Unsaturated Zone Hydrology for Scientists and Engineers

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Physical Properties and Characteristics of Soils

INTRODUCTION

The role of physical properties in unsaturated zone processes—in relation to water content, contaminant transport, porosity, bulk density, saturation, and so on—cannot be overstated. A sampling of the areas within this subject that have been extensively researched might include the rate of water intake in soils (Bertrand 1965), the relation of exchangeable cations to physical properties of soils (Baver 1928), the infiltration of water in layered soils (Bruce and Whisler 1973), soil aeration (Lemon 1962; Lemon and Wiegand 1962), mechanical stress on root growth (Barley 1962; Russell 1952; Wild 1988), X-ray diffraction techniques for identifying soil minerals (Whittig 1986), and many others. This chapter will present a succinct discussion of the mineralogical composition of soils and various clays, soil profile development, and texture, and various physical characteristics and relations of soil constituents. A short list of suggested readings is given at the end of the text for the reader who desires greater detail or additional background on soils and their properties, followed by a complete list of references.

2.1 MINERALOGICAL COMPOSITION

Porous media, of which soil is an example, consist of three separate phases: solid, liquid, and gas. Due to the simultaneous presence of more than one phase, a porous medium is a heterogeneous system (also called a disperse, polyphase system), with each phase varying both mechanically and chemically from the others. Other examples of porous media include sedimentary rocks, stratigraphic columns, glacial deposits, gravel fills, mine tailings, and manmade media such as greenhouse mixes used in the propagation of various flora.

Soils are formed by weathering of rocks on the earth's surface, which causes their decomposition through physical and chemical deterioration. The physical processes involved, determined by climate, include freezing and thawing, wetting and drying, flowing water, moving ice, and abrasion by sand particles moved by wind or water. The small, loose rock particles that result from physical weathering become the parent material of soil. Solubilization, hydration, oxidation, reduction, precipitation, leaching, and other physicochemical processes further decompose the minerals that make up the rock particle. Because of the presence of populations of microbes and other fauna, a continual microbial and biochemical decomposition of some parts of the parent material continues as organic matter is added through the decay of animals and plants. The original rocks become the soil as it is seen today.

The mineralogical composition of soil is determined by the rocks or parent material from which it is formed. Other soil-forming factors include climate, biospheric parameters, effects of topography, and time. The solid phase of a soil includes sand, silt, and clay fractions commonly expressed as a percentage of the whole: 30 percent sand, 35 percent clay, and 35 percent silt, for example, for a total of 100 percent. The sand and silt fractions consist mainly of primary minerals that do not normally occur in the clay fraction. The most important and well-known of these primary minerals are quartz and feldspars.

Soils are usually deposited or formed in layers conditioned by geographic location, climate, and the formation processes that these layers have undergone. The accumulated layers in any given soil are called a profile. Each layer (called a *horizon*) within this profile has unique morphological characteristics.

Quartz is present in nearly all soils and makes up 50 to 90 percent of the sand and silt fractions. The quartz crystal is composed of linked silica tetrahedra in which the four oxygen atoms present are shared with adjacent silica tetrahedra, leaving a lack of cleavage planes. Feldspars are aluminosilicates of calcium, potassium, and sodium. Similar to quartz, the structure of feldspar is made up of silica- and aluminum-containing tetrahedra linked together by the sharing of each oxygen atom between adjacent tetrahedra. The result is a three-dimensional tectosilicate structure. The common minerals of granite and gneiss rocks are composed of potassium feldspars that include adularia, microcline, orthoclase, and others.

Accessory minerals developed during soil formation occur in the sand and silt fractions. Depending on the soil type, these minerals can be present in appreciable amounts and include amphiboles, pyroxenes, apatite, olivines, magnetite, ilmanite, and tourmaline. Occasionally, carbonate and sulphur-bearing minerals occur in soils in the form of dolomite and gypsum, respectively.

QUESTION 2.1

Are there situations in which only two of the three phases might occur in a soil?

Clays

Clays are colloidal and crystalline in nature and are the active mineral portion of soils. Clay minerals are commonly formed from the soluble products of primary minerals, and are usually referred to as secondary minerals. Most clays are crystalline and have a definite, repeating arrangement of atoms, dominated by layered planes of oxygen atoms. Silicon and aluminum atoms bond the oxygen atoms together through two kinds of bonds: ionic (attraction of positively and negatively charged atoms) and covalent (sharing of paired electrons). Three or four planes of oxygen atoms, with intervening silicon, aluminum, or other atoms (depending on clay type), make up one layer (see figures 2.1 and 2.2).

The crystal lattices of most clay minerals are composed of two structural units, namely tetrahedra and octahedra. The tetrahedral units consist of a silica tetrahedron in which the silicon atom is equidistant from four ionically bonded oxygen atoms at the corner of the structural unit. The planes of oxygen atoms held together by the silicon atoms are tetrahedrally oriented, and are referred to as silica tetrahedral sheets. These tetrahedral groups are arranged in a hexagonal network. The octahedral units consist of two sheets of closely packed oxygen atoms or hydroxyl units bound at the edges to an aluminum, iron, or magnesium atom that shares the oxygen atoms (see figure 2.1). Clays in which the oxygen and other atoms are less regularly oriented are called *amorphous* materials.

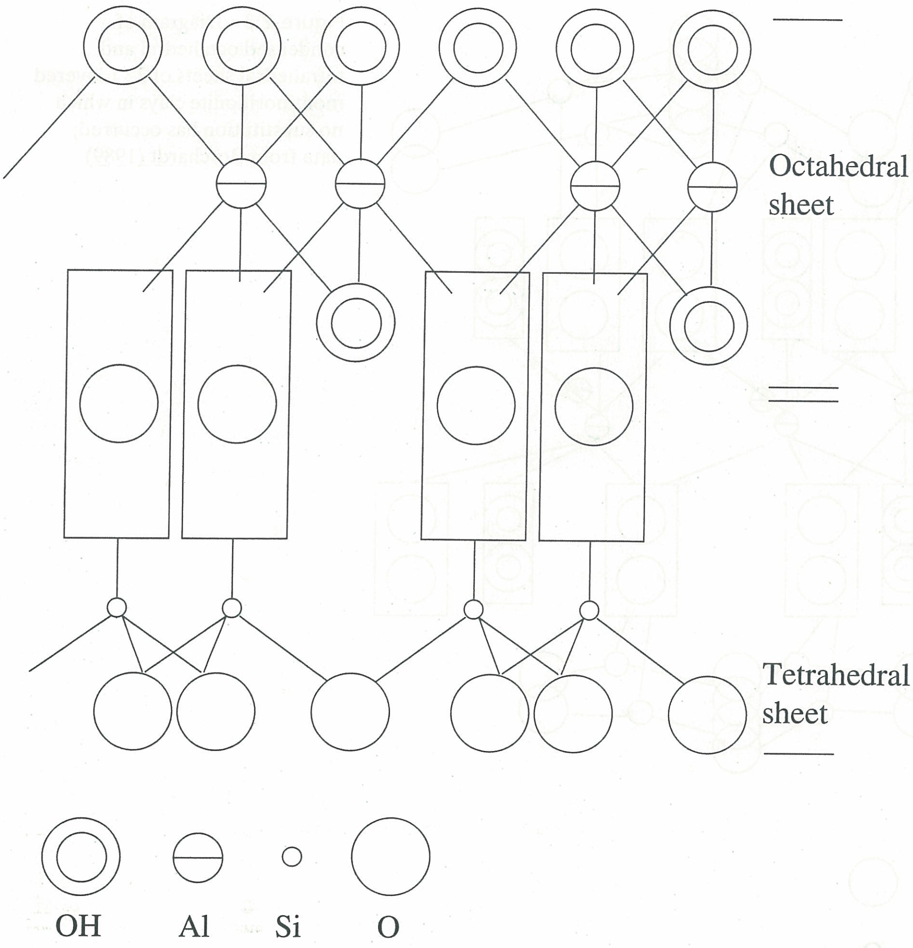


Figure 2.1 Diagram of condensed octahedral and tetrahedral sheets of clays (1 : 1 type); data from Mott (1988)

Silicate Clays

The most common clays are discussed here, and in the next sub-section. The silicate clays represent well-structured materials. Generally, clays are referred to by the designations 1:1, 2:1, 2:2, and so on. This refers to the ratio of tetrahedral silica to octahedral alumina sheets in a layer of clay. In some clays, a silicon ion can be substituted by aluminum, and vice versa. Likewise, any ion of similar size to aluminum may also substitute, giving the clay very different properties, due to the ionic charge differences between ions. These ions may include iron, magnesium, and zinc.

When clay is examined closely under a microscope, the structure appears very much like a deck of cards. Each "card" represents a layer of clay that is an exact replica of the next layer, with each layer being held together by van der Waal's–London forces (*as explained in chapter 3*). This "deck" can both expand and/or shrink with the addition or loss of water. The amount of expansion between each card or layer and, thus, the thickness of the total number of layers, depends on the ion of substitution; this is *discussed* in section 3.6. Formation of layer after layer forms a *tactoid;* that is, an aggregated platelet. These tactoids are formed primarily by ion exchange in which calcium typically begins to segregate in the interlayer

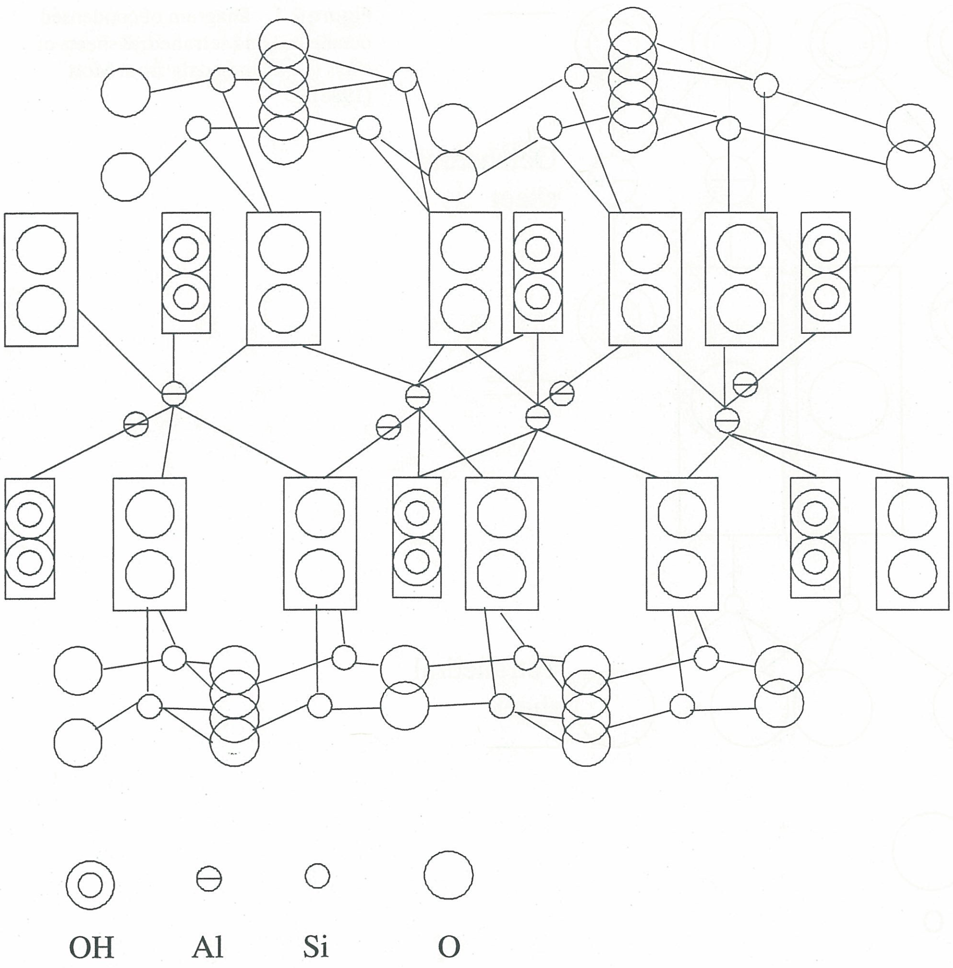


Figure 2.2 Diagram of condensed octahedral and tetrahedral sheets of 2 : 1 layered montmorillonite clays in which no substitution has occurred; data from Borchardt (1989)

regions and sodium ions on the external surfaces of the tactoid. Depending on the characteristics of the exchanging ions, tactoids can range in thickness from 1 to over 1000 Å. Additionally, selectivity coefficients play a major role in tactoid formation. A detailed discussion of tactoids is beyond the scope of this text. The reader is referred to McBride (1989) for in-depth discussions of tactoids and the surface chemistry of clay minerals.

Amorphous Clays

Amorphous clays are mixtures of alumina and silica that have not formed a well-oriented crystalline structure. Iron oxides and mixtures of other weathered oxides may also be part of this mixture. Such clays are usually found in areas where copious amounts of weathered products are present but have not had sufficient time or the proper conditions for crystal formation.

Amorphous clays are most common in soils that have formed from volcanic ash, which is also a porous medium. These clays are unique in that their charge originates from hydroxyl ions (OH^-) on the clay surface, which can either gain or lose a hydrogen ion (H^+); thus, the clays can be either positively or negatively charged. They are referred to as variable charged clays, and are more difficult to manage agronomically than other clays. Descriptions of the most important varieties of amorphous clays follow.

Kaolinite is a 1:1 silicate mineral commonly occurring in highly weathered, well-drained soils where leaching has removed Ca^{2+}, Mg^{2+}, K^+, and Na^+. It is more common than montmorillonite in acidic soils. It occurs as a crystal of hexagonal shape having a triclinic symmetry. Kaolinite clays have such strong hydrogen bonding that almost no substitution of silicon or magnesium for aluminum can occur. This is actually due to the bonding between the O layer of the tetrahedral surface of the clay platelet to the OH ions on the octahedral surface of the adjoining platelet, which prevents swelling in the presence of water. The net negative charge is low, resulting in a low cation exchange capacity (CEC) of 3–15 eq kg^{-1}, which varies with pH. The thickness of layers is 7.2 Å. In addition to hydrogen bonding, the smallest hydrated diameter for cations is 7.6 Å for potassium; thus, cations can only attach to the edge or face and not to the interlayer. This explains the lack of swelling and low CEC. Tactoids for kaolinite are 500 Å thick, which results in a low specific area (80,000 $m^2\ kg^{-1}$).

Illite is a 2:1 young (limited weathering) clay that has a structure similar to micas, consisting of two tetrahedral layers separated by an octahedral layer, and is sometimes called a hydrous mica. Illite is characterized by high substitution of Si^{-4} by Al^{-3}, or Fe^{-3}, in the tetrahedral layers. The resulting large negative charge on the tetrahedral surfaces is balanced by sorption of a potassium ion, which fits into the "oxygen hole" in the tetrahedral lattice. This same potassium ion also attaches to an oxygen hole in an adjoining platelet or interunit layer, resulting in adjacent platelets being so tightly bound that water cannot penetrate between them. Consequently, illite has low swelling ability, and also is an important source of potassium in agriculture. The structure is weakened to an extent by the incorporation of water between some lamellae; that is, the exchange of potassium by other cations during the weathering process results in some interlayer hydration, which weakens the structure of the clay aggregate. Thus, the potassium ions locked within the lamellae are slowly released through the course of the growing season. Illite has an intermediate CEC (15–40 eq kg^{-1}). Illite is found in soils high in primary minerals that are not extensively weathered, and may occur in the same environment as montmorillonite. The layer thickness of illite (from top to top) is 10 Å, with a tactoid thickness of 50–300 Å and a medium specific surface area of 80,000–120,000 $m^2\ kg^{-1}$.

Vermiculite is a 2:1 intermediate-age clay that is similar in structure to illite, and is also characterized by substitution of silicon by aluminum in the tetrahedral layers, but the resulting charge imbalance is compensated by interlayer sorption of hydrated magnesium ions, rather than potassium ions. The large hydrated radius of the magnesium ions results in larger interlayer separation. As a result, this is a moderately swelling clay, with a high CEC (120–150 eq kg^{-1}) and a high specific surface area (600,000 $m^2\ kg^{-1}$). The structure of vermiculite is composed of six water molecules in octahedral coordination with magnesium. Layer thickness is 14 Å, which collapses to 10 Å with potassium saturation; tactoids are 50 Å thick.

Montmorillonite (also termed a smectite) is a 2:1 intermediate-age clay, in which water easily penetrates between the layers, causing free swelling of the clay particle. Thus, minerals of montmorillonite or smectite isomorphous series are termed freely expandable layer silicates (other examples are beidellite, nontronite, and saponite). These minerals occur as very small particles ranging in diameter from about 0.01 to 1 μm. Interlayer spacing ranges from about 12 to 18 Å, depending on the exchangeable cation species present and the degree of hydration. Complete drying causes a spacing of less than 10 Å, whereas total hydration can literally "float" the layers apart so that they are independent of each other. The general structural formula is $X_{0.8}(Al_{0.3}Si_{7.7})Al_{2.6}Fe_{0.9}{}^{+3}Mg_{0.5})\cdot O_{20}(OH)_4\cdot nH_2O$; however, without consideration to lattice substitutions, the theoretical formula, according to Grim (1953), is $Si_8Al_4\cdot O_{20}(OH)_4\cdot nH_2O$ (interlayer).

Montmorillonite is comprised of clay platelets made up of one aluminum octahedral layer sandwiched between two silicon tetrahedral layers. Charge imbalances in montmorillonite arise due to substitution of aluminum by divalent ions in the octahedral layer and silicon–aluminum substitution in the tetrahedral layer. This results in a net negative charge that is offset by sorbed cations to the surface of the tetrahedral layers. Both tetrahedral and octahedral layers are combined, so that the tips of the tetrahedrons in each silicon layer and one of the hydroxyl (OH) layers of the octahedral layer form a common layer. The atoms which are common to both the tetrahedral and octahedral layers are then O instead of OH. In continous layers—that is, where O layers of each unit tactoid are adjacent to O layers of neighboring units—weak bonds are formed that cause excellent cleavage planes (Grim 1953). Figure 2.2 is an example of the structure of montmorillonite in which no substitution has occurred.

Formation can occur directly from parent material in poorly drained areas, or indirectly from weathering of illite → vermiculite → montmorillonite. Montmorillonite is characteristically present in soils known as vertisols. The bonding between units varies depending on the adsorbed cation. Montmorillonite has a high CEC (80–100 eq kg^{-1}), and usually falls into one of three types: sodium, calcium, or aluminum montmorillonites. There is no bonding between layers in sodium montmorillonite. Bonding occurs in calcium montmorillonite due to a thinner electrical double layer (discussed in chapter 3), and there is more bonding in aluminum montmorillonite due to a higher valence of the charge species. The layer thickness is 10–20 Å, with a tactoid thickness of 10–100 Å and a high specific surface area (700,000–800,000 $m^2\ kg^{-1}$), resulting in a highly reactive soil. Bentonite is an impure deposit of montmorillonite, and is commonly used as a low-permeability layer beneath landfills and for sealing around scientific instruments in the field, as well-drillers mud, and is also used in cosmetics.

Chlorite is a 2:2 type clay. In chlorite, two silica tetrahedra, an alumina octahedron, and a magnesium octahedral sheet make up the structure. Chlorite is a highly weathered clay that may form from vermiculite or montmorillonite. Some substitution in the tetrahedral and octahedral layers may occur. Consequently, chlorites do not swell upon wetting and have a low CEC (20–40 eq kg^{-1}). Layer spacing is 14 Å, with tactoids 100–1000 Å thick and a medium specific surface area (80,000 $m^2\ kg^{-1}$). Primary chlorites are unstable in acidic weathering environments; thus, they are almost always found in more recent deposits, such as aridisols and mollisols. While we have classified chlorite as a 2:2 clay, there is some disagreement on this matter. A more detailed discussion can be found in *Barnhisel and Bertsch (1989).*

Sesquioxides are mixtures of aluminum ($Al(OH)_3$), iron hydroxides ($Fe(OH)_3$), and iron oxides (Fe_2O_3) that form under conditions of intense weathering and heavy rainfall. They occur most extensively in warm, humid, well-drained soils in tropical regions and are generally termed unstructured clays. Most of the silica and some alumina have been washed from the soil; as a consequence, these clays are not sticky compared to montmorillonitic types. These clays can be either amorphous or crystalline. Sesquioxide soil types are easily identified, due to the characteristic red and yellow colors developed through the weathering process. Sesquioxides do not shrink or swell; however, they do coat large soil particles and form stable aggregates. Due to the high content of iron and aluminum oxides within the clay, it has a very high specific surface area and can readily adsorb, thus fixing nutrients such as applied phosphorous and forming insoluble phosphates. Soils containing sesquioxide clays are sometimes referred to as oxide minerals or hydrous oxides and contain such minerals as gibbsite and laterites, including hematite and goethite.

Allophanes are another hydrous oxide soil type worth mentioning. They are amorphous aluminosilicate gels having silicon in a tetrahedral coordination, and the metal ion in

TABLE 2.1 Summary of Physical Characteristics for Various Silicate Clays

Silicate clays	Type	Degree of weathering	Substitution	Swelling	CEC (eq kg^{-1})	Layer thickness (Å)	Tactoid thickness (Å)	s* (m^2 g^{-1})
Chlorite	2 : 2	high	some	none	20–40	14	100–1000	80
Illite	2 : 1	limited	high	low	15–40	10	50–300	80–120
Kaolinite	1 : 1	high	none	low	3–15	7.2	500	80
Montmorillonite	2 : 1	intermediate	high	high	80–100	10–20	10–100	700–800
Vermiculite	2 : 1	intermediate	high	moderate	120–150	14	50	600

*Multiply by 1000 to obtain m^2 kg^{-1}

octahedral coordination. They are stable and highly porous and as such are very permeable. Extensive leaching has removed most of the primary minerals and nutrients, leaving the soil infertile. Allophanes also have a high specific surface area, due to the presence of large amounts of aluminum and iron, which results in a high phosphorous-fixing capacity. Allophanes are essentially noncrystalline aluminosilicate compounds that are associated with weathered volcanic ash. They are members of a series of naturally occurring minerals that have been classed as hydrous aluminum silicates. These silicates have a wide range of chemical composition that is characterized by the presence of Si-O-Al bonds, short range order, and a differential thermal analysis curve that displays a low temperature endotherm and high temperature exotherm. Since these criteria are relatively strict, allophanes are limited to a small sector of the total spectrum of noncrystalline (and also paracrystalline) aluminosilicates which have been developed by the weathering of volcanic ash, pumice, and other soil materials. See table 2.1 for a summary of the physical characteristics of various clays.

QUESTION 2.2

What is the primary binding (attractive) force between adsorbed cations and a clay platelet?

QUESTION 2.3

Why do clay particles have different specific surface areas (assuming that they have equal planar extensions)?

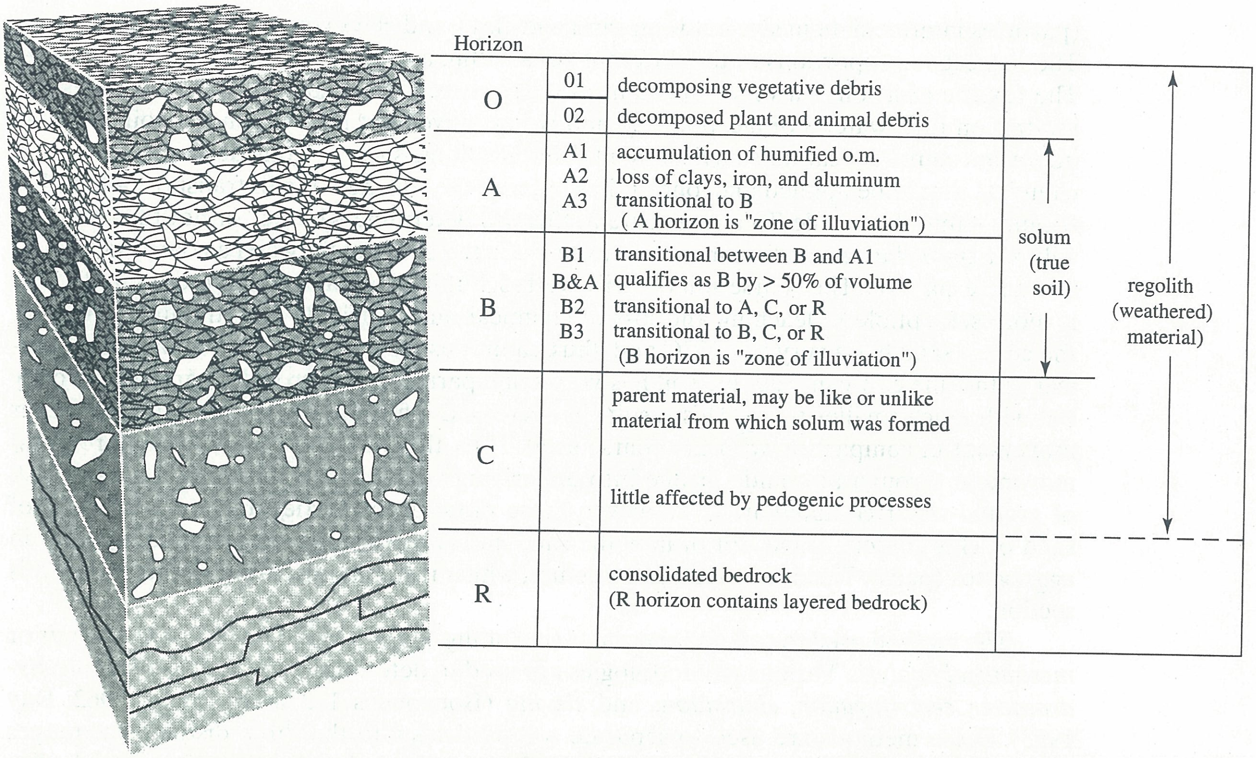
2.2 SOIL PROFILES

The clays and minerals discussed in the previous section, organic matter and other solids, along with air and water are the essential elements that compose the soil. Soil formation results in the natural development of layers within a profile. Each of these layers is composed of structural units consisting of various fractions of sand, silt, clay, and organic matter. When formed in place by pedogenic processes, the layer is called a horizon. The soil formation process may take many thousand years or as little as a few decades. The development of each distinguishable horizon is due to physical, chemical, and biological disintegration in the weathering processes that have caused organic matter accumulation, colloid transport, and the deposition of clays, humus, and soluble minerals such as carbonates and gypsum. Most soil horizons noticeably differ from each other in color, content of organic matter, texture, and structure or chemical properties. Horizons are parallel to the soil surface; if a vertical cross-section were exposed, say, along the side of a pit, each horizon could be clearly seen and the differences between horizons would be immediately noticeable.

Horizon development within the soil profile has occurred under four broad categories: addition, removal, transport, and transformation of soil material. The horizons of a soil are usually characterized by soil morphologists into four or five major designations, each composed of subdesignations labeled by Arabic numerals or letters (*U.S. Soil Conservation Service, 1985*). The major horizon designations (from the soil surface downward) consist of: (1) the O horizon, which is characterized by organic matter lying on the surface; (2) the A horizon, which is considered the mineral horizon and may have lost appreciable amounts of clay, iron, and aluminum oxides by eluvial (washed-out) processes; and (3) the B horizon, which contains illuvial (washed-in) concentrations of clay, sesquioxides, and other colloids and materials transported within the profile. The remaining two major designations are: (4) the C horizon, which is considered the parent material and has been affected very little by ongoing biological activity; and (5) the R horizon, which is consolidated bedrock underlying the C horizon. The C horizon does not necessarily have to be the parent material, because the soil profile could have been formed by wind deposition (aeolian) or water sedimentation (alluvial) processes. Likewise, the R horizon may not be like the parent material from which the overlying horizon was formed. If this is the case the R is preceded by a Roman numeral denoting the horizons' lithologic discontinuity. The O horizon is not always present, in which case the A becomes the upper horizon.

Three other horizon designations that are not always present in soil are the AB, A&B, and AC. Within the A and B horizons are subdesignations: A1, A2, A3, B1, B&A, B2, and B3. The subdesignations may not all be present in a particular soil, depending on the developmental processes and geographic location of the soil. The AB horizon is a transitional horizon between A and B, which has the upper part dominated by properties of A and the lower part dominated by properties of B. The A&B could qualify as an A2 subdesignation, except that part of this horizon, constituting less than 50 percent by volume, would be classified as a B horizon soil. The AC horizon is transitional between A and C; it has properties of both A and C but is not dominated by either. This designation is used only where there is no clearly discernible B horizon. An example of a soil profile is illustrated in figure 2.3, along with a description of the major horizon designations.

A "typical" soil profile cannot be illustrated, because there is so much variety from one geographic region to the next and, for that matter, even within a single region. A soil profile consists of a succession of horizons, due primarily to the combination of soil-forming factors under which the profile develops. Climate is usually the dominant factor in soil formation, although many soil-forming processes are interrelated. The detectable layers of clays, organic matter, and salts that have moved downward into the profile are a result of both the amount and pattern of precipitation. Soil profiles tend to develop faster under warm, humid, forested conditions, where there is adequate water to transport soil colloids, and where there are considerable amounts of organic matter to decompose. Other climate-related factors that have a direct effect on soil formation are temperature and the types and amounts of vegetation present. For example, soil profiles that develop under forests have more horizons than profiles that develop under grasslands. Likewise, profiles in arid climates exhibit little horizon development due to lack of organic matter, absence of leaching, and inadequate soil protection by sparse vegetation (against erosion by water and wind). For an illustration of soil profiles developed in different geographic regions and under various climatic regimes, the reader is referred to Hausenbuiller (1978).



Horizon				
O	01	decomposing vegetative debris		regolith (weathered material)
	02	decomposed plant and animal debris		
A	A1	accumulation of humified o.m.	solum (true soil)	
	A2	loss of clays, iron, and aluminum		
	A3	transitional to B		
		(A horizon is "zone of illuviation")		
B	B1	transitional between B and A1		
	B&A	qualifies as B by > 50% of volume		
	B2	transitional to A, C, or R		
	B3	transitional to B, C, or R		
		(B horizon is "zone of illuviation")		
C		parent material, may be like or unlike material from which solum was formed		
		little affected by pedogenic processes		
R		consolidated bedrock (R horizon contains layered bedrock)		

Figure 2.3 Representation of a complete soil profile (left) with horizon descriptions (right); data from Hillel (1982)

QUESTION 2.4

What are the five major processes involved in physical weathering in soil profile formation?

QUESTION 2.5

Decomposition of soil is usually referred to as chemical weathering. What are the two major processes associated with this phenomenon?

QUESTION 2.6

Associated with chemical weathering is structural change caused by chemical composition. What are the three major chemical processes affecting structural change?

2.3 SOIL TEXTURE

Soils are composed of particles of varying size, from large rocks to microscopic clays. However, the texture of soils is limited (by definition) to the fine earth fraction of the sand, silt, and clay sized particles. These different size groups, called *separates,* are sands (which usually consist of quartz, but may also be fragments of feldspar, mica, and heavy minerals), silts (particles intermediate in size between sand and clay), and clays (the smallest size fraction). The respective proportions of the different sizes within a specific soil determine the *texture*. The texture of a soil is a very important physical characteristic, because it determines the infiltration rate, water storage, porosity, and water movement rate within the soil. This has important agronomical and environmental implications, such as considerations of how a chemical should be applied and, once it has entered the soil, how it will transport and redistribute within the profile. To a large extent, this will determine whether or not the chemical will be a hazard to the environment. An example is the contrast between a coarse sandy soil and a fine clay soil. The coarse soil is easily wetted, but it also drains much more rapidly and is more susceptible to leaching and loss of chemicals and nutrients than the fine soil. Also, the coarse sand has a very low CEC and, thus, cannot usually adsorb as much organic chemical as the fine clay can. The fine soil has very small particles that provide a greater porosity but with much smaller pores. This results in decreased infiltration and retardation of water movement in comparison with the coarse sand. Thus, there is the potential for more rapid movement through the sandy profile into underlying aquifers, and therefore a greater risk of ground-water contamination. However, these physical properties are not a function of sand or clay content alone, but of how the sand and clay fractions are grouped together in aggregates (peds). This is termed soil structure, which is briefly discussed at the end of this section.

The method used to sort soil separates (including soils) is called *particle size analysis* or *mechanical analysis.* Various methodologies are used to determine size, such as *pipetting, hydrometer, centrifugation, elutriation,* and *sieving* (Bouyoucos 1927a&b; Chepil 1962; Day 1953). These methods are used to separate soil particles into the three distinct size ranges previously mentioned: sands, silts, and clays. Standardized classifications for particle size analysis (shown in figure 2.4) have been established by the United States Department of Agriculture (USDA), United States Public Roads Administration (USPRA), British Standards Institute (BSI), Massachusetts Institute of Technology (MIT), U.S. Bureau of Soils (USBS), and the International Soil Science Society (ISSS). However, while each of these

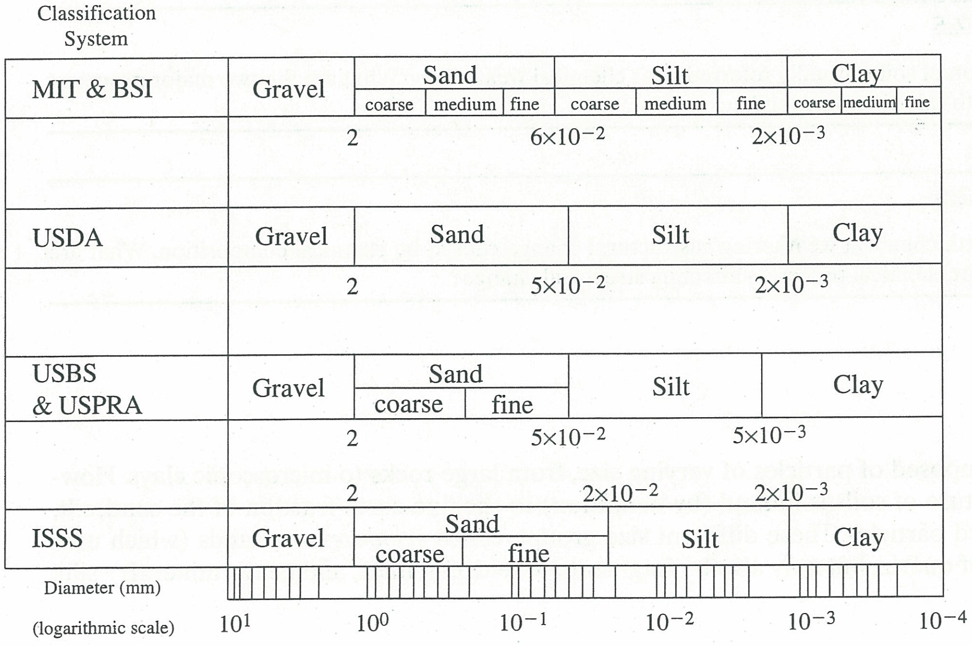


Figure 2.4 Soil classification system based on standards of the Massachusetts Institute of Technology (MIT), British Standards Institute (BSI), U.S. Department of Agriculture (USDA), U.S. Bureau of Soils (USBS), U.S. Public Roads Administration (USPRA), and International Soil Science Society (ISSS)

agencies has a standardized classification for soil separates, they vary slightly. This is especially true in the silt and sand separates.

While there is some confusion concerning which system to use, and what the upper and lower limits of particle size are, most members within a particular society or organization tend to use the same classification system; for example, agronomists and soil scientists who belong to the American Society of Agronomy (ASA) and Soil Science Society of America (SSSA) tend to use the USDA classification system. Members of other organizations, such as the American Society of Civil Engineers (ASCE), might use a different classification system.

An additional separate or fraction is important in some soils. This is gravel, or stones of size > 2mm (particle size < 2mm is classified as soil material). If this gravel or stony fraction makes up a significant weight of a soil being studied, and influences the physical and chemical processes such as transport and cation exchange, this size fraction should be reported as well.

The combination of all three fractions constitutes the matrix of the soil, sometimes referred to as its *structure.* Soil structure describes how the various size fractions are grouped together into stable aggregates. Aggregates are secondary units composed of soil particles bound together by organic substances, iron oxides, clays, carbonates, or silica. Natural aggregates that vary in their water stability are called *peds,* while the term *clod* refers to a soil mass broken into virtually any shape by artificial means, such as plowing, earth moving, or tillage. If a soil ped is broken apart, a broken piece of this ped is called a *fragment.* Another term

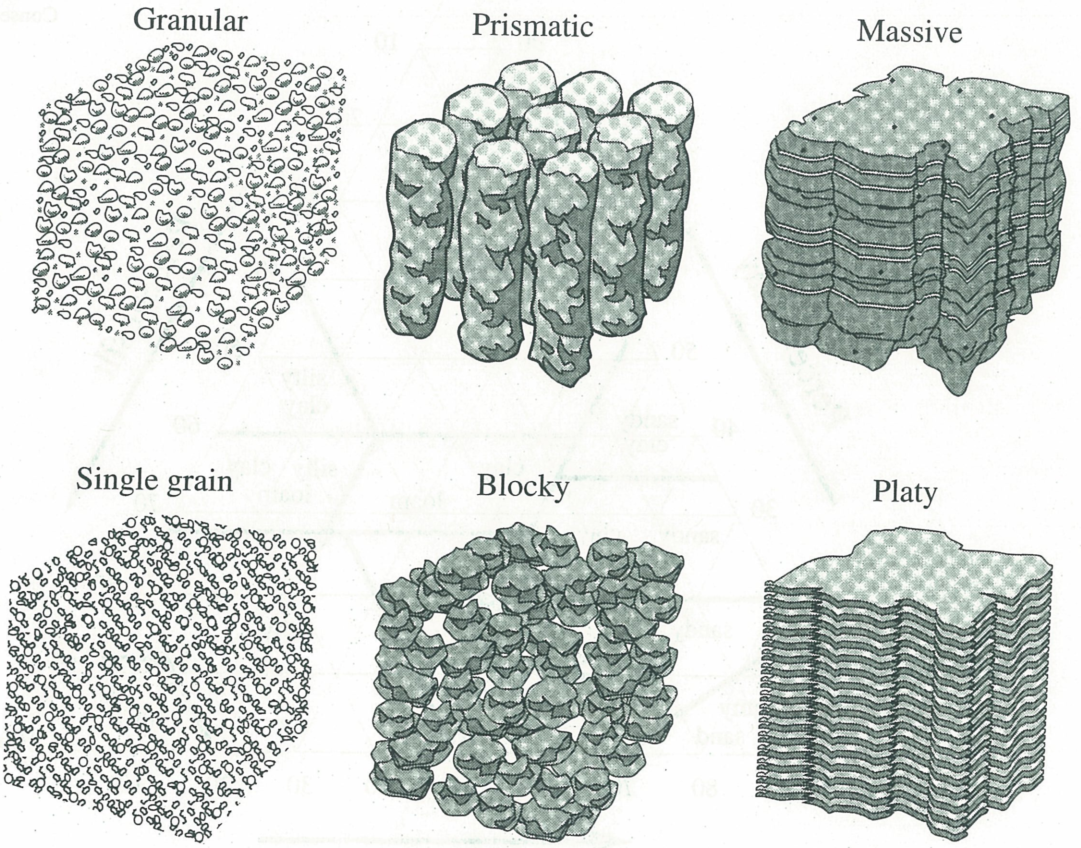


Figure 2.5 Depictions of soil structures; data from USDA, Agriculture Information Bulletin, No. 199, 1959

often confused with ped is *concretion,* which is a coherent mass formed within soil by chemical precipitation from percolating ground water. An example of this phenomenon is "iron stone" formation in the southern Coastal Plains near Tifton, Georgia. Sometimes referred to as "shot," these concretions become very hard. Upon crushing these pellet shaped concretions, it is observed that they are very hard on the outside with a small content of normally appearing soil on the inside.

As mentioned, structure determines the rate of infiltration of precipitation into the soil profile. Generally, as structure changes from granular or blocky to massive or platy, infiltration decreases. Examples of media structure (shape and arrangement of peds) can be seen in figure 2.5.

2.4 SOIL CLASSES

The texture class of a soil is determined by the proportion of the three size fractions. The soil texture triangle in figure 2.6 depicts the various texture classes a soil can be assigned, based upon the size fraction ratios. The texture triangle is based mainly on the USDA classification, with clay size < 0.002 mm, silt 0.002–0.05 mm, and sand 0.05–2.0 mm. Use of the triangle is

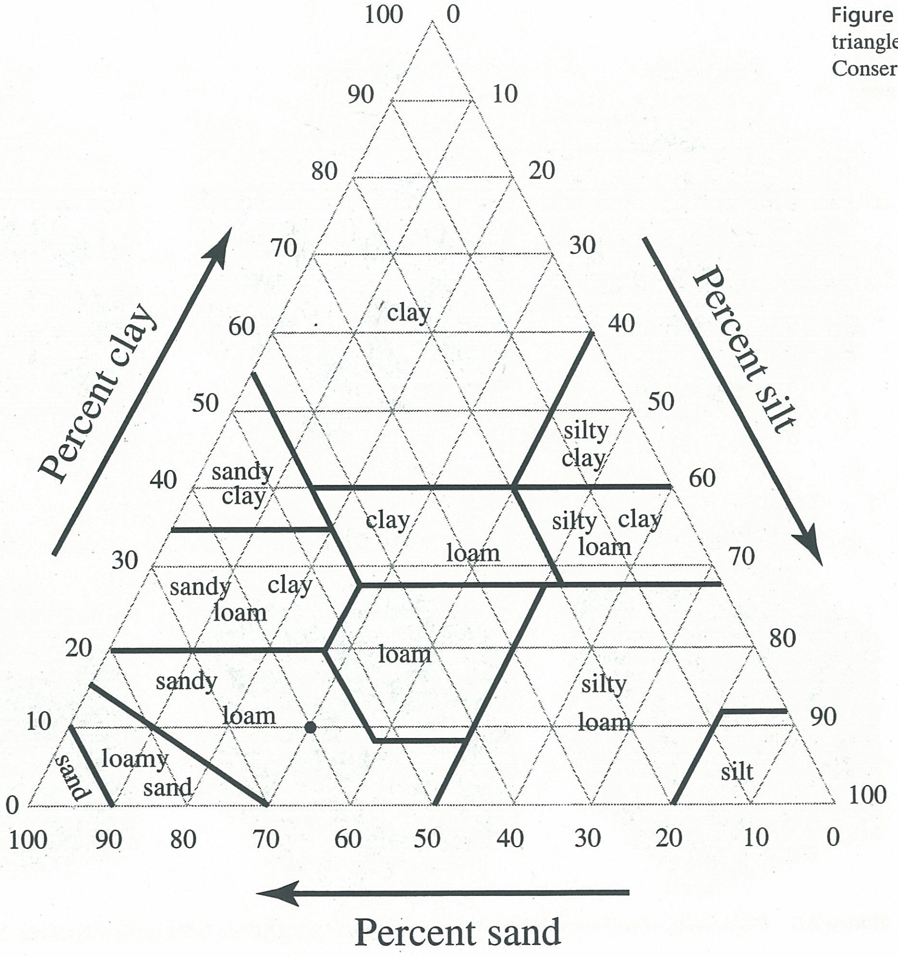


Figure 2.6 Soil texture triangle. Source: U.S. Soil Conservation Service

fairly straightforward. Suppose one had a soil that was 30% silt, 60% sand, and 10% clay; which texture class would it belong to? Proceeding clockwise from the apex of the triangle, we see that the right side of the triangle is 0–100% silt, the bottom is 0–100% sand (from right to left), and the left side is 0–100% clay (from bottom left corner back to the apex). Begin by finding the line on the right side that coincides with 30% silt (parallel to 70% clay), and intersect this line with the 60% sand line (parallel to 0% clay at the bottom of the triangle). By intersecting these two lines with the line that coincides with 10% clay from the left, we see that this soil is a sandy loam (see "black dot" in lower left corner of figure 2.6).

Examination of the texture triangle shows that finer soils (clays) are located at the top of the triangle, coarser soils (sands) in the lower left corner, and intermediate soils (silts) in the lower right corner. The center of the triangle is composed of clay loam and loam soil types. These two soil types, especially the clay loam, have an even ratio of fine, intermediate, and coarse size fractions. Their infiltration rates, CEC, specific surface area, and other physical characteristics will be between those soils that border them on either side; that is, a clay loam will conduct more water than a silty clay loam, but less than a sandy clay loam. Agriculturally, this clay loam would be a preferable medium under most circumstances, because it would retain more moisture and have better aeration and drainage than other soils, and would also have an ample nutrient supply.

QUESTION 2.7

If a soil contains approximately 20% clay, and the remaining fractions are equal amounts of sand and silt, what is its texture?

2.5 PARTICLE SIZE ANALYSIS

Particle size distribution is related to the physical and chemical activity of soils, and represents a stable soil characteristic. As discussed in section 2.8, particle size distribution indirectly relates to specific surface area. Because many chemical and physical properties of porous media are associated with surface activity, particle size analysis is the standard means for characterizing and classifying fractions. The proportioning of solids into different size ranges is termed *particle size analysis* or *mechanical analysis.*

Soil solids are composed of discrete units. To quantitatively determine separate particle size, the solids must be separated into these discrete units by both chemical and physical means. The complete particle size analysis of a medium involves two main processes: *dispersion,* the chemical or physical separation of a medium mass into its discrete units by the removal of cementing agents (such as organic matter, carbonates, and oxides) and subsequent disaggregation of clays; and *fractionation,* the grading or separating of particles into different particle size groups. Prior to dispersion, removal of organic matter and other cementing agents is accomplished by the addition of hydrogen peroxide in the case of organic matter, hydrochloric acid for carbonates, and oxalic acid–sodium sulphide for the ferric state of iron and iron and aluminum oxides. After removal of the cementing agents, *deflocculation* (separation of clay particles) is performed by the addition of a chemical dispersing agent and some form of mechanical agitation.

Mechanical agitation is usually accomplished by shaking or stirring and, occasionally, ultrasonic vibration. Sodium hexametaphosphate [$Na_3(PO_4)_6$] is commonly used as the dispersing agent. The sodium monovalent cation replaces the polyvalent cations adsorbed on porous media, thereby breaking interparticle linkages. Activity of the polyvalent cations is also reduced by precipitation by phosphorus. Adsorbed sodium cations raise the electronegativity of colloids until these particles repel each other, and complete dispersion or deflocculation occurs (this causes a thickening of the electrical double layer). Care must be taken to insure complete dispersion. Incomplete dispersion results in small particles, flocs, or aggregates settling as larger (sand) or intermediate (silt) sized particles that may bias results of analysis by causing lower values for both silt and clay.

After complete dispersion of a sample, fractionation is used to separate the primary particles into their respective size groups. Separation of coarse particles can be achieved by passing them through a nest of graded sieves to a particle diameter > 0.05 mm, which corresponds to a 270-mesh sieve. This sieving process can be performed by hand or with a mechanical shaker. The sieve nest is stacked in descending order, so that the coarsest sieve is on top, progressing to the finest sieve on the bottom. Sieving can be done under wet or dry conditions. Standard sieve and mesh sizes used for fractionation are listed in table 2.2.

Sedimentation is required for classification of particles finer than 0.05 mm. In 1851, Sir George Gabriel Stokes formulated what has become known as *Stokes' law,* which is used in the calculation of fine particle fractions by sedimentation, in both the micropipette and hydrometer methods of particle size analysis. The theory of Stokes' law is that the resistance offered by a liquid of given density to the fall (terminal velocity) of a spherical particle is proportional to the square of the particle's radius and not to its surface area. The velocity of a particle falling without resistance will increase due to acceleration by gravitational forces. However, a particle falling through a fluid of given viscosity will encounter resistance, and the terminal velocity of the particle will be slowed as stated by Stokes' law (due to the product of the particle's radius and initial velocity and the viscosity of the fluid).

The resistance, or drag force (F_d), acting upon an individual particle as denoted by Stokes' Law, is given by

$$F_d = 6\pi\eta r v \tag{2.1}$$

where η is fluid viscosity (kg m^{-1} s^{-1}), and r (m) and v (m s^{-1}) are radius and velocity of the soil particle, respectively. In the cgs unit system, F_d is expressed in dynes (g cm^2/sec^{-2}). The force of gravity (F_g) acting upon the soil particle is expressed as

$$F_g = \frac{4}{3}\pi r^3(\rho_s - \rho_f)g \tag{2.2}$$

TABLE 2.2 Sieve Series Specifications

Size of sieve (μ)	Sieve number (mesh per in.)	Sieve opening (mm)	Nominal wire (diameter, mm)
4000	5	4.000	1.370
2000	10	2.000	0.900
1190	16	1.190	0.650
1000	18	1.000	0.525
840	20	0.840	0.510
500	35	0.500	0.315
250	60	0.250	0.180
210	70	0.210	0.152
177	80	0.177	0.131
149	100	0.149	0.110
74	200	0.074	0.053
53	270	0.053	0.037
37	400	0.037	0.025

where $(4/3)\pi r^3$ is the particle volume (cm^3), ρ_s is the particle density ($\mathrm{g\ cm}^{-3}$), ρ_f is fluid density ($\mathrm{g\ cm}^{-3}$), and g is gravitational acceleration ($\mathrm{cm\ s}^{-2}$). Setting $F_d = F_g$ and solving for v, we obtain Stokes law:

$$v = \frac{2}{9}\frac{r^2(\rho_s - \rho_f)}{\eta}g \tag{2.3}$$

However, we are usually interested in both the time of settling and the particle diameter. Rewriting equation 2.3 in terms of diameter rather than radius, we have

$$v = \frac{d^2(\rho_s - \rho_f)g}{18\eta} \tag{2.4}$$

Because $v = h/t$ (distance over time), we can substitute this relation into equation 2.4, and solve for the time (t) of settling to obtain

$$t = \frac{18h\eta}{d^2(\rho_s - \rho_f)g} \tag{2.5}$$

A rearrangement of equation 2.5 will yield the diameter of the particle, as follows.

$$d = \left[\frac{18h\eta}{tg(\rho_s - \rho_f)}\right]^{1/2} \tag{2.6}$$

Assuming a constant temperature (for constant liquid and particle densities), the right hand side of the equation is a constant (k) except for time, and can be rewritten as

$$d = \frac{k}{t^{1/2}} \tag{2.7}$$

Several limitations must be assumed for Stokes' law to reflect reality: (1) The particles should be large enough (> 0.0002 mm) to be unaffected by Brownian (thermal) motion of the fluid molecules by which they are surrounded. (2) The particles should be spherical, smooth, and rigid. (3) Free fall of the particles should be free from hindrance by other particles, that is, the solution must be sufficiently dilute to prevent one particle from interfering with another. (4) Laminar flow must exist around each settling particle. Practical limitations arise from the fact that the viscous drag on individual particles begins to increase nonlinearly with velocity, once the Reynolds number becomes about 1.0 (Schlicting 1968). This results in a particle of larger diameter falling more slowly than predicted by Stokes' law. Gibbs, Matthews, and Link (1971) show the specific nature of these effects for soil-type porous media; their results also indicate that soil particles larger than 100 μm do not follow Stokes' law and, therefore, 100 μm should be the particle size limitation when using Stokes' law. Particles larger than 2 mm should be removed by sieving, since they fall rapidly and can create turbulent flow. (5) Particles should be of the same density. The average density of most soil particles is 2.65 $\mathrm{g\ cm}^{-3}$, while that of iron oxides and other minerals may exceed 5 $\mathrm{g\ cm}^{-3}$. Additionally, hydrated particles will settle faster than dehydrated ones. (6) The temperature of the liquid must remain constant, since fluid viscosity is temperature dependent. Increases or decreases in temperature can increase or decrease terminal velocity, causing error in results.

Essentially, all of these parameters change with each soil sample. Particles are not smooth but have jagged edges, the shapes are not spherical but may be rod shaped or disk shaped or flat. Hence, the data gathered from particle size analysis are approximations, although they serve as a useful index for comparing one media to another.

Stokes' law applies to both the pipette and hydrometer methods. For these methods, the portion of sample collected in the sieve pan (< 50 μm) is suspended in a cylinder of water in a constant-temperature bath. After the suspension thermally equilibrates, the cylinder is stirred thoroughly, and time is measured beginning at the end of stirring. Regardless of the method used, a reference depth below the suspension surface, h_1 (generally 10 cm), is selected. Equation 2.5 is used to solve for time of settling to this point. Particles larger than a given size will all have settled by a selected sampling time, so the fluid above the reference point contains only materials of cumulative grain size smaller than this given size. By selecting sampling times for incrementally smaller grain sizes, concentrations of particles of the size increment may be determined by difference. For the hydrometer method, the percent sand is obtained directly from the first hydrometer reading (nominally 40 seconds), the percent silt is obtained from a two-hour corrected hydrometer reading, and the percent clay is obtained by difference (see equation 2.9 and the surrounding discussion). The pipette method utilizes direct sampling. A small subsample of the fluid in question is taken by inserting a pipette into the sol at a depth h_1 and time t. As with the hydrometer method, all coarser particles than those sampled for at the measured depth and time will have settled beyond that point. The soil fraction of interest in the pipette method is the clay fraction ($< 2\ \mu$m). Settling time is dependent on temperature as well. When using the pipette method, care must be used so that the suction applied to the pipette is not sufficient to disturb the solution surrounding its tip.

Hydrometers come in two types: specific gravity, and gram per liter. The specific-gravity hydrometer measures the percentage of sample remaining in suspension at the reference point discussed previously. This percentage may be calculated by the following equation:

$$P = \left[\frac{V}{W}\frac{\rho_s}{\rho_s - \rho_f}\right](R_a - \rho_f)100 \tag{2.8}$$

where P = percentage media remaining in suspension at the reference point; V = volume of suspension (cm^3); W = weight of oven dry soil (g); ρ_s = particle density (g cm^{-3}); ρ_f = fluid density (g cm^{-3}); and R_a = corrected hydrometer reading. Corrections to the reading obtained must be applied due to presence of dispersing agent, temperature, and position of meniscus. The dispersing agent changes the specific gravity of the deionized or distilled water used; hydrometers are usually calibrated in the factory at 20 °C, thus changes in temperature will affect the reading, due to change in the viscosity of the solution; and the hydrometer is designed to be read at the bottom of the meniscus (if the reading cannot be taken at the bottom of the meniscus a correction factor must be applied). These correction factors require the recalibration of the hydrometer for present conditions.

The gram-per-liter hydrometer directly reads the concentration of suspension at the reference point in gL^{-1}. The percentage of soil remaining in suspension can be obtained by

$$P = \frac{Ra}{W}100 \tag{2.9}$$

where a = correction factor due to particle density (see table 2.3). For vessels less than ten times the particle diameter the following equation should be used:

$$v = \frac{2}{9\eta}\frac{r^2(\rho_s - \rho_f)g}{\left(1 + 8\frac{r}{R_a}\right)\left(1 + 3\frac{r}{L}\right)} \tag{2.10}$$

where L = point of measurement within hydrometer.

Two additional methods for measuring soil texture are centrifugation and elutriation. Centrifugation is similar to the two previously discussed sedimentation methods, except that

TABLE 2.3 Correction Factor for Varying Particle Density

Particle density (kg m^{-3}) ρ_s	Correction factor a
2950	0.94
2900	0.95
2850	0.96
2800	0.97
2750	0.98
2700	0.99
2650	1.01
2600	1.02
2550	1.03
2500	1.04
2450	1.05

the particles are forced to settle much more rapidly through the suspension. Modification of Stokes' Law allows calculation as follows.

$$\frac{dR_r}{dt} = \frac{R_r\omega^2(\rho_s - \rho_f)d^2}{18\eta} \tag{2.11}$$

where R_r = radius of rotation; t = time; and ω = angular velocity.

The radius of rotation (R_r) is taken as the distance from the center of the centrifuge to the surface of the suspension within the tube. Integrating equation 2.11 and solving for time gives

$$t = \frac{18\eta \ln \dfrac{R_{r_2}}{R_r}}{\omega^2 d^2(\rho_s - \rho_f)} \tag{2.12}$$

where R_{r_2} = depth to point of measurement. As with equation 2.10, this can also be solved for particle diameter (d). The centrifugation method is used where particle size is < 50 nm. This size particle will be affected by Brownian motion, and might not settle if not for the added centrifugal force.

The final method to be discussed is that of elutriation. Soil fractions are determined by letting particles wash out in a flowing current of water at a constant velocity (v_1). Water is passed through a vertical cylinder from bottom to top. A velocity is chosen, which will carry off any particle whose velocity of fall in still water is less than the set velocity. When the flowing water is clear and free of particles, the velocity is increased to v_2 and the process repeated. An easier way to perform this analysis is to pass the porous media through a series of flasks which decrease in size with each flask. In this manner each flask contains different particle sizes.

As a final note on the determination of texture, there is a quick, albeit "dirty" method for rapid determination in the field. This is by feel, a method in which many soil scientists have been trained. It involves molding soil in the hand and rubbing it between the thumb and index fingers, under both moist and dry conditions. Based on the length and size of the ribbon which can be formed, the soil texture is determined. With some soil (such as sands) no ribbon will be formed, and individual particles will feel gritty; the soil will also lack any plasticity. As clay and silt contents increase, there will be a smoother feeling and longer ribbons can be formed. This method is mostly used in soil mapping and surveying. A trained individual can be very adept at judging soil texture by feel. In fact, the American Society of Agronomy has both regional and national student competitions each year that utilize this method.

QUESTION 2.8

The pipette method of texture analysis is being used to determine the time required for a particle ($d = 2\ \mu\text{m}$) to settle to a depth of 10 cm. Calculate t and v using Stokes' law. Assume $\eta = 1.002 \times 10^{-2}\ \text{g cm}^{-1}\ \text{s}^{-1}$ and $\text{g} = 981\ \text{cm s}^{-2}$.

QUESTION 2.9

Calculate F_d and F_g acting upon the above particle. Assume the same properties as in the previous question, and that $\rho_f = 1.39\ \text{g cm}^{-3}$ and $\rho_s = 2.65\ \text{g cm}^{-3}$.

2.6 SOIL AS A PHASE SYSTEM

A phase is one part of a system, uniform throughout in physical properties and chemical composition. It is separated from other homogeneous parts of the system by boundary surfaces. Systems that exist in nature can be mono- or polyphasic. As described in the introduction to chapter 1, soil is a three-phase system composed of solid, liquid, and gas. The solid phase is composed of mineral and organic matter that make up the soil matrix; the liquid phase (or "soil water") is composed of water and dissolved substances; and the gas phase (or "soil air") contains several gases in dynamic equilibrium. A phase within a system can be homogeneous or heterogeneous (whether gaseous, liquid, or solid). If the chemical composition within the phase does not vary from one location to the next the phase is homogeneous, else it is heterogeneous; it can be homogeneous physically but heterogeneous chemically, or vice versa. An example of this would be ice and water, which is chemically homogeneous but physically heterogeneous.

A system can be completely heterogeneous. Within the soil system, the interfaces between solid, liquid, and air are continually changing, resulting in extensive interactions of adsorption, ion exchange, displacement, surface tension, and friction between each phase or combination of phases. It is important to note here that the solid phase of soils is relatively stable in comparison to the very dynamic liquid and gaseous phases. Generally, the greater the interaction, the larger the interfacial area existing between the phases. This will become more apparent with the discussion of specific surface area in section 2.8. A volume composition denoting the three phases in a typical loam soil is shown in figure 2.7.

2.7 DENSITY AND VOLUME–MASS–WEIGHT RELATIONS

Particle Density

The density of grains composing porous media is known as particle density (or mean particle density). Particle density is defined as the mass of solids per unit volume, and can be obtained from the following equation.

$$\rho_s = \frac{M_s}{V_s} \tag{2.13}$$

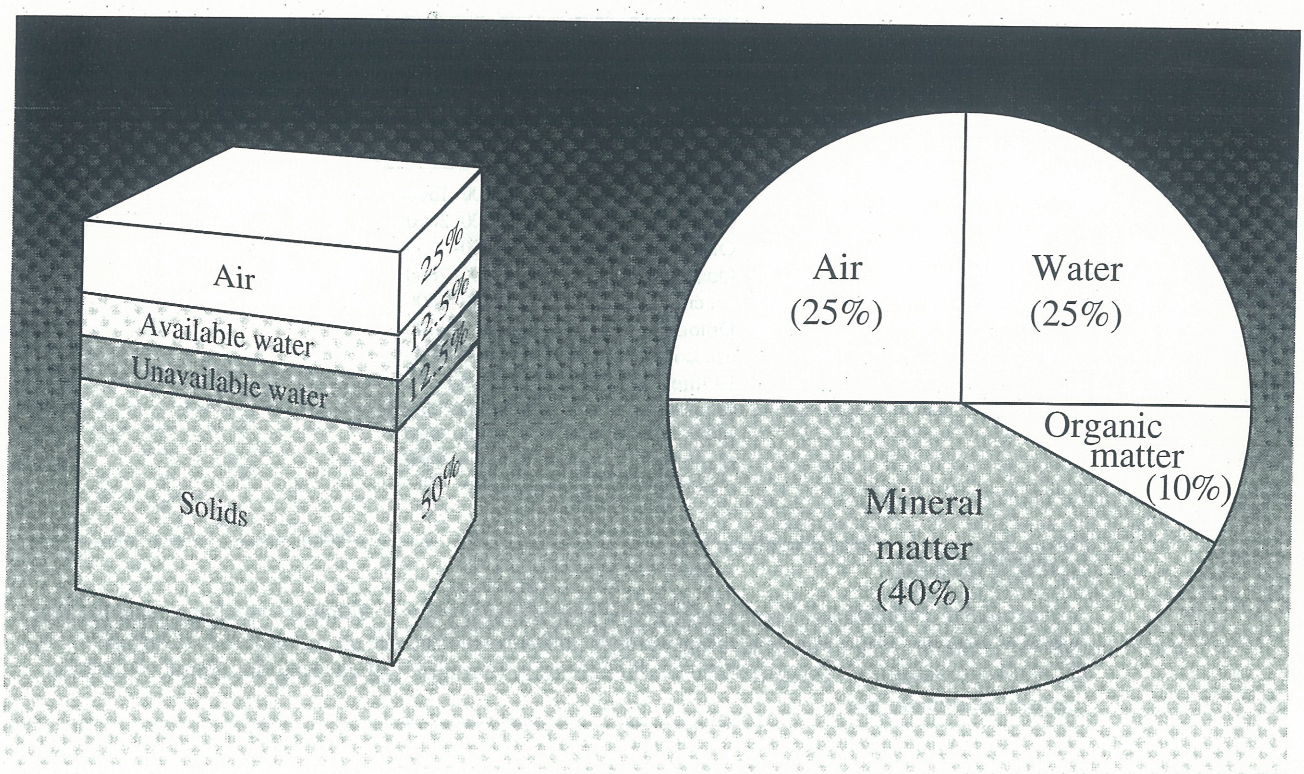


Figure 2.7 Cubical and pie-graph representations, illustrating a typical composition for a loam soil

where ρ_s = particle density (kg m^{-3}); M_s = mass of the solids (kg); and V_s = volume of the solids (m^3). Particle densities for various mineral constituents are given in table 2.4. The particle density of most quartz-type minerals such as sands, which are very common, is about 2650 kg m^{-3}; nonetheless, particle density for soils is quite variable. Tropical soils and other media high in iron and additional heavy minerals have a higher particle density, whereas soils containing substantial amounts of organic matter have lower particle densities.

The most common method of measuring particle density is the use of a pycnometer. The mass is obtained by weighing the sample; the volume is obtained from the mass and density of water displaced within the apparatus. Particle density from the pycnometer is calculated as

$$\rho_s = \frac{\rho_f(m_s - m_a)}{(m_s - m_a) - (m_{sw} - m_w)} \tag{2.14}$$

where ρ_f = fluid density (g cm^{-3}); m_s = mass of pycnometer filled with oven-dry soil sample (g); m_a = mass of pycnometer filled with air (g); m_{sw} = mass of pycnometer filled with soil and water (g); and m_w = mass of pycnometer filled with water (g). Because soil particles vary in density, the data obtained from the pycnometer is a weighted average that will yield a mean particle density. Occasionally, density may be expressed as specific gravity, which is the ratio of the density of the material compared to water at atmospheric pressure and a temperature of 4 °C, rendering the value dimensionless. Other methods for determining density have also been developed, such as using radioactive isotopes (Bernhard and Chasek 1953).

TABLE 2.4 Particle Density of Various Soil Constituents

Mineral constituents	Particle density (kg m^{-3})
Humus	1300–1500
Clay	2200–2600
Orthoclase	2500–2600
Quartz	2500–2800
Calcite	2600–2800
Dolomite	2800–2900
Muscovite	2700–3000
Biotite	2800–3100
Apatite	3200–3300
Pyrite	4900–5200
Hematite	4900–5300

QUESTION 2.10

A hydrologist obtains a soil sample of 76 g (after oven drying) from a field site, which is placed in a pycnometer (100 cm^3 volume). Deaerated water is added to the sample to drive off all air at 20 °C. The mass of water and soil is 144 g. **(a)** Why was it necessary to remove all air from the sample? **(b)** What is the particle density $^b\rho_s$?

Void Ratio

The void ratio, *e*, is defined as the ratio between the volume of voids and the volume of solids. It is expressed as

$$e = \frac{(V_a + V_w)}{V_s} \tag{2.15}$$

This relates the fractional volume of soil pores to the volume of solids, rather than to the total volume of soil as in the definition of porosity, discussed next. The volume is composed of $V_a + V_w$, which are the volumes of air and liquid phases, respectively, and V_s, which is the volume of solids. Since porosity is usually expressed as ϕ, one will often see $V_a + V_w$ used when discussing void ratio.

QUESTION 2.11

An intact soil core (30 cm diameter × 40 cm height) is obtained from a research facility in Nevada. The core is weighed and placed in an oven to dry, after which it is reweighed. Water loss is 9766 cm^3 and ρ_b (dry bulk density) = 1.30 g cm^{-3}. Assume a particle density of 2.65 g/cm^3. **(a)** Determine V_a and V_w. **(b)** What is the void ratio?

Porosity

Soils are composed of mixtures of discrete large and small particles that may be loose single grains or bound in the form of aggregates, but the quantity of smaller particles and the aggregate size within a given soil has a marked effect on porosity (Diamond 1970; Klute 1986; Marshall 1958; Page 1948). A soil physicist is mainly concerned with unconsolidated soil, in which porosity depends on the packing of the grains and aggregates, their size distribution, shape, and arrangement. An engineer, on the other hand, may be concerned with consolidated media, where the porosity depends upon the degree of particle cementation within the materials. Porosity is expressed as a volume percent, ranging in most soils from 30–60%. Like many other media properties, soil porosity is governed more by particle size than any other parameter. Generally, the smaller the particle size the smaller the pores, but the greater the porosity; hence, coarse soils have a lower porosity than fine soils. However, particle sorting also has an affect, so this tendency is not absolute.

Porosity is expressed by

$$\phi = \frac{V_f}{V_t} = \frac{(V_a + V_w)}{(V_s + V_a + V_w)} \tag{2.16}$$

To understand the manner in which porosity depends upon structure and sorting arrangement (mode of packing), we should first consider the normal packings of uniform rods or spheres. Figure 2.8 illustrates simple models of porous media and their equivalent porosity. It can be seen that the least compact arrangement of uniform spheres is achieved through cubical packing ($\phi = 48\%$), and the most compact arrangement is represented by a rhombohedredal packing in which each sphere is tangential to twelve neighboring spheres ($\phi = 26\%$). Other packings of intermediate arrays fall somewhere between these two. Because the packings are of uniformly sized spheres, porosity is independent of particle size. If particles vary in size, porosity is dependent on particle size and distribution.

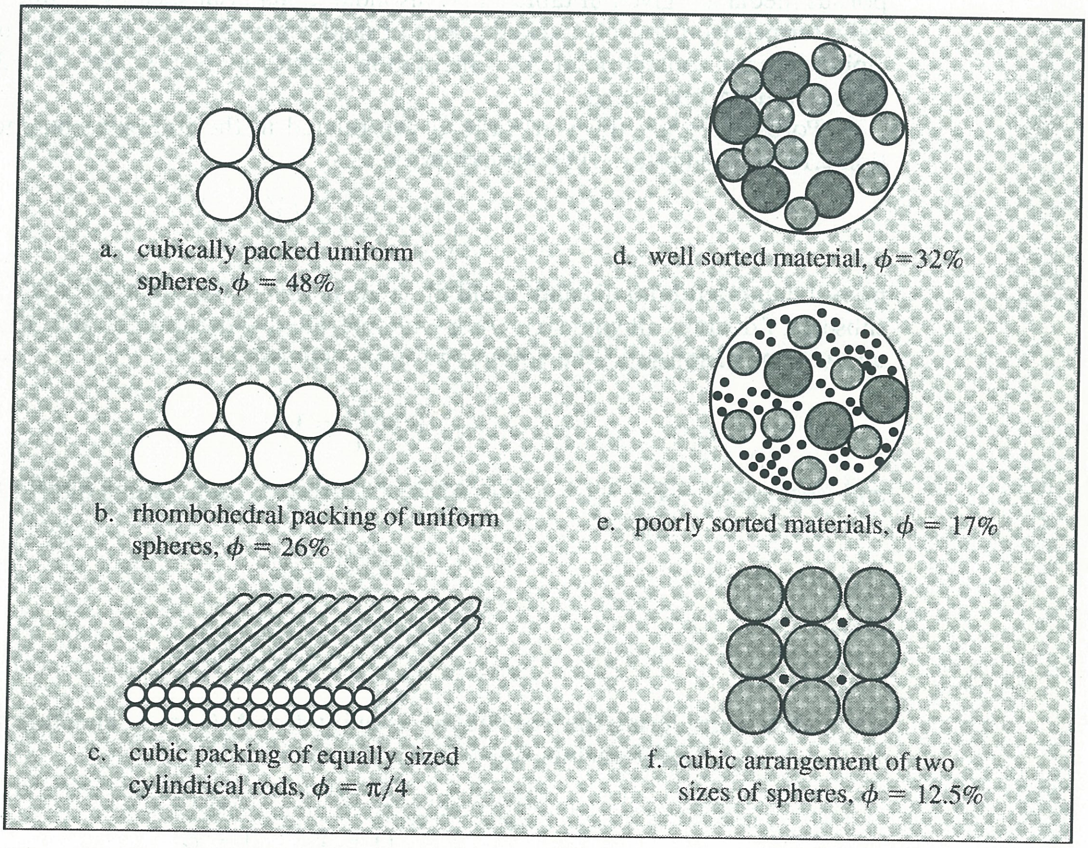


Figure 2.8 Representations of various packings and arrangements for porous media, showing porosity (ϕ) for each arrangement

TABLE 2.5 Bulk Density and Porosity Values of Selected Soils

Porous media	Bulk density* (kg m^{-3})	Porosity value (%)
Limestone and shale[†]	2780	1–20
Sandstone	2130	10–20
Gravel and sand	1920	30–35
Gravel	1870	30–40
Fine to medium mixed sand	1850	30–35
Uniform sand	1650	30–40
Medium to coarse mixed sand	1530	35–40
Silt	1280	40–50
Clay	1220	45–55
Ideal soil (even particle size distribution)	1310	50–60
Peat ("muck")	0.65–1.1	60–80

Sources: Data from Bear 1972; Day 1965; and Gill 1979

*Values for bulk density reflect average values, and are not specific for a given media type

[†] Includes dolomite

Other factors that affect the porosity of soils are compaction, consolidation, and cementation. Since compacting (overburden) forces vary with depth, porosity will also vary with depth (this is especially true of clays and shales). Typical porosity values for common porous media are given in table 2.5. Consolidated material such as sedimentary rock is initially formed from loosely packed grains of sand, which become cemented by chemical precipitates at the contact point. As cementation progresses, pore space is filled and the porosity is significantly reduced.

Porosity, ϕ, and the void ratio, e, are related in the following manner: Given equation 2.15 the void ratio, e, may be expressed as

$$e = \left(\frac{V_t - V_s}{V_s}\right) = \left(\frac{V_t}{V_s} - 1\right) \tag{2.17}$$

by assuming that both air and liquid fill voids in a medium; then, $V_a + V_w = V_f$, where the subscript f denotes all void space (for example, filled with liquid or air). Thus, in the case of unit volume of soil solid, the void ratio can be expressed as

$$e = \frac{V_f}{1} = V_f \tag{2.18}$$

and, because $V_t = 1 + V_f = 1 + e$ (where 1 represents unit volume of solid), then

$$\phi = \frac{e}{(1 + e)} \tag{2.19}$$

and

$$V_f + V_s = 1; \qquad V_s = 1 - V_f \tag{2.20}$$

As a result, considering the soil volume as unity, $\phi = \phi_v/1 = \phi_v$; and, due to equation 2.15, $V_f = \phi = eV_s$; thus, (remembering that $\phi = (V_t - V_s)/V_t$)

$$e = \frac{\phi}{V_s} = \frac{\phi}{(1 - V_f)} = \frac{\phi}{1 - \phi} \tag{2.21}$$

Typically, porosity is the measure preferred by hydrologists and soil scientists, whereas void ratio is usually used in geochemical engineering. Porosity also has a relation with bulk density (discussed in the following subsection), $\phi = [1 - (\rho_b/\rho_s)]$.

Several methods have been developed for the determination of porosity (Bear 1972; Klute 1986). This involves finding V_t, and either V_s or V_f. Here, V is volume and the subscripts t, s, and f refer to total, solid, and fluid respectively. Most of these methods are termed direct methods. The total volume of a sample can be determined by simple measurement if the sample was obtained with a cylindrical or cubical sampling device. However, any device or sample of irregular shape may also be used for determination of V_t (total volume). The most common method for determining V_t is to measure the volume of a fluid displaced by an immersed sample, taking care that the fluid does not penetrate the pores within the sample. The most common direct methods for measuring porosity are: **(1)** *Pycnometer*—V_t is determined first, then the sample is dried and crushed to remove all pores and placed in a pycnometer, where V_s is determined by fluid displacement. A pycnometer is filled with the sample and mercury, and then with mercury alone, to obtain V_t by weight difference (Klute 1986). **(2)** *Mercury Injection*—The sample is placed in a chamber filled with mercury at atmospheric pressure. The volume of mercury displaced gives V_t. Next, the pressure within the chamber is gradually increased, to force the mercury into the pores. This will yield not only V_f (volume of fluid forced into pores), but also a capillary pressure curve. If very high pressures are required, however, this method will not be suitable (Klute 1986). **(3)** *Compression Chamber*—Two chambers are connected by a valve. The sample is placed in the first chamber, and the gas is evacuated from the second chamber. A valve between chambers is then opened to allow isothermal expansion of the gas remaining in the first chamber. Considering an initial pressure of p_1 and a final pressure of p_2, we have

$$(V_1 - V_s)p_1 = (V_1 - V_s + V_2)p_2 \tag{2.22}$$

or

$$V_s = V_1 - \left(\frac{p_2}{p_1 - p_2}\right)V_2 \tag{2.23}$$

where V_1 = volume of the first chamber; V_2 = volume of the second chamber; V_s = volume of solid. As a result, the porosity $\phi = 1 - V_s$. **(4)** *Gas Expansion*—This method is based on Boyle–Mariotte's gas law, and is carried out at constant temperature. A sample is placed in one of two chambers. The first chamber is at V_1, with initial pressure of P_1. Gas is allowed to expand isothermally into the second chamber (by opening the interconnecting valve), which had previously been evacuated and now has a pressure of p_0. The first chamber equilibrates to a pressure of p_2 and, from Boyle-Mariotte's law, we have

$$\frac{p_1(V_1 - V_s)}{Z(p_1)} + \frac{p_0 V_2}{Z(p_0)} = \frac{p_2(V_1 + V_2 - V_s)}{Z(p_2)} \tag{2.24}$$

where $Z(p)$ is the compressibility factor. **(5)** *Washburn–Bunting Porosimeter*—V_f is determined by measuring the volume of air extracted at atmospheric pressure from a sample, by creating a partial vacuum in the porosimeter. A partial volume is obtained by manipulating mercury in the porosimeter reservoir.

An example of an indirect method is that developed by Norel (1967), based on the absorption of radioactive particles by a fluid allowed to saturate a sample. Because of the radiation involved, this has not gained much popularity.

QUESTION 2.12

Using equation 2.21 and the void ratio obtained from question 2.11, what is ϕ for the core described in question 2.11?

Bulk Density

The bulk density of a medium is usually expressed in terms of dry soil, and is the ratio of media mass to total volume, which includes all three phases of the soil system: $\rho_b = V_s\rho_s + V_w\rho_w + V_s\rho_s$; V_α is the volume fraction of phase α, where the phase is either liquid, solid, or gas (Davidson, Biggar, and Nielsen 1963; Gardner and Calissendorff 1967). Average bulk density values for selected media are listed in table 2.5. Since the volume fraction is defined as a ratio of the volume of a phase to the total volume of media, the sum of the ratios of all phases represents that total volume and, thus, equals 1. Dry bulk density, expressed in kg m^{-3}, is calculated as follows:

$$\rho_b = \frac{V_s\rho_s}{V_t} = \frac{M_s}{V_t} \tag{2.25}$$

where M_s = dry mass of porous media in kg (usually considered "dry" after oven drying at 105 °C for twenty-four hours). Bulk density can range from 1220 kg m^{-3} in clay soil to 1850 kg m^{-3} in sands. Bulk density, like porosity, is affected by the uniformity of packing, amount of compaction, size distribution, and clay type. Bulk density is always lower than particle density, because bulk density considers both air and liquid fractions that have large volumes compared to the density of a single particle; it is also dependent on clay and water content in shrink–swell porous media. Particle density is independent of water content. A wet soil cannot be completely compacted and remain porous. The bulk density will still be moderately lower than particle density, even under extreme compaction.

QUESTION 2.13

Calculate ρ_b of 1 m^3 of soil assuming volume fractions of solid, liquid, and gas of 0.5, 0.34, and 0.16, respectively. Assume ρ_l = 1.0 g cm^{-3} and ρ_g = 1.3 kg m^{-3}.

QUESTION 2.14

Assume the soil in question 2.13 is dry. **(a)** What is the contribution of each of the phases to bulk density? **(b)** What is the bulk density?

QUESTION 2.15

Consider the relation of the bulk density of a soil to all three phases, as previously discussed. **(a)** Derive an equation for a dry soil analogous to this. **(b)** With this answer, express the difference between a dry bulk density (ρ_b) and a wet bulk density (ρ_w).

Water Content

Water content, or wetness of a soil, is the volume or mass of water V_w occupying space within the pores (Bruce and Whisler 1973; Davidson, et al. 1969; Gardner 1986). Water content is typically expressed in one of three ways: mass basis, volume or depth basis, and percent or degree saturation.

Mass wetness (sometimes referred to as gravimetric water content) is determined by extracting a soil sample, oven-drying it (generally 105 °C for 24 hours), and determining the amount of water lost through the drying process. Oven drying is necessary to remove hygroscopic water adhering to particles that cannot be removed by air drying. This is sometimes a significant amount, depending on clay content and specific surface area; however, not all hygroscopic water can be removed. Mass wetness is described by

$$w = \frac{V_w \rho_w}{V_s \rho_s} \tag{2.26}$$

where V_s and V_w are the volume fractions of the solid and liquid phases, ρ_s is the density of the solid phase, and ρ_w and ρ_b are the wet and dry bulk densities.

Volumetric water content is generally more useful for field and laboratory studies, because it is the form in which soil water content (as a fractional basis, e.g., 0.34, 0.48, etc.) is reported—in the results from gamma attenuation, neutron probe, and time domain reflectometry (TDR) water content measuring devices. Like mass wetness, it is generally reported as a percentage, but is compared to total volume rather than the volume of solids present. Because both the V_w and V_t units of measure are cm^3, they cancel in the equation, leaving θ unitless. Volumetric water content is given by

$$\theta = \frac{V_w}{V_t} \tag{2.27}$$

Volumetric water content may also be obtained by its relation to mass wetness and bulk density, as follows.

$$\theta = w\left(\frac{\rho_b}{\rho_w}\right) \tag{2.28}$$

where ρ_w = density of water at measured temperature. A diagrammatic representation of the volume–mass relation within soil is shown in figure 2.9.

Equivalent depth of water is a measure of the ratio of depth of water per unit depth of porous media, described by volumetric water content:

$$d_w = \theta d_t \tag{2.29}$$

where d_w = equivalent depth of soil water if it were extracted and ponded over the surface (cm), and d_t = total depth of soil under consideration (cm). For example, water ponded on one hectare to a depth of 2.54 cm has a volume of 254 m^3. Depth of water is a very important concept in agricultural and irrigation practices.

Degree of saturation, s, expresses the volume of water in relation to the volume of pores. In arid-zone soils that remain perpetually dry, saturation can be as low as 0%; but it may be as high as 100% in lowland or wetland soils that remain constantly moist. However, most soils above the water table will never reach 100% saturation because of air entrapped in "dead end" pores. Degree of saturation is expressed by

$$s = \frac{V_w}{V_a + V_w} \tag{2.30}$$

(For groundwater reservoirs, entrapped air will normally dissolve within several months, resulting in water that is supersaturated with N_2 and air.)

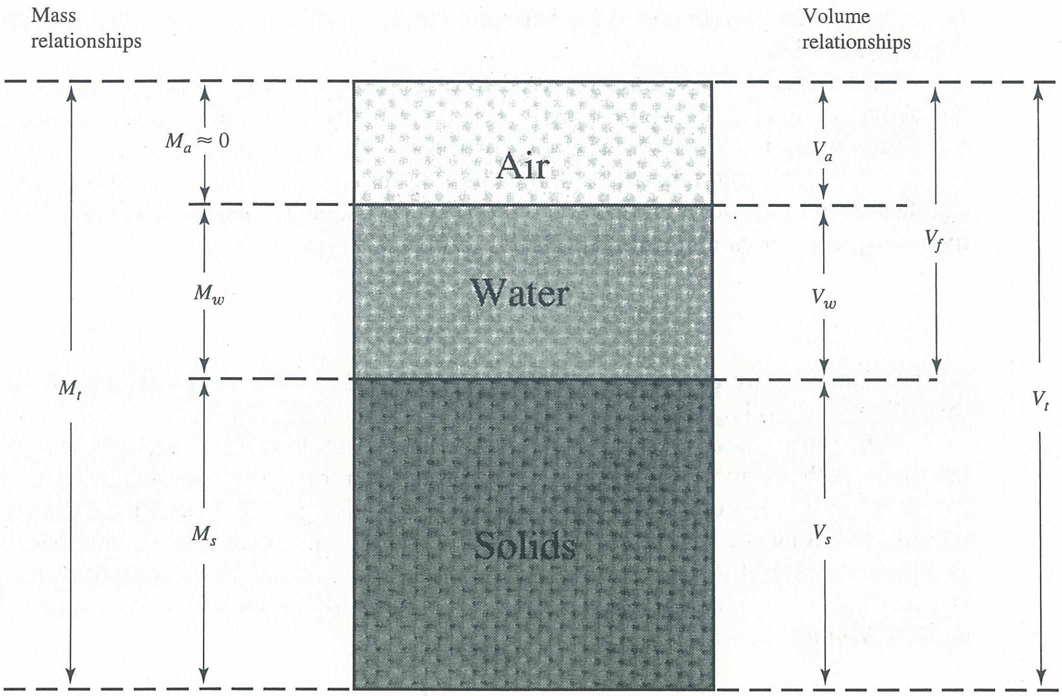


Figure 2.9 Diagrammatic representation of volume–mass relation within porous media

QUESTION 2.16

A soil physicist decides to apply water to a research area to investigate the movement of a conservative tracer. If $\rho_b = 1.37$ g cm^{-3} and gravimetric water content (w) is 0.19, what is the volumetric or volume fraction of water (θ) of the research area?

QUESTION 2.17

The scientist in question 2.16 wants to move the chemical tracer to a depth of 40 cm. **(a)** What volume of water must be applied per m^2 of soil surface? **(b)** If the scientist uses 23% less water than needed, to what depth would the tracer move? Assume that the tracer will only move to the maximum depth of the wetting front, although in field situations there would be great variability.

QUESTION 2.18

A soil sample is collected with an initial weight of 136 g. After oven-drying at 105 °C for 24 h, it weighs 118 g. Assuming $\rho_b = 1.32$ g cm^{-3} calculate: **(a)** w. **(b)** θ.

Air-Filled Porosity

We have discussed void ratio and porosity, which consider both air volume and water volume that occupy soil pores. The term air-filled porosity is used to specify the volume ratio of the air-filled pores. Air-filled porosity is expressed as

$$V_f = \frac{V_t - V_s - V_w}{V_t} \tag{2.31}$$

where V_f = volume fraction of air occupying soil pores. Equation 2.31 is a measure of the relative air content of soils. As such, it is an index of soil aeration, which is important in an agricultural setting. For example, an unsaturated zone hydrologists seeking to determine phase transport of herbicides would consider that air-filled porosities $< 10\%$ are not conducive to gas transport (since the soil is near saturation at that point). Porosity is also related negatively to the degree of saturation s by $V_f = \phi - s$ (remember that, in this case, $\phi = (V_a + V_w)/V_t$).

QUESTION 2.19

A 1 m^3 volume of porous media has $\theta = 0.39$ (see equation 2.27). **(a)** What is the air-filled porosity V_f ? **(b)** If ρ_b (the oven-dry bulk density) = 1.32 g cm^{-3}, what is V_a? **(c)** What causes the difference between **(a)** and **(b)**? **(d)** What is the difference between the void ratio and the air-filled porosity of a porous media?

2.8 SPECIFIC SURFACE AREA

The specific surface area (s) of a soil is generally defined as the total interstitial surface area of pores (A_s) per unit bulk volume (V_t), per unit mass (M_s), or per unit volume of platelets (V_s) (Bower and Goertzen 1959; Carter, Mortland, and Kemper 1986). In equation form, they appear as follows.

$$s_b = \frac{A_s}{V_t}; \quad s_m = \frac{A_s}{M_s}; \quad s_v = \frac{A_s}{V_s} \tag{2.32}$$

where s_b, s_m, and s_v = specific surface area per unit bulk volume of soil, per unit mass, and per unit volume of particles, respectively; A_s = total surface area per specified unit; V_t = bulk volume of soil; M_s = mass of solid particles; and V_s = volume of solid portion.

Specific surface area is normally expressed in square meters per gram of soil ($m^2\ g^{-1}$). The physical behavior of soil—water retention, hysteresis, or cation exchange, for example—is highly dependent upon the surface area. Particle size as well as shape are the parameters that have the greatest influence on the surface area of a porous media. Clays and organic colloids have very small particles; because of this, they have a high specific surface area compared to sands. This is especially true for clays with platy, disk, or rodlike shapes that have a far greater surface area per unit volume than spherical sand particles of the same volume. Not only do clay particles have a greater outer surface area but, as discussed previously, clays that shrink and swell (such as montmorillonite) expose a large internal surface area. The surface area of a typical sand is about 1 $m^2\ g^{-1}$, whereas that of montmorillonite is about 800 $m^2\ g^{-1}$. Mathematically, we can compare the surfaces of a sphere and of a thin rectangular particle to their volumes. For a sphere, we obtain

$$s_v = \frac{A_s}{V_s} = \frac{\pi d^2}{\frac{\pi}{6} d^3} = \frac{6}{d} \tag{2.33}$$

Relating this to a mass basis to obtain proper units, we get

$$s_m = \frac{s_v}{\rho_s} \tag{2.34}$$

Assuming $\rho_s = 2.65$ g cm^{-3}, and the diameter of a spherical particle is 0.05 mm, the surface area is 0.045 m^2 g^{-1}. For a thin rectangular particle such as a clay, the relation is

$$s_v = \frac{A_s}{V_s} = \frac{2L^2 + 4L^1 l}{L^2 l} = \frac{2(L + 2l)}{L^1 l} \tag{2.35}$$

Assuming that l is insignificant in the numerator (in this case, l is the thickness of the clay particle and $l << L$), due to properties of addition, we have $s_v = 2/l$ with units cm^2 cm^{-3}.

$$s_m = \frac{s_v}{\rho_s l} = \frac{2}{(2.65)l} \tag{2.36}$$

Further, assuming that $L = 1$ cm and the thickness (l) of the rectangular particle is 10 Å (10×10^{-10} m), the surface area will equal 750 m^2 g^{-1}. The specific surface area for selected soils are given in table 2.6.

The specific surface area of a soil is probably one of its most important properties. For unsaturated zone hydrologists investigating contaminant transport, the surface area plays a direct role in the movement of chemicals within the profile: how chemicals are retained and exchanged, released and retarded, by the soil surface. The greater the surface area of the soil, the more reactive it will be—especially in the presence of nonconservative chemicals and tracers. In such studies, surface area plays a very important role, and is a more useful index than texture for describing soil.

Specific surface area can be measured by various methods, the most common of which is gas adsorption as developed by Brunauer, Emmett, and Teller (1938). Nitrogen is adsorbed onto the soil surface in a monomolecular layer at low temperature (near the boiling point of liquid nitrogen). The adsorption of the gas takes place when a concentration gradient of the gas exists at the surface of the solid. The gas molecules are attracted by their degree of polarity to the force fields of the atoms at the solid surface.

TABLE 2.6 Specific Surface Area of Common Soil Types

Porous media	Surface area (m^2 kg^{-1})
Montmorillonite	700,000–800,000
Clay soil	150,000–250,000
Silty clay loam	120,000–200,000
Silt loam	50,000–150,000
Illite	80,000–120,000
Loam	50,000–100,000
Kaolinite	80,000
Sandy loam	10,000–40,000
Silt soil	5,000–20,000

Sources: Data from Gill 1979; Hillel 1980; and Carter, Mortland, and Kemper 1986

As Brunauer, Emmett, and Teller described the method, at sufficiently low gas pressure, the amount of nitrogen adsorbed per unit area is related to pressure, temperature, and heat of adsorption. This relation is given as follows.

$$\sigma = K_1 p \exp\left(\frac{Q_h}{RT}\right) \tag{2.37}$$

where σ = adsorption per unit area (mol m^{-2}); K_1 = constant (mol/N); P = gas pressure (Pa); Q_h = heat of adsorption (cal mol^{-1}); R is the gas constant (1.987 cal mol^{-1} K^{-1}); and T = temperature (K).

Equation 2.37 indicates that as pressure increases, the adsorption increases, and as temperature increases, adsorption decreases. The Langmuir equation describes this phenomena more precisely, because the soil surface has a finite area. It is given as follows.

$$\frac{P}{V} = \frac{1}{K_2 V_m} + \frac{P}{V_m} \tag{2.38}$$

where P = pressure; V = volume of adsorbed gas per gram of soil; K_2 = constant (1/Pa); and V_m = volume of adsorbed gas forming a monomolecular layer on adsorbent. Plotting P/V versus P will yield a straight line with a slope of $1/V_m$. Assuming a monomolecular layer, the specific surface area of the soil can be obtained by multiplying the cross-sectional area by the number of molecules in V_m.

As described by Brunauer, Emmett, and Teller (1938), and now known as the BET equation, the heat of nitrogen adsorption decreases with each successive layer of molecules adsorbed; however, it is assumed that after the first layer, subsequent layers are equal to the heat of condensation of the gas. The BET equation calculates the number of molecules in a monomolecular layer of adsorbent as follows.

$$\frac{P}{V(P_0 - P)} = \frac{1}{V_m C} + \frac{(C-1)P}{V_m C P_0} \tag{2.39}$$

where P_0 = the pressure of gas required for a monolayer saturation at experimental temperature, C = $\exp[(Q_{h1} - Q_{h2})/RT]$; Q_{h1} = heat of adsorption of the first layer of adsorbed gas, and Q_{h2} = heat of liquification of gas. For this equation, V_m can be obtained by plotting $P/V(P_0 - P)$ versus P/P_0. Because of the heterogeneity of the active surfaces of most adsorbents, the Langmuir equation is not strictly observed; the BET equation is most useful for pressures (P/P_0) of 0.05 to 0.45.

Other methods for determination of specific surface area, though not commonly used, include a statistical method developed by Chalky, Cornfield, and Park (1949); a heat of wetting method; a fluid flow method developed by Kozeny (1927); and an adsorption method using cetyl pyridinium bromide as the adsorbate. Several of these methods were compared by Brooks and Purcell (1952). A method that has gained attention is one in which ethylene glycol is added to a soil and allowed to evaporate until the rate of loss decreases and equilibrium is presumed to be reached, at which time it is assumed that only a monomolecular layer of gas remains on the surface. A more extensive review of this method, in which the adsorbate has been replaced by ethylene glycol monoethyl ether, may be found in Carter, Mortland, and Kemper (1986).

QUESTION 2.20

Calculate the specific surface area (s) of 35 g of soil that has an average particle diameter (d) of 20 nm.

SUMMARY

In this chapter we have discussed the mineralogical composition of a porous medium consisting of solid, liquid, and gaseous phases, and the various processes that affect that composition—such as solubility, hydration, oxidation, reduction, leaching, and so on. Also discussed were how soil profiles and horizons develop, and the colloidal and crystalline nature and properties of soils. Various types of silicate and amorphous clays were described in some detail. We illustrated the various fractions of soil (specifically, sand, silt, and clay) and how they are separated. The analysis of soil fractions is critical to many environmental reports and research investigations, because it is these fractions which primarily determine how a chemical will adsorb, degrade, transport, and contaminate a given site (contaminant transport is covered in detail in chapter 10). We presented precise density, volume, mass, and weight relations between soil phases that describe characteristics of various media. In particular, the specific surface area is very important in understanding physical behavior, because water retention, hysteresis, cation exchange, adsorption, and so on, are highly dependent on surface area. Chapter 3 will focus on the microscopic properties associated with soil particles—such as the electrical double layer, ion exchange, hydration, flocculation, and cracking.

ANSWERS TO QUESTIONS

2.1. When all media pores are filled with liquid, only two phases would exist (liquid and solid). Also, when a soil is very dry, all measurable liquid has been removed, leaving only the solid and gaseous phases.

2.2. The primary force is the electrostatic force, in which the cations are electrically attracted to soil particles. This is discussed in further detail in chapter 3.

2.3. Clays are not spherical; in general, they have a platelike shape. As plate thickness (δ) increases, s decreases. Thus, a stacking of clay layers occurs within crystals, which forms thicker particles. This is the most logical reason for differences in s among clay minerals.

2.4. The major processes involved in physical weathering are freezing and thawing, heating and cooling, wetting and drying, grinding actions, and microbial activities leading to decomposition.

2.5. The two major processes are *hydrolysis* and *carbonation.* In hydrolysis, minerals react with water to form hydroxides, such as orthoclase feldspar with water:

$$KAlSi_3O_8 + H_2O \rightarrow HAlSi_3O_8 + KOH$$

Hydrolysis is one of the most important processes in soil profile development and change. Carbonation is also important, because carbonic acid (H_2CO_3) dissolves minerals more readily than water alone, and also forms more soluble bicarbonates.

2.6. The three major chemical processes affecting structural change are: (1) *hydration,* in which water changes mineral structure and makes the mineral softer, more stressed, and more easily decomposed; (2) *oxidation,* where a chemical combining with oxygen changes in oxidation number of the element (usually increasing it), also increasing element volume and making it softer; and (3) *reduction,* which usually takes place in zones of decreased aeration that results in electrically unstable compounds, more soluble compounds, or compounds which are stressed more internally, all of which cause more rapid decomposition.

2.7. By examining the texture triangle (figure 2.6), one may determine that a mixture of 20% clay, 40% sand, and 40% silt is classified as a loam.

2.8. Using equation 2.5, we obtain $t = [18(10\ \text{cm})(1.002 \times 10^{-2}\ \text{g cm}^{-1}\ \text{s}^{-1})/(2 \times 10^{-4}\ \text{cm})^2 (2.65\ \text{g cm}^{-3} - 1.00\ \text{g cm}^{-3})(981\ \text{cm s}^{-2})] = 2.79 \times 10^4\ \text{s} = 7.7\ \text{h}$.

2.9. First, solve for v using equation 2.3. Assume $\rho_s = 2.65\ \text{g cm}^{-3}$ and $\rho_f = 1.39\ \text{g cm}^{-3}$; then, $v = 2/9[(2 \times 10^{-4}\ \text{cm}/2)^2(2.65\ \text{g cm}^{-3} - 1.39\ \text{g cm}^{-3})(981\ \text{cm s}^{-2})/1.002 \times 10^{-2}\ \text{g cm}^{-1}\ \text{s}^{-1})] =$

2.74 cm s^{-1}. Using this answer and equations 2.1 and 2.2, $F_d = 6\pi(1.002 \times 10^{-2}\ \text{g cm}^{-1}\,\text{s}^{-1}) \times (2 \times 10^{-4}/2)(2.74\ \text{cm s}^{-1}) = 5.18 \times 10^{-5}\ \text{g cm s}^{-2}$; and $F_g = 4/3\pi(2 \times 10^{-4}\ \text{cm}/2)^3(2.65\ \text{g cm}^{-3} - 1.39\ \text{g cm}^{-3})(981\ \text{cm s}^{-2}) = 5.18 \times 10^{-9}\ \text{g cm s}^{-2}$.

2.10. **(a)** Without driving off the air, the mass of soil and water in the pycnometer will be too small, resulting in a larger volume of solid than is actually present. This will yield a ${}^b\rho_s$ which is smaller than it should be. **(b)** Mass of solids = 76 g, mass of water + solids = 144 g, mass of water = (144 g − 76 g) = 68 g (this would also be the volume of water, assuming $\rho_f = 1.0\ \text{g cm}^{-3}$). The volume of solids = (100 − 68) cm^3 = 32 cm^3, thus, ${}^b\rho_s = M_s/V_t = 76\ \text{g}/32\ \text{cm}^3 = 2.38\ \text{g cm}^{-3}$.

2.11. **(a)** Because we know the water loss, $V_w = 9766/V_t = 0.34$. **(b)** The total mass of soil is $28.724 \times 10^3\ \text{cm}^3 \times 1.30\ \text{g cm}^{-3} = 37.34$ kg. To determine V_s, we must determine the total volume of soil and then divide it by the total volume of the intact core (28,724 cm^3). Thus, $(37.34\ \text{kg}/2650\ \text{kg m}^{-3})(1 \times 10^6\ \text{cm}^3\ \text{m}^{-3}) = 14{,}091\ \text{cm}^3$ soil. Divide this by V_t to obtain $V_s = 0.49$. Since all phases sum to unity, $V_g = (1 - V_s - V_w) = (1 - 0.49 - 0.34) = 0.17$. Hence, from equation 2.15, the void ratio = (0.17 + 0.34)/(0.49) = 1.04. This is a midrange value, since void ratios vary between 0.30 to 2.0.

2.12. Using the first part of the equation, $\phi = (1.04)(0.49) = 0.51$. This is slightly higher than one would expect, but has been measured in highly expanding soils.

2.13. Assuming standard values for ρ_a, the $\rho_b = (V_s\rho_s + V_w\rho_w + V_g\rho_g) = (0.50 \times 2650 + 0.34 \times 1000 + 0.16 \times 1.3)\text{kg m}^{-3} = 1665.21\ \text{kg m}^{-3}$.

2.14. **(a)** The contribution would be 1325 kg, 340 kg, and 0.21 kg for the solid, liquid, and gas phases, respectively. There is very little contribution by the liquid and gas phases. **(b)** $\rho_b = 1665\ \text{kg m}^{-3}$.

2.15. **(a)** Because the soil is dry, the liquid phase is removed, leaving $\rho_b = (V_s\rho_s + V_g\rho_g)$. **(b)** Considering a wet soil $\rho_w = (V_s\rho_s + V_w\rho_w + V_g\rho_g)$, by setting this equal to the value for ${}^b\rho_s$ in part (a), we obtain $\rho_w - \rho_s = (V_w\rho_w - \rho_g(\phi - V_g))$.

2.16. By assuming $\rho_w = 1.0\ \text{g cm}^{-3}$, we can use equation 2.30 to obtain $\theta = 0.19(1.37\ \text{g cm}^{-3}/1.00\ \text{g cm}^{-3}) = 0.26$.

2.17. **(a)** The scientist wants water to penetrate to a 40 cm depth, so a volume of 40 cm^3 of water will be required for each cm^2 of soil. Using the relation $\phi = 1 - (\rho_b/\rho_s)$, we have $\phi = 1 - (1.37/2.65) = 0.48$. Because the liquid portion $\theta = 0.26$ (from question 2.16), $V_g = 0.22$. Thus, if we want only the gas-filled portion of soil to be filled with water, $0.22 \times 40 = 8.8\ \text{cm}^3$ of water would be required to move the wetting front to the desired depth. However, because infiltrating water will tend to displace resident water in the soil pores through piston-type displacement, we would need to replace the entire porosity of the soil with new, infiltrating water. The required volume would be $0.48 \times 40 = 19.2\ \text{cm}^3$ of water. **(b)** If 23% less water is applied 0.77×40 cm (water needed) = 30.8 cm, which would be the depth of maximum water movement.

2.18. **(a)** $w = M_w/M_s = (136 - 118)/(118) = 0.15$. **(b)** From equation 2.29, $\theta = 0.15(1.32\ \text{g cm}^{-3}/1.0\ \text{g cm}^{-3}) = 0.20$.

2.19. **(a)** θ is the liquid-filled porosity (V_w). Assuming an ideal soil (which is not necessarily the case under field conditions, but which will serve for illustration), $V_s = 0.50$, and $V_s + V_w + V_a = V_t$; thus, $V_a = (1 - 0.50 + 0.39) = 0.11$. **(b)** Using the relation $V_a = (1 - \rho_b/\rho_s)$, then $\phi_a = (1 - 1.32\ \text{g cm}^{-3}/2.65\ \text{g cm}^{-3}) = 0.50$. Accounting for the ϕ_l portion in the pores, $V_a = (0.50 - 0.39) = 0.11$. **(c)** In this instance there is no difference. Differences that might occur could be due to several factors: ρ_s as given in the general relation as 2.65 g cm^{-3} is usually true only of mineral soils, and actual particle densities in other soil types are somewhat lower; spatial variability (because any value obtained would be an average for the area selected); and accuracy and dependability of laboratory techniques. **(d)** The void ratio is defined as the ratio between the volume of voids and the volume of solids, whereas porosity is defined as the volume of void space per bulk volume of soil.

2.20. Assuming a spherical particle, $s = s_v/\rho_s = ((6/d)/2.65\ \text{g cm}^{-3}) = (2.26\ \text{cm}^2/2 \times 10^{-6}\ \text{g}) = 1.13 \times 10^6\ \text{cm}^2\ \text{g}^{-1} = 113\ \text{m}^2\ \text{g}^{-1}$. Thus, total surface area for the sample = $(35\ \text{g})(113\ \text{m}^2\,\text{g}^{-1}) = 3.96 \times 10^3\ \text{m}^2$. If we assume a platelet shape, $s = s_v/\rho_s l$ (where l is the thickness of the platelet) = $[2/(10^{-10}\ \text{m})(2.65\ \text{g cm}^{-3})(10^6\ \text{cm}^3/\text{m}^3)] = 754.7\ \text{m}^2\ \text{g}^{-1}$. This gives a total surface area for the sample of $26.415 \times 10^3\ \text{m}^2$.

ADDITIONAL QUESTIONS

2.21. Consider a 1 m^3 sample of a soil that contains 38% pore space.

(a) What is the mass of water required to saturate it?
(b) What is the mass of air when the soil is completely dry?
(c) What is the mass of the solid phase?
(d) What is the mass of the soil when saturated?
(e) What would be the mass of soil if the volume fraction of air were 16%?

2.22. An intact soil core (30 cm diameter by 40 cm high) requires 11,500 cm^3 of water to saturate it. What is the porosity of the core?

2.23. To determine particle density (ρ_s), a scientist takes 88 g of granular (dry) soil at random and puts it into a flask > 100 cm^3 in volume. Next, the scientist adds deaerated water at 25 °C to drive out all air in the soil. The scientist then fills the bottle to exactly 100 cm^3. The mass of water and soil is 163 g.

(a) What is ρ_s?
(b) Why was it necessary to drive out all the air?

2.24. A 100 cm^3 soil sample has a wet mass of 183 g. After drying at 105 °C, its mass is 141 g.

(a) Find ρ_w and ρ_b.
(b) Find V_s, V_w, and V_g.
(c) Is the soil well aerated?
(d) Are the values obtained for ϕ true for every part of the soil sample?

2.25. Express w in terms of ρ_w and ρ_b.

2.26. Express θ in terms of ρ_w and ρ_b.

2.27. A soil scientist obtains a soil sample that has a $\rho_w = 1820$ kg/m^3 and $\rho_b = 1528$ kg/m^3. Calculate the following.

(a) The volume fraction of water (θ).
(b) The wetness (w).
(c) Air filled porosity (V_g).

2.28. Soil samples are collected with an extraction device that has a volume = 90 cm^3, mass = 120 g, mass + moist soil = 243 g, mass + oven-dry weight = 207 g, mass + soil after heating to 900 °C = 167 g. Calculate ρ_w, ρ_b, w, θ, V_{min}, V_{org}, V_s, V_g, V_T, and ρ_s. What is the significance of ρ_s after heating to 900 °C?

2.29. Calculate the specific surface area of a soil that has 80% fine sand (average $d = 0.1$ mm); 10% silt (average $d = 10$ μm); 10% illite clay (average $a = b = 200$ nm and $\delta = 5$ nm). (The density of the solid phase for sand and silt is 2660 kg/m^3, and for the illite clay is 2750 kg/m^3.)

2.30. Consider the situation in question 9.

(a) What is the contribution of each of the texture separates to the specific surface area of the soil?
(b) What is the texture class of this soil?
(c) To which component can most of the physical and chemical properties of the soil be attributed?

2.31. Assume that the average volumetric water content in the top meter of a soil is 0.17, a rainfall of 2.5 cm occurs, and none of the rain goes below 1 m in the soil.

(a) How much water (L) falls on an area 1 ha in size?
(b) How much water (L) is in the soil before the rain?
(c) How much water (L) is in the soil after the rain?
(c) What is θ after the rain?

2.32. Assume the same situation as in the previous problem, but with a 3 cm rain.

(a) What is d_w (cm) in the top meter of soil before the rain?
(b) What is d_w (cm) after the rain (in the top meter)?
(c) What is θ in the soil after the rain?

2.33. If a mass of soil is immersed in water, will it become completely saturated?