2 Physical Properties of Forest Soils

Physical properties of forest soils develop under natural conditions by the influence of permanent vegetation over a long period of time. Physical properties of forest soils may be almost permanent properties unless modified by harvesting operations, shifting cultivation, and forest fires. Important physical properties of forest soils include texture, structure, porosity, density, aeration, temperature, water retention, and movement. The physical properties of forest soils affect every aspect of soil fertility and productivity. Soil physical properties determine the ease of root penetration, the availability of water and the ease of water absorption by plants, the amount of oxygen and other gases in the soil, and the degree to which water moves both laterally and vertically through the soil. Soil physical properties also influence the natural distribution of forest tree species, growth, and forest biomass production. However, soil physical properties are largely controlled by the size, distribution, and arrangement of soil particles.

2.1 Soil Particles

Soils are composed of variously sized particles. There are two types of particles—primary particles and secondary particles. Individual discrete particles are called primary particles and their aggregates are known as secondary particles. Particles greater than 2 mm diameter are known as gravels which include pebbles (2–7.5 cm), cobbles (7.5–25 cm), stones (25–60 cm), and boulders (>60 cm). Although particles larger than 2 mm are less common and they hardly affect soil fertility and productivity, many productive forests have developed on gravelly or stony soils. Primary particles with the maximum "effective diameter" of 2 mm are classified into three categories—sand, silt, and clay. Effective diameter, which is the diameter of a sphere that has a velocity of fall in a liquid medium equal to the particle in question, has been considered because soil particles are not all spherical. The dimensions of the different categories of soil particles differ between the United States Department of Agriculture (USDA) and the International Soil Science Society (ISSS) classification systems (Table 2.1).

Table 2.1 Dimensions of sand, silt, and clay particles

ISSS System		USDA System	
Soil particles	Diameter (mm)	Soil particles	Diameter (mm)
Coarse sand	2.00–0.20	Very coarse sand	2.00–1.00
Fine sand	0.20–0.02	Coarse sand	1.00–0.50
Silt	0.02–0.002	Medium sand	0.50–0.25
Clay	<0.002	Fine sand	0.25–0.10
		Very fine sand	0.10–0.05
		Silt	0.05–0.002
		Clay	<0.002

Soil separates do not differ in size alone; there are important differences in their physical behavior and mineralogical makeup.

Characteristics of Soil Particles

Sand: Sand particles are mainly fragments of quartz and some feldspars and mica. They have little surface area exposed (0.1 $m^2\ g^{-1}$ specific area). Sand particles are visible to the naked eye, gritty in feeling, having little or no capacity to hold water or nutrients, and bind other particles. They are loose when wet, very loose when dry. Sand does not absorb water and does not exhibit swelling and shrinkages, stickiness and plasticity.

Silt: Silt particles are also fragments of primary minerals. Most silt particles are not visible to the naked eye, but can be seen through an ordinary microscope. They feel smooth when wet and like talcum powder when dry. They have low to medium capacity to attract water, nutrients and other particles. Because of adhering film of clay, they exhibit some plasticity, cohesion, adhesion and absorption and can hold more amount of water than sand but less than clay.

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Clay: Clay particles are mainly secondary minerals such as kaolinite, smectite, vermiculite, illite, chlorite, hydrated oxides of Fe and Al, etc. Clay particles can be seen by an electron microscope and have large surface area (10–1000 $m^2\ g^{-1}$). They have electrical charges, both negative and positive, on their surfaces. Because of these properties, clays have high water and nutrient holding capacity and they participate in chemical reactions in the soil.

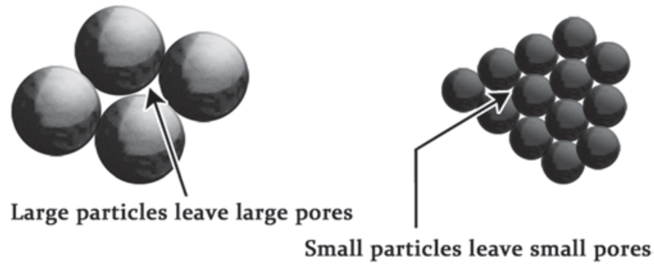

Fig. 2.1 Pores left by soil particles of different sizes

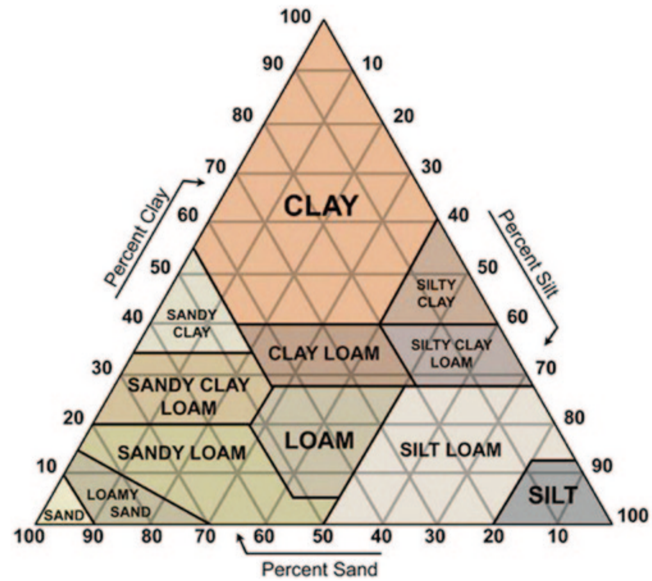

Fig. 2.2 Soil textural triangle devised by the USDA

2.2 Soil Texture

Soil texture refers to the degree of fineness or coarseness created by the close packing of variously sized particles together in a soil. It is determined by the relative proportion of sand, silt, and clay in a soil. Soils are rarely composed of a single size class of particles; they are mixtures of different size classes. However, one or two size classes usually dominate the physical behavior of the soil. The soil is then named after the name of that soil separate. Thus, a sandy soil displays the properties of sand particles. When a soil equally exhibits the properties of sand, silt, and clay, then it is called loam (approximately 40 % sand, 40 % silt, and 20 % clay). A higher proportion of sand in loam produces sandy loam. In this way, 12 textural classes have so far been identified. They are (from coarse to fine) sand, loamy sand, sandy loam, loam, silt loam, silt, sandy clay loam, clay loam, silty clay loam, sandy clay, silty clay, and clay. Soil texture is not usually changed by management practices. Soil texture is inherited from the parent materials and it originates through weathering and pedogenic processes, including recrystallization, eluviation, and illuviation. It may, however, be altered by erosion, deposition, truncation, landfill, etc.

Sand particle is coarse, clay is fine, and silt is medium. When sand particles are packed together (i.e., if the soil is sandy), they leave larger gaps or "pores" between them (Fig. 2.1). On the other hand, gaps between clay particles are small. Larger pores, the macropores, generally accommodate air and smaller pores, whereas the micropores contain water. Thus, the coarseness or fineness of soil determines its air–water relationships. Texture is one of the most important physical properties of soil that affects its fertility and productivity. Actually, the whole soil environment is regulated by soil texture.

Soil texture is determined in the laboratory by a technique based on the velocity of fall of a particle in a liquid medium, which is proportional directly to the square of the radius of the particle and inversely to the viscosity (a fluid's internal resistance to flow) of the liquid (Stokes' Law, Hillel 1980). Percentages of sand, silt, and clay are determined by either "pipette method" or "hydrometer method" (Day 1965). Soil textural class names can be obtained from the "USDA textural triangle" (Fig. 2.2), if the percentages of any two size fractions are known. For example, lines for 40 % sand and 20 % clay intercept in the area demarcated as loam. Experienced foresters and soil survey personnel have learned the technique of determining soil texture in the field fairly accurately by feeling a lump of soil between their fingers. Readers are referred to Thien (1979) who has provided a flowchart for the determination of texture by the feel method. (For this section, the reader may further consult Foth 1990 and Brady and Weil 2002.)

2.2.1 Suitable Forest Tree Species for Different Soil Textures

A tree species may occur on a variety of soil textures in their natural habitat, but their best growth occurs in the most suitable sites. For example, black cherry (*Prunus serotina* Ehrh.) grows well on a wide variety of soils throughout its range in eastern North America, if summer growing conditions are cool and moist. On the Allegheny Plateau, black cherry develops well on all soils except for the very wettest and very driest (Hough 1965). Site quality does not differ between soils developed from glacial till and from residual parent material. It tolerates a wide range of soil drainage and grows about the same on well-drained sites as on somewhat poorly drained sites but shows rapid loss in productivity with increasingly wetter conditions (Becker et al. 1977; Cerutti et al. 1971). Cottonwood (*Populus deltoides*) does well in a heavy cold damp soil (Chittendon 1956; Duke 1983) but thrives best on moist well-drained, fine sandy loams or silts close to streams (Duke 1983). On the other hand, European larch (*Larix decidua*) grows better in fertile soil consisting of loam, sandy loam, or silty loam. Shade and poorly drained conditions are not well tolerated (IUCN 2006). Hackberry (*Celtis occidentalis*) grows in many soils, and although principally a bottomland tree, it is frequently found on limestone outcrops or limestone soils (usually fine textured). In western Nebraska, hackberry grows on the north side of sand dunes and in river valleys (Eyre 1980).

Sandy soils are coarse textured soils being loose and friable; they absorb water rapidly and drain it quickly, and can be worked easily in both moist and dry conditions. They are called "light textured soils." They are usually poor in fertility and suffer from water scarcity. Sandy soils are widely distributed in the tropics occupying most of arid and semiarid areas. For instance, the total estimated extent of Arenosols is 900 million hectares, mainly in Western Australia, South America, South Africa, Sahel, and Arabia (FAO 2006). These soils are characterized by a low soil organic carbon, a low cation exchange capacity, a high risk of nutrient leaching, a low structural stability, and a high sensitivity to erosion and to crusting (Pieri 1992; Sanchez and Logan 1992). Some tree species requiring low water and nutrients can grow there; some even prefer sandy soils but growth of most tree species in sandy soil is stunted. The following is a list of tree species that can be found naturally growing well on sandy soils, or can be grown successfully on sandy soils.

List of trees suitable for sandy soils

Common name	Scientific name
Amur Maple	*Acer ginnala*[a]
Black cherry	*Prunus serotina*[a]
Black locust	*Robinia pseudoacacia*[a]
Box Elder	*Acer negundo*[a]
Canadian spruce	*Picea alba*[a]
Cedar wattle	*Acacia elata*[b]
Heath-leaved banksia	*Banksia ericifolia*[b]
Chinese juniper	*Juniperus chinensis*[a]
Coastal she-oak	*Casuarina equisetifolia*[b]
European white birch	*Betula alba*[a]
Gray birch	*Betula populifolia*[a]
Hedge maple	Acer campestre[a]
Hop tree	*Ptelea trifoliate*[a]
Large tooth aspen	*Populus grandidentata*[a]
Monarch Birch	*Betula maximowicziana*[a]
Norway spruce	*Picea excels*[a]
Pignut	*Carya glabra*[a]
Pitch pine	*Pinus rigida*[a]
Quaking aspen	*Populus tremuloides*[a]
Red cedar	*Juniperus virginiana*[a]
Red swamp banksia	*Banksia occidentalis*[b]
Scarlet oak	*Quercus coccinea*[a]
Scots pine	*Pinus sylvestris*[a]
Sour cherry	*Prunus cerasus*[a]
Swiss mountain pine	*Pinus Montana*[a]
Tatarian Maple	*Acer tataricum*[a]
Tree of Heaven	*Ailanthus glandulosa*[a]
White pine	*Pinus strobes*[a]
White poplar	*Populus alba*[a]

[a] http://www.backyardgardener.com/tree/indexlist5.html
[b] Evans (2010) Native Australian trees suited to sandy soils

Loam soils are medium textured soils having generally more nutrients and humus than sandy soils, and have better infiltration and drainage than clay soils. Loams are desirably porous and retain sufficient water, nutrients, and air. Unless adapted in extreme textures, most tree species do well in sandy loam to clay loam soils for adequate water, air, and nutrients. To mention a few—Balsam fir (*Abies balsamea*), Basswood (*Tilia americana*), and Hackberry (*Celtis occidentalis*) (USDA NRCS 1995); Bitternut hickory (*Carya cordiformis*) (Voss 1985), European larch (*Larix decidua*) (Matras and Paques 2008), Gamar (*Gmelina arborea*) (Onyekwelu et al. 2006), Mahagoni (*Swietenia mahagoni*), and Garjan (*Dipterocarpus turbinatus*) (Hossain 2012); and Red maple (*Acer rubrum*) (Abrams 1998), Red oak (*Quercus rubra*) (Celine et al. 1996), Red pine (*Pinus resinosa*) (Leaf et al. 1971), Teak (*Tectona grandis* (Zanin 2005) White ash (*Fraxinus americana*) and White oak (*Quercus alba*) (USDA NRCS 1995).

Clay soils are generally compact and stiff; sticky when wet and hard when dry and require much energy to work in both wet and dry conditions. They are called "heavy textured soils." Clay soils retain large amount of water but drain very slowly. They are often waterlogged. They are fertile but plants usually suffer from oxygen stress in clay soils. Suitable tree species for clay soils are listed as follows:

List of trees suitable for clay soils

Common name	Scientific name
Amur corktree	*Phellodendron amurense*[c]
Bitternut hickory	*Carya cordiformis*[c]
Black ash	*Fraxinus nigra*[b]
Butternut	*Juglans cinerea*[c]
Common hackberry	*Celtis occidentalis*[b]
Common honey locust	*Gleditsia triacanthos*[c]
Cottonwood	*Populus deltoides*[c]
Crabapple	'Prairie Fire' Malus spp[a]
European alder	*Alnus glutinosa*[c]
European larch	*Larix deciduas*[c]

(continued)

Common name	Scientific name
Golden Black Spruce	Picea mariana[a]
Green ash	*Fraxinus pennsylvanica*[b]
Hawthorn	*Crataegus species*[c]
Japanese Plume Cedar	Cryptomeria japonica[a]
Kentucky coffee tree	*Gymnocladus dioicus*[c]
Norway maple	*Acer platanoides*[c]
River Birch	Betula nigra[b]
Shagbark hickory	*Carya ovate*[c]
Silver maple	*Acer saccharinum*[b]
Southern Magnolia	*Magnolia grandiflora*[a]
Tamarack	*Larix laricina*[c]

[a] http://voices.yahoo.com/trees-heavy-clay-soil-4924142.html. Accessed 7 June 2013

[b] Ware (1980) Selecting trees for clay soils. Metro Tree Impr Alliance (METRIA) Proc. 3:102–106

[c] http://www.extension.umn.edu/yardandgarden/ygbriefs/h-claytrees.html. Accessed 7 June 2013

2.2.2 Soil Texture and Species Distribution

According to Ramade (1981), soil texture governs most of the properties of the soil, its permeability, its capacity to retain water, its degree of aeration, its ability to make the nutrients stored in the clay–humus complex available to plants, its ability to withstand mechanical working of the top soil, and, finally, its ability to support a permanent plant cover. In the Lapland of Finland, Sepponen et al. (1979) demonstrated that pine stands are found on more coarse-grained soils than spruce stands. Jha and Singh (1990) suggested that soil texture is an important factor in the constitution and distribution of dry tropical forest communities. Soil texture was found to be largely responsible for the distribution of hardwood species within an old growth forest in the Sandhills region of southeastern USA. The higher clay content (and presumably higher moisture and fertility) of the upland soils have allowed for the development of a forest type that closely resembles the oak-hickory forests (Gilliam et al. 1993). Levula et al. (2003) found an obvious relation between soil particle-size distribution and tree species composition in a forest stand of Finland. The basal area (BA) of pine and the pine–spruce BA ratio correlated positively, and the BA of spruce negatively, with the mean particle size of the B1 horizon. Also, the coarse sand and gravel fraction in the B1 horizon correlated positively with the BA of pine and negatively with the BA of spruce. Scull and Harman (2004) studied species distribution and site quality in forests of southern lower Michigan, USA. They observed that poorly drained sites (depending on whether they were sandy or loamy) supported either pine communities (on the coarser sandy sites) or broad-leaved communities [ash, elm (*Ulmus* spp.), red maple, sugar maple, and yellow birch] mixed with the needle-leaved species. Well-drained sandy sites supported several different vegetation mixes of pine with northern pin oak (*Quercus ellipsoidalis*) and black oak (*Q. velutina*) depending on whether the sand was aeolian, outwash, or morainic. Well-drained loamy soils were broken down into two groups. Sugar maple and basswood were most distinctive of silt loams, whereas clay loams supported populations of maple, elm, yellow birch, and occasionally, hemlock (*Tsuga canadensis*), and balsam fir. Four soil types have been identified in a 52-ha forest dynamics plot in Bornean mixed dipterocarp forest (ranked by increasing fertility and moisture: sandy loam, loam, fine loam, and clay). The distributions of 73 % of tree species in the plot are significantly aggregated on one of these soil types (Russo et al. 2005).

2.3 Soil Structure

Soil structure is the arrangement of soil particles into units of different sizes and shapes. These units are called peds or aggregates and the processes of formation of peds are collectively called aggregation. According to Lal (1991), soil structure refers to the size, shape, and arrangement of solids and voids, continuity of pores and voids, their capacity to retain and to transmit fluids and organic and inorganic substances, and ability to support vigorous root growth and development. Peds differ from “clods” and “concretions”; clods or chunks are artificially formed (such as by plowing) hard soil mass. Concretions are hard lumps produced by the precipitation of dissolved substances (usually iron and manganese oxides). In some soils and sediments, the particles are not aggregated but remain separated; such soils are called single grained, such as some sandy soils. Most soils are structured soils. In some soils such as heavy clays, all the particles adhere together. Structure of these soils is called massive.

Soil structure determines pore-size distribution, which affects water flow and erosion potential (White 1985), microbial and faunal behavior (Edwards and Bremner 1967; Elliott et al. 1980; van Veen et al. 1984), and organic matter dynamics (Campbell and Souster 1982; Tisdall and Oades 1982; Schimel et al. 1985; Schimel 1986). Aggregation may affect nutrient turnover by controlling microbial predation (van Veen et al. 1984) and by protecting organic matter from microbial degradation (Young and Spycher 1979; McGill et al. 1981; Van Veen and Paul 1981; Voroney et al. 1981; Tisdall and Oades 1982). Aggregates influence microbial community structure (Hattori 1988), limit oxygen diffusion (Sexstone et al. 1985), regulate water flow, determine nutrient adsorption and desorption (Linquist et al. 1997; Wang et al. 2001), and reduce run-off and erosion (Barthes and Roose 2002). All of these processes have profound effects on soil organic matter dynamics and nutrient cycling (Six et al. 2004).

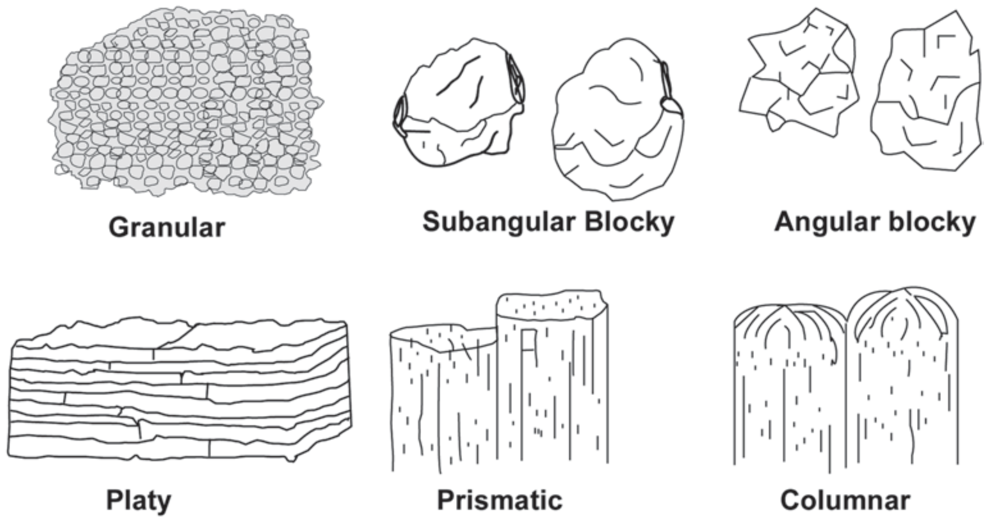

Fig. 2.3 Approximate shapes of different soil structures

Soil structure is classified based on shapes of peds into types, on size of peds into classes, and on distinctness and stability of aggregates into grades. There are four types of soil structure—spheroidal (granular and crumb), block-like (angular blocky and sub-angular blocky), plate-like, and prism-like (prismatic and columnar) (Fig. 2.3). On the basis of size of aggregates, soil structure is divided into five classes: very fine, fine, medium, coarse, and very coarse. Grades of soil structure are identified on the basis of the visibility in a horizon and stability of aggregates. Stability of aggregates refers to their resistance to destruction by water. Three structural grades are identified: weak (poorly formed, indistinct peds which are not durable), moderate (moderately well-developed peds which are fairly durable), and strong (very well-formed peds which are quite durable and distinct) (Hillel 1980).

2.3.1 Soil Structure Formation

Aggregation results from complex interactions of many factors including the environment, soil management, plant influences, and soil properties, such as mineral composition, texture, soil organic matter, pedogenic processes, microbial activities, exchangeable ions, nutrient reserves, and moisture availability (Kay 1998). There are several mechanisms of aggregation. Aggregates are formed in stages, with different bonding mechanisms dominating at each stage (Tisdall and Oades 1982). According to Duiker et al. (2003), aggregation results from the rearrangement of particles, flocculation, and cementation. Clay particles are electrically charged particles. They carry both negative and positive charges, but there are, generally, higher number of negative charges on their surfaces. They attract cations. When clay particles adsorb and get satisfied largely with monovalent cations like Na^+ and K^+, they remain dispersed in a liquid medium. Clay particles may come closer in a suspension only when they are flocculated. Flocculation (a process wherein colloids come out of suspension in the form of flock or flakes by the addition of a clarifying agent) is effected by replacing monovalent cations on colloidal surfaces by polyvalent cations like Ca^{2+}, Mg^{2+}, Fe^{2+}, Fe^{3+}, Al^{3+}, etc. Flocculation of clay particles forms clay flocks which are cemented together to form microaggregates. Microaggregates are bound together to produce macroaggregates. However, flocculation alone is not enough for aggregation; it involves a combination of different processes such as hydration, pressure, dehydration, etc. and requires cementation of flocculated particles. Well-known natural organic cements are the polysaccharides—levans and dextrans, polyuronides, etc. They stabilize the aggregates when they are dehydrated. Dehydrated oxides of Fe and Al bind soil particles firmly together. Colloidal clays are strong cementing agents; their cementation depends on their mineralogical composition, specific surface, and the state of deflocculation and dehydration.

Sand and silt particles are inert materials. They can come closer but cannot hold themselves together because they do not possess the power of adhesion and cohesion. Clays form coatings on them and cement several sand and silt particles into a larger unit. Soil particles are bound mechanically by plant roots and fungal hyphae. There are also various chemical compounds which act as cements and gums in soils. These materials include flocculated clays, hydrated oxides of Fe and Al, organic compounds produced by root exudation, microbial synthesis and excretion and organic matter decomposition, polyvalent cations, etc. (Osman 2013). Thus, soil particles aggregate as a result of physical binding by roots and hyphae, gluing by bacterially produced polysaccharides, and physical–chemical interactions between silicate clay surfaces and functional groups of partially decomposed organic matter (Aspiras et al. 1971; Aringhieri and Sequi 1978; Tisdall and Oades 1979; Sorensen 1981).

2.3.2 Soil Structure and Tree Growth

Soil structure, which modifies soil texture, exerts important influences on the edaphic conditions and the environment (Bronick and Lal 2005). Soil structure regulates pore size, number of pores, and distribution of pores and total porosity of the soil. Thus, retention and movement of soil water including infiltration, permeability, percolation, drainage, leaching, etc. all depend on soil structure. Soil structure affects tree growth by influencing root distribution and the ability to take up water and nutrients (Pardo et al. 2000). Soil structure facilitates oxygen and water infiltration and can improve water storage (Franzluebbers 2002). Microbial processes like organic matter decomposition, mineralization, stabilization, nitrification, and nitrogen fixation are influenced by soil structural conditions. Passioura (1991) reviewed literature on the relationship of soil structure and plant growth and suggested that the root–soil contact, which influences water and nutrient uptake, depends on soil structure. Aggregation may affect nutrient turnover by controlling microbial predation (Van Veen et al. 1984) and by protecting organic matter from microbial degradation (Voroney et al. 1981).

2.4 Density, Porosity, and Compaction

Density is defined as the mass per unit volume. There are two kinds of soil density—bulk density and particle density. Bulk density is estimated by dividing the mass (dry weight, usually after oven drying at 105 °C until constant weight) by total volume of soil (volume of solid + volume of pores) and particle density is estimated by dividing mass by the volume of the solids.

If a core of soil has a total volume V_T, which is the sum of the volume of solids, V_S and volume of pores, V_P and if the dry weight of the soil is W_S, then

$$\text{Bulk density, } D_b = \frac{W_s}{V_T}, \text{ and}$$

$$\text{Particle density, } D_p = \frac{W_s}{V_T}$$

Bulk density can be determined by the core method. Standard stainless steel cores with sharp edges may be used. The cores are pushed into the soil with the help of a wooden hammer at soft strokes so that the natural structure is not disturbed. After collection, the core with the soil is dried in an oven at 105 °C for 24 h and the dry weight of the soil is divided by the volume of the core to obtain bulk density. Particle density may be measured by a pycnometer (Pall and Mohsenen 1980).

Bulk density in surface soils in oak and pine forests of Indian Central Himalaya was found to range between 1.24 and 1.94 Mg m^{-3} (Jina et al. 2011). Bulk density of soil varies due to organic matter content, texture, compaction, and porosity of the soil. Histosols have very low bulk density (about 0.70 Mg m^{-3}; Brady and Weil 2002) and particle density (< 1.0 Mg m^{-3}). Loose soils have lower bulk densities and compact soils have higher bulk densities. Coarse textured or sandy soils tend to have a higher bulk density. Average bulk density of a loam-textured mineral soil is taken to be 1.3 Mg m^{-3}. Forest soils are more porous soils and have lower bulk density than cultivated soils. Bulk density is an important physical property. It is needed in order to convert soil chemical data, routinely expressed as mass concentrations, into amounts per unit area or unit volume (Viro 1951; Tamminen 1991). As it is an index of soil compaction, bulk density affects site productivity (Strong and La Roi 1985; Gale and Grigal 1987) and tree growth (Hamilton and Krause 1985; Froehlich et al. 1986).

Problem

The internal diameter of a soil core is 4.2 cm and the length of the core is 6 cm. You have collected a full core field moist soil, weighed and dried in an oven at 105 °C for 24 h. Weight of the empty core (W_1) is 160 g, weight of core + field moist soil (W_2) is 320 g, weight of core + dry soil (W_3) is 275 g. Calculate field moisture content and bulk density.

Solution

Volume of core, $V = \pi r^2 L = 3.14 \times 2.1^2 \times 6 = 83.08$ cm^{-3}
Weight of moist soil, $W_4 = W_2 - W_1 = 320 - 160 = 160$ g
Weight of dry soil, $W_5 = W_3 - W_1 = 275 - 160 = 115$ g
Weight of moisture $W_6 = W_2 - W_3 = 160 - 115 = 45$ g
% field moisture content $= (W_6 \div W_5) \times 100 = 39\,\%$, and
Bulk density $= W_5 \div V = 115 \div 83.08 = 1.38$ g cm^{-3} $= 1.38$ Mg m^{-3}.

Bulk density may differ in different depths of soil. In forested mineral soils of central and northern Finland, Tamminen and Starr (1994) observed that bulk density rapidly increased with depth in the surface but remained uniform at depths > 20 cm. This was related with distribution of organic matter and compaction. Bulk density tends to increase with depth primarily due to the lack of organic matter and aggregation.

Density of dominant soil forming silicate minerals ranges from 2.60 to 2.75 Mg m^{-3}. Therefore, particle density of mineral soils usually remains within this range. Mineral soils are considered to have an average particle density of 2.65 Mg m^{-3}. Bulk density and particle density values are used to estimate the porosity (the percentage of soil volume that is void) according to the following relationship:

$$\text{Porosity} = \left\{1 - \frac{\text{Bulk density}}{\text{Particle density}}\right\} \times 100$$

Spaces in soil that are not occupied by solids are called pores and the total volume of pores is the porosity. Soil pores are filled with either air or water, or both depending on the soil moisture content, and condition of rainfall and irrigation. Roots and soil organisms, both macro and microflora and fauna, occupy these pores. There are small or large pores depending on the texture and structure. Large pore spaces allow fast infiltration and percolation of water through a soil. On the other hand, small pores have strong attractive forces to hold water in the pore. Large pores, also called macropores (>0.75 mm), accommodate air and conduct water into the soil where they fill the micropores (<0.75 mm). Clay soils have numerous micropores and hold large quantities of water, but since they have few macropores, they produce very slow infiltration rates. The pores in the clays may be so small and hold water so tenaciously that the water is not available to plants. Sandy soils have numerous macropores, but few micropores. Total porosity of soil is of little practical significance. The number and proportion of macropores and micropores and distribution of different size classes of the pores seem to be more important. However, porosity of a soil is an index of its state of compaction.

Problem

A soil has a bulk density of 1.40 Mg m−3. Calculate the weight of soil of 1 ha area and 20 cm depth. The soil has 0.23 % total N (w/w), calculate nitrogen content per hectare basis. Taking particle density as 2.65 Mg m−3, calculate its porosity.

Solution

Volume of 1 ha-20 cm soil is $= 10{,}000 \times 20/100 = 2000\ m^3$

Bulk density is $= 1.40$ Mg m^{-3}

Weight of 1 ha-20 cm soil is $= 2000 \times 1.4 = 2800$ Mg (2,800,000 kg)

Total N $= 0.23\,\% = 0.23 \times 2800 \div 100 = 6.44$ Mg/ha-20 cm.

Particle density is $= 2.65$ Mg m^{-3}

So, porosity is $= [(1 - 1.40\} \div 2.65] \times 100 = 47\,\%$

Soil compaction is the physical consolidation of the soil by an applied force that destroys soil structure, compresses soil volume, increases bulk density, reduces porosity, and limits water and air movement (Osman 2013). Stunted growth of trees, badly formed tree roots, standing water, and physically dense soil are some signs of forest soil compaction. Texture, structure, organic matter, and water are important soil factors that determine susceptibility of a soil to compaction. Soils made up of particles of about the same size compact less than soil with a variety of particle sizes. Hard, dense, low-organic matter soils suffer more from compaction than loose, friable, high-organic matter soils. A dry soil is not easily compacted; soil compaction increases with the increase in soil water content until a moist condition is reached. Wet soils are also less susceptible to compaction.

When soil is compacted, soil strength is increased and total porosity is reduced at the expense of the large pores. Thus, volumetric water content and field capacity are increased, while air content, water infiltration rate, and saturated hydraulic conductivity are decreased, if a soil is compacted. Compacted soils offer physical impedance to root extension, which affects water and nutrient absorption. Soil compaction caused by harvesting operations can affect future regeneration and growth of trees (Sakai et al. 2008). Available soil water content is reduced when a soil is compacted (Rockich et al. 2001). When this occurs, nutrients and water may become limited because the plant's demands exceed the ability of the root system to access these resources (Froehlich and McNabb 1984). Nutrients have limited mobility in compacted soil (Unger and Kaspar 1994); so nutrient deficiencies may seriously limit tree growth there.

Forest soils may be compacted by grazing animals and by expanding roots of the trees themselves, but more noticeably by vehicles used for a range of mechanized forest operations. Harvesting machinery may be very heavy and, combined with the pushing and pulling and lifting of logs, may exert large pressures on the soil. Some harvesting operations may greatly disturb the soil (Greacen and Sands 1980). Compaction of forest soils may be caused by:

1. trafficking by heavy equipment during felling, forwarding, skidding, and site preparation operations;
2. dragging action of logs as they are moved from the stump to the landing; and
3. slash disposal and the creation of planting or seeding sites during site preparation.

The main forces causing compaction of forest soils are mostly heavy machinery used during timber harvesting and mechanical site preparation (Greacen and Sands 1980; Froehlich and McNabb 1984). Soil compaction is a major factor affecting forest production (Mckee et al. 1985; Firth and Murphy 1989). Froehlich (1973) concluded that 50 % of the harvested area is disturbed and 25 % can be considered as compacted in western USA forest under tractor logging. In British Columbia, over 20 % of the harvested area has been compacted (Utzig and Walmsley 1988). Nowadays, forestry machineries are becoming heavier and more damaging on forest soil compaction. Deep rutting strip off the top layers of soil from the ground and push it beside skid trail, thus resulting in heavy soil erosion (Zeleznik et al. 2009). Mechanization of some forest operations has greatly intensified over the last decade, particularly in plantations, and forest managers are currently expressing concern about the compaction of forest soils and its consequences. Approximate contact pressures of a range of machines used in the logging of forests are mentioned below (Greacen and Sands 1980):

Vehicle	Pressure (kPa)
Flexible tracked skidder	30–40
Crawler tractor	50–60
Rubber-tyred skidder	55–85
Forwarder with rear bogie	85–100
Forwarder with single rear axle	105–125

Recovery of compacted forest soils is variable depending on the severity of the compaction and local conditions. Most studies, however, indicate that recovery from compaction and its effects are a long-term affair. Severely compacted soils may require up to 40 years or more to recover naturally, according to Hatchell and Ralston (1971). Froehlich and McNabb (1984) state that the effects of soil compaction should be assumed to persist for several decades on forest sites.

2.4.1 Effect of Soil Compaction on Forest Tree Growth

Soil compaction affects ecosystem stability and site productivity (Froehlich 1979; Wronski and Murphy 1994; Kuan et al. 2007). Soil compaction is particularly severe on temporary access areas such as forest landings and skid trails. These areas may be unproductive unless soil rehabilitation is carried out. Tree species growing on compacted soil show reduced root elongation rate (Whalley et al. 1995) and reduced height growth (Greacen and Sands 1980; Ares et al. 2007; Bulmer et al. 2007).

Soil compaction affects seed germination and seedling emergence of forest trees (Taylor 1971) and natural regeneration in forests (Sokolovskaya et al. 1977). Complex interactions occur among soil strength, water, nutrient availability, and aeration due to compaction. Roots may fail to penetrate smaller pores than their diameter. Because compaction both increases soil strength and decreases the number of macropores, the rate of root elongation and therefore root length is reduced. Zyuz (1968) reported abundant pine roots in soils of strength less than 1,700 kPa. Root penetration was restricted above 2,500 kPa. The penetration of roots of Radiata Pine (*Pinus radiata*) into sandy soil in South Australia was found to be greatly restricted at soil strengths greater than 3,000 kPa (Sands et al. 1979). Soil compaction decreases growth of forest trees; however, this is not always true and not of all tree species. Growth reduction has been reported in economically significant forest tree species such as *Pinus radiata* (Potter and Lamb 1974), *Pinus elliottii* (Haines et al. 1975), *Pinus taeda* (Duffy and McClurkin 1974), *Pinus ponderosa* (Trujillo 1976), *Pinus nigra* var. *mararima* (Berben 1973), *Picea abies* (Sokolovskaya et al. 1977), and *Pseudotsuga menziesii* (Berben 1973). Lodgepole pine and white spruce seedlings were grown on four compacted forest soils of Alberta, Canada. Significant reduction in seedling growth (maximum root depth, maximum root depth in soil core, total weight, shoot weight, root weight, stem diameter, shoot height, seedling survival, and shoot weight: root weight ratio) was observed with increased bulk density (Corns 1988).

2.5 Soil Air

Well-drained forest soils are generally porous, and the large pores contain air if they are not saturated with water. Approximately 25 % volume of a loam-textured surface soil is filled with air under field condition. In dry soils, water from many micropores is evaporated and these micropores may also be filled with air for some time. On the other hand, air in the pores may be expelled from the soil by water immediately after rainfall or irrigation. Then the macropores are also filled, although temporarily, with water. After cessation of rain or irrigation, say some hours or days later, gravity pulls this water into the groundwater table, leaving the spaces for air. Therefore, soil air is a dynamic component of the soil.

Soils get air from the atmosphere, for which soil air has a composition closely related to the atmospheric air. The atmosphere consists of nitrogen (78.084 % by volume) and oxygen (20.9476 %) with smaller quantities of water vapor (variable), carbon dioxide (CO_2) (0.0314 %), inert gases such as argon (0.934 %), and some other rare gases. The concentration of oxygen in soil is less because of utilization by living organisms, roots, and chemical reactants. The concentration of CO_2 in soil air is higher because of biological respiration. The soil CO_2level in active soil reaches 0.15 % at 15 cm, but can reach 5 % and affect the pedosphere chemistry (Adl 2003). O_2 content in well-aerated soils may be about 20 %. It is reduced as the water content increases and the rate of exchange of gas between soil and atmosphere decreases. In some cases, oxygen can be as low as 5–10 %. If the soil remains saturated with water or waterlogged for a long period, say several days, O_2 content may completely be depleted. However, the composition of soil air is highly variable. It depends on the ease of exchange of air between the atmosphere and the soil. The more readily the exchange occurs, the closer is the composition of soil and atmospheric air. Renewal of air in soil is not so rapid because of the discontinuity of soil pores and the obstruction offered by soil water.

Soil texture, structure, and porosity affect the amount and composition of soil air and its renewal. Light textured soil or sandy soil contains much higher air than heavy soil. Diffusion of air occurs slowly in a waterlogged soil. Under prolonged waterlogging, the O_2 in soil air is diminished and the CO_2 concentration highly increases. In waterlogged soils, some volatile gases like H_2S, CH_4, and C_2H_4 may accumulate and exert toxic effects on plant roots. Soil microorganisms utilize O_2 for respiration and decomposition of organic matter and produce CO_2. Hence, soils rich in organic matter contain higher percentage of CO_2 in soil air.

2.5.1 Aeration

Aeration refers to the process of exchange of air and its constituent gases between soil and the atmosphere. Aeration takes place by two physical processes: mass flow and diffusion.

Mass Flow Mass flow occurs when the entire mass of air moves from one place to another, particularly from the soil to the atmosphere and from the atmosphere to the soil under conditions of fluctuating air pressure. Air is driven away from or into the soil pores by this pressure. Movement occurs due to atmospheric pressure changes, soil temperature changes, entry of water following rain and irrigation, evaporation of water from pores, and driving force of wind near soil surface. However, mass flow is much less important than diffusion, except perhaps in layers at or very near the soil surface.

Diffusion Diffusion is the predominant process of soil aeration. In diffusion, individual gas constituent moves separately under the influence of partial pressure gradient. When partial pressure of CO_2 in soil air increases due to root and microbial activity, molecules of CO_2 diffuse from soil to the atmosphere. When O_2 in soil air is consumed for respiration in soil, its partial pressure is reduced and a partial pressure gradient is created between the soil and the atmosphere; then O_2 diffuses into the soil. Generally, there is net diffusion of CO_2 into soil and net diffusion of O_2into the atmosphere. Through diffusion, each gas moves in a direction determined by its own partial pressure. The diffusion of gases in soil is regulated by numbers of air-filled pores and their continuity, both decreasing with increasing soil water content, and the tortuosity of air-filled pores which increase with soil water content.

2.5.2 Effect of Soil Air on Forest Tree Growth

Root respiration is an essential physiological function of a tree related to its healthy growth. The energy released in root respiration is necessary for mineral uptake, synthesis of new protoplasm, and maintenance of cell membranes. In the absence of an adequate supply of free O_2, root respiration is very inefficient and does not release enough energy to maintain essential root functions, particularly mineral uptake (Rosen and Carlson 1984). Richards and Cockroft (1974) observed that root growth was inhibited when air space in soil was reduced to less than 15%. Root growth was negligible when air space dropped to 2%. Growth of trees in a number of investigations, including *Pinus sylvestris*, *Picea abies*(Youngberg 1970), *Pinus contorta*(Lees 1972), and citrus (Patt et al. 1966), was found to be severely reduced when the O_2 content dropped below 10%.

Tree roots generally function well at oxygen levels above 10% (Kozlowski 1985). At lower concentrations, roots will cease growth and lose their selective permeability (Costello et al. 1991). Metabolic processes are altered and cellular metabolites may accumulate to toxic levels. By-products of anaerobic respiration (lactate, pyruvate, and ethanol) accumulate to toxic levels (Levitt 1980). Also, some root pathogens become active under anaerobic conditions. For example, Oak root fungus, *Armillaria mellea*, may colonize root tissues killed by oxygen stress, and from that food base, invade healthier portions of the root system (Wargo 1981). Likewise, accumulation of ethanol in roots may stimulate attack by *Armillaria* (Wargo and Montgomery 1983). Low oxygen in soil reduces height growth, leaf growth, cambial growth, and reproductive growth of trees that grow on poorly aerated soils, although the amount of growth reduction varies widely among species. Some tree species can tolerate low O_2 supply by (1) producing hypertrophied lenticels which assist in aeration of the stem and release of toxic compounds and (2) growing new roots to replace loss of original roots under anaerobic conditions (Kozlowski 1986). Not only do flooded or compacted forest soils have limited oxygen in the root zone, but vast areas of heavy textured soils may also be poorly aerated (Kramer and Kozlowski 1979). It has been estimated that the soil air to a depth of 1 m does not contain enough O_2 to maintain respiration of plants and microorganisms for more than a few days (Drew 1983).

2.6 Soil Temperature

Soil temperature is an important physical property that regulates most of the physical, chemical, and biological processes of the soil, and the physiological processes of soil organisms and forest plants. Soil temperature has tremendous ecological impacts through evaporation, transpiration, organic matter decomposition, CO_2 emission due to soil respiration, and permafrost thawing. In forest ecosystems, soil temperature regulates microbial transformations of nitrogen (Plymale et al. 1987) and sulfur (Strickland and Fitzgerald 1984), and other nutrients. It controls decomposition of organic matter and formation of humus (Waide et al. 1988), as well as root growth, activity, and respiration (Marshall and Waring 1985).

2.6.1 Factors Affecting Forest Soil Temperature

Temperature of forest soil is influenced by a number of factors, namely meteorological conditions (i.e., solar radiation and air temperature), site topography, soil water, organic matter, texture, and the area of surface covered by litter and canopies of plants (Paul et al. 2004). Since soils obtain heat mainly from the solar radiation, and the amount and intensity of solar radiation depend on geographic, climatic, edaphic, and topographic conditions, soil temperature varies widely among forest types. For example, tropical forest soils are warmer than temperate and tundra soils.

Plant cover, which reduces albedo, can be extremely important to the energy balance of high-latitude ecosystems. For example, Hall et al. (1996) found that jack pine-spruce forests in Saskatchewan and Manitoba had an average albedo of 0.08 compared to a value of 0.8 for bare snow. It has also been shown that the tussock growth form of *Eriophorum vaginatum* improves the soil thermal regime in Alaskan tundra (Chapin et al. 1979). The insulating effect of vegetation arises partly from development of a layer of organic detritus on the soil surface, which reduces daily and seasonal temperature fluctuations. Breshears et al. (1998) reported that winter soil temperatures under patches of woody plants (*Pinus edulis* and *Juniperus monosperma*) in semiarid woodland were warmer than interpatch areas due to the buildup of a litter layer. The insulating properties of litter result from the relatively low thermal conductivity of organic matter compared to mineral soil and the large proportion of air spaces.

The hydrologic cycle has an important role in controlling soil temperature. In a bare soil, color and moisture content mainly determine the albedo (ratio of incident radiation to reflection from surface) (Hanks and Ashcroft 1980). Dark soils absorb more energy than lighter ones, and wetting the soil effectively darkens it. Snow cover has a dominant effect on albedo; the albedo of snow- and ice-covered surfaces ranges from 0.45 to 0.90 (Strahler and Strahler 1983). Snow also acts as an effective insulator which retards the loss of heat from the soil. Bertrand et al. (1994) and Stadler et al. (1996) observed an inverse relationship between depth of snowpack and depth of soil freezing at mid-latitudes. Soil freezing can be prevented entirely under very deep snow (Isard and Schaetzl 1995).

Under field condition, soil temperature can be estimated from air temperature with enough precision (Soil Survey Staff 1975). To get mean annual soil temperature (MAST) in tropical regions, 2.5 °C is added to the air temperature in the soil–atmosphere interface. Since the canopy and the forest vegetation and the litter layers have insulating and buffering effects, this relation may not hold good for forest soils. Experimental studies have shown that vegetation canopies can lower soil temperature during growing season significantly and reduce MAST(Qashu and Zinke 1964; Cermak et al. 1992). Qashu and Zinke (1964) concluded that lower soil temperatures under oak or pine canopies may result from a lower rate of soil warming during springtime. Soils are not good conductors of heat; and, therefore, subsoils are cooler than the surface soil. For the same amount of solar radiation, the soil is heated more and faster than the air in the soil–atmosphere interface. Therefore, surface soil temperature is generally 1–5 °C higher than the air in the immediate soil surface. The surface soil temperature fluctuates with seasons more than that of the subsoil. There may be a depth in the soil where soil temperature fluctuates very little with seasons.

Soil temperature depends on soil cover. Generally, forested sites are cooler than field sites in the summer and warmer in the winter. Forest canopy intercepts solar radiation and acts as an insulator. Clear cutting forests, that is, removing the vegetative cover, increases soil temperature. Spittlehouse and Stathers (1990) observed that maximum daily soil temperature (1-cm depth) in a clear-cut exceeded 50 °C compared to 16 °C in an adjacent mature, western hemlock-fir forest in coastal British Columbia. Forests regulate soil temperature by influencing (1) air temperature, (2) amount and intensity of precipitation reaching the soil, (3) depth and duration of snow cover, (4) wind velocity, (5) shade of living cover, (6) thickness of the forest floor, (7) water content of the soil, (8) humus content of the soil, (9) color of the soil, and (10) structure of the soil. Denuded sites and fields showed greater temperature variation in the winter than forests. Jeffrey (1963) observed that canopy covers of different forest tree species differ in trapping and absorbing solar radiation. Mueller (2002) studied three forest types and observed that a mountain White Pine–Hemlock–Hardwood forest, with boreal components in a valley flat at Ramsey's Draft in Virginia, had growing season soil temperatures lower than those of Oak- Hickory Ridge and Mesic Slope Forests in the Shenandoah Valley. Some early season temperature differences may be as great as 4 °C. Balisky and Burton (1995) suggested that percent canopy cover was the primary control on soil temperature in high-elevation Engelmann Spruce (*Picea engelmannii* Parry ex. Engelm.)-subalpine fir forests.

Leaf litters on the forest floor act as insulators, keeping the soils warmer in the winter by trapping radiant heat. This blanketing effect also decreases the diurnal range in soil temperature especially in the spring. During the summer, forest cover lowers the maximum soil temperature when compared to the fields. The forest floor is a highly efficient insulating layer maintaining low soil temperatures and often a shallow permafrost table (Dyrness 1982). In several experiments, removal of the forest-floor layer was found to exert profound changes in the thermal regime of the soil. For example, forest-floor removal resulted in an active layer 132 cm thick (compared to 47 cm in an undisturbed area) after 3 years (Viereck and Dyrness 1979). Brown et al. (1969) reported similar results after tractor removal of surface organic matter; they stress that melting permafrost often leads to serious problems with erosion. By far, the greatest increases in depth to permafrost were measured on the plots that had the entire forest-floor layer mechanically removed. In these plots, the average increase in active-layer thickness was almost sevenfold (from about 20–137 cm) after 4 years.

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