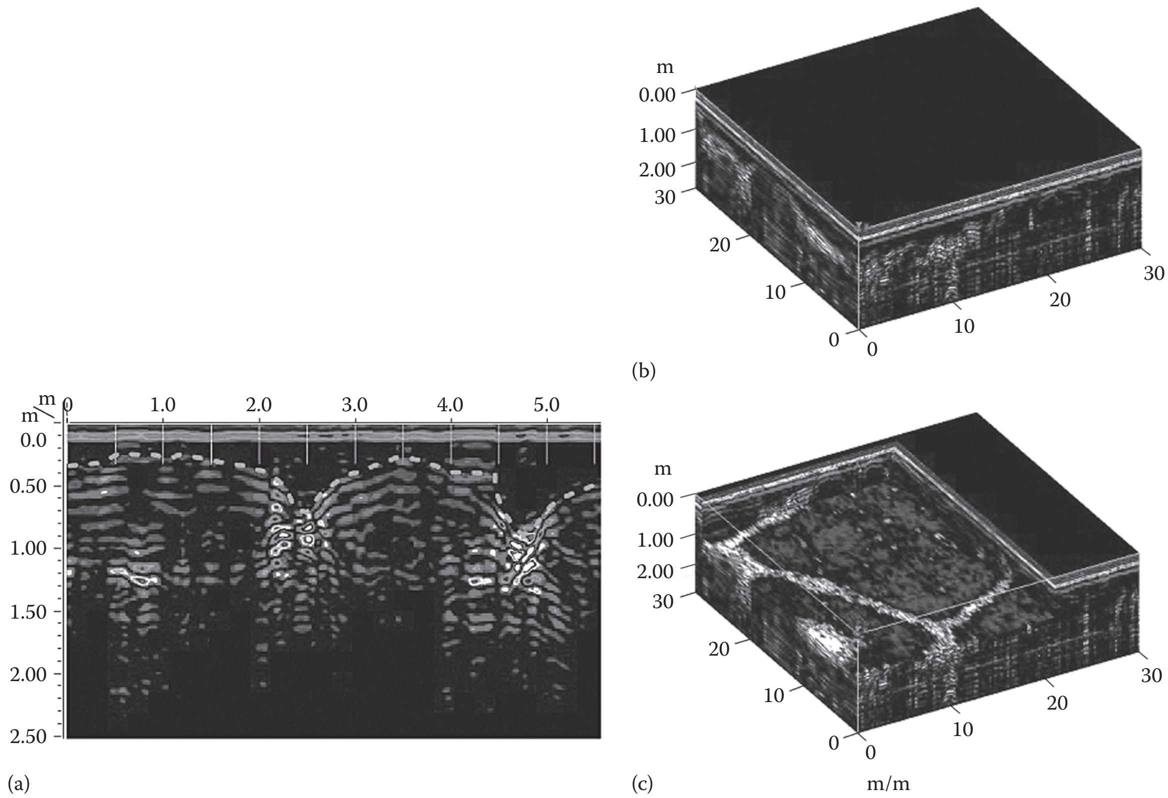


FIGURE 39.5 A 2D radar record and 3D GPR pseudo-images from a 30 × 30 m grid site in Wisconsin. (a) The segmented line in the 2D radar record shows the contact separating a sand mantle from the underlying clayey lacustrine deposits. (b and c). The 3D pseudo-images show the geometry of the ice-wedge pseudomorphs. (From Doolittle, J.A., and F.E. Nelson. 2009. Characterizing relict cryogenic macrostructures in mid-latitude areas of the USA with three-dimensional ground-penetrating radar. Permafrost Periglac. Process. 20:257–268.)

textured, more radar opaque material filling the ice-wedge voids. These closed segments were interpreted to have formed as portions of the finer-textured, ice-rich lacustrine deposits, which were susceptible to consolidation and deformation, closed during thawing and prevented infilling with superjacent sands.

39.3.3 Applications

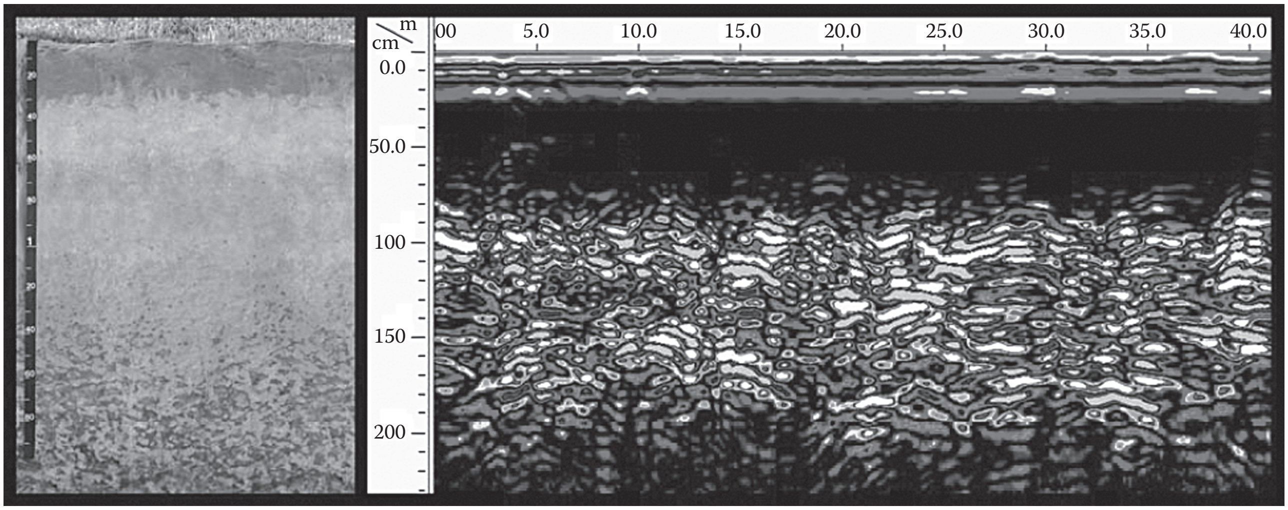
39.3.3.1 Use of GPR in Soil Survey

Ground-penetrating radar provides data on the presence, depth, lateral extent, and variability of many diagnostic subsurface horizons that are used to classify soils (Collins et al., 1986; Doolittle, 1987; Schellentrager et al., 1988; Puckett et al., 1990). In the United States, GPR is principally used as a quality control tool to verify the taxonomic composition of soil map units, document the presence and depth to soil horizons and features, and assess spatial and temporal variations in soil properties.

GPR was first used in Florida in 1979 to identify and determine depths to diagnostic subsurface horizons used to classify and map soils (Benson and Glaccum, 1979; Johnson et al., 1979; Collins, 2008). Since that time, GPR has been used extensively for soil survey quality control in Florida because of the ubiquity of sandy soils with favorable physiochemical properties for electromagnetic wave propagation and the presence of contrasting soil horizons (Schellentrager et al., 1988).

Johnson et al. (1979) working in sandy soils with well-expressed horizons, observed that radar interpreted depths of selected horizon boundaries were within ±2.5–5.0 cm of the measured depths. Asmussen et al. (1986) observed an average difference of 19.2 cm between the radar interpreted and measured depths to argillic (Bt) horizons, which ranged in depth from about 20 to 450 cm. Rebertus et al. (1989) observed that the difference between the interpreted and measured depths to a discontinuity, which ranged in depth from 0 to about 230 cm, was less than 15 cm in 94% of the observations. Collins et al. (1989) observed an average difference of 6 cm between the interpreted and measured depths to bedrock, which ranged in depth from about 80 to 240 cm. Simeoni et al. (2009) reported an accuracy of ±10 cm between the interpreted and measured depths to B horizon, which ranged in depth from about 25 to 100 cm. For organic soils, Rosa et al. (2008) reported a mean maximum difference of 32 cm between measured and GPR interpreted depths of peat, which ranged in thickness from 0 to 8 m. Differences were attributed to surface and subsurface irregularities, and spatial variations in peat moisture contents and bulk densities (Rosa et al., 2008).

In a GPR evaluation of Fuquay soils (loamy, kaolinitic, thermic Arenic Plinthic Kandiudults) in northern Florida, the sandy mantle overlying loamy marine deposits is virtually free of reflectors (Figure 39.6). The underlying, loamy sediments, though relatively transparent to GPR, contain a large number



(a) (b)

FIGURE 39.6 Radar record (b) showing contact of sandy deposits with the underlying, more heterogeneous, loamy marine sediments from an area of Fuquay soils (a) in Florida.

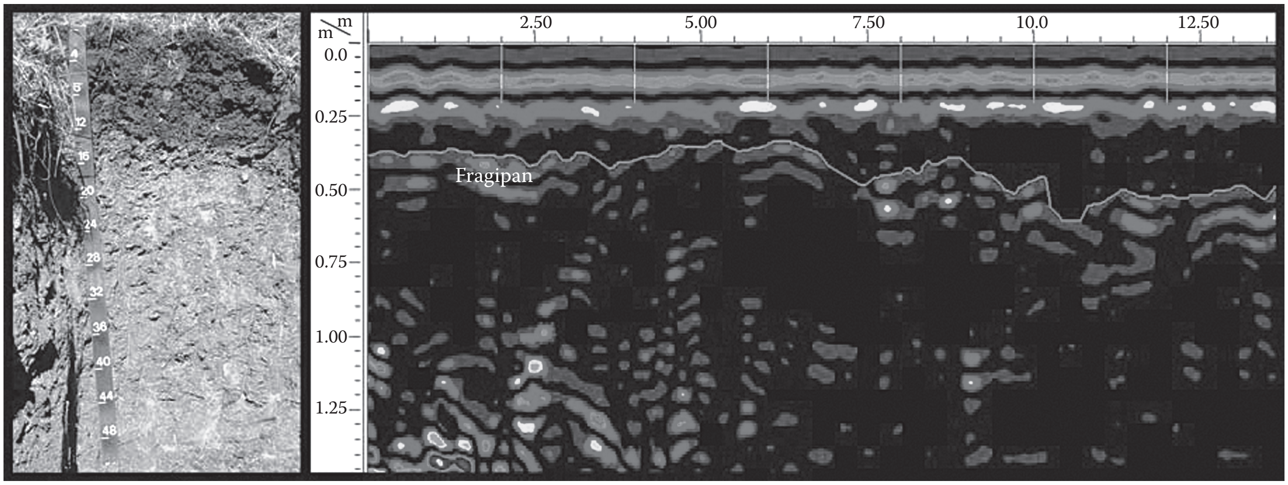
of reflectors, which aid identification of the contact between the two materials. The large number of reflectors of varying sizes, shapes, orientations, and reflected signal amplitudes suggests heterogeneous soil materials, which may be alternating layers of loamy and sandy materials with pockets of ironstone nodules that are evident in the lower part of the Fuquay photograph (Figure 39.6a). As over 1 m of penetration was achieved with the 200 MHz antenna through these loamy materials, the clay fraction is assumed to be dominated by low-activity clay minerals (e.g., kaolinite, gibbsite, and goethite) as implied in the taxonomic classifications of Fuquay soils (Kandiudults).

Ground-penetrating radar has been used to estimate the depth to argillic horizons (Asmussen et al., 1986; Collins and Doolittle, 1987; Doolittle, 1987; Truman et al., 1988; Doolittle and Asmussen, 1992), spodic horizons (Collins and Doolittle, 1987; Doolittle, 1987; Burgoa et al., 1991), placic horizons (Lapen et al., 1996), and buried palaeosols (Chapman et al., 2009). These diagnostic horizons generally have well-defined upper boundaries that have abrupt increases in bulk density, clay (argillic horizon), organic C complexed with Fe and Al (spodic horizon), or cemented Fe, Mn, or Fe-humus complexes (placic horizon). GPR has also been used to determine the thickness of albic horizons (E horizons) and evaluate the depth, lateral extent, and continuity of duripans, petrocalcic, and petroferric horizons (Doolittle et al., 2005), fragipans (Olson and Doolittle, 1985; Lyons et al., 1988; Doolittle et al., 2000), ortstein layers (Mokma et al., 1990a), and traffic pans (Raper et al., 1990). In these reports, the diagnostic horizons evaluated have distinct properties that enhanced their detection with GPR including cementation and induration (duripan, ortstein, petrocalcic, and petroferric horizons) and/or high bulk density (fragipans and traffic pans).

Ground-penetrating radar has been used to infer marked changes in soil color associated with abrupt and contrasting changes in organic carbon contents (Collins and Doolittle, 1987). GPR has also been used to infer the concentration of clay lamellae (Farrish et al., 1990; Mokma et al., 1990b; Tomer et al., 1996) and plinthite (Doolittle et al., 2005) in soils. Steelman and Endres (2009) used GPR to monitor soil freezing and the seasonal development of surficial, seasonally frozen soil layers. In areas of permafrost, GPR has been used to estimate the thickness of the active layer (Doolittle et al., 1990b) and to identify ice wedges (Hinkel et al., 2001).

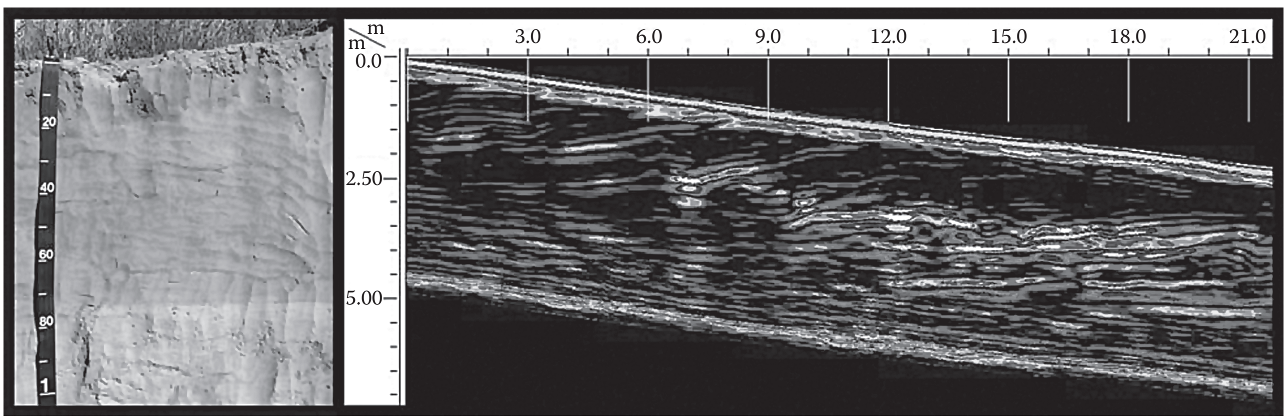
In evaluations of a landscape in Pennsylvania containing Erie (fine-loamy, mixed, active, mesic Aeric Fragiaquepts) soils that have a fragipan and Fremont (fine-loamy, mixed, semiactive, acid, mesic Aeric Endoaquepts) soils that lack a fragipan, the radar record (collected with a 400 MHz antenna) has a seemingly continuous interface that grades from well to very poorly expressed (identified by a continuous line) at depths ranging from about 25 to 53 cm below the soil surface (Figure 39.7). Ground truth observations indicated that in areas where this interface had higher-amplitude reflections (lighter colors), the soil had a fragipan (Erie). In contrast, where the interface was denoted by lower amplitude reflections (black or darker colors) the soil had fragic properties but lacked a true fragipan and was considered to be Fremont. Multiple radar traverses across the landscape indicated that Fremont was the more abundant soil in the area.

Radar records of deep, sandy, excessively drained Valentine soils (mixed, mesic Typic Ustipsamments) in the Nebraska Sand Hills indicates thin, distinct, discontinuous, inclined to wavy bands throughout the soil (Figure 39.8b). These bands correspond to thin (<7.5 cm thick) layers of oriented clay on or bridging grains (lamellae) that are common in these sandy soils (Soil Survey Division Staff, 1999; Figure 39.8a). The radar record in Figure 39.8b has been surface normalized to correct the radar record for changes in elevation to produce graphic displays that



(a) (b)

FIGURE 39.7 Radar record (b) showing a continuous subsurface interface with intermittent fragic properties in an area of Erie soils (a) in western Pennsylvania.



(a) (b)

FIGURE 39.8 Surface normalized radar record (b) from an area of Valentine soils (a) in western Nebraska with discontinuous, inclined-to-wavy, thin bands of lamellae.

more closely resemble the topography of the landform and the soil horizons or features.

In areas of Valentine soils, the penetration depth of GPR typically ranges from about 2 to 5 m, which is less than expected for a very deep, excessively drained sandy soil. The presence of small amounts of clay in the lamellae, however, increases signal attenuation and reduces penetration depths (Harari, 1996; Olhoeft, 1998). In some soils, clay contents of only 5%–10% have been reported to reduce penetration depths to less than 1 m (Walther et al., 1986). Additionally, soils, such as Valentine, with an abundance of smectite or vermiculite, and associated relatively high CEC are more attenuating to GPR than soils with low-activity clay minerals.

In many upland areas, rock fragments and irregular or weathered bedrock surfaces limit the effectiveness of conventional methods to examine soil profiles and determine the depth to bedrock. In these areas, GPR is more reliable and effective than traditional soil surveying tools for determining the depth to bedrock and the composition of soil map units based on soil-depth criteria (Collins et al., 1989; Schellentrager and Doolittle, 1991). As seen in radar record shown in Figure 39.9b, the soil/bedrock interface can provide an abrupt, well-expressed, and easily identifiable reflector.

In some soils, however, coarse fragments in the overlying soil, irregular bedrock surfaces, fracturing, and the presence of saprolite make the identification of the soil/bedrock interface on radar records more ambiguous. Even with these limitations, GPR can be an effective tool for evaluating bedrock depths (Collins et al., 1989; Davis and Annan, 1989; Gerber et al., 2010) and changes in rock type (Davis and Annan, 1989), characterizing internal bedding, cleavage and fracture planes (Holloway and Mugford, 1990; Stevens et al., 1995; Toshioka et al., 1995;

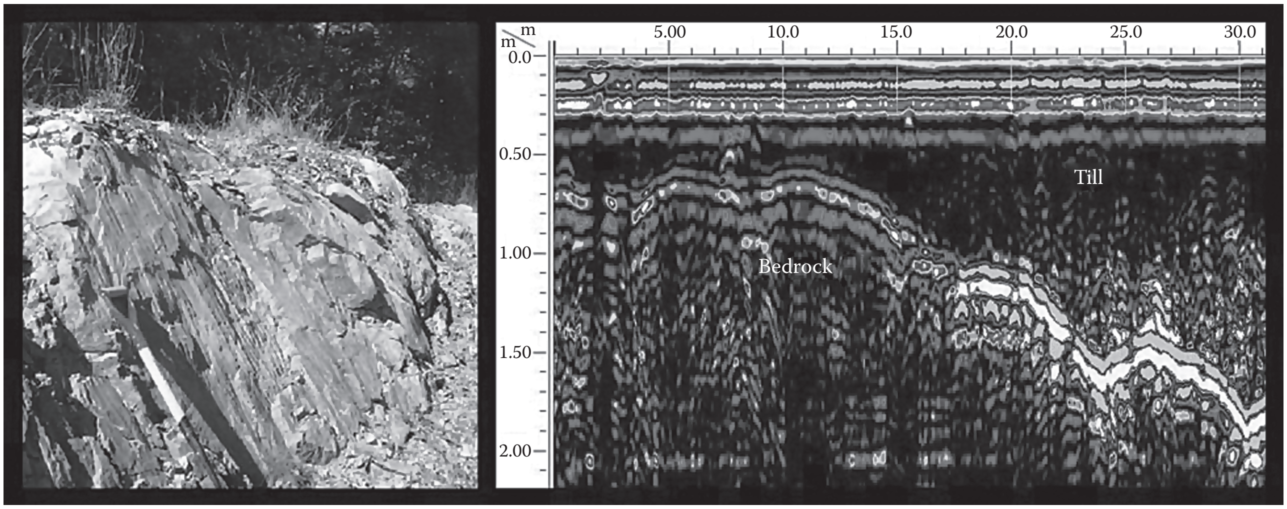


FIGURE 39.9 Radar record (b) collected with a 400 MHz antenna showing soil/bedrock interface in area of Monson soils (a) in central Maine.

Lane et al., 2000; Grasmueck et al., 2004; Nascimento da Silva et al., 2004; Porsani et al., 2005), and identifying cavities, sinkholes, and fractures in limestone (Barr, 1993; Pipan et al., 2000; Al-fares et al., 2002).

One example of the use of GPR over bedrock is a radar record from an area of Monson soils (loamy, isotic, frigid Lithic Haplorthods) in central Maine (Figure 39.9). This shallow, somewhat excessively drained soil formed in a thin mantle of glacial till over slate, phyllite, or schist. The phyllite shown in the photograph contains a large number of narrow, closely spaced, vertical fractures, which cannot be effectively resolved with the 200 MHz antenna used to collect the 2D radar record. Because of scattering losses, attenuation, wavelength scale heterogeneities, and geometric constraints, the number of fractures interpreted from radar data is an order of magnitude less than the number observed in outcrops (Lane et al., 2000). In addition, fractures with large dip angles reflect very little radar energy back to the antenna and are not accurately imaged because of spatial aliasing distortion (Lane et al., 2000). Fractures filled with water or saturated materials produce higher-amplitude reflections than air-filled or unsaturated fractures (Lane et al., 2000).

Ground-penetrating radar also has been used to improve soil-landscape models and soil map unit design on glacial-scoured uplands (Doolittle et al., 1988), former periglacial environments (Doolittle and Nelson, 2009; Gerber et al., 2010), wetland catenas (Lapen et al., 1996), and coastal plain sediments (Rebertus et al., 1989; Puckett et al., 1990). The use of GPR to improve soil-landscape models, however, has been limited.

Ground-penetrating radar has been extensively used in the investigation of peatlands. Compared with traditional coring methods, GPR provides a more rapid and efficient means for estimating the thickness and characterizing the subsurface topography of organic deposits (Ulriksen, 1980, 1982). GPR has been used to estimate the thickness and volume of organic deposits (Ulriksen, 1982; Shih and Doolittle, 1984; Tolonen et al., 1984; Collins et al., 1986; Worsfold et al., 1986; Welsby, 1988; Doolittle et al., 1990a; Pelletier et al., 1991; Hanninen, 1992; Turenne et al., 2006), to distinguish layers having differences in degree of humification and volumetric water content (Ulriksen, 1982; Tolonen et al., 1984; Worsfold et al., 1986; Chernetsov et al., 1988; Theimer et al., 1994; Lapen et al., 1996), and to classify organic soils (Collins et al., 1986). Lowe (1985) used GPR to estimate the number of logs and stumps buried in peatlands. More recently, GPR has been used in peatlands to study subsurface piping (Holden et al., 2002), permafrost (Moorman et al., 2003), and subsurface deposits and hydrostratigraphy (Comas et al., 2004, 2005; Kettridge et al., 2008; Lowry et al., 2009). Peatlands often display considerable anisotropy in composition, moisture content, and bulk density (Warner et al., 1990), and such differences have allowed separation of organic layers that differ in degree of humification, bulk density, and dielectric permittivity (Tolonen et al., 1982; Chernetsov et al., 1988; Hanninen, 1992; Nobes and Warner, 1992; Theimer et al., 1994; Mellett, 1995; Comas et al., 2005; Lowry et al., 2009). The successful identification of interfaces resulting from differences in water content and degree of humification, however, has not been universal (Wastiaux et al., 2000; Sass et al., 2010).

Figure 39.10 is a radar record obtained with a 70 MHz antenna in a ponded area of Carlisle (euic, mesic Typic Haplosaprists) soils in northwestern Rhode Island. On this radar record, the upper 80–90 cm is plagued by parallel bands of background noise caused by reflected signals from the air/ice and ice/water interfaces. Changes in reflective patterns and signal amplitudes are evident at a depth of about 90 cm, which represents the water/organic matter interface. On this radar record, the organic/mineral soil interface is easily identified (a segmented

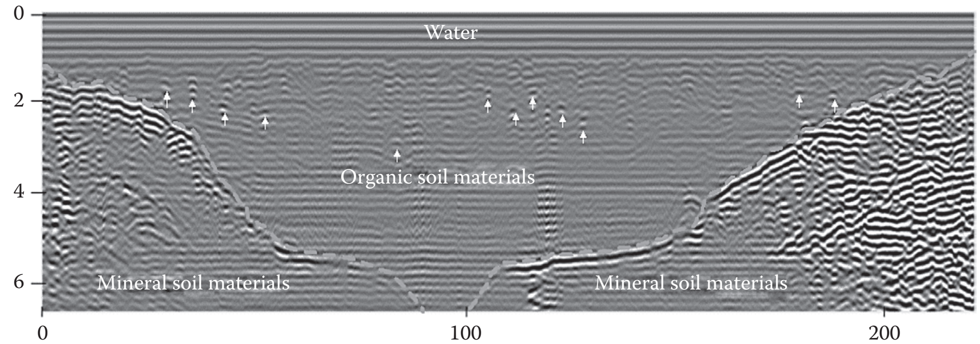


FIGURE 39.10 Radar record collected with a 70 MHz antenna over an inundated peatland in Rhode Island. A segmented line is used to highlight the organic/mineral soil interface. Arrows indicate locations of some buried logs, larger roots, or stumps. Scales are in meter.

line denotes the approximate location of this interface). This interface can be traced with reasonable confidence to a depth of about 675 cm (the maximum depth scanned). Buried logs and stumps produce hyperbolic reflections (indicted by arrows) within the organic materials.

Profiling depths of greater than 8 m have been reported in peatlands with low electrical conductivity (Ulriksen, 1980; Lowe, 1985; Worsfold et al., 1986; Theimer et al., 1994; Comas et al., 2005; Lowry et al., 2009), but GPR does not provide similar penetration depths in all organic soils. Penetration depths are largely affected by the specific conductance of the pore water (Theimer et al., 1994; Comas et al., 2005; Lowry et al., 2009). GPR is ineffective in coastal peatlands that are tidally influenced, contain sulfidic materials, and/or have high electrical conductivities. In noncoastal areas, GPR is generally more effective in acidic than alkaline peatlands. Lower frequency (<200 MHz) antennas are typically more effective for evaluating peatlands. In higher latitudes, peatlands are often surveyed during winter months, when surface layers are frozen and snow covered and can be easily traversed on foot or with snowmobiles or tracked vehicles. In lower latitudes, grass and reed covered peatlands have been successfully surveyed with airboats. Pelletier et al. (1991) described the use of helicopters to survey extensive peatlands in remote areas of Ontario.

39.3.3.2 GPR Evaluations of Groundwater, Soil Moisture, and Preferential Flow

Ground-penetrating radar has been used extensively for hydropedological investigations. In areas of coarse-textured soils, GPR has been used to chart water table depths among monitoring wells and into nearby areas (Sellmann et al., 1983; Davis et al., 1984; Wright et al., 1984; Shih et al., 1986; Johnson, 1987; Truman et al., 1988; Bohling et al., 1989; Smith et al., 1992; Iivari and Doolittle, 1994; Doolittle et al., 2006a). In addition, GPR has been used to design hydrologic models (Violette, 1987; Taylor and Baker, 1988), define recharge and discharge areas (Johnson, 1987; Bohling et al., 1989), predict groundwater flow patterns (Steenhuis et al., 1990; Iivari and Doolittle, 1994; van Overmeeren, 2004; Doolittle et al., 2006a), and delineate near-surface hydrologic conditions (Beres and Haeni, 1991; van Overmeeren, 1998). Collins et al. (1994) used GPR to identify and size subsurface cavities that influence preferential flow in a karst landscape; this information was used to improve contaminant transport models.

Radar records from an area of sandy Oakville (mixed, mesic Typic Udipsamments) soils formed in aeolian dunes in northwestern Indiana indicated depth to the water table ranging from 1.1 to 6.4 m (Figure 39.11; surface elevation has been normalized). Bedding planes within the dune and contrasting strata can

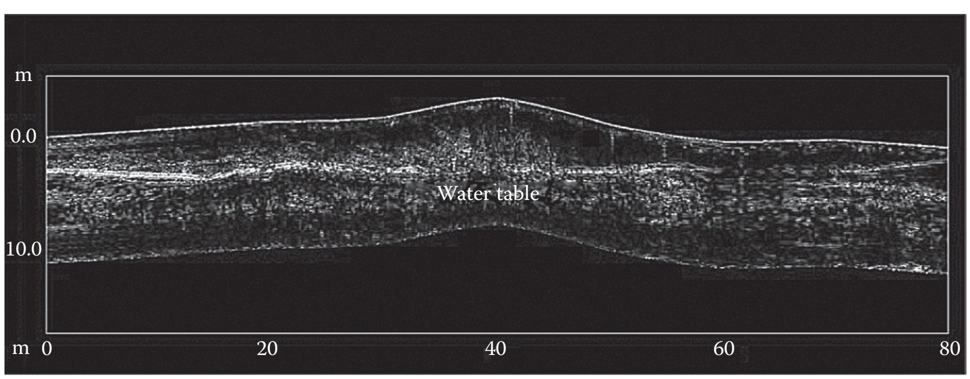


FIGURE 39.11 Surface-normalized radar record obtained with a 200 MHz antenna across a low aeolian dune in northwestern Indiana showing a water table as high-amplitude linear reflections, which stretch across the image.

be observed on the radar record. Reflections from the interior of dunes are largely produced by differences in moisture contents and differences in density and grain size between strata (Schenk et al., 1993; Harari, 1996).

At intermediate (field or catchment) scales, GPR has been used to map spatiotemporal variations in SWC (Lesmes et al., 1999; Huisman et al., 2001, 2002, 2003; Hubbard et al., 2002; Galagedara et al., 2003; Grote et al., 2003). A GPR method known as ground wave analysis has been used to measure the SWC of surface layers (Huisman et al., 2001). The ground wave is the portion of the energy that travels directly from a transmitting to a receiving antenna through the upper part of the soil. The velocity of the ground wave is strongly dependent on the water content of the upper few centimeters of the soil (Huisman et al., 2002). The volume of soil evaluated in ground wave analysis is determined by the antenna separation, width of the antenna, and depth of ground wave influence (Huisman et al., 2002). The depth of ground wave influence is typically 10–50 cm but is not well defined and decreases with increasing SWC (Huisman et al., 2002).

Ground-penetrating radar has been used to study preferential pathways in soils (Vellidis et al., 1990; Kung and Donohue, 1991; Kung and Lu, 1993; Boll et al., 1996; Tomer et al., 1996; Steenhuis et al., 1998). These studies have been conducted on sandy soils and under conditions favorable to GPR evaluations. However, Gish et al. (2002) used GPR data to locate and characterize soil layers that control subsurface flow at field scales in medium- and fine-textured soils. Using high-frequency antennas (450 or 500 MHz), these workers used GPR to identify depth to restrictive soil layers and the locations of discontinuities or breaks in these layers.

39.4 Electromagnetic Induction

39.4.1 Overview of Electromagnetic Induction

39.4.1.1 Principles of Operation

Electromagnetic induction methods involve the use of instruments known as ground conductivity meters (GCM), which consist of a transmitter coil and either one or two receiver coil(s) spaced at a set distance(s) apart. Ground conductivity meters operate by generating an alternating electrical current, which is passed through the transmitter coil. This alternating electrical current generates a time-varying electromagnetic field above the surface; the primary electromagnetic field. The primary electromagnetic field induces eddy currents that flow through the soil and generate a secondary electromagnetic field, which propagates through the ground. The secondary field is proportional to the ground current and is used to calculate the soil's "apparent" or "bulk" electrical conductivity. The amplitude and phase of these two electromagnetic fields are measured by the receiver coil(s).

Apparent conductivity (EC_a) is a weighted, average measurement for a column of earthen materials to a specific depth (Greenhouse and Slaine, 1983) and is a measure of the soil's ability to conduct electrical current. Variations in EC_a result from changes in the electrical conductivity of earthen materials, which is influenced by the type and concentration of ions in solution, the amount and type of clay in the soil matrix, the volumetric water content, and the temperature and phase of the soil water (McNeill, 1980). The EC_a of soils will increase with increases in soluble salts, water content, and/or clay contents (Rhoades et al., 1976; Kachanoski et al., 1988).

Even though interpretations are complicated by the interaction of multiple soil properties, EC_a has been used to infer and map variations in soils and soil properties at different spatial scales and levels of resolution. Variations in EC_a have been associated with changes in particle-size distribution, clay mineralogy, bulk density, CEC, salinity, plant nutrients, organic C concentration, and water content. Presently, EC_a mapping is recognized as one of the most valuable methods in agriculture for measuring the spatial variability of soils and soil properties at field and landscape scales (Corwin, 2008; Lück et al., 2009).

Electromagnetic induction is not suitable for use in all soil investigations. Generally, EMI has been most successful in areas where subsurface properties are reasonably homogeneous, the effects of one property (e.g., clay, water, or salt content) dominates over other properties, and variations in EC_a can be directly associated with changes in the dominant property affecting EC_a (Cook and Walker, et al., 1992). A weakness of this interpretative process is "equivalence" when several interacting properties can produce similar EMI responses. As an example, the simultaneous interaction of soil depth, clay, and moisture contents can lower the relationship of EC_a with any one of these properties alone. An increase in clay content will produce higher EC_a. Spatially, however, the increase in EC_a associated with an increase in clay can be offset by reduced soil depth over more electrically resistive materials and/or lower water content, both of which lower the measured EC_a. These three interacting and spatially varying soil properties can result in EC_a measurements that are difficult to associate with one soil property (e.g., clay content) alone. Because of equivalence, tacit knowledge, onsite soil sampling, and data analysis are required to decipher the exact site-specific causes for the variability in EC_a (Sommer et al., 2003).

Relationships between EC_a and soil properties can vary over short distances and foster inconsistent or ambiguous results (Carroll and Oliver, 2005). The effectiveness of EMI as a soil-mapping tool will depend upon the degree to which differences in physiochemical soil properties that affect EC_a correspond to differences in soil types. If a strong and meaningful relationship is established between soils and EC_a, field-scale EC_a mapping can be used to identify distinct zones, each with relatively homogenous soils and soil properties (Johnson et al., 2001).

39.4.1.2 Depth of Penetration

The depth of penetration of the electromagnetic current in the soil is dependent on the geometry (spacing and orientation) of the coils, frequency of the induced current, height above the soil surface, and the conductivity of the profiled material(s). The orientation of the transmitter and receiver coil axis (with respect to the ground surface) affects the response from materials at different depths (McNeill, 1980). In the horizontal dipole orientation (HDO) or perpendicular geometry (PRP), GCMs are more sensitive to surface materials. In the vertical dipole orientation (VDO) or horizontal coplanar geometry (HCP), GCMs are more sensitive to deeper materials.

In most studies, a single GCM is used. The choice of GCM will depend on the objectives of the survey, soil conditions, and the desired depth of penetration (Lück et al., 2009). In the United States, the most widely used GCM in agriculture is the EM38 meter (Geonics Limited, Mississauga, Ontario) and the Dualem-1S meter (Dualem, Inc., Milton, Ontario), which have maximum theoretical penetration depths of 1.5 m. For these GCMs, the depth of penetration is considered "geometry limited" rather than "skin depth limited" because penetration depth is a function of the distance between the transmitter and receiver coils (McNeill, 1996). For geometry limited GCMs, larger intercoil spacings and lower operating frequency result in greater penetration depths. For multifrequency GCMs developed by Geophex Limited (Raleigh, NC) and GSSI (Salem, NH), the depth of penetration is dependent upon the operating frequency and the EC_a of the profiled material(s). With these GCMs, the depth of penetration is considered "skin depth limited" rather than "geometry limited" (Won, 1980, 1983; Won et al., 1996). Skin depth represents the maximum depth of penetration for a GCM operating at a specified frequency and sounding a medium of known conductivity. The skin depth (D) can be estimated using the following equation (McNeill, 1996):

$$D = \frac{500}{(sf)^{-2}}, \tag{39.6}$$

where

s is the ground conductivity (mS m^{-1})

f is the frequency (kHz)

According to Equation 39.6, skin depth is inversely related to the operating frequency and the conductivity of the earthen materials. Low-frequency signals have longer periods of oscillation, lose energy less rapidly and, therefore, travel farther than high-frequency signals. Multifrequency sounding allows multiple depths to be profiled with one pass of the sensor (Won et al., 1996).

For all GCMs, surface and shallow layers contribute more to the overall response than deeper layers. de Jong et al. (1979) and Slavich (1990) reported that the actual depth of penetration is not fixed, but will vary depending on the EC_a of the profiled material(s). In general, GCMs will see deeper in electrically resistive than in conductive soils (Won et al., 1996).

Each instrument has characteristic depth sensitivity response curves. The theoretical depth of penetration has been arbitrarily defined as the depth at which 70% of a GCM's cumulative response is obtained in the overlying soil volume (Saey et al., 2008). Although contributions to the measured response come from all depths within the theoretical depth of penetration, the largest contribution to the response comes from the "depth of observation," which is the depth that has the greatest contribution to the measured EC_a (Roy and Apparao, 1971). As noted by Roy and Apparao (1971), for all GCM, the depth of observation is often considerably shallower than the theoretical penetration depth specified by manufacturers. In saline soils, the high conductivity of surface and near-surface soil horizons contributes greatly to the depth-weighted response of GCMs and limits the depth of observation. In areas of highly conductive soils, manufacturer's specifications or skin depth estimations may provide fairly accurate estimates of the theoretical depth of penetration, but greatly overestimate the depth of observation.

The response of GCMs (EC_a) is directly proportional to the electrical conductivity of the soil under conditions of low induction numbers. These conditions (low induction numbers) are met when the intercoil spacing is less than the skin depth. Conditions of low induction numbers are typically satisfied in soils having relatively low conductivities (<100 mS m^{-1}) (McNeill, 1980). In saline soils, however, the low induction numbers condition is not fulfilled, and EC_a is not directly proportional to the electrical conductivity of the soil.

39.4.1.3 Temperature Effects

Soil electrical conductivity is an electrolytic process that is a function of the concentration and mobility of the dissolved ions present in the soil solution and on the surfaces of soil particles, and thus, is dependent on the temperature and phase of the soil water (Allred et al., 2005). Air temperature does not directly affect soil electrical conductivity, but does contribute to the instrument "drift" reported in some studies (Allred et al., 2005). Variations in soil temperature, however, do affect EC_a. As soil temperatures increase, the viscosity of water decreases, the mobility of dissolved ion increases, and, as a consequence, EC_a increases (McNeill, 1980; Allred et al., 2005). Seasonal fluctuations in soil temperature are greatest in the upper part of soil profiles and have the greatest affects on EC_a. Brevik et al. (2004) observed that daily changes in soil temperatures within the upper 10 cm do not significantly influence EC_a. Results from studies conducted by Allred et al. (2005) suggest a soil temperature threshold of 8°C, below which there is a substantial reduction in measured EC_a, especially when the soil is frozen. Apparent conductivity measurements are unreliable when frozen layers are present in soils (McKenzie et al., 1989).

Electrical conductivity increases at a rate of approximately 1.9% per °C increase. McKenzie et al. (1989), working on saline soils, stressed the importance of correcting the soil temperature to a standard temperature for more accurate conversions of EC_a to saturated paste electrical conductivity (EC_e). When comparing results from EMI surveys of the same area that are completed at different times of the year, the effect of temperature on EC_a is important and customarily expressed at a reference temperature of 25°C. The EC_a can be adjusted to a reference EC at 25°C (EC_{25})

using the following equation (United States Salinity Laboratory Staff, 1954):

$$EC_{25} = f_t EC_t, \tag{39.7}$$

where

f_t is a temperature conversion factor (see Table 15 in United States Salinity Laboratory Staff, 1954)

EC_t is the EC_a measured at a particular temperature

Approximations for the temperature conversion factor are also available in polynomial form (Stogryn, 1971; Rhoades et al., 1999; Wraith and Or, 1999) or other temperature conversion equations such as the one modified from Sheets and Hendrickx (1995) by Corwin and Lesch (2005):

$$EC_{25} = 0.4470 = 1.4034 \exp\left(-\frac{T}{26.8150}\right). \tag{39.8}$$

In most studies, only a single soil temperature is measured. A common concern is the depth at which the measurement should be made. Lück et al. (2009) suggested that because the maximum sensitivity of the EM38 meter operated in the vertical dipole orientation is between depths of 30–50 cm, temperature measurements should be made within this depth interval.

Changes in air temperature, humidity, and atmospheric electricity (spherics) affect the stability or drift of GCMs (Sudduth et al., 2003). Sudduth et al. (2001) reported instrument drifts as great as 3 mS m^{-1} for the EM38 meter. The causes for this drift are not entirely due to variations in air temperature but are a function of instrument instability over time (Sudduth et al., 2001). Instrument drift is more noticeable and has a greater impact on interpretations in areas of low EC_a. Suggested methods to minimize instrument drift include the use of a calibration traverse line (Sudduth et al., 2001), ample instrument warm-up time, shading the meter, and conducting surveys when temperatures are below 40°C (Robinson et al., 2004).

39.4.1.4 Instruments

An increasing number of commercially available GCMs are available (see Figure 39.12). Ground conductivity meters commonly used in soil investigations include the Dualem-1S and Dualem-2 meters (Dualem, Inc.); the EM31, EM38, EM38DD, and EM38-MK2 meters (Geonics Limited); GEM-2 sensor (Geophex Limited); and Profiler EMP-400 (Geophysical Survey

FIGURE 39.12 Selected EMI equipment. Pictured are the Veris Soil EC 3100 mapping system (a); the Profiler EMP-400 multifrequency sensor (b); the EM38 meter (c); and the Dualem-2 meter (d).

Systems, Inc.). The dual-geometry configuration of Dualem meters, the dual dipole orientations of the EM38DD meter, and the dual receiver–transmitter spacings of the EM38-MK2 meter allow simultaneous measurement of EC_a over two separate depth intervals. The aforementioned GCMs all have standard connectors for GPS communication and support data loggers and proprietary software, which are used to record and integrate EC_a and GPS data. Some GCMs, such as the multifrequency Profiler EMP-400, come with internal GPS.

Another sensor that is commonly used in precision agriculture and soil investigations is the towed-array resistivity [ER] unit (Figure 39.12a). This ER sensor measures the potential gradient resulting from the injection of electrical current into the soil through coulter electrodes. Soundings and profiles are made with an arrangement of current and potential electrodes known as arrays. The depth of penetration is "geometry limited" and dependent upon the spacing and the arrangement of the electrodes. Arrays are connected to a field computer or data logger and integrated with GPS. Compared with GCMs, towed-array resistivity units do not experience instrument drift, and are less susceptible to spherics and engine noise interference. However, towed-array resistivity units are invasive and require good ground contact. As a consequence, the use of towed-array resistivity units is restricted in cultivated fields and over rocky or frozen grounds.

In the United States, several towed-array resistivity units are manufactured by Veris Technologies (Salina, KS). Towed-array resistivity units used in agriculture include the Veris Soil EC 2000, 3100, and 3500 mapping systems. These mapping systems utilize direct-contact, coulter electrodes that are arranged in a modified Wenner configuration (Lund et al., 2000).

Each of the aforementioned GCM and ER instruments has distinct operational advantages and disadvantages (Sudduth et al., 2003). Comparative studies have generally revealed close similarities between EC_a data collected with different instruments (Sudduth et al., 1999, 2003; Doolittle et al., 2001, 2002a; Saey et al., 2008). However, differences in sensor calibration, depth sensitivity curves, and volume of soil material measured will affect measurements and result in slightly different EC_a values. In comparative studies using different GCM and ER units, the highest correlations were obtained with sensors having similar depth sensitivity curves (Sudduth et al., 1999, 2003). In general, differences in EC_a maps produced from data collected with different GCMs or ER units have been more noticeable over soils with highly contrasting layers (Sudduth et al., 2003).

39.4.2 Field Methodology

The speed, economy, and simplicity of EMI make this geophysical method attractive in agriculture. In the past, grids were laid out across survey areas and EC_a measurements were obtained at discrete stations, typically represented by the grid intersections. Presently, the synergism of EMI, GPS, and field data recording systems has enabled the rapid collection, storage, and processing of larger data sets and the more comprehensive coverage of sites. The speed and ease at which data are georeferenced and recorded greatly reduces survey time and makes practical the surveying of large areas. Kitchen et al. (2003, 2005) discuss the integration of EMI and GPS technologies to improve interpretations.

Integrated systems have been used at various spatial scales to map soil EC_a using either point or continuous sampling methods. Point sampling methods measure EC_a at specific stations, which are often located within a predetermined grid or at random points within a survey area. With this method, establishment of a grid and the positioning of measurement points are more time-consuming than the actual collection of EC_a data. As a consequence, compared with continuous sampling methods, the number of observation is often less for surveys conducted using point sampling methods. However, because of the greater time spent at each measurement point, a greater number of EC_a and position measurements can be collected and averaged. This procedure is known as "stacking" and is used to reduce unwanted background noise and improve data quality, but its use increases the time needed to complete the EMI survey. With the point sampling method, the operator typically moves sequentially along grid or traverse lines and measures EC_a in two dipole orientations or geometries. This method has been used most often for EMI surveys of small research areas (typically less than 40 ha) (Hendrickx et al., 1992; Khakural et al., 1998; Lesch et al., 1998; James et al., 2003; Jung et al., 2005).

Electromagnetic induction surveys of very large areas (300–10,000 km^2) have required larger grid intervals with more widely spaced stations. For surveys of very large areas, the spacing between measurement points has ranged from 0.5 (Williams and Arunin, 1990) to 5 km (Williams and Baker, 1982). Apparent conductivity data will be more generalized in surveys of very large areas as the short-range variability in soils and soil properties is masked by the coarse grid scale.

Continuous, simultaneous recording of both GPS and EC_a data in field computers as the GCM or ER is moved along traverse lines (continuous sampling) produces a greater number of measurements, provides more complete coverage of sites, and is less time-consuming than point sampling methods. Continuous sampling methods, however, may result in reduced positional accuracy, and GCMs must remain stable and in a fixed orientation when operated in the continuous mode. Consequently, either two separate surveys are required to collect EC_a data in both dipole orientations, or a GCM with a dual-geometry configuration, dual receiver–transmitter spacings, or multiple frequencies must be used to simultaneously collect measurements over different depth intervals.

Depending on the size of the survey area and the resources available, either a mobile or pedestrian survey is used to collect EC_a data. For large, open areas or whenever the total number of observations exceeds 1600 data points, Freeland et al. (2002) recommended the use of mobile over pedestrian surveys. In large open fields, mobile surveys provide more comprehensive coverage, greater acquisition efficiency, and less operator fatigue than pedestrian surveys (Cannon et al., 1994; Chen et al., 2000; Johnson et al., 2001; Corwin et al., 2003; Greve and Greve, 2004; Corwin et al., 2006). Cannon et al. (1994) reported a fivefold increase in productivity when mobile survey methods were used instead of pedestrian survey methods. However, in rough terrain and under unfavorable field conditions, mobile EMI surveys are impractical and pedestrian surveys must be carried out.

For continuous sampling methods, the inline (along traverse lines) and lateral (distance between traverse lines) sampling intensities are more critical to map quality (precision and accuracy of boundaries) than the applied interpolation technique (e.g., kriging, inverse distance) (James et al., 2003; Farahani and Flynn, 2007). In general, slower speeds of advance, smaller traverse line spacing, and faster data acquisition rates lead to higher sampling intensities and map quality (Farahani and Flynn, 2007). These workers found that map quality decreased only slightly as the interval between traverse lines was increased from 2.5 to 50 m. As the interval between traverse lines was increased to greater than 80 m, however, map quality declined very rapidly as map delineations became highly distorted and incorrectly classified. King et al. (2001) considered a traverse line spacing of about 24 m reasonable for larger fields. Using these examples as guidelines, the final decision on the most suitable traverse spacing should be tempered by tacit knowledge of the soils and expert site assessments.

39.4.3 Applications

Apparent conductivity maps have been used to identify soil and management zones and to predict differences in soils, soil properties, and crop yields. Differences in EC_a have been associated with spatial changes in one or more specific soil physiochemical properties. Electromagnetic induction has been used to assess and map soil salinity, assist site-specific management, characterize variability in soil physiochemical properties, and direct soil sampling.

39.4.3.1 Soil Salinity Assessment

Methods commonly used to appraise soil salinity include: (1) visual observation of crop appearance (Soil Survey Division Staff, 1993), (2) sampling and measurement of the saturated paste extract (EC_e) (United States Salinity Laboratory Staff, 1954), (3) Wenner array or four-electrode sensor (Halverson et al., 1977; Nadler, 1981), (4) EMI (Corwin and Rhoades, 1982, 1984, 1990; Williams and Baker, 1982; Wollenhaupt et al., 1986; McKenzie et al., 1989; Rhoades et al., 1989a, 1989b), and (5) multi- and hyperspectral imagery (Corwin, 2008).

In soil surveys, visual crop observations and saturated paste extract method are the most commonly used methods to distinguish and map phases of soil salinity (United States Salinity Laboratory Staff, 1954; Soil Survey Division Staff, 1993). While visual observations are adequate for salinity mapping, these interpretations provide only a qualitative measure of soil salinity and are dependent on the presence of plant cover and surface salts. The saturated paste extract method is commonly accepted as the most accurate measure of soil salinity (McNeill, 1985), but it requires considerable resources for field sampling and laboratory analysis. Thus, only a limited number of soil samples can be collected and analyzed. Because of the limited sample numbers and the high spatial variability of soil salinity, analysis of saturated paste extracts is of limited value for the assessment of soil salinity across large management units or for soil surveys (Corwin, 2008).

In the early 1970s, the USDA-ARS Salinity Laboratory began to use bulk soil electrical conductivity as a field measure of soil salinity (Rhoades and Ingvalson, 1971), and techniques to evaluate salinity with EMI were subsequently developed (de Jong et al., 1979; Rhoades and Corwin, 1981;Williams and Baker, 1982). This early research was directed principally toward the vertical profiling of salinity through the root zone (Corwin, 2008). Several empirical relationships were developed but were site specific and could not be extrapolated and used at other sites without calibration (Corwin, 2008).

A major challenge in using EMI to map soil salinity has been the conversion of apparent conductivity (EC_a) into the conductivity of the saturated paste extract (EC_e); the most commonly used measure of soil salinity. A number of models have been developed that relate EC_a to EC_e (Wollenhaupt et al., 1986; McKenzie et al., 1989; Rhoades et al., 1989a, 1989b; Corwin and Rhoades, 1990; Slavich, 1990; Cook and Walker, 1992; Lesch et al., 1995a, 1995b; Johnston et al., 1997). Depth-weighted calibration models (Wollenhaupt et al., 1986; McKenzie et al., 1989) are useful in mapping relative spatial differences in soil salinity within units of management, but do not assess salinity variation with depth in the profile (Cassel et al., 2009). Rhoades et al. (1989a, 1989b) developed empirical equations and coefficients for determining soil salinity at different depths across units of management using a dual pathway parallel conductance model. This deterministic model relies on estimates of water content, texture, bulk density, and soil temperature to estimate soil salinity from EC_a data. Multilinear regression models for determining soil salinity at different soil depth intervals from EC_a data were later developed (Lesch et al., 1992, 1995a, 1995b). These stochastic models are independent of the salinity profile shape, but require the collection and analysis of soil samples for calibration (Cassel et al., 2009). Unfortunately, models relating EC_a to EC_e are both time dependent and site specific (Lesch et al., 1998). As a consequence, calibration equations and modeled results usually cannot be extrapolated to other sites (Cassel et al., 2009). To potentially reduce the number of samples needed for model calibration, directed soil sampling models have developed to characterize spatial variations in soil salinity across units of management (Lesch et al., 1992).

In areas of saline soils, of the properties that influence EC_a, the concentration of soluble salts typically have the greatest effect on EC_a (van der Lelij, 1983; Johnston et al., 1997). Williams and Baker (1982) estimated that 65%–70% of the variation in EC_a in saline soils can be explained by changes in the concentration of soluble salts alone, and EC_a can provide reasonably accurate estimates of soil salinity (Williams and Baker, 1982; van der Lelij, 1983; Diaz and Herrero, 1992). Although the estimates of salinity levels are reasonably accurate, they may not meet accuracy standards for specific objectives (Rhoades et al., 1989a; Johnston et al., 1997).

39.4.3.2 Soil Property Prediction, Site-Specific Management, and High-Intensity Soil Surveys

Apparent conductivity is a helpful surrogate for characterizing the variability of soil properties that are too difficult, time-consuming, or expensive to measure by other means (Jaynes, 1995; Stafford, 2000; Corwin et al., 2008). As apparent conductivity is a function of several interacting soil properties, the use of EC_a to characterize the variability of a soil physiochemical property is not always straightforward and uncomplicated. Effective use of EMI to predict soil properties requires sampling and analysis of a limited number of ground-truth observations for site-specific calibration of EC_a to the property of interest.

In nonsaline soils, EC_a is largely affected by water content, clay content, mineralogy, CEC, and organic matter content, with reported correlations ranging from 0.4 to 0.8 (Farahani et al., 2005). Electromagnetic induction has been used extensively to assess depths to argillic horizons, claypans, fragipans (Stroh et al., 1993; Sudduth and Kitchen, 1993; Doolittle et al., 1994; Sudduth et al., 1995; Mueller et al., 2003), and bedrock (Zalasiewicz et al., 1985; Palacky and Stephens, 1990; Bork et al., 1998). Apparent conductivity has been used as a surrogate measure of SWC (Kachanoski et al., 1988, 1990; Sheets and Hendrickx, 1995; Khakural et al., 1998; Waine et al., 2000; Mueller et al., 2003; Hezarjaribi and Sourell, 2007; Huth and Poulton, 2007; Tromp-van Meerveld and McDonnell, 2009), clay content (Williams and Hoey, 1987; Mueller et al., 2003; Sommer et al., 2003; King et al., 2005; Weller et al., 2007; Cockx et al., 2009; Harvey and Morgan, 2009), CEC (Triantafilis et al., 2009), exchangeable Ca and Mg (McBride et al., 1990), soil pH (Bianchini and Mallarino, 2002), soil organic carbon (Jaynes, 1996; Johnson et al., 2001), field-scale leaching rates of solutes (Slavich and Yang, 1990), herbicide partition coefficients (Jaynes et al., 1994), and available N (Eigenberg et al., 2002). In these reported studies, EC_a was either directly related to the soil property under investigation or the property (such as soil organic carbon) was associated with a property (moisture and clay content) that the sensor measures. The strength and significance of the relationships among EC_a and other properties, however, can vary both within and among individual fields (Bekele et al., 2005; Carroll and Oliver, 2005), and often EC_a has a complex relationship with different, interacting soil properties (Farahani et al., 2005).

Site-specific management adjusts farming practices to measured variations in soil properties by dividing farmlands into management zones that have different seeding and chemical requirements (Mulla, 1993). Electromagnetic induction and ER have been used as accurate, fast, and inexpensive means of mapping soils and soil properties at a level of resolution that meets the requirements of site-specific management (Jaynes et al., 1995; Sudduth et al., 1995; Lund et al., 2000; Bianchini and Mallarino, 2002; Corwin and Lesch, 2003; Godwin and Miller, 2003; Adamchuk et al., 2004; Hedley et al., 2004; Corwin et al., 2008). As large, high-resolution data sets can be collected from mobile EMI platforms equipped with GPS, EMI is well suited to surveying of large units of management (Freeland et al., 2002; Adamchuk et al., 2004).

Spatial EC_a data identifies areas with reasonably homogenous soils (Doolittle et al., 1996; Frogbrook and Oliver, 2007) and often corresponds well with soil patterns across the landscape. These data can provide information on the distribution of soils at different scales of resolution; from field and landscape scales (Khakural et al., 1998; King et al., 2005; Weller et al., 2007) to regional inventories (Brevik and Fenton, 2003; Harvey and Morgan, 2009). Soil surveys incorporating EC_a data may have greater detail than those prepared with conventional methods due to the spatial intensity and abundance of data (Jaynes, 1995; Hedley et al., 2004; Doolittle et al., 2008) and may more accurately identify small included areas of dissimilar soils (Fenton and Lauterbach, 1999).

Kravchenko (2008) and Kravchenko et al. (2002) used EC_a to distinguish soils belonging to different soil drainage classes. Several studies have associated changes in EC_a with changes in the clay content of soils (King et al., 2005; Weller et al., 2007; Harvey and Morgan, 2009). In Coastal Plain landscapes, Anderson-Cook et al. (2002) used EC_a to classify soil series with an accuracy of greater than 85%. Doolittle et al. (2008) used EC_a to identify and delineate soils to improve the quality of a high-intensity soil survey in Illinois, but independent observations showed only slight improvements in the taxonomic purity of the map units when EC_a data were used. Inclusion of EC_a data in the survey effort, however, resulted in refinement of map units and increased confidence in mapping decisions.

Maps of EC_a are frequently compared with soil maps to better understand and interpret the distribution and extent of soils within delineations, adjust the placement of soil boundary lines, and refine soil maps. An example of this qualitative methodology is an EMI survey that was conducted in south-central Illinois to determine the distribution of sodium-affected soils; Natrudalfs and Natraqualfs. In this landscape, Natrudalfs and Natraqualfs are intermingled with non-sodium-affected Argiudolls, Argiaquolls, and Endoaqualfs. Areas of the sodium-affected soils are variable in size, distribution, and shape;

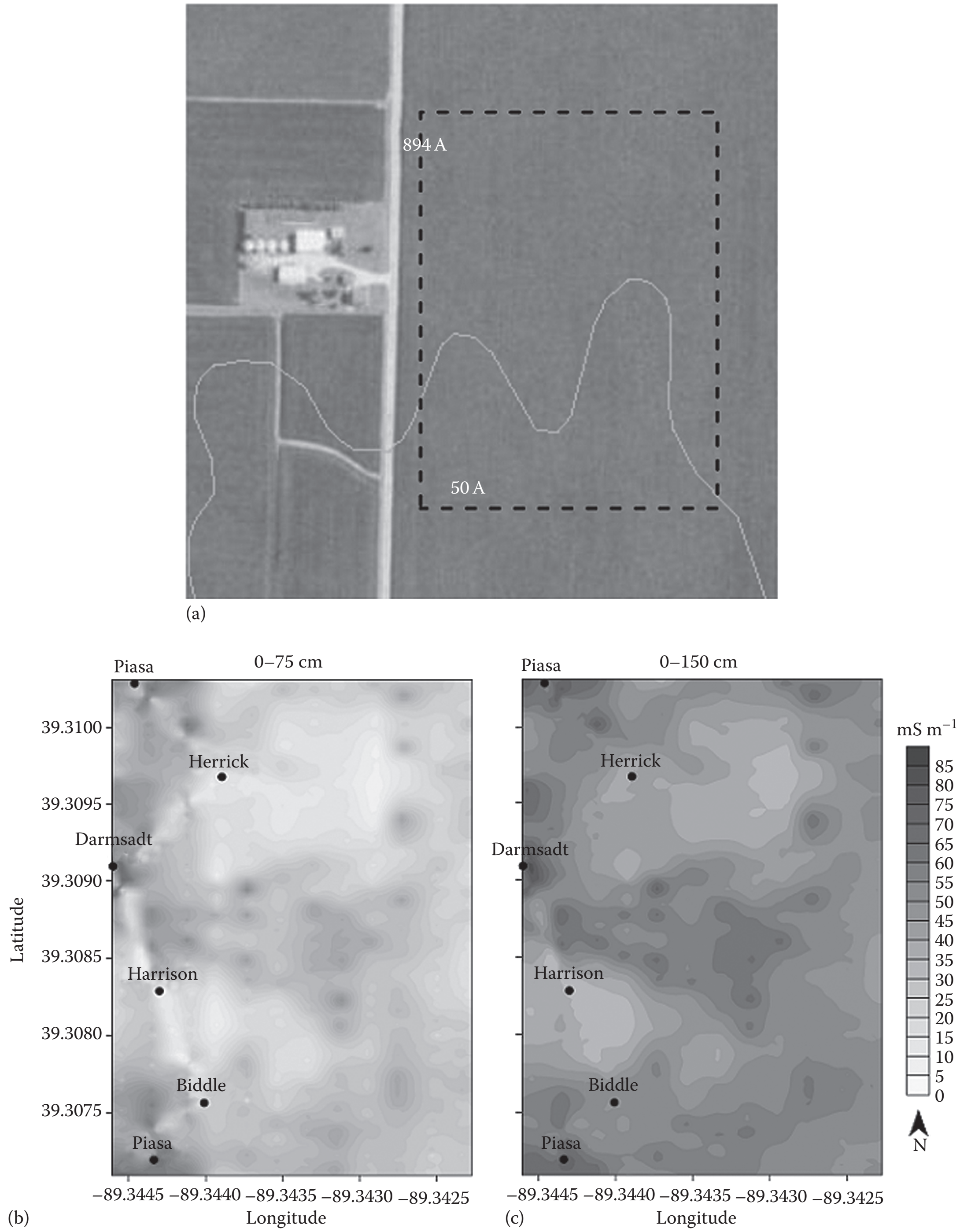


FIGURE 39.13 EMI survey of a site in south-central Illinois. The location is outlined on the soil map (Soil Survey Staff, 2009) as shown in (a). Spatial distributions of EC_a collected with the EM38-MK2-2 meter operated in the VDO are shown in the two plots: one for the shallower sensing 50 cm (b) and one for the deeper sensing 100 cm (c) intercoil spacings.

and their extent in individual soil polygons is unknown. The study site (enclosed by rectangle on the soil map in Figure 39.13a) contains two relatively large delineations of Herrick–Biddle–Piasa silt loams, 0%–2% slopes (894A), and Virden silty clay loam, 0%–2% slopes (50A). Of the named soils, only Piasa is a sodium-affected soil (Table 39.1).

The EMI survey of the area (Figure 39.13b) indicated that EC_a ranged from 0.6 to 84.0 mS m^{-1} and increased with depth and proximity to the water table. From this survey, sites for soil description were located from visual examination of prominent EC_a zones, and the soil identified at each observation point (Figure 39.13b). The soils identified at the three observation sites in areas

TABLE 39.1 Taxonomic Classifications of the Soil Series Identified at the Central Illinois Site

Soil Series	Taxonomic Classification
Biddle	Fine, smectitic, mesic Aquic Argiudolls
Darmstadt	Fine-silty, mixed, superactive, mesic Aquic Natrudalfs
Harrison	Fine-silty, mixed, superactive, mesic Oxyaquic Argiudolls
Herrick	Fine, smectitic, mesic Aquic Argiudolls
Oconee	Fine, smectitic, mesic Udollic Endoaqualfs
Piasa	Fine, smectitic, mesic Mollic Natraqualfs
Virden	Fine, smectitic, mesic Vertic Argiaquolls

of EC_a > 60 mS m^{-1} (in VDO) were identified as sodium-affected soils (Darmstadt and Piasa). In contrast, the sites in areas with EC_a < 55 mS m^{-1} (in VDO) were identified as the non-sodium-affected soils (Harrison, Biddle, and Herrick) (Table 39.1).

39.4.3.3 Directed Sampling

Software has been developed by the USDA-ARS Salinity Laboratory (Riverside, California) to select optimal soil sampling locations at field scales based on EC_a data (Lesch et al., 1995a, 1995b, 2000; Lesch, 2005). The ESAP (EC_e Sampling, Assessment, and Prediction) software though designed to predict EC_e from EC_a data can be used to predict other physiochemical properties as well. The program employs prediction-based sampling and modeling as a cost-effective alternative to geostatistical techniques, which are more sample intensive (Eigenberg et al., 2008). The ESAP methodology relies on high-density EC_a data to direct low-density soil sampling to calibrate suitable predictive equations. One of ESAP's components is the response surface sampling design (RSSD), which employs a directed-sampling approach to locate calibration sample points from observed locations and magnitudes of EC_a data (Lesch, 2005; Corwin et al., 2006; Eigenberg et al., 2008).

The ESAP-RSSD was used in a salinity assessment of a 16 ha field in Grand Forks County, North Dakota (Figure 39.14) to identify 20 optimally located sample points to characterize variability and to predict soil salinity across this unit of management. Because EC_a of measurements in the EM38DD HDO mode (shallow mode, ~0–75 cm) were less than those in the VDO mode (deep mode, ~0–150 cm), soils in this area were persumed to have a normal salt profile, that is, salt concentration increases with depth. The locations of the 20 RSSD-selected sampling sites and their numerical identifier are shown in Figure 39.14a.

After laboratory data were available for soil layers sampled at each of the 20 sample points, the ESAP calibrate program was used to convert EC_a data into EC_e data using a stochastic calibration model (Lesch et al., 1992, 1995a, 1995b, 2000). This stochastic method uses multilinear regression models to predict EC_e from EC_a data and provide a spatial representation of soil EC_e across the area (Figure 39.15).

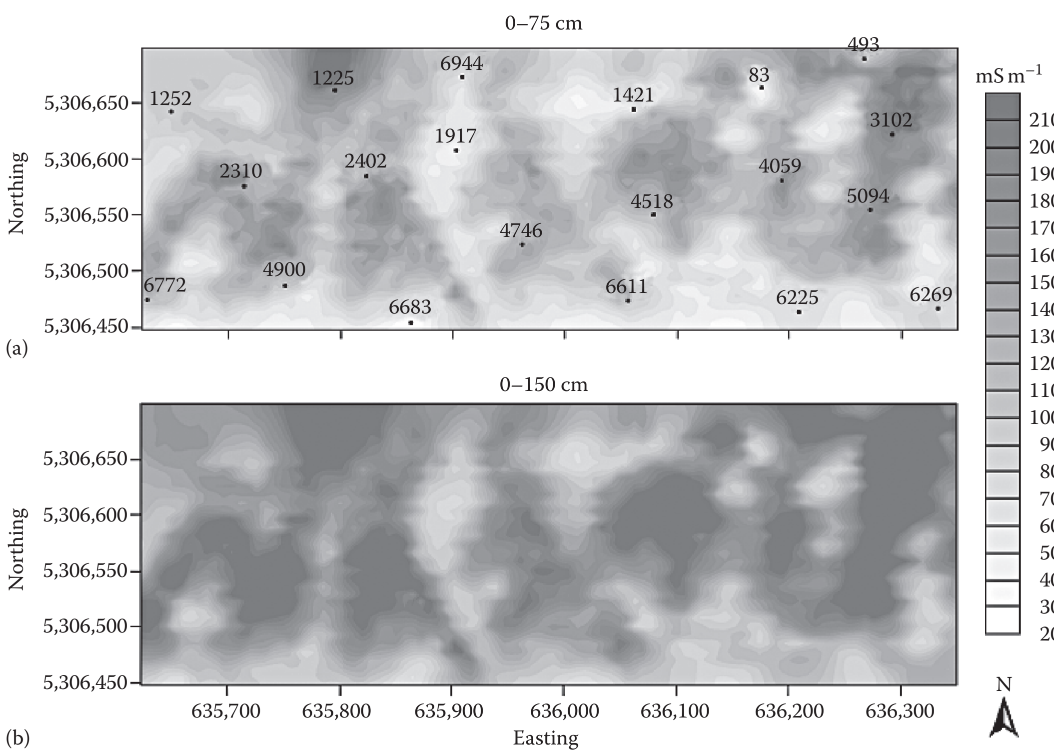


FIGURE 39.14 EC_a data from an area of saline Bearden soils in North Dakota. Data were collected with an EM38DD meter operated in the shallower sensing HDO (a) and the deeper sensing VDO (b). Numbers on the upper plot identify the locations of 20 sampling sites determined by the RSSD program of the ESAP software suite.

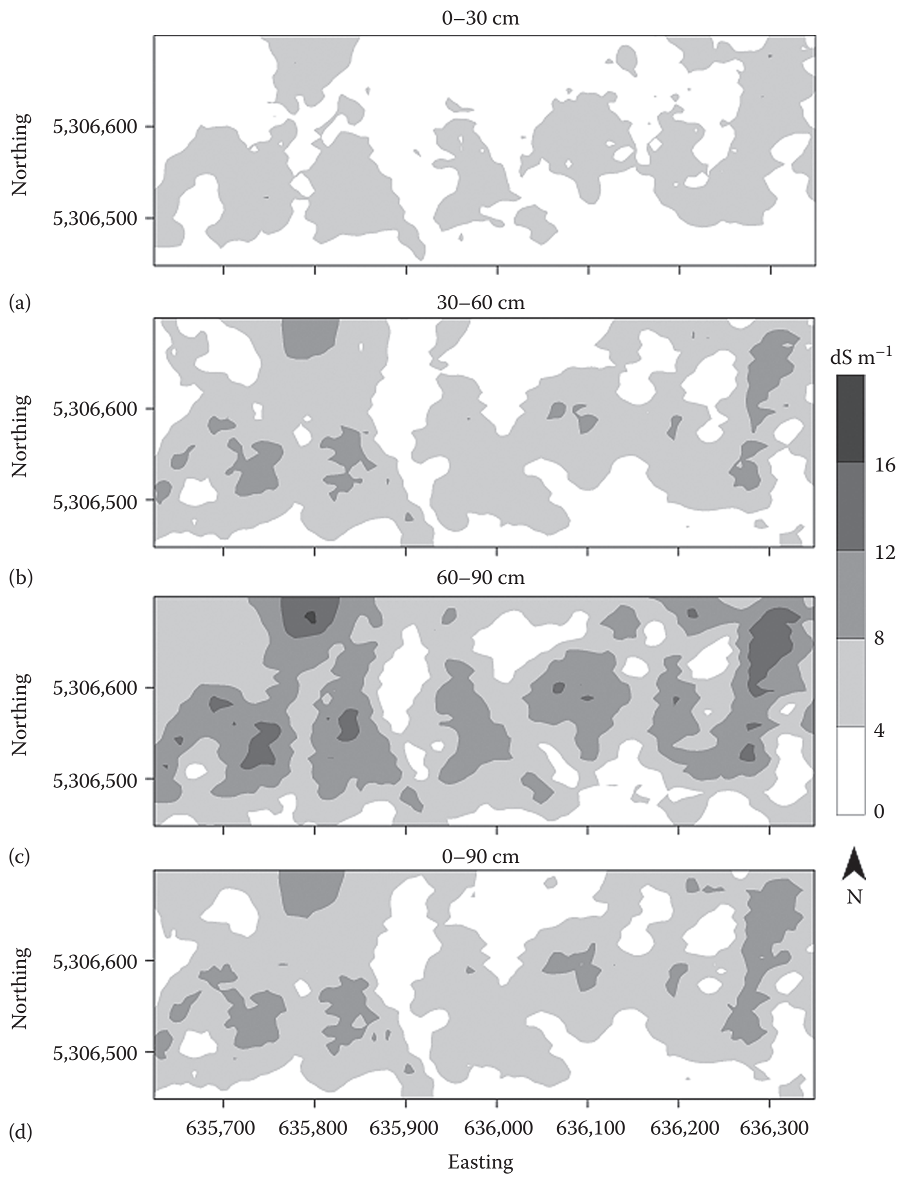


FIGURE 39.15 Distribution of estimated EC_e for different depth intervals ((a) 0–30 cm, (b) 30–60 cm, (c) 60–90 cm, and (d) 0–90 cm) within the Bearden saline site (Figure 39.14). Estimates from the stochastic model in the ESAP calibrate program.

TABLE 39.2 Basic statistics for a field of saline Bearden soils in Grand Forks County, North Dakota. Data predicted for different soil depth intervals using the ESAP stochastic model. All EC_e values are expressed in dS m^{-1}

	0–30 cm	30–60 cm	60–90 cm	0–90 cm
Number	7040	7040	7040	7040
Minimum	0.00	0.00	1.08	0.38
25th percentile	2.39	3.28	4.60	1.36
75th percentile	4.76	6.47	9.28	6.83
Maximum	9.34	11.67	16.73	12.58
Mean	1.60	2.13	8.78	2.20

Within the area evaluated, mean predicted EC_e ranged from 1.6 to 8.78 dS m^{-1} in the 0–30 and the 60–90 cm depth intervals, respectively (Table 39.2).

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