1. SOIL PHYSICS AND SOIL PHYSICAL CHARACTERISTICS

SOIL SCIENCE

To begin, we define *soil* as the weathered and fragmented outer layer of the earth's terrestrial surface. It is formed initially through disintegration, decomposition, and recomposition of mineral material contained in exposed rocks by physical, chemical, and biological processes. The material thus modified is further conditioned by the activity and accumulated residues of numerous microscopic and macroscopic organisms (plants and animals). In a series of processes that may require hundreds or even thousands of years and that is called *soil genesis*, the loose debris of rock fragments is transmuted into a more or less stable, internally ordered, actively functioning natural body. Ultimately, this culminates in the formation of a characteristic *soil profile*, which resembles a layer cake. We can visualize the soil profile as a composite living body, in the same way that we think of the human body as a distinct organism, even though in reality it is an ensemble of numerous interdependent and symbiotically coordinated groups of organelles, cells, organs, and colonies of myriad organisms.

Soil science is the study of the soil in all its ramified manifestations and facets: as a central link in the biosphere, as a medium for the production of agricultural commodities, and as a raw material for industry and construction. As such, it shares interests with geology, sedimentology, terrestrial ecology, and geobotany as well as with such applied sciences as agronomy and engineering. Because of its varied interests and concerns, soil science itself is commonly divided into several subdivisions, including pedology (soil formation and classification), soil chemistry, soil mineralogy, soil biology, soil fertility, and soil mechanics. However, such distinctions are often arbitrary, because in fact all of the environmental sciences are inextricably interconnected.

SOIL PHYSICS

Soil physics is one of the major subdivisions of soil science. It seeks to define, measure, and predict the physical properties and behavior of the soil, both in its natural state and under the influence of human activity. As physics deals in general with the forms and interactions of matter and energy, so soil physics deals specifically with the state and movement of matter and with the fluxes and transformations of energy in the soil. On the one hand, the fundamental study of soil physics aims at understanding the mechanisms governing such processes as terrestrial energy exchange, the cycles of water and of transportable materials, and the growth of plants in the field. On the other hand, the practical application of soil physics aims at the proper management of the soil by means of cultivation, irrigation, drainage, aeration, improvement of soil structure, control of infiltration and evaporation, regulation of soil temperature, and prevention of erosion.

Soil physics is thus both a basic and an applied science, with a very wide range of interests. The study of soil science in general and of soil physics in particular is driven not only by the innate curiosity that is our species' main creative impulse, but also by urgent necessity. The intensifying pressure of population and development has diminished the soil resources of our small planet and has led to their unsustainable use and degradation in too many parts of the world.

Since the soil is not an isolated medium but is in constant dynamic interaction with the larger environment, soil physics is an aspect of the more encompassing field of environmental physics (sometimes called biospheric physics) and of the overall science of geophysics.

The early soil physicists were interested primarily in the engineering and the agricultural aspects of their discipline, hence their research focused on the soil as a material for construction or as a medium for the production of crops. Recent decades have witnessed an increasing emphasis on the environmental aspects and applications of soil physics. Consequently, research in soil physics has expanded its scope to include phenomena related to natural ecosystems and to processes affecting the quality of the environment. Processes occurring in the soil are now seen to affect the entire terrestrial environment, including local and regional climates, the natural food chain, biodiversity, and the fate of the voluminous waste products of our civilization (among which are many pathogenic and toxic agents).

Increasingly, the main concern of soil physics has shifted from the laboratory to the field and from a restricted one-dimensional view to an expansive three-dimensional view interfacing with the domains of sister disciplines such as meteorology and climatology, hydrology, ecology, and geochemistry. The larger domain of soil physics now encompasses greater complexity and variability in space and time, the treatment of which requires reliance on stochastic as well as deterministic methods. Consequently, the science is becoming ever more interesting and relevant.

The task of soil physics is made difficult by the enormous and baffling intricacy of a medium containing myriad mineral and organic components, all irregularly fragmented and variously associated in a geometric pattern that is so complex and labile as to challenge our imagination and descriptive powers. Some of the solid material consists of crystalline particles, while some is made up of amorphous gels that may coat the crystals and modify their behavior. The solid phase in the soil interacts with the fluids, water, and air that permeate soil pores. The entire soil is hardly ever in equilibrium as it alternately wets and dries, swells and shrinks, disperses and flocculates, hardens and softens, warms and cools, freezes and thaws, compacts and cracks, absorbs and emits gases, adsorbs and releases exchangeable ions, dissolves and precipitates salts, becomes acidic or alkaline, and exhibits aerobic or anaerobic conditions leading to chemical oxidation or reduction.

THE SOIL PROFILE

The most obvious part of any soil is its surface zone. Through it, matter and energy are transported between the soil and the atmosphere. The surface may be smooth or pitted, granular or crusted and cracked, friable or hard, level or sloping, vegetated or fallow, mulched or exposed. Such conditions affect the processes of radiation and heat exchange, water and solute movement, and gaseous diffusion.

Important though the surface is, however, it does not necessarily portray the character of the soil as a whole. To describe the latter, we must examine the soil in depth. We can do this by digging a trench and sectioning the soil from the surface downward, thus revealing what is commonly termed the *soil profile*.

The soil profile typically consists of a succession of more-or-less distinct strata. These may result from the pattern of deposition, or sedimentation, as can be observed in wind-deposited (*aeolian*) soils and particularly in water-deposited (*alluvial*) soils. If, however, the strata form in place by internal soil-forming (*pedogenic*) processes, they are called *horizons*.

The top layer, or *A horizon*, is the zone of major biological activity and is therefore generally enriched with organic matter and darker in color than the underlying soil. Here, plants and animals and their residues interact with an enormously diverse and labile multitude of microorganisms, such as bacteria, protozoa, and fungi, millions of which can be found in a mere handful of topsoil. In addition, there are usually varied forms of macroorganisms (including earthworms, arthropods, and rodents) that burrow in the soil. The A horizon is generally the most fertile zone of the soil, but it is also the zone most vulnerable to erosion by water and wind (especially if it is denuded of vegetative cover or its protective residues).

Underneath the A horizon is the *B horizon*, where some of the materials (e.g., clay or carbonates) that are leached from the A horizon by percolating water tend to accumulate. The B horizon is often thicker than the A horizon. The pressure of the overlying soil tends to reduce the porosity of the deeper layers. In some cases, an overly dense or indurated B horizon may inhibit gas exchange, water drainage, and root penetration.

Underlying the B horizon is the C *horizon*, which is the soil's parent material. In a soil formed of bedrock in situ (called a *residual soil*), the C horizon consists of the weathered and fragmented rock material. In other cases, the C horizon may consist of alluvial, aeolian, or glacial deposits.

The character of the profile depends primarily on the climate that prevailed during the process of soil formation. It also depends on the parent material, the vegetation, the topography, and time. These five variables are known as the *factors of soil formation* (Fanning and Fanning, 1989). Mature soils are such that have been subjected to those factors for a sufficient length of time so that full profile development has taken place. The A, B, C sequence of horizons is clearly recognizable in some cases, as, for example, in a typical zonal soil (i.e., a soil associated with a distinct climatic zone), such as a *podzol* (also known as a *spodosol*). In other cases, no clearly developed B horizon is discernible, and the soil is then characterized by an A,C profile. In a recent alluvium, hardly any profile differentiation is apparent.

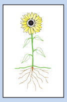
The typical development of a soil and its profile, called *pedogenesis* (Buol et al., 2003), can be summarized: It begins with the physical disintegration of an exposed rock formation, which provides the soil's parent material. Gradually, the loosened material is colonized by living organisms. The consequent accumulation of organic residues at and below the surface brings about the development of a discernible A horizon. That horizon may acquire an aggregated structure, stabilized to some degree by gluelike components of the organic matter complex (known as *humus*) resulting from the decomposition of plant and animals residues. Continued weathering (decomposition and recomposition) of minerals may bring about the formation of clay. Some of the clay thus formed tends to migrate downward, along with other transportable materials (such as soluble salts), and to accumulate in an intermediate zone (the B horizon) between the surface zone of major biological activity and the deeper parent material of the so-called C horizon.

Important aspects of soil formation and profile development are the twin processes of *eluviation* and *illuviation* ("washing out" and "washing in," respectively), wherein clay and other substances emigrate from the overlying eluvial A horizon and accumulate in the underlying illuvial B horizon. The two horizons come to differ substantially in composition and structure.

Throughout these processes, the profile as a whole deepens as the upper part of the C horizon is gradually transformed, until eventually a quasi-stable condition is approached in which the counterprocesses of soil formation and soil erosion are more or less in balance. In the natural state, the A horizon may have a thickness of 0.1–0.5 meters. When stripped of vegetative cover and pulverized or compacted by tillage or traffic, this horizon may lose half or more of its original thickness within a few decades.

In arid regions, salts such as calcium sulfate and calcium carbonate, dissolved from the upper part of the soil, may precipitate at some depth to form a cemented *pan* (sometimes called *caliche*, from the Spanish word for "lime"). Erosion of the A horizon may bring the B horizon to the surface. In extreme cases, both the A and B horizons may be scoured off by natural or human-induced erosion so that the C horizon becomes exposed and a new cycle of soil formation may then begin. In other cases, a mature soil may be covered by a new layer of sediments (alluvial or aeolian) so that a new soil may form over a "buried" old soil. Where episodes of deposition occur repeatedly over a long period of time, a sequence of soils may be formed in succession, thus recording the pedological history — called the *paleopedology* — of the region (including evidence of the climate and vegetation that had prevailed at the time of each profile's formation).

Numerous variations of the processes described are possible, depending on local conditions. The characteristic depth of the soil, for instance, varies from location to location. Valley soils are typically deeper than hillslope soils, and the depth of the latter depends on slope steepness. In places, the depth of the profile can hardly be ascertained, because the soil blends into its parent material without any distinct boundary. However, the zone of



BOX 1.1 Soil Physics and the Environment

As human populations have grown and living standards have risen, the requirements for agricultural products have increased enormously. More land has been brought under cultivation, including marginal land that is particularly vulnerable to degradation by such processes as erosion, depletion of organic matter and nutrients, pollution, waterlogging, and salination. Other forms of land use — towns, roads, factories, airports, feedlots, waste disposal sites, and recreational facilities — usurp ever more land.

Consequently, the domain remaining for natural ecosystems has shrunk and been divided into smaller enclaves, to the detriment of numerous species. The mutual checks and balances that have long sustained the rich diversity of life on earth are now threatened by the human appetite for resources and the wanton way they are used and their waste products discarded. The task therefore is to supply human needs in ways that are sustainable locally and do not damage the larger environment.

Two alternative approaches have been proposed to prevent further destruction of the remaining natural ecosystems and to relieve pressure on fragile marginal lands. One way is to restrict human activities to choice areas, where production can be intensified. This calls for optimizing all production factors so as to achieve maximum efficiency in the utilization of soil, water, energy, and other necessary inputs (e.g., nutrients and pest control measures). The problems are that soil processes are difficult to control, and, because the soil is an open system in constant interaction with its surroundings, the time-delayed and space-removed consequences of soil processes are difficult to predict. Full control of agricultural production can ultimately be achieved only in enclosed spaces such as greenhouses or in confined fields.

Another approach is to devise more naturalistic modes of production that are compatible with the environment and do not require the drastic isolation of production from neighboring ecosystems. This agro-ecosystem approach is exemplified by the trends toward organic farming and agroforestry.

Either way, the physical attributes and processes of the soil are of prime importance. Physical factors strongly affect whether the soil is to be cool or warm, anaerobic or aerobic, wet or dry, compact or highly porous, hard or friable, dispersed or aggregated, impervious or permeable, eroded or conserved, saline or salt free, leached or nutrient rich. All these, in turn, determine whether the soil can be a favorable or unfavorable medium for various types of plants and other living organisms as well as for alternative modes of production; in short, whether the soil can be managed productively and sustainably while neutralizing — rather than transmitting — environmental pollutants.

biological activity seldom extends below 2–3 meters and in many cases is shallower than 1 meter.

SOILS OF DIFFERENT REGIONS

Each climatic zone exhibits a characteristic group of soils. In the humid tropics, there is a tendency to dissolve and leach away the silica and to accumulate iron and aluminum oxides. As a result, the soils are typically colored red, the hue of oxidized iron. Chunks or blocks excavated from such soils and dried in the sun may harden to form bricks; hence these soils are called *laterites*, from the Latin word *later*, meaning "brick." On the other hand, soils of the humid cool regions often exhibit an A horizon consisting of a thin surface layer darkened by organic matter and underlain by a bleached, ashlike layer; in turn, this overlies a clay-enriched B horizon. These soil were called *podzols* by the early Russian pedologists (from the Russian words *pod* = "ground" and *zola* = "ash") and are known as *spodosols* in the current American system. Soils that are poorly drained ("waterlogged") may exhibit conditions of chemical reduction in the profile, indicated by streaks of discoloration ("mottling").

In contrast with the soils of humid regions, from which nearly all readily soluble salts have been leached, the soils of arid regions tend to precipitate the moderately soluble salts of calcium and magnesium. (Especially prevalent are accumulations of calcium carbonate [lime] and calcium sulfate [gypsum].) Under certain conditions, arid-zone soils may even accumulate the more readily soluble salts of sodium (NaCl and Na_2CO_3) and of potassium. Such soils are prone to excessive salinity, in extreme cases of which they become practically sterile. Irrigated soils in poorly drained river valleys of arid regions are particularly liable to undergo the process of salination. When such soils are leached of excess salts, they must be treated with soil amendments (such as gypsum) to replace the exchangeable sodium ions with calcium, lest the sodium ion cause dispersion of the clay and the breakdown of soil structure.

An outstanding soil formed naturally in some intermediate semihumid to semiarid regions (e.g., Ukraine, Argentina, and the prairie states of North America — the so-called "corn belt" of the United States) is the soil classically called *chernozem* (Russian for "black earth") with its unusually thick, humus-rich, fertile A horizon.

Whereas Russian pedologists were the first to develop a universal classification system of soils over a century ago, other schools of pedology have since offered alternative taxonomies, claimed to be more detailed and comprehensive. Notable among these is the systems offered by the U.N. Food and Agriculture Organization and the one developed by the U.S. Soil Survey. Each system recognizes hundreds of soil types and their variants.

By way of illustration, we present a hypothetical soil profile in Fig. 1.1. This is not a "typical" soil, for among the myriad of differing soil types recognized by pedologists no single type can be considered typical. Our figure is only meant to suggest the sort of contrasts in appearance and structure among different horizons that may be encountered in a soil profile. Pedologists classify soils by their mode of formation (genesis) and recognizable properties (see Fig. 1.2).

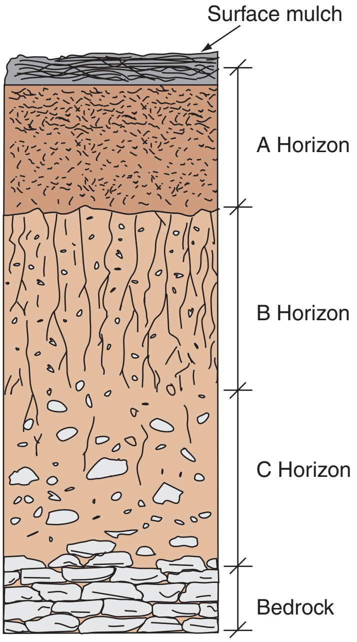
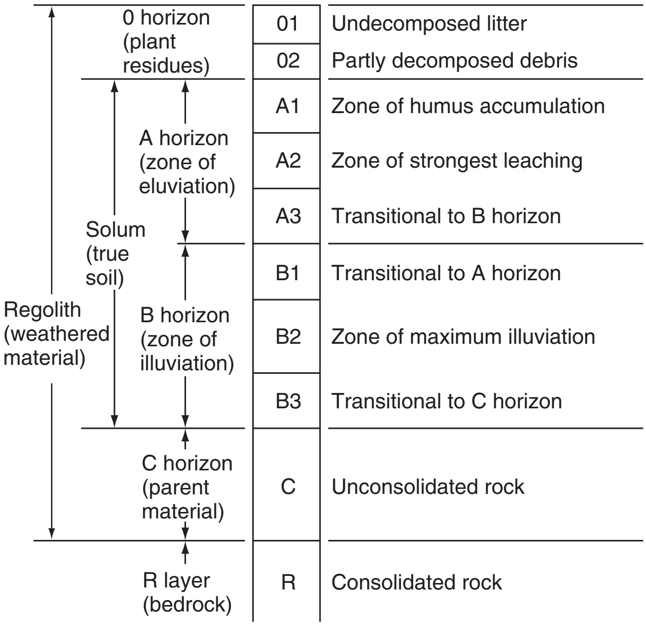


Fig. 1.1. Schematic representation of a hypothetical soil profile. The A horizon is shown with an aggregated crumblike structure, the B horizon with columnar structure, and the C horizon with incompletely weathered rock fragments.



Regolith (weathered material)		0 horizon (plant residues)	01	Undecomposed litter
			02	Partly decomposed debris
	Solum (true soil)	A horizon (zone of eluviation)	A1	Zone of humus accumulation
			A2	Zone of strongest leaching
			A3	Transitional to B horizon
		B horizon (zone of illuviation)	B1	Transitional to A horizon
			B2	Zone of maximum illuviation
			B3	Transitional to C horizon
		C horizon (parent material)	C	Unconsolidated rock
		R layer (bedrock)	R	Consolidated rock

Fig. 1.2. Descriptive terminology for soil profile horizones.

SOIL AS A DISPERSE THREE-PHASE SYSTEM

The term *system* refers to a group of interacting, interrelated, or interdependent elements constituting an integrated entity. Since all of nature is in fact an integrated entity, defining any domain within nature to be a "system" is an admittedly arbitrary exercise. We choose to do so for reasons of convenience, since our own limitations prevent us from dealing with the entire complexity of nature all at once. However, the part of nature on which we may wish to focus our attention at any moment is necessarily a subsystem inside a larger system, with which our system interacts continuously. The modes of interaction generally include transfers or exchanges of materials and of energy.

Systems may vary widely in size, shape, and degree of complexity. They may consist of one or more substances and of one or more phases. The simplest system is one that is comprised of a single substance that has the same physical properties throughout. An example of such a system is a body of water consisting entirely of uniform ice. Such a system is homogeneous. A system comprised of a single chemical compound can also be heterogeneous if that substance exhibits different properties in different regions of the system. A region that is internally uniform physically is called a *phase*. A mixture of ice and liquid water, for instance, is chemically uniform but physically heterogeneous, because it includes two phases. The three ordinary phases in nature are the solid, liquid, and gaseous phases.

A system containing several substances may also be monophasic. A solution of salt and water, for example, is a homogeneous liquid. A system of several substances can also be heterogeneous. In a heterogeneous system the properties may differ not only between one phase and another, but also between the internal parts of each phase and the interfaces over which the phase comes into contact with one another. Interfaces exhibit specific phenomena resulting from the interaction of the phases. The importance of such phenomena (including adsorption, surface tension, and friction) in determining the behavior of the system as a whole depends on the magnitude of the interfacial area per unit volume of the system.

Systems in which at least one of the phases is subdivided into numerous minute particles, which together present a very large interfacial area per unit volume, are commonly called *disperse systems*. Colloidal sols (including aerosols), gels, emulsions, and aerosols are examples of disperse systems.

The soil is a heterogeneous, polyphasic, particulate, disperse, and porous system, with a large interfacial area per unit volume. (A handful of clay, for instance, may have an internal surface area of several hectares, a hectare being equal to 2.5 acres!). The disperse nature of soil and its consequent interfacial activity give rise to such phenomena as adsorption of water and chemicals, capillarity, ion exchange, swelling and shrinking, dispersion and flocculation.

The three phases of ordinary nature are represented in the soil as follows: the solid phase forms the *soil matrix*; the liquid phase is the water in the soil, which always contains dissolved substances, so it should properly be called the *soil solution*; and the gaseous phase is the *soil atmosphere*. The solid matrix of the soil consists of particles that vary in chemical and mineralogical composition as well as in size, shape, and orientation. It also contains amorphous substances, particularly organic matter, which is attached to the mineral grains

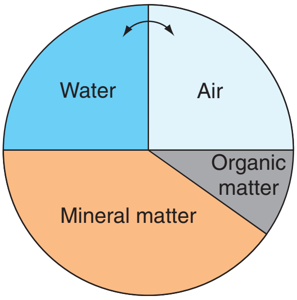


Fig. 1.3. Schematic composition (by volume) of a medium-textured soil at a condition considered optimal for plant growth. Note that the solid matter constitutes 50% and the pore space 50% of the soil volume, with the latter divided equally between water and air. the arrows indicate that water and air are related so that an increase in one is associated with a decrease in the other.

and may bind them together in assemblages called *aggregates*. The organization of the solid components of the soil determines the geometric characteristics of the pore spaces in which water and air are transmitted and retained. Finally, soil water and soil air vary in composition, both in time and in space.

The relative proportions of the three phases in the soil are not fixed but vary continuously, depending on such variables as weather, vegetation, and management. Figure 1.3 presents the hypothetical volume composition of a medium-textured soil at a condition considered approximately optimal for plant growth.

BOX 1.2 The Concept of Representative Elementary Volume

Some soil properties (for example, temperature) can be measured at a point, whereas other properties are volume dependent. Suppose we wish to measure some volume-dependent soil property, such as porosity. If our sample is very small, say, the size of a single particle or pore, the measured porosity may vary between zero and 100 percent, depending on the point at which we make our measurement (whether at a particle or at a pore). If we measure the porosity repeatedly at several adjacent points, the results will fluctuate widely. However, if we increase the scale or volume of the each sample so as to encompass within it both particles and pores, the fluctuations among repeated measurements at adjacent locations will diminish. If we keep enlarging the sample progressively, we will eventually obtain a consistent measurement of the soil's average porosity. The minimal volume of sample needed to obtain a consistent value of a measured parameter has been called the *representative elementary volume* (REV) (Bear, 1969). Obviously, the REV becomes larger in soils that are strongly aggregated (as well as in soils that are cracked or otherwise heterogeneous) than in more uniform soils.

The problem with the REV concept is that different parameters may exhibit different spatial or temporal patterns, so the REV for one parameter or property may differ from those for other parameters. That is to say, each property may have its own characteristic scale. Even more serious may be the failure of the REV concept in the case of "structured" fields, i.e., in fields that vary systematically in one direction or another. In such fields, increasing the size of the sample measured may not produce a consistent value at all.

VOLUME AND MASS RELATIONSHIPS OF SOIL CONSTITUENTS

Let us consider the volume and mass relationships among the three phases of the soil, and define some basic parameters that can help to characterize the soil physically. Figure 1.4 is a schematic depiction of a hypothetical soil in which the three phases have been separated and stacked one atop the other for the purpose of showing their relative volumes and masses.

In the figure, the masses of the phases are indicated on the right-hand side: the mass of air M_a, which is negligible compared to the masses of solids and water; the mass of water M_w; the mass of solids M_s; and the total mass M_t. (These masses can also be represented in terms of their weights, being the product of each mass and gravitational acceleration.) The volumes of the same components are indicated on the left-hand side of the diagram: volume of air V_a, volume of water V_w, volume of pores $V_f = V_a + V_w$, volume of solids V_s, and the total volume of the representative sample V_t.

On the basis of this diagram, we can now define terms that are generally used to express the quantitative interrelations of the three primary soil phases.

Density of Solids (Mean Particle Density) ρ_s

$$\rho_s = M_s / V_s \tag{1.1}$$

In most mineral soils, the mean mass per unit volume of solids is about 2600–2700 kg/m^3. This is close to the density of quartz, which is generally the most prevalent mineral in the coarsest fraction of the soil. Some of the minerals composing the finest fraction of the soil have a similar density. However, the presence of iron oxides and of various other “heavy” minerals (generally defined as those having a density exceeding 2900 kg/m^3) increases the average value of ρ_s, whereas the presence of low-density organic matter generally

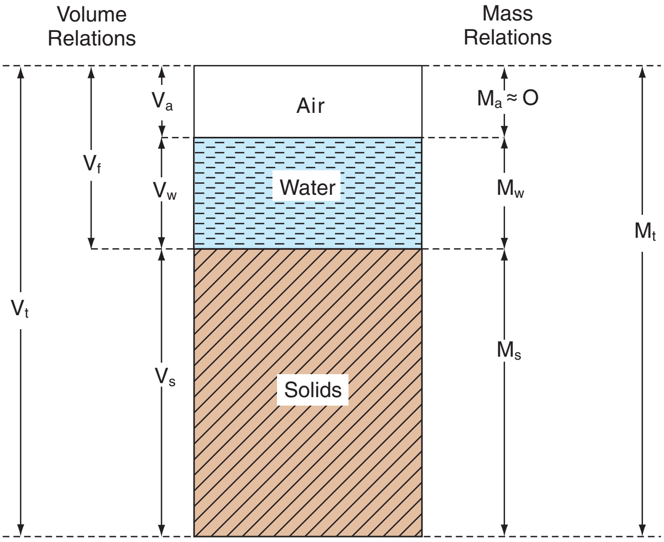


Fig. 1.4. Schematic diagram of the soil as a three-phase system.

lowers the mean density of the solid phase. Sometimes the density is expressed in terms of the *specific gravity*, σ_g, which is the ratio of the density of any material to that of water at 4°C and at atmospheric pressure. The latter density is about 1000 kg/m^3, so the specific gravity of the solid phase in a typical mineral soil is about 2.65, a value that is numerically (though not dimensionally) equal to the density expressed in the cgs. system of units (g/cm^3).

Dry Bulk Density ρ_b

$$\rho_b = M_s / V_t = M_s / (V_s + V_a + V_w) \qquad (1.2)$$

The dry bulk density expresses the ratio of the mass of solids to the total soil volume (solids and pores together). Obviously, ρ_b is always smaller than ρ_s. If the pores constitute half the volume, then ρ_b is half of ρ_s, namely, about 1300–1350 kg/m^3, equivalent to a bulk specific gravity (i.e., the ratio of the soil's bulk density to the density of water at standard conditions) of 1.3–1.35. The bulk specific gravity of sandy soils with a relatively low volume of pores may be as high as 1.6, whereas that of aggregated loams and clay soils may be below 1.2. Whereas the mean particle density is typically constant, the bulk density is highly labile. It is affected by the structure of the soil, that is, its looseness or degree of compaction, as well as by its swelling and shrinkage characteristics. The latter depend both on clay content and water content. Even in extremely compacted soils, however, the bulk density remains appreciably lower than the density of the solid matter, since the particles can never interlock perfectly. However much the pore space can be reduced by compaction, it can never be eliminated.

Total (Wet) Bulk Density ρ_t

$$\rho_t = M_t / V_t = (M_s + M_w)/(V_s + V_a + V_w) \qquad (1.3)$$

This is an expression of the total mass of a moist soil per unit volume. As such, this parameter depends more strongly than does the dry bulk density on soil wetness or water content.

Dry Specific Volume v_b

$$v_b = V_t / M_s = 1/\rho_b \qquad (1.4)$$

The volume of a unit mass of a dry soil (the reciprocal of the dry bulk density) serves as another useful index of the degree of looseness or compaction of a soil body.

Porosity f

$$f = V_f / V_t = (V_a + V_w)/(V_s + V_a + V_w) \qquad (1.5)$$

Porosity is an index of the relative pore space in a soil. Its value generally ranges from 0.3 to 0.6 (30–60%). Coarse-textured soils tend to be less porous than fine-textured soils, though the mean size of individual pores is greater in the former. In clayey soils, the porosity is highly variable because the soil alternately swells, shrinks, aggregates, disperses, compacts, and cracks. As generally defined, the term *porosity* refers to the volume fraction of pores, and this value should be equal, on average, to the areal porosity (the fraction of pores in a representative cross-sectional area) as well as to the average lineal porosity (the fractional length of pores along a straight line passing through the soil in any direction). However, the total porosity reveals nothing about the sizes or shapes of the various pores in the soil.

Void Ratio *e*

$$e = V_f/V_s = (V_a + V_w)/(V_t - V_f) \tag{1.6}$$

The void ratio is also an index of the fractional pore space, but it relates that space to the volume of solids rather than to the total volume of the soil. As such, it ranges between 0.3 and 2. The advantage of this index over the preceding one is that in the case of e any change in pore volume affects only the numerator of the defining equation, whereas in the case of f such a change affects both the numerator and the denominator. Void ratio is the index generally preferred by soil engineers, while porosity is more frequently used by agronomists.

Sample Problem

Prove the following relation of porosity to particle density and bulk density:

$$f = (\rho_s - \rho_b)/\rho_s = 1 - \rho_b/\rho_s$$

Substituting the definitions of f, ρ_s, and ρ_b, we can rewrite the equation as

$$V_f/V_t = 1 - (M_s/V_t)/(M_s/V_s)$$

Simplifying the right-hand side, we obtain

$$V_f/V_t = 1 - (V_s/V_t) = (V_t - V_s)/V_t$$

But since $V_t - V_s = V_f$, we have

$$V_f/V_t = V_f/V_t$$

Soil Wetness (Water Content)

The water content of a soil can be expressed in various ways: relative to the mass of solids, or to the total mass, or to the volume of solids, or to the total volume, or to the volume of pores. The various indexes are defined as follows.

Mass Wetness *w*

$$w = M_w/M_s \tag{1.7}$$

This is the mass of water relative to the mass of dry soil particles. The standard definition of *dry soil* refers to a mass of soil dried to equilibrium (in practice, over a 24-hour period) in an oven at 105°C, though a clay soil may still contain an appreciable amount of water at that state. Mass wetness is sometimes expressed as a decimal fraction but more often as a percentage. A sample of soil dried in “ordinary” air at ambient temperature (rather than in an oven) will generally retain several percent more water than if dried in the oven. Similarly, an oven-dry soil exposed to “ordinary” air will gradually gain appreciable moisture. This phenomenon results from the tendency of the soil’s clay fraction to adsorb moisture from the air, a property known as *hygroscopicity*. The amount thus adsorbed depends on the type and content of clay in the soil as well as on the humidity of the ambient atmosphere. The water content at saturation (when all pores are filled with water) is also higher in clayey than in sandy soils. In different soils, w can range between 25% and 60%, depending on bulk density. In the special case of organic soils, such as peat or muck soils, the saturation water content on the mass basis may exceed 100%.

Volume Wetness θ

$$\theta = V_w/V_t = V_w/(V_s + V_f) \qquad (1.8)$$

The volume wetness (often termed *volumetric water content*) is generally computed as a percentage of the total soil volume. At saturation, therefore, it is equal to the porosity. In sandy soils, θ at saturation is on the order of 40%; in medium-textured soils it is approximately 50%; and in clayey soils it can approach 60%. In the last, in fact, the volume of water at saturation can exceed the porosity of the dry soil, since clayey soils swell upon wetting. The use of rather than θ to express water content is often more convenient because it is more directly applicable to the computation of fluxes and water volumes added to soil by rain or irrigation and to quantities extracted from the soil by evaporation and transpiration. The volume ratio is also equivalent to the depth ratio of soil water, that is, the depth of water per unit depth of soil.

Sample Problem

Prove the following relation between volume wetness, mass wetness, bulk density, and water density ($\rho_w = M_w/V_w$):

$$\theta = w\rho_b/\rho_w$$

Again, we start by substituting the respective definitions of w_v, w_m, ρ_b, and ρ_w:

$$V_w/V_t = [(M_w/M_s)(M_s/V_t)]/(M_w/V_w)$$

Rearranging the right-hand side, we get

$$V_w/V_t = (V_w M_w M_s)/(M_w M_s V_t) = V_w/V_t$$

Water Volume Ratio v_w

$$v_w = V_w/V_s \qquad (1.9)$$

For swelling soils, in which porosity changes markedly with wetness, it may be preferable to refer the volume of water present in a sample to the invariant volume of particles rather than to total volume. At saturation, v_w is equal to the void ratio e.

Degree of Saturation *s*

$$s = V_w/V_f = V_w/(V_a + V_w) \tag{1.10}$$

This index expresses the water volume present in the soil relative to the pore volume. Index s ranges from zero in a completely dry soil to unity (100%) in a saturated soil. Complete saturation, however, is hardly ever attainable in field conditions, since some air is nearly always present. In a relatively dry soil the air phase occupies a continuous space, whereas in a very wet soil air may be occluded or encapsulated in the form of discontinuous bubbles.

Air-Filled Porosity (Fractional Air Content) f_a

$$f_a = V_a/V_t = V_a/(V_s + V_a + V_w) \tag{1.11}$$

This is a measure of the relative content of air in the soil and as such is an important criterion of soil aeration. It is related negatively to the degree of saturation s (i.e., $f_a = f - s$). The relative volume of air in the soil may also be expressed as a fraction, a, of the pore volume (rather than of the total soil volume). Thus,

$$a = V_a/V_f = V_a/(V_a + V_w) \tag{1.12}$$

Sample Problem

A sample of moist soil with a wet mass of 1.0 kg and a volume of 0.64 liters (6.4×10^{-4} m^3) was dried in the oven and found to have a dry mass of 0.8 kg. Assuming the typical value of particle density for a mineral soil (2650 kg/m^3), calculate the bulk density ρ_b, porosity f, void ration e, mass wetness w_m, volume wetness θ, water volume ratio v_w, degree of saturation s, and air-filled porosity f_a.

Bulk density: $\rho_b = M_s/V_t = 0.8 \text{ kg}/6.4 \times 10^{-4} \text{ m}^3 = 1250 \text{ kg/m}^3$

Porosity: $f = 1 - \rho_b/\rho_s = 1 - 1250/2650 = 1 - 0.472 = 0.528$

Alternatively, $f = V_f/V_t = (V_t - V_s)/V_t$

and since $V_s = M_s/\rho_s = 0.8 \text{ kg}/2650 \text{ kg/m}^3 = 3.02 \times 10^{-4} \text{ m}^3$

hence $f = (6.4 - 3.02) \times 10^{-4} \text{ m}^3/6.4 \times 10^{-4} \text{ m}^3 = 0.528 = 52.8\%$

Void ratio: $e = V_f/V_s = (V_t - V_s)/V_s = (6.4 - 3.02) \times 10^{-4} \text{ m}^3/3.02 \times 10^{-4} \text{ m}^3 = 1.12$

Mass wetness: $w = M_w/M_s = (M_t - M_s)/M_s = (1.0 - 0.8) \text{ kg}/0.8 \text{ kg} = 0.25 = 25\%$

Volume wetness: $\theta = V_w/V_t = 2.0 \times 10^{-4} \text{ m}^3/6.4 \times 10^{-4} \text{ m}^3 = 0.3125 = 31.25\%$

(*Note*: $V_w = M_w/\rho_w$, wherein $\rho_w \approx 1000 \text{ kg/m}^3$)

Alternatively, $\theta = w\rho_b/\rho_w = 0.25(1250 \text{ kg/m}^3/1000 \text{ kg/m}^3) = 0.3125$

Water volume ratio: $v_w = V_w/V_s = 2.0 \times 10^{-4} \text{ m}^3/3.02 \times 10^{-4} \text{ m}^3 = 0.662$

Degree of saturation: $s = V_w/(V_t - V_s) = 2.0 \times 10^{-4} \text{ m}^3/(6.4 - 3.02) \times 10^{-4} \text{ m}^3 = 0.592$

Air-filled porosity: $f_a = V_a/V_t = (6.4 - 2.0 - 3.02) \times 10^{-4} \text{ m}^3/6.4 \times 10^{-4} \text{ m}^3 = 0.216$

Sample Problem

What is the equivalent depth of water contained in a soil profile 1 m deep if the mass wetness of the upper 0.4 m is 15% and that of the lower 0.6 m is 25%? Assume a bulk density of 1200 kg/m^3 in the upper layer and 1400 in the lower layer. How much water does the soil contain, in cubic meters per hectare of land?

Recall that $\theta = w(\rho_b/\rho_w)$, where $\rho_w = 1000$ kg/m^3

Volume wetness in the upper layer: $\theta_1 = 0.15(1200/1000) = 0.18$

Equivalent depth in upper 0.4 m $= 0.18 \times 0.4$ m $= 0.072$ m $= 72$ mm

Volume wetness in lower layer: $\theta_2 = 0.25(1400/1000) = 0.35$

Equivalent depth in lower 0.6 m $= 0.35 \times 0.6$ m $= 0.21$ m $= 210$ mm

Total depth of water in 1-m profile $= 0.072$ m $+ 0.210$ m $= 0.282$ m

Volume of water contained in 1-m profile per hectare
(1 ha $= 10^4$ m^2) $= 0.282$ m $\times$ 1000 m^2 $= 2820$ m^3