of trees in storms, and they typically have relief of as much as 0.5 to 1 m and a diameter of 1 to 3 m. This disturbance creates a characteristic pit and mound surface morphology and complicates the characterization of boreal soils (Fisher and Miller, 1980).

The major effect of humans on most boreal forests has been through the ignition of fire, but increasingly large portions of the boreal forests are being harvested for wood products. Very little of the boreal forest realm has been cleared and plowed for agriculture.

SUMMARY

Forest soils are variable at all spatial scales—from beneath one tree to another, across slopes and landforms, and across major gradients in climate. Some generalizations are possible, but not many; on average, tropical soils are no poorer or richer than temperate soils. The major types of forests around the world have some correspondence to soil types, but again, the variations in soil types within a forest type are striking. Major challenges and opportunities in understanding and managing forest soils depend on local details rather than on regional generalizations.

CHAPTER 4

Physical Properties of Forest Soils

Soil physical properties profoundly influence the growth and distribution of trees through their effects on soil moisture regimes, aeration, temperature profiles, soil chemistry, and even the accumulation of organic matter. Some physical properties can be altered intentionally by management, including draining wet soils and plowing subsoil to break up hardpans. The most widespread result of management on the physical properties of soils may be inadvertent soil compaction during harvesting operations.

The physical properties of soils constrain soil water properties, chemical reactions, and especially biologic activity. The importance of overall physical soil features is clear in a 4-year-old plantations of *Eucalyptus grandis*, where the volume was twice as great on ridgetop soils as on toeslope soils (Figure 4.1), primarily as result of differences in soil depth. The major physical properties of soils include texture (size distribution of particles), structure (aggregation of particles), porosity, stoniness, and depth.

In this chapter we examine the most important physical features of soils, discussing the major ranges of conditions that are common in forests and the effects of management activities.

SOIL TEXTURE IS FUNDAMENTAL

Soil can be conveniently divided into three phases: solid (including mineral and exchange phases), liquid, and gas. The solid phase makes up approximately 50 percent of the volume of most surface soils and consists of a mixture of inorganic and organic particles varying greatly in size and shape. The size distribution of the mineral particles determines the texture of a given soil. Schemes for classifying soil particle sizes have been developed in a number of countries. The classification used by the USDA, based on diameter limits in millimeters, is outlined in Table 4.1.

Mineral soils are usually grouped into three broad textural classes—sands, silts, and clays—and combinations of these class names are used to indicate mixtures (Figure 4.2).

The most important differences in soil texture relate to the surface areas of particles of different sizes. Medium-sized sand has a diameter of 0.25 to

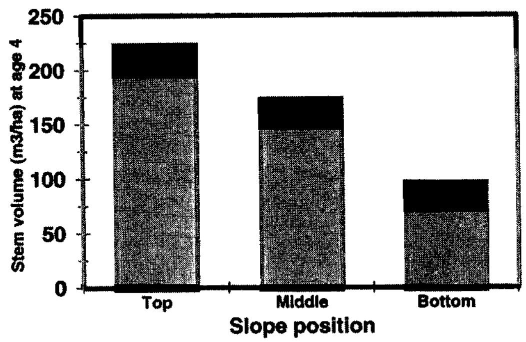

Figure 4.1 The stem volume of *Eucalyptus grandis* at age 4 years was twice as great on deep Oxisol (lateritic) soils at the top of a slope in Argentina as on shallower soils at the bottom of the slope (data from Dalla-Tea and Marco, 1996).

0.50 mm, about 7000 grains per gram, and a surface area of 0.013 m^2/g (Barber, 1995). Silt has a diameter of 0.002 to 0.050 mm, about 20 million particles per gram, and a surface area of 0.09 m^2/g. Clay particles are less than 0.002 mm in size, with 400 billion particles per gram, and a surface area of 1 to 10 m^2/g. The surface area per gram spans a range of three orders of magnitude, with dramatic effects on water potential, organic matter binding, cation exchange, and overall biotic activity.

Cobbles and **gravels** are fragments larger than 2 mm in diameter. They are not included in the particle-size designations because they normally play a minor role in agricultural soils. However, coarse particles are common in forest soils (up to 80 percent or more of some mountain soils). Major properties of coarse fragments are recognized by adjectives that indicate modification of the texture name (Table 4.2).

A moderate amount of rock in a fine-textured soil may favor tree growth. Coarse fragments may increase penetration of air and water, as well as the rate of soil warming in spring. Nevertheless, a coarse skeleton dilutes the soil and can be detrimental to tree growth if it occupies a large volume in sandy soils, because of low water-holding capacity and exchange capacity. Coarse-textured materials contribute little to plant nutrition.

TABLE 4.1 USDA Classification of Soil Particle Sizes

Name of Soil	Diameter Limits, mm
Sand	0.05–2.0
Very coarse	1.0–2.0
Coarse	0.5–1.0
Medium	0.25–0.5
Fine	0.10–0.25
Very fine	0.05–0.10
Silt	0.002–0.05
Clay	Less than 0.002

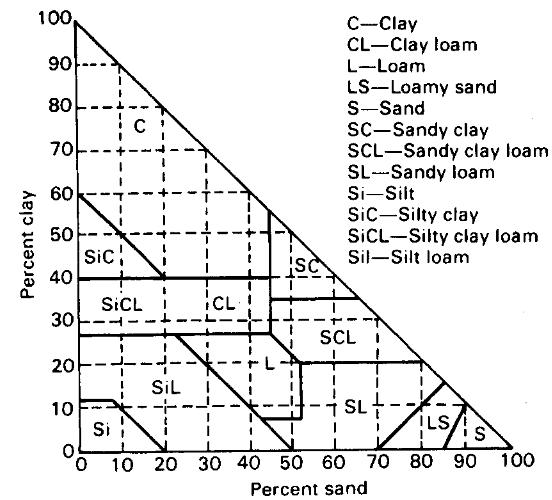

Figure 4.2 Soil textural triangle: percentage of clay and sand in the main textural classes of soils; the remainder of each class is silt.

TEXTURE INFLUENCES TREE GROWTH

Soil texture has major effects on forest growth, but these effects are indirect, manifested through the effect of texture on features such as water-holding

TABLE 4.2 USDA Classification Scheme for Coarse Fragments

Shape and Kind of Fragment	Size and Name of Fragment		
Rounded fragments	2 mm to 8 cm in diameter	8–25 cm in diameter	>25 cm in diameter
	Gravelly	Cobbly	Stony or bouldery
Thin, flat fragments	2 mm to 15 cm in length	6–38 cm in length	>38 cm in length
	Channery	Flaggy	Stony or bouldery

capacity, aeration, and organic matter retention. For example, deep, coarse, sandy soils often support low-productivity stands of pines, cedar, scrub oak, and other species that cope well with moisture and nutrient stress. Some sands have lenses of finer-textured materials, and variations in texture strongly influence overall soil properties (such as water retention).

SOIL STRUCTURE MODERATES EFFECTS OF SOIL TEXTURE

Soil structure refers to the aggregation of individual mineral particles and organic matter into larger, coarser units. This aggregation generally reduces bulk density (megagrams of soil per cubic meter), and increases water movement and aeration. Field descriptions of soil structure usually give the type or shape, class or size, and degree of distinctness in each horizon of the soil profile. The following types of structure are recognized by the U.S. Natural Resource Conservation Service: platy, prismatic, columnar, angular blocky, subangular blocky, granular, crumb, single grain, and massive. The size class of aggregates ranges from very fine to fine, medium, coarse, and very coarse. Structural classes are determined by comparing a representative group of peds with a set of standardized diagrams. Grade is determined by the relative stability or durability of the aggregates and by the ease of separating one from another. Grade varies with moisture content and is usually determined on nearly dry soil and designated by the terms weak, moderate, and strong. The complete description of soil structure, therefore, consists of a combination of the three variables, in reverse order of that given above, to form a type, such as moderate fine crumb.

Important drivers of soil aggregation include mineral chemistry, salts, clay skins, oxide coatings, growth and decay of fungal hyphae and roots, freezing and thawing, wetting and drying, and the activity of soil organisms (especially earthworms). Soil texture has considerable influence on the development of aggregates. The absence of aggregation in sandy soils gives rise to a single-grain structure, whereas loams and clays exhibit a wide variety of structures. Soil animals, such as earthworms and millipedes, favor the formation of crumb structure in the surface soil by ingesting mineral matter along with organic materials, producing casts that provide structure to the soil. The intermediate products of microbial synthesis and decay are effective stabilizers, and the cementing action of the more resistant humus components that form complexes with soil clays gives the highest stability.

Soil aggregation is also influenced by different tree species. In an experiment with 35-year-old plots with different tree species, the average size of aggregates ranged from 1.5 mm under white pine to 2.1 mm under Norway spruce (Scott, 1996). Across the species, average aggregate size increased with increasing fungal mass ($r^2 = 0.66$) and declining bacterial biomass ($r^2 = 0.72$; bacterial biomass declined as fungal mass increased, $r^2 = 0.87$).

The influence of tree species on soil aggregation was also apparent from a "lysimeter" experiment in California in which 50 m^3 chambers were filled with soil and planted to Coulter pine or oak (Graham and Wood, 1991). After 40 years, soils under the influence of pine lacked earthworms, had developed a clay-depleted A horizon, and had accumulated enough clay in the B horizon to qualify as an argillic horizon. Soils under oak (*Quercus dumosa*, a shrub) developed a 7-cm A horizon (90 percent of which was earthworm casts) enriched in humus and clay relative to the underlying C horizon. In this experiment, the plant species affected earthworm activity, which dominated structural development of the soil.

BULK DENSITY ACCOUNTS FOR THE COMPOSITION OF MINERALS, ORGANICS, AND PORE SPACE

Bulk density is the dry mass (of <2mm material) of a given volume of intact soil in Megagrams per cubic meter (which also equals kilograms per liter). Well-developed soil structure increases pore volume and decreases bulk density. The particle density of most mineral soils varies between the narrow limits of 2.60 and 2.75, but the bulk density of forest soils varies from 0.2 in some organic layers to almost 1.9 in coarse sands. Soils high in organic matter have lower bulk densities than soils low in this component. Soils that are loose and porous have low mass per unit volume (bulk density), while those that are compacted have high values. Bulk density can be increased by excessive trampling by grazing animals, inappropriate use of logging machinery, and intensive recreational use, particularly in fine-textured soils.

Increases in soil bulk density are generally harmful to tree growth for the same reasons that structure affects soil properties. Compacted soils have higher strength and can restrict penetration by roots. Reduced aeration in compacted soils can depress the activities of roots, aerobic microbes, and animals. Reductions in water infiltration rates are common when soils become compacted, and anaerobic conditions may develop in puddled areas.

The importance of differences in soil bulk density (and soil strength) for tree growth is illustrated by an extensive characterization of variation in soil bulk density across a stand of *Eucalyptus camaldulensis* on an Inceptisol in the savanna region of central Brazil (Figure 4.3). Where the bulk density of the 0- to 20-cm soil averaged 1.25 kg/L, stem volume of the 4-year-old trees was 25 m^3/ha (Gonçalves et al., 1997). Where the bulk density averaged 1.06 kg/L, stem volume was more than three times greater (90 m^3/ha).

Pore volume refers to that part of the soil volume filled by water or air. The proportions of water and air change over time, and soil water content drastically affects soil aeration. Gas molecules diffuse about 10,000 times faster through air than through water. Coarse-textured soils have large pores, but their total pore space is less than that of fine-textured soils (although puddled clays may have even less porosity than sands). Because clay soils have greater total pore space than sands, they are normally lighter per unit volume (have

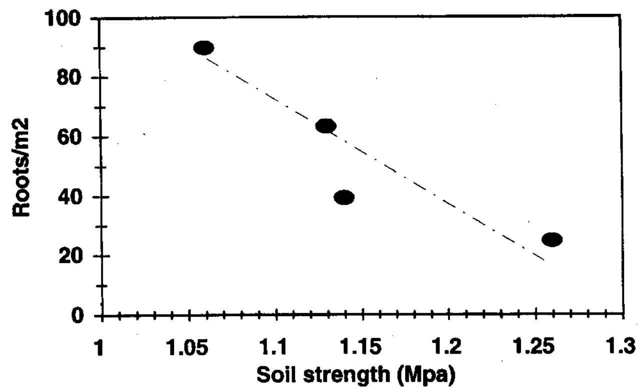

Figure 4.3 Variations across a stand in bulk density of an Inceptisol in the Brazilian savanna strongly affected the volume of wood accumulated by *Eucalyptus camaldulensis* at age 4 years (data from Pereira, 1990, cited in Gonçalves et al., 1997).

lower bulk density), but differences in soil structure can override the basic influence of texture on pore volume and bulk density.

Pore volume is conveniently divided into **capillary** and **noncapillary pores.** Soils with a high proportion of capillary (small-diameter) pores generally have high moisture-holding capacity, slow infiltration of water, and perhaps a tendency to waterlog. By contrast, soils with a large proportion of noncapillary (large-diameter) pores generally are well aerated, and have rapid infiltration and low moisture-retaining capacity. The pore size distribution of a soil may be broad, with large portions of capillary and noncapillary pores, particularly where soil animals and old root cavities provide large-radius pores in clayey soils.

Sandy surface soils have a range in pore volume of approximately 35 to 50 percent, compared to 40 to 60 percent or higher for medium- to fine-textured soils. The amount and nature of soil organic matter and the activity of soil flora and fauna influence pore volume and soil structure. Pore space is reduced by compaction and generally varies with depth. Some compact subsoils may have no more than 25 to 30 percent pore space. Tree species can also change the distribution of pore sizes in soils; soils under Norway spruce have been reported to have more pore space than adjacent soils under beech (Nihlgård, 1971).

The pore volume of forested soils is normally greater than that of similar soil used for agricultural purposes because continuous cropping results in a reduction in organic matter and macropore spaces (Figure 4.4). Porosity of most forest soils varies from 30 to 65 percent.

Soil aggregates are generally more stable under forested conditions than under cultivated conditions. Continued cultivation tends to reduce aggregation in most soils through mechanical rupturing of aggregates and by a reduction

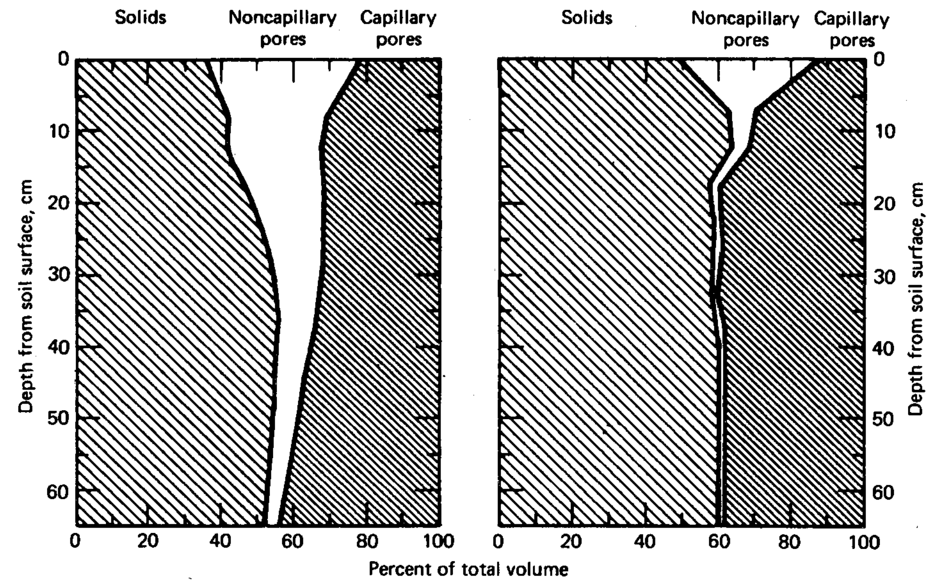

Figure 4.4 Porosity as measured in the surface 60 cm of Vance Soil (Typic Hapludult) in the South Carolina Piedmont: (left) in an undisturbed mixed hardwood forest and (right) in abandoned farmland (Hoover, 1949).

in organic matter content and associated cementing action of microbial exudates and fungal hyphae (Figure 4.5).

LIFE IN THE SOIL DEPENDS ON THE SOIL ATMOSPHERE

Soil air is important primarily as a source of oxygen for aerobic organisms, including tree roots. Soil air composition, like air volume, is constantly changing in a well-aerated soil. Oxygen is used by plant roots and soil microorganisms, and carbon dioxide is liberated in root respiration and by aerobic decomposition of organic matter (Figure 4.6).

Gaseous exchange between the soil and the atmosphere above it takes place primarily through **diffusion**. Consumption of oxygen by respiration in the soil leads to a gradient from high carbon dioxide in the ambient air into the low-oxygen soil air. The oxygen content of air in well-drained surface soils seldom falls much below the 20 percent found in the atmosphere, but oxygen deficits are common in poorly drained, fine-textured soils. Under these conditions, gas exchange is very slow because of the small, water-filled pore spaces. In very wet soils, carbon dioxide concentrations may rise to 5 or 6 percent and oxygen levels may drop to 1 or 2 percent by volume (Romell, 1922). However, oxygen is not necessarily deficient in all wet soils in spite of the absence of

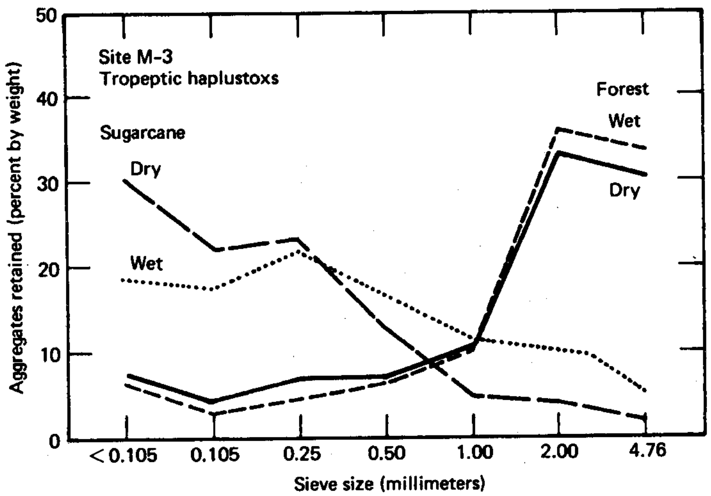

Figure 4.5 Stability of aggregates of an Oxisol used for forestry and for sugarcane in Hawaii (Wood, 1977). The forest soil has much larger aggregates that do not pass through a 1-mm sieve, whereas most of the sugarcane soil aggregates are much smaller. Reproduced from *Soil Science Society of America Journal* 41, no.1 (1977):135, with permission of the Soil Science Society of America.

voids. If the soil water is moving, it may have a reasonably high content of oxygen brought in through **mass flow** of water. Soils saturated with stagnant water are low in oxygen, and they are very poor media for the growth of most higher plants. Soil air usually is much higher in water vapor than is atmospheric air, and it may also contain a higher concentration of such gases as methane and hydrogen sulfide, formed during organic matter decomposition.

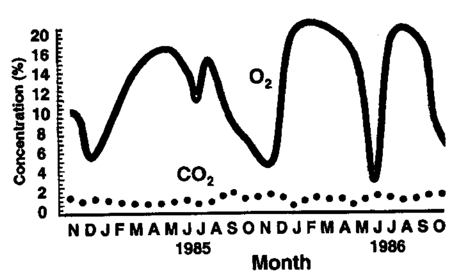

Figure 4.6 Trends in soil oxygen (solid line) and carbon dioxide (dotted line) concentrations at 20-cm depth in gleyic podzol (Spodosol) developed in glacial till in northern Sweden (data from Magnusson, 1992). During saturated periods, oxygen concentration dropped below 5 percent, and carbon dioxide concentration varied between 1 and 2.5 percent.

Oxygen concentrations in soil air as low as 2 percent are generally not harmful to most trees for short periods. Some *Alnus*, *Taxodium*, *Nyssa*, and *Picea* species can thrive at low levels of soil oxygen. Seedling root growth of many species is reduced by an oxygen content of 10 percent or lower. Any restriction in gas exchange between the tree roots and the atmosphere will eventually result in the accumulation of carbon dioxide.

SOIL STRUCTURE CAN BE HARMED BY INAPPROPRIATE MANAGEMENT ACTIVITIES

Compaction by heavy equipment or repeated passages of light equipment compresses the soil mass and breaks down surface aggregates, decreasing the macropore volume and increasing the volume proportion of solids. Reductions in air diffusion and water infiltration often combine with increases in soil strength in compacted soils. Compaction occurs more frequently on moist soils than on dry soils (because water lowers soil strength, "lubricating" soil particles) and more often on loamy-clay soils than on sandy soils. The effects of compaction may be less permanent in fine-textured soils (especially those containing considerable shrink-swell clays) than in some coarse-textured soils because of swelling and shrinking as a result of wetting and drying.

The use of large machines in forest harvesting is the primary driver of increases in soil bulk density and strength and decreases in pore volume. For example, 10 passes with a rubber-tired skidder substantially increased soil strength in a harvesting unit in Australia (Figure 4.7), although in this case, the amount of soil compaction may not have been great enough to affect growth.

A careful greenhouse study illustrates the pieces of the soil compaction puzzle (Simmons and Pope, 1988). When soils at −0.3 MPa were compacted to increase bulk density from 1.25 to 1.55 mg/m^3, the air-filled porosity declined from 36 percent to 20 percent, and soil strength increased from 2.3 to 4.1 MPa. When these soils were moister (with soil water potential near field capacity—0.01 MPa), air-filled porosity declined from 22 percent in the 1.25 mg/m^3 soil to 6 percent in the 1.55 mg/m^3 soil. The wetter soils had much lower soil strength (0.4 to 0.8 MPa, too low to impede root growth), but the most compacted soil had poor root growth because of anaerobic conditions that resulted from the low air-filled pore volume.

Under severe compaction, soils may puddle. Puddling results from the dispersal of soil particles in water and the differential rate of settling, which permits the orientation of clay particles so that they lie parallel to each other. The destruction of soil structure by this method may result in a dense crust that has the same effect on soil conditions as a thin, compacted layer. The crust is most common on soil surfaces where the litter has been removed by burning or by mechanical means. Reduced germination and increased mortality rates of loblolly pine seedlings have been observed on soils compacted or puddled by logging equipment.

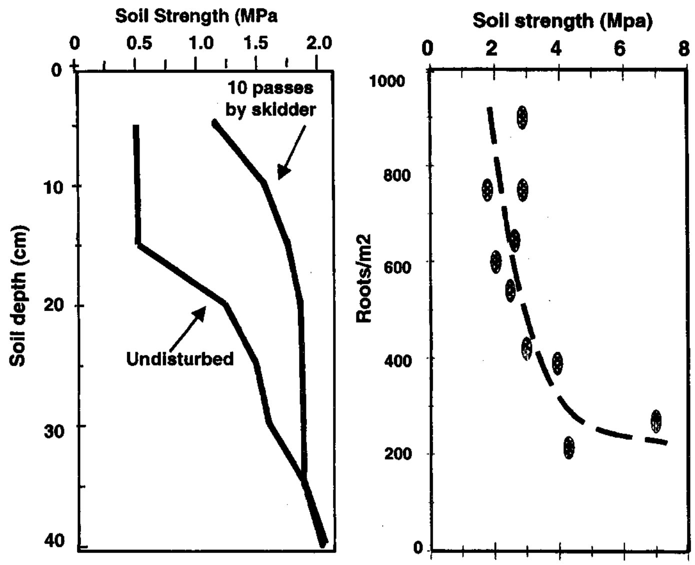

Figure 4.7 Soil strength increased substantially after 10 passes with a rubber-tired skidder in Australia, and root penetration declined as root strength increased. However, the domain over which these effects occurred did not overlap; the skidder did not increase soil strength enough to reduce root growth (modified from Gracen and Sands, 1980).

SOIL COLOR INCORPORATES EFFECTS OF ORGANIC MATTER AND OTHER FACTORS

Color is an obvious characteristic of soils that is used to differentiate soil horizons and classify soils. In many parts of the world, soils may be described as Red and Yellow Podzolics, Brown Earths, Brown Forest soils, and Chernozens (black soils).

Soil color depends on pedogenic processes and the parent material from which the soil was derived. Most soil minerals, such as quartz and feldspars, are light in color. Color is generally imparted by small amounts of colored materials, such as iron, manganese, and organic matter. Red colors generally indicate ferric (oxidized iron) compounds associated with well-aerated soils. Yellow colors may signify intermediate aeration. Ferrous (reduced iron) compounds of blue and gray colors are often found under reduced conditions associated with poorly aerated soils. Mottling (marbling of colors) often indicates a zone of alternately good and poor aeration. Manganese compounds and organic matter produce dark colors in soils. Color intensity is often used to estimate organic matter content. It is not a foolproof system, however, because the pigmentation of humus is less intense in humid zones than in arid

regions. Brown colors predominate in slightly decomposed plant materials, but more thoroughly decomposed amorphous material is nearly black.

Color itself is of no importance to tree growth, but color may indicate several important characteristics of soil. These characteristics include geologic origin and degree of weathering of the soil material, degree of oxidation and reduction, content of organic material, and leaching or accumulation of such chemical compounds as iron, which may greatly influence soil quality.

Dark-colored surface soils absorb heat more readily than light-colored soils but because of their generally higher content of organic matter, they often have higher moisture content. Therefore, dark soils may warm less rapidly than well-drained, light-colored soils. Soil color influences the temperature of bare soils, but it has less effect on the temperature of soils beneath forest canopies. Soil color becomes important after fires, when removal canopy combines with blackening of the soil surface to increase temperature (see Chapter 10).

SOIL TEMPERATURE INFLUENCES BIOTIC AND ABIOTIC PROCESS RATES

Soil temperature is a balance between heat gains and losses. Solar radiation is the principal source of heat, and losses are due to radiation, conduction, and convection (and evaporation of water in some cases). When well-developed canopies are present, the temperature of upper soil layers varies more or less according to the temperature of the air immediately above it. Temperature fluctuations in deeper soil layers are moderated. In the absence of a forest canopy, topsoil temperature may be much higher than air temperature because of direct absorption of solar radiation.

Soil temperatures generally decrease with increases in elevation, although postharvest soil temperatures may be extreme at high elevations on sunny days. Soil temperatures in winter are commonly warmer beneath snowpacks than are air temperature (Figure 4.8). Conifer forests may also have cooler soils than hardwood deciduous species (Figure 4.8) for several reasons. The higher leaf area of conifers may intercept more light and reduce solar heating of the soil, and may reduce convective heat losses as well. Conifer canopies also intercept more snow in winter, providing shallower snow layers on the ground to insulate the soil from frigid winter air. Temperatures also tend to decrease with latitude, and seasonal swings become more pronounced (Figure 4.8).

Aspect (direction of slope) influences soil temperature. Soils receive more solar radiation on south-facing slopes in the Northern Hemisphere and on north-facing slopes in the Southern Hemisphere. Higher temperatures typically lead to greater rates of evapotranspiration, so the heating effect on the soil may be amplified by drying.

The tree canopy and forest floor moderate extremes of mineral soil temperatures. They protect the soil from excessively high summer temperatures by intercepting solar radiation, and they reduce the rate of heat loss from the soil

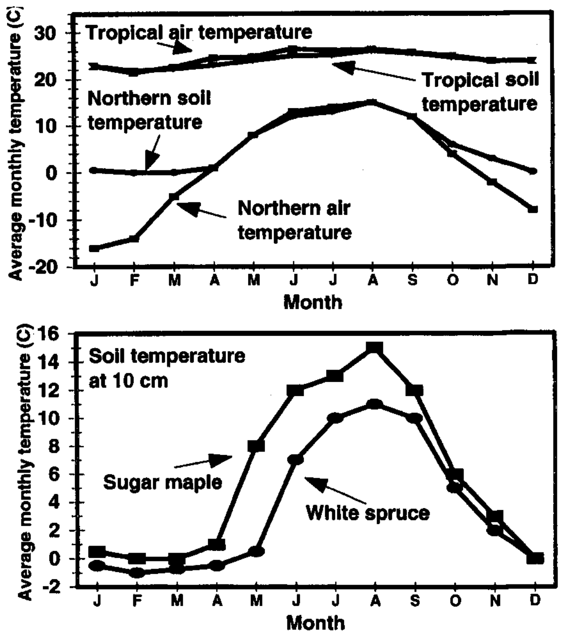

Figure 4.8 Seasonal traces of temperatures for air and upper soil at the Luquillo Experimental Forest in Puerto Rico (tropical site) and at Isle Royale, Michigan, for the northern forests of sugar maple and white spruce. The upper graph provides air temperature and soil temperature for the tropical, closed-canopy forest, which shows little seasonal variation and no difference between air and soil temperatures. The soil in the sugar maple stand remains warmer throughout the winter than the air because of the insulating snow layer. The lower graph contrasts soil temperatures under sugar maple and white spruce; spruce soils are colder because of less snow cover in winter and perhaps because of lower light penetration of the canopy (Luquillo data from X. Zou, personal communication; Isle Royale data from Stottlemyer et al., 1998).

during the winter. Forest cover may influence the persistence of frost in soils in cold climates; however, freezing generally occurs earlier and penetrates deeper in bare soil than in soil under a forest cover.

The effect of forest canopies on reducing frosts has long been attributed to canopies' absorbing longwave radiation emitted by the soil and then reradiating some of the energy back to the soil in a form of the greenhouse effect. Direct measurements of radiation budgets have not supported this story (Löfvenius, 1993); it appears that the beneficial effects of canopies on soil temperatures at night derive from the effects of canopies on air movement. Air at the surface of bare soils may become very cold as a result of radiative heat loss to the cold night sky. The presence of even a few trees per hectare can reduce the development of such cold air layers by intercepting some of the air currents higher above the ground and creating turbulent airflow that prevents stratification.

Part of the ameliorating effect of the forest cover on soil temperature is due to the forest floor (O horizon), as shown for a clay soil in Figure 4.9. Organic layers have low thermal conductivity, and they lower maximum summer temperatures and raise minimum winter temperatures. Diurnal fluctuations are also dampened.

Snow also insulates soils because of its low thermal conductivity. A porous snow covering of about 45 cm was sufficient to prevent soil freezing in frigid northern Sweden (Beskow, 1935), and a covering of 20 to 30 cm was sufficient to prevent freezing in southern Sweden. Winter soil temperatures may often be higher at northern locations and high elevations, where the snow cover is thick. Sartz (1973) reported that aspect influences soil freezing in Wisconsin because the direction of the slope affects the amount of snow accumulation. Therefore, depending on the conditions, frost may be deeper on southern slopes than on northern slopes (Table 4.3).

Specific heat and **conductance** of heat are inherent properties of soils that influence swings in daily temperature. Specific heat refers to the number of Joules needed to raise a unit mass of water by a defined number of degrees; raising the temperature of 1 kg (or 1 L) of water from 15° to 16°C requires an input of 4.2 MJ of energy. Conductance refers to the movement or penetration of thermal energy into the soil profile. Both specific heat and conductance are influenced somewhat by texture, and especially by soil water content and organic matter content. Water has high specific heat and high conductance.

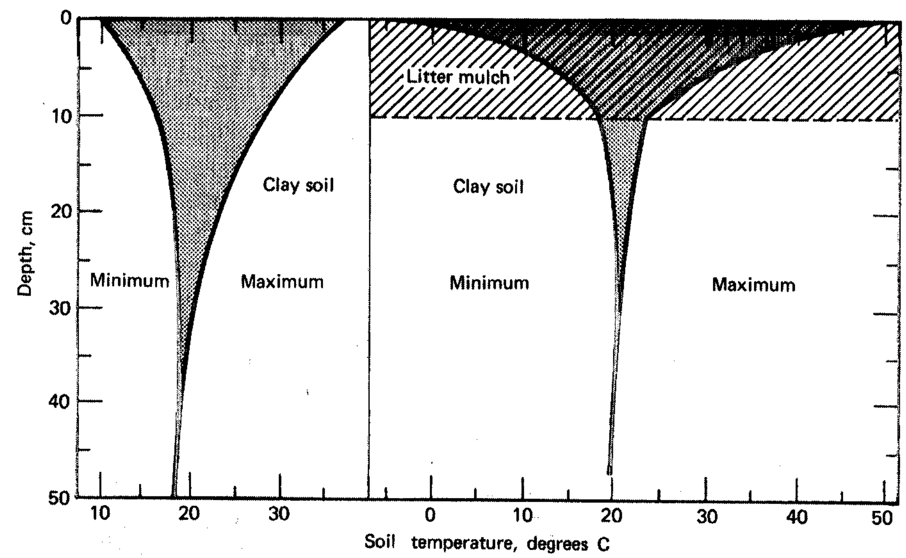

Figure 4.9 Daily temperature variations with depth for (left) an unmulched and (right) a mulched clay soil where the litter mulch has a lower thermal conductivity (Cochran, 1969).

TABLE 4.3 Snow and Frost Depths on North and South Slopes in Wisconsin (in cm)

	Snow		Frost	
Year and Date	North	South	North	South
1970 (February 25)	25	0	8	11
1971 (February 25)	55	25	0	0
1972 (March 1C)	45	25	16	23
1973 (March 01)	5	0	21	12

Source: Sartz (1973).

Wet soils are slower to change their temperature for two reasons: it takes a lot of energy to warm the water, and meanwhile the water transmits heat deeper into the soil. Organic matter has low conductance and impedes the movement of thermal energy.

Favorable soil temperature is essential for the germination of seeds and the survival and growth of seedlings. In some cold climates, a dense canopy may keep soil temperature so low that it delays germination and slows up seedling development. On the other hand, removal of the forest cover may result in increases in soil temperature to lethal levels for germinating seedlings. Soil surface temperatures vary among substrates, with high temperatures in dark residues (especially charred residues) and lower temperatures in moist, decayed logs (Figure 4.10). Clearcut areas have different radiation and wind regimes, which tend to accentuate daily swings between low nighttime and high daytime temperatures (Figure 4.11).

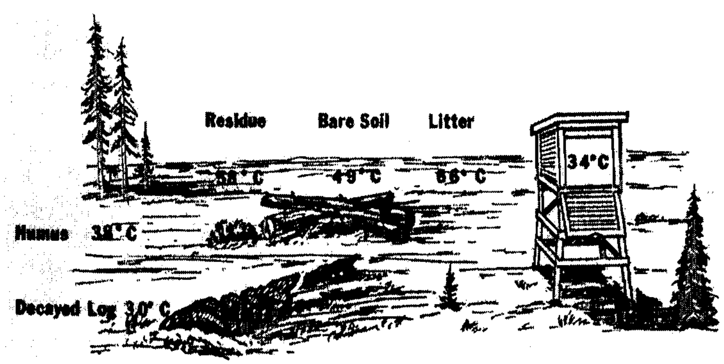

Figure 4.10 Surface temperatures of substrates in a clearcut area in the mountains of Montana (Hungerford, 1980).

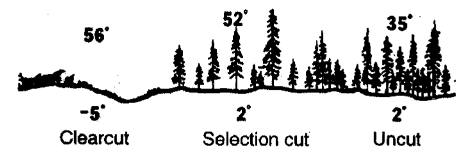

Figure 4.11 Average surface temperatures in uncut, partially cut, and clearcut units in the mountains of Montana (Hungerford, 1980). Night temperatures in the clearcut area fall below 0°C as a result of reduced turbulence without trees, which allows air at the soil surface to supercool; daytime temperatures reach extreme levels under high-altitude sunshine.

Root growth is also affected by soil temperatures, as shown by Lyford and Wilson (1966). They found that day-to-day variations in growth rates of red maple root tips closely paralleled variations in surrounding soil temperatures.

SOIL WATER IS PART OF THE HYDROLOGICAL CYCLE

The supply of moisture in soils largely controls the types of tree that can grow, and the distribution of forests around the world relates to patterns in precipitation and soil moisture. Water is essential to the proper functioning of most soil and plant processes. Carbon uptake by plants requires the moist interiors of leaves to be exposed to relatively dry air. Given the concentration gradient between wet leaves and dry air, about 1000 molecules of water are lost for every molecule of carbon dioxide acquired. In addition to serving the metabolic needs of the plant, water is critical for many functions in the soil. Water is a solvent and a medium of transport of plant nutrients, the medium for the action of the exoenzymes produced by soil microbes, and an important influence on soil temperature and aeration.

The retention and movement of water in soils and plants involves energy transfers. Water molecules are attracted to each other in a polymer-like grouping, forming an open tetrahedral lattice structure. This asymmetrical arrangement results from the dipolar nature of the water molecule. Although water molecules have no net charge, the hydrogen atoms sit to one side of the oxygen atom, giving the molecule a partial positive charge on one side and a partial negative charge on the other. As a result, the hydrogen atoms of one molecule attract the oxygen atom of an adjacent molecule. This hydrogen bonding (along with partial covalent bonding) accounts for the forces of adhesion, cohesion, and surface tension that largely regulate the retention and movement of water in soils. **Adhesion** refers to the attractive forces between soil surfaces and water molecules. At the water-air interface, **surface tension** may be the only force retaining water in soils. It results from the greater attraction of water molecules for each other (**cohesion**) than for the air.

Water with high free energy tends to move toward a zone of low free energy—from a wet soil to a dry soil and from the upper soil to the lower soil. The amount of movement depends on the differences in the energy states between the two zones; these differences are referred to as differences in potential. The potential of water in soil has three components. **Matric potential** is the attraction of water to soil surfaces; small pores hold water more tightly than larger pores do. Water that is low in dissolved ions tends to move into areas with more concentrated salts, which represents the **osmotic potential.** Water also tends to move toward the center of the Earth because of the **gravitational potential.** The total soil water potential is the combined effects of matric, osmotic, and gravitational potentials. In physics, matter tends to move from zones of higher potential to zones of lower potential, and the unit for potential is the Pascal ($=1$ Newton/m^2; 25 years ago, the units commonly used in the United States were bars or atmospheres, which equal about 100 kPa, or 0.1 MPa). Common soil water potentials range from near zero for very wet soils to -3.0 MPa (MegaPascal) or lower. The water potential at the bottom of a lake may have a positive potential because the mass of water above it has a positive pressure (or head).

Investigators have long attempted to devise useful equilibrium points or constants for describing soil moisture. Such terms as **field capacity** and **permanent wilting point** have found their way into soils literature over the years. Most of these terms deal with hypothetical concepts and do not apply equally well to all soil conditions. Nevertheless, they are employed commonly and may be of use.

Field capacity describes the amount of water held in the soil after gravity has drained most of the water that is easily drained. Field capacity is difficult to determine accurately. Factors such as soil texture, structure, and organic matter content influence measurements. Soil layers of different pore size markedly influence water flow through a profile, greatly affecting field capacity. Field capacity may be defined as the water content and water potential after 1 or 2 days of draining following full rewetting of the soil, or 0.03 MPa for silt loam soils and 0.01 to 0.001 MPa for sandier soils.

Permanent wilting point is a classic term for the soil moisture potential at which plants remain permanently wilted, even when water was added to the soil. Just as field capacity has been widely used to refer to the upper limit of soil water storage for plant growth, the permanent wilting point is used to define the lower limit. The use of a test plant, such as a sunflower, is the most widely accepted method for determining this soil water potential, but it has no particular relevance to trees.

Soil water-holding capacity is a useful term that refers to the quantity of water held within soils between the freely drained level of field capacity and some arbitrary potential beyond which plant uptake becomes minimal. For example, the Commerce silt loam holds about 45 percent moisture (water mass as a percentage of equivalent soil dry mass) at field capacity and about 10 percent moisture at -2 MPa, for a water-holding capacity of 35 percent of the

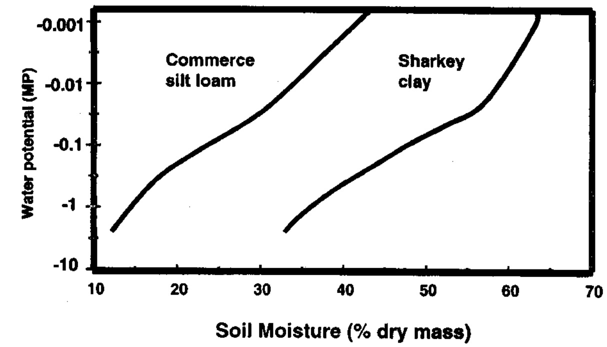

Figure 4.12 Relations of soil water potential to soil moisture content in Sharkey clay and Commerce silt loam (modified from Bonner, 1968).

soil's dry mass (Figure 4.12). Clays hold more water than silt loams, but they hold it more tightly. The Sharkey clay holds about 65 percent water at field capacity and 35 percent at -2 MPa, for a water-holding capacity of 30 percent of the dry soil mass. At 35 percent moisture, roots obtain water easily from the silt loam (0.01 MPa) but not from the clay (-2.0 MPa).

Soil water potential also affects the activities of microbes. A variety of studies have shown that carbon dioxide formation (an integrated measure of microbial activity) is reduced by half as soil water potential declines from saturated to -0.2 MPa (Sommers et al., 1980). Nitrogen mineralization also declines as soil moisture declines. One study in Kenya found that gross Nitrogen mineralization dropped from 2.2 to 0.08 μg/g daily as soil water potential fell from -0.06 to -5.9 MPa (Pilbeam et al., 1993).

WATER FLOW IN UNSATURATED SOILS DEPENDS ON WATER CONTENT AND HYDRAULIC CONDUCTIVITY

The availability of soil water to plants depends on its potential and on the hydraulic conductivity of the soil. In saturated soils, water uptake by trees is not limited by the rate of water movement through soils. As water drains from the soil, macropores empty and water is present only in capillary pores, which hold water with strong negative potential and also retard the flow of water. The rate at which water can move through a soil, **hydraulic conductivity**, is related to water-filled pore size and water potential. For example, conductivity at 0.02 Mpa would be about 10,000 times greater than at -1.0 MPa. At very negative potentials, water in sands is held only at points of contact between the relatively large particles. Under these conditions, there is no continuous water film and no opportunity for liquid movement. Layers of sand in a profile of fine-textured soil, therefore, can inhibit downward movement of water in a fashion similar to that of compact clay or silt pans.

Layering or stratification of materials of different texture is common in soils as a result of differences in the original parent material or pedogenic development. These factors can result in silt or clay pans or lenses of sand or gravel. Layers of fine-textured materials over coarse sands and gravel, as well as the reverse situation, are common in glacial outwash sands in glaciated terranes. Many Ultisols (highly weathered soils with low base saturation) have sandy A horizons underlain by clay-rich B horizons. Changes in texture throughout a profile tend to slow water movement whether the change down through the profile is from coarse to fine (the fine horizon has lower saturated conductivity) or from fine to coarse (water cannot leave the fine layer and enter the coarse layer except near saturation).

FACTORS AFFECTING INFILTRATION AND LOSSES OF WATER

The term infiltration is generally applied to the entry of water into the soil from the surface. Infiltration rates depend mostly on the rate of water input to the soil surface, the initial soil water content, and internal characteristics of the soil (such as pore space, degree of swelling of soil colloids, and organic matter content). Overland flow happens only when the rate of rainfall exceeds the infiltration capacity of the soil. Because of the sponge-like action of most forest floors and the high infiltration rate of the mineral soil below, there is little surface runoff of water in mature forests. Overland flow is rarely a problem in undisturbed forests even in steep mountain areas.

When rainfall does exceed infiltration capacity, the excess water accumulates on the surface and then flows overland toward stream channels. This surface flow concentrates in defined stream channels and causes greater peak flows in a shorter time than water that infiltrates and passes through the soil before reappearing as streamflow. High-velocity overland flow can erode soils. Soil compaction by harvesting equipment, disturbances by site preparation, and practices that reduce infiltration capacity and cause water to begin moving over the soil surface are of great concern to forest managers. Special care is warranted on shallow soils and those that have inherently low infiltration rates. Soils that are wet because of perched water tables or because of their position along drainage ways are unusually susceptible to reduction in porosity and permeability by compaction.

The litter layer beneath a forest cover is especially important in maintaining rapid infiltration rates. This layer not only absorbs several times its own mass of water, but also breaks the impact of raindrops, prevents agitation of the mineral soil particles, and discourages the formation of surface crusts.

The incorporation of organic matter into mineral soils, artificially or by natural means, increases their permeability to water as a result of increased porosity. Forest soils have a high percentage of macropores through which large quantities of water can move—sometimes without appreciable wetting of the soil mass. Most macropores develop from old root channels or from

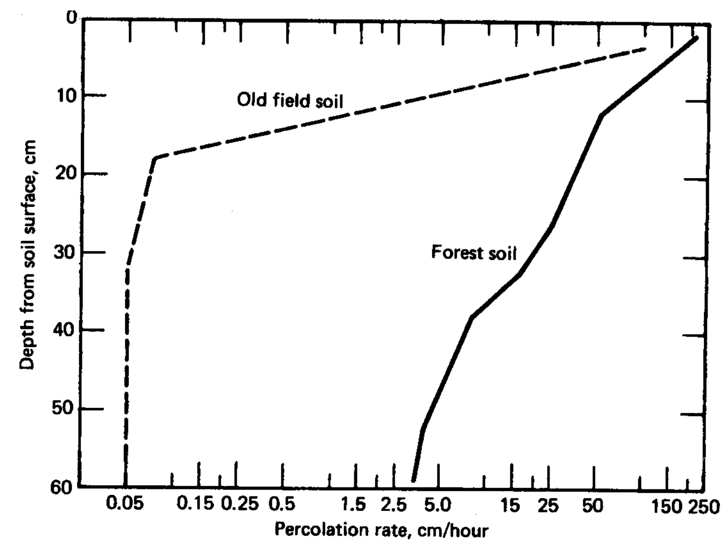

Figure 4.13 Comparative percolation rates by soil depth for a forest soil and an adjacent old-field soil in the South Carolina Piedmont (Hoover, 1949).

burrows and tunnels made by insects, worms, and other animals. Some macropores result from structural pores and cracks in the soil. As a consequence of the better structure, higher organic matter content, and presence of channels, infiltration rates in forest soils are considerably greater than in similar soils subjected to continued cultivation (Figure 4.13). Wood (1977) found that water infiltration rates were higher on 14 of 15 forest sites than on adjacent sites used for pastures or for pineapple or sugarcane crops in Hawaii. In this study, lower bulk densities and greater porosities were found in forest-covered soils than in nonforested soils.

The presence of stones increases the rate of water infiltration in soils because the differences in expansion and contraction between stones and the soil produce channels and macropores. However, stones reduced the retention capacity of soils for 40 Oregon sites (Figure 4.14).

Burning of watersheds supporting certain types of vegetation may result in a well-defined hydrophobic (water-repellent) layer at the soil surface (see Chapter 10), which may increase erosion rates. This heat-induced water repellency results from the vaporization of organic hydrophobic substances at the soil surface during a fire, with subsequent condensation in the cooler soil below.

Evaporation from surface soils increases with increases in the percentage of fine-textured materials, moisture content, and soil temperature and decreases with increases in atmospheric humidity. Evaporation losses are also influenced by wind velocity. Heyward and Barnette (1934) found that the upper soil layers of an unburned longleaf pine forest contained significantly more moisture than

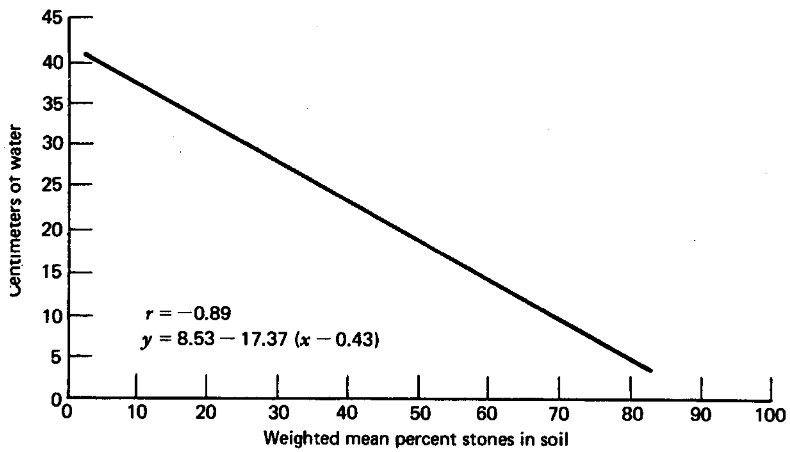

Figure 4.14 Retention storage capacity in the surface 120 cm of 40 soil profiles representing four soil series in Oregon as a function of average stone content (Dyrness, 1969).

did similar soils in a burned forest because of the mulching effect of the O horizon in the unburned areas.

Fine-textured soils have higher retention capacity for water than sands, and can store larger amounts of water following storm events. The effective depth of the soil (rooting depth) and the initial moisture content are factors that also influence the amount of rainwater that can be retained in soils for later use by plants.

The condition of a forest has a strong influence on soil water as a result of differences in water use and water input. Clearcutting a mixed conifer forest in Montana led to substantially wetter soils throughout the year (Figure 4.15). The difference was related to lower water use in the absence of trees but also to greater snow input (Figure 4.16). Snow that accumulates in conifer canopies

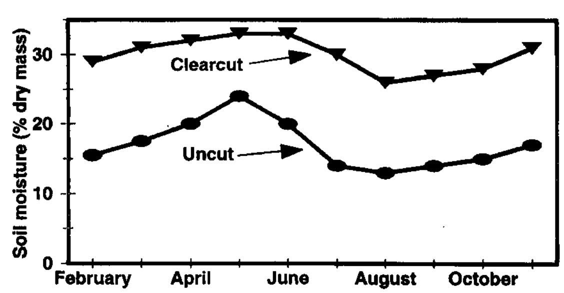

Figure 4.15 Seasonal course of soil moisture in a clearcut and an uncut forest in the mountains of Montana (data from Newman and Schmidt, 1980).

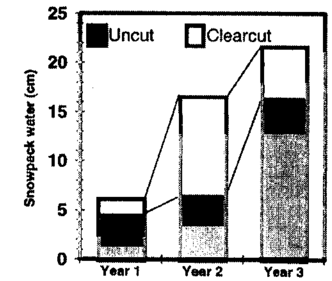

Figure 4.16 The water content of the snowpack across 3 years was substantial higher in clearcut areas than in adjacent mature conifer forests in the mountains of Montana (data from Newman and Schmidt, 1980).

may sublimate without reaching the soil surface, lowering the input of water to the soil relative to clearcut areas, with less sublimation (or interception loss).

Differences in interception loss can also be important among tree species that support substantially different leaf areas. For example, Helvey and Patric (1988) contrasted the interception loss for a watershed dominated by a mixed-age, mixed-species hardwood forest with a watershed planted with a single-age stand of white pine. Average annual interception loss from the mixed-hardwood stand (250 mm/yr) was less than half of the loss from the white pine stand (530 mm/yr). Soil leachates were more concentrated under the pine stand, due in part to the reduced quantity of water leaching through the soils (data in Johnson and Lindberg, 1992).

TREES MAY GET WATER FROM THE CAPILLARY FRINGE

The upper surface of the zone of saturation in a soil is called the groundwater table. Extending upward from the water table is a zone of moist soil known as the **capillary fringe**. In fine-textured soils, this zone of moisture may approach a height of 1 m or more, but in sandy soils it seldom exceeds 25 to 30 cm. In some soils, the height of the water table fluctuates considerably between wet and dry periods. In a forested soil in New England (Lyford, 1964), water tables were highest during late autumn, winter, and early spring; gradually lower in late spring and early summer; and higher again in the autumn. In the flatwoods of the southeastern coastal plain of the United States, the water table may be above the soil surface during wet seasons and fall to a depth of 1 m or more during dry seasons.

Soil depth often reflects the volume of growing space for tree roots above some restricting layer, and in wet areas this restricting layer may be a high water table. The water table may be a temporary (perched) zone of saturation above an impervious layer following periods of high rainfall.

Large seasonal fluctuations in the depth to the water table are harmful to root development for many species. Roots that develop during dry periods in the portion of the profile that is later flooded may be killed under the induced anaerobic conditions. A reasonably high water table is not detrimental to tree growth as long as there is little fluctuation in its level. In an experiment in the southeastern coastal plain of the United States, slash pines grown for 5 years in plots where the water table was maintained at 45 cm by a system of subsurface irrigation and tile drains were 11 percent taller than trees in plots where the water table was maintained at 90 cm, but they were 60 percent taller than trees in plots where the water table was allowed to fluctuate normally (White and Pritchett, 1970).

The primary benefits derived from drainage of wet soils may result more from an increase in the nutrient supply than from an increase in the soil oxygen supply. The improved nutrition is brought about by an increase in the volume of soil available for root exploitation, and by faster mineralization of organic compounds. For example, fertilizer experiments in nutrient-poor wet soils of the coastal plain indicated that tree growth was increased as much by draining as by fertilization (Pritchett and Smith, 1974).

Trees obtain moisture from the water table or the capillary fringe within reach of their deep roots even when the water table is at a considerable depth. Thus the water tables of some wet soils are lowered to a greater extent where trees are grown than where grasses and other shallow rooted plants are grown. Wilde (1958) reviewed a number of reports of general rises in the soil water table following removal of a forest cover from level areas in temperate and cold regions. The increase in the water table level results from a reduction in water losses from evaporation on leaf surfaces (interception loss) and from transpiration of water from within leaves.

TREES REQUIRE GREAT QUANTITIES OF WATER

Tree roots absorb vast quantities of water to replace that lost by transpiration and that used in metabolic activities. Under favorable conditions, this loss can be as much as 6 mm/day (60,000 L/ha) during summer. Trees use 500 to 1000 kg of water for each kilogram of dry matter produced. Tree roots can effectively exploit soil moisture even when the soil moisture content is relatively low. The mycelia of mycorrhizal fungi may play an important role in the extraction of soil, water and nutrients.

In spite of the efficiency of most tree root systems, the distribution of trees is controlled to a great extent by the water supply. Wherever trees grow, their development is limited to some degree by either too little or too much water. Even in relatively humid areas of the temperate zone, forest soils may be recharged to field capacity only during the dormant season or for a brief time after very wet growing periods.

PHYSICS IN A LANDSCAPE CONTEXT CAN DETERMINE SOIL CHEMISTRY

The physical properties described above need to be placed in a landscape context. Most forest soils are affected by their context on landscapes. Ridgetop soils may be excessively well drained. Midslope sites receive water and nutrients from upslope sites but lose some water to downslope sites. Lower-slope positions may receive more water and nutrients than leach away, and in some cases may become saturated and experience periods of low redox potential. These landscape issues are sometimes lumped together under the term topography, and many generalizations may be offered. We stress that landscape position (or topography) is very important in forest soils but that broad generalizations may not be helpful. For example, the sequence of soil moisture supply, nutrient supply, and overall fertility commonly increases from ridges, to midslopes, to toeslopes in relatively young soils on the sides of hills, mountains, and valley sides. Other situations, such as the *Eucalyptus* plantation on an Oxisol in Argentina, show exactly the reverse pattern (Figure 4.1).

Rather than try to establish generalizations about the role of landscapes in soil physics, we present some of the biogeochemistry of a single landscape sequence (catena) for a boreal forest in Sweden. The same biogeochemical factors are integrated in different ways at other sites, but this case study illustrates the sorts of features that emerge in landscape contexts.

The Betsele transect is located in northern Sweden in the valley of the Umeå River (Table 4.4, Figure 4.17). The transect is 100 m long, descending a gentle

TABLE 4.4 Stand and Soil Characteristics Along a 90-m Transect of Soils Developed in Sandy Till in Sweden

Feature	Upper End	Lower End
Dominant tree	Scots pine	Norway spruce
Site index (m at 100 years)	17	28
Basal area (m^2/ha)	22	32
Forest floor morphology	Thin Oa (L), indicating little soil faunal mixing Thick Oe (F) Thin Oi (H)	Thick Oi (H) only, indicating major soil fauna activity
Forest floor solution		
pH	3.8	6.4
NO_3 (μmol/L)	10	160
PO_4 (μmol/L)	300	8
Forest floor Extractable Fe + Al (mg/g)	2	35

Source: Data from Giesler et al. (1998).

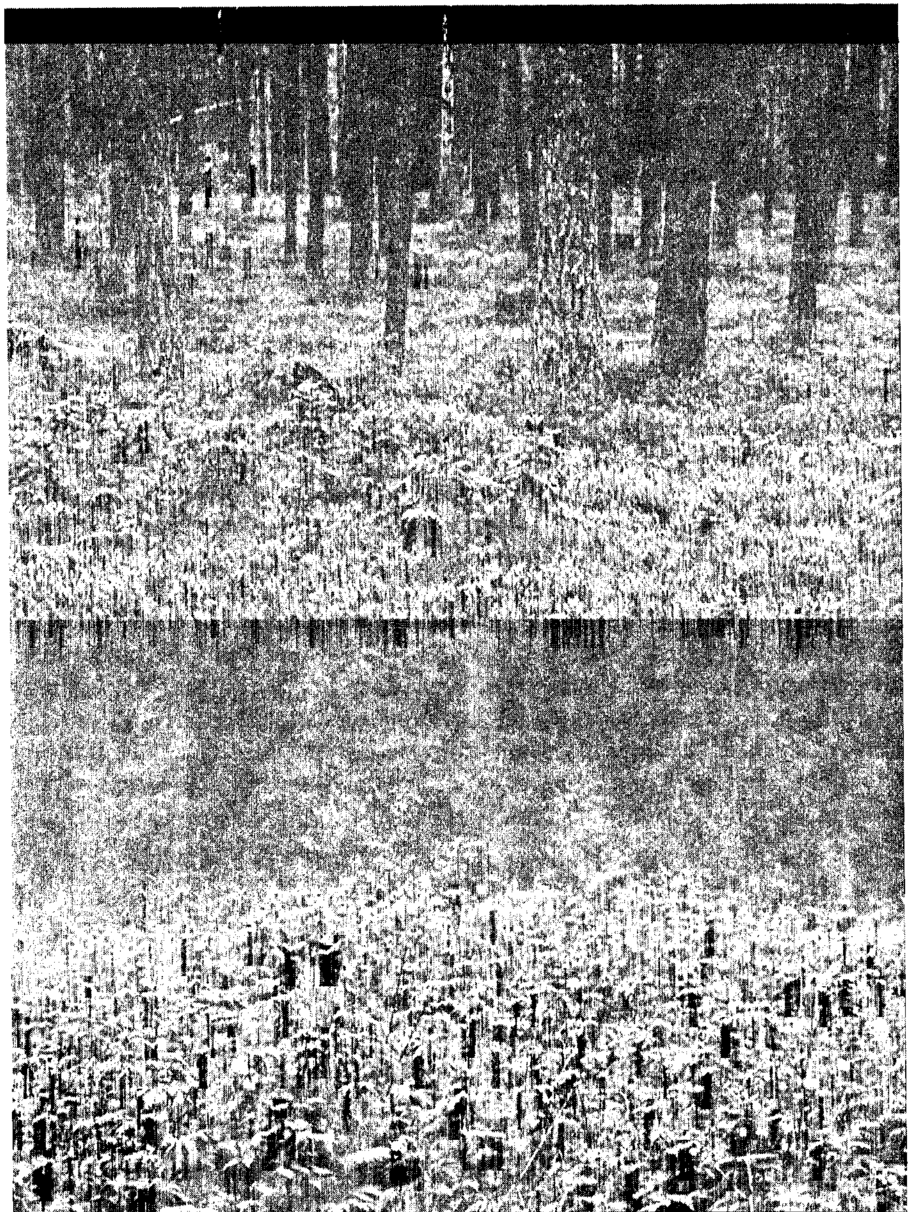

Figure 4.17 Endpoints of a landscape sequence from well-drained (upper) soils with Scots pine, Norway spruce, and *Vaccinium* to a discharge area (that receives groundwater inputs) with Norway spruce and a diverse, tall herbaceous layer (photos from Peter Högberg).

(2 percent) slope of soils formed from sandy till deposits. The upper end of the transect is dominated by a 125-year-old stand of Scots pine with an understory of dwarf shrubs (mostly *Vaccinium*). The lower end is dominated by a Norway spruce stand of the same age, with a variety of tall herbs in the understory. The upper end is much less productive (site index of 17 m at 100 years) than the lower end (site index 28 m at 100 years).

The key driver of the differences in soil chemistry and overall soil fertility is soil water. The upper end of the transect is excessively drained; water leaches downward out of the soil. The lower end is in a position to receive the "discharge water" from the upslope ecosystems, providing abundant water supply and changing nutrient cycles. The major biogeochemical implications are that the upper ecosystems acidify and are nitrogen limited, and the lower ecosystems receive abundant alkalinity (acid-neutralizing capacity) from the discharge water and have high nitrogen availability. One surprising feature is the low (and limiting) concentrations of phosphorus at the lower slope position. Here, mixing of the organic litter with mineral soil brings phosphorus into contact with iron and aluminum, allowing sorption and removal of phosphorus from the soil solution (similar to the earthworm story in Paré and Bernier, 1989; see Chapter 9). Alternatively, Giesler et al. (1998) suggest that iron and aluminum enrichment of the forest floor at the toeslope may have resulted from dissolved inputs, particularly iron, which may have high solubility during anaerobic periods under saturated conditions. Overall, the authors conclude that the 3-unit gradient in soil pH was driven primarily by the differences in base saturation (caused by input of alkalinity into discharge water) (see Chapter 5) rather than by differences in acid quantity or strength.

SUMMARY

The physics of soils is fundamental to soil temperature, water relations, chemistry, and the life that depends on the soil. Differences in soil physics among soils, or within a soil over time, have major influences on tree growth. Soil texture refers to the size of mineral particles in the soil, and soil structure concerns the three-dimensional conglomerations of mineral particles and organic matter. The pore space within soils is important in influencing water infiltration, soil atmosphere composition, and ease of penetration by roots. Forest management operations may affect soil structure, especially by altering soil pore space and bulk density. Soil temperature regimes differ by regions and management activities, driving differences in the rates of soil processes (such as decomposition) and in the species composition and growth of forests. Water is held under negative potential within freely draining soils, and the characteristics of water retention and release as a function of water potential determine the availability of water to plants. Forest harvest reduces water losses through interception/evaporation and transpiration, increasing water yield from forest lands and the average soil water content. The physics of soils is often influenced by adjacent areas where upslope ecosystems supply water, chemicals, and sometimes material that affect both the biota and long-term development of soil structure and other properties.

CHAPTER 5

Soil Chemistry and Nutrient Uptake

The chemistry of soils is fascinatingly complex, involving inorganic reactions between solid phases (including minerals, mineral surfaces, and organic matter), the liquid phase (near surfaces and in the bulk soil solution), and an incredible diversity of soil organisms. In this chapter, we focus on the chemistry of the soil solution as the pool of chemicals available for plant use. The soil solution is a dilute soup of dozens of chemicals, with the soup supplied with chemicals from atmospheric inputs and from the solid phases, and depleted by processes of plant uptake and movement into the solid phases (Wolt, 1994). The broader aspects of biogeochemistry determine the amount of flow (or flux) of chemicals through the soil solution on an annual scale. In this chapter, we focus on shorter-term issues of what comprises the soil solution, how the soil solution is buffered by relatively rapid interactions with the solid phases, and how the soil solution provides nutrients for plant uptake.

MAJOR SOIL ANIONS INCLUDE CHLORIDE, SULFATE, BICARBONATE, AND SOMETIMES NITRATE

Soil solutions contain dissolved chemicals, and many of these chemicals carry negative charges (anions) or positive charges (cations). The total charge of anions in forest soils commonly ranges between 100 and 500 $\mu mol_c/L$ (micromoles of charge per liter, or 10^{-6} moles of charge per liter), and maintenance of electroneutrality requires an equivalent concentration of cations. Figure 5.1 illustrates the anion composition of soil solutions from A horizons in six temperate forests. The Norway spruce forest in Norway has a soil solution dominated by chloride and sulfate. The chloride shows a moderate influence of inputs of salt from the ocean, and the sulfate concentrations indicate substantial deposition of sulfur from polluted air. The loblolly pine soil in North Carolina, in the United States, has some chloride, a large amount of sulfate, and some bicarbonate or alkalinity. As described later, soils contain high partial pressures of CO_2 as a by-product of respiration, and the CO_2 dissolves into the soil solution to form carbonic acid. Depending on the pH of the soil solution, some of the carbonic acid may dissociate to form H^+ and bicarbonate (HCO_3^-). Both the Norway spruce and loblolly pine soils have carbonic acid