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Introduction to Soil Science

(oktatási segédlet)

This educational guide is prepared for foreign students studying through the ERASMUS scholarship partnership at the Faculty of Forestry at the University of Sopron.

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1. The Soil as Part of Our Ecosystem

"Soils are crucial to life on earth. To a great degree, the quality of the soil determines the nature of plant ecosystems and the capacity of land to support animal life and society. As human societies become increasingly urbanized, fewer people have intimate contact with the soil, and individuals tend to lose sight of the many ways in which they depend upon soils for their prosperity and survival. Indeed, the degree to which we are dependent on soils is likely to increase, not decrease, in the future.

Soils will continue to supply us with nearly all of our food (except for what can be harvested from the oceans). How many of us remember, as we eat a slice of pizza, that the pizza's crust began in a field of wheat and its cheese began with grass, clover and corn rooted in the soils of a dairy farm? Most of the fiber we use for lumber, paper and clothing has its roots in the soils of forests and farmland. Although we sometimes use plastics and fiber synthesized from fossil fuel as substitutes, in the long term we will continue to depend on terrestrial ecosystems for these needs.

In recent times, biomass grown on soils is likely to become an increasingly important feedstock for fuels and manufacturing as the world's finite supplies of petroleum are depleted during the course of this century.

One of the stark realities of the 21st century is that the human population that demands all of these products, will increase by several billion, while the amount of soil available to provide them will not increase. In fact, this resource base is even *decreasing* because of soil degradation and urbanization. Thus, to survive as a species, we will have to greatly improve the efficiency and sustainability with which we manage our soil resources.

The art of soil management is as old as civilization. As we meet the challenges of this century, new understanding and new technologies will be needed to protect the environment and, at the same time, produce food and biomass to support society. The study of soil science has never been more important for foresters, natural resource managers, farmers, engineers and ecologists alike.

1.1 Functions of Soils in Our Ecosystems

In any ecosystem, whether your backyard, a farm, a forest, or a regional watershed, soils have **five key roles** to play.

First, soil supports the growth of higher plants, mainly by providing a medium for plant roots and supplying nutrient elements that are essential to the entire plant. Properties of the soil often determine the nature of the vegetation present and, indirectly, the number and types of animals (including people) that the vegetation can support.

Second, soil properties are the principal factor controlling the fate of water in the hydrologic system. Water loss, utilization, contamination, and purification are all affected by the soil.

Third, the soil functions as nature's recycling system. Within the soil, waste products and dead bodies of plants, animals, and people are assimilated, and their basic elements are made available for reuse by the next generation of life.

Fourth, soils provide habitats for a myriad of living organisms, from small mammals and reptiles to tiny insects to microscopic cells of unimaginable numbers and diversity.

Fifth, in human-built ecosystems, soil plays an important role as an engineering medium. Soil is not only an important building material in the form of earth fill and bricks (baked soil material), but provides the foundation for virtually every road, airport, and house we build.

1.1.1 Medium for Plant Growth

The soil provides several things for a growing plant. First, the soil mass provides **physical support**, anchoring the root system so that the plant does not fall over.

Plant roots depend on the process of respiration to obtain energy. Since root respiration, like our own respiration, produces carbon dioxide (CO_2) and uses oxygen (O_2), an important function of the soil is **ventilation** – allowing CO_2 to

escape and fresh O_2 to enter the root zone. This ventilation is accomplished via the network of soil pores.

An equally important function of the soil pores is to absorb water and hold it where it can be used by plant roots. As long as plant leaves are exposed to sunlight, the plant requires a continuous stream of water to use in cooling, nutrient transport, turgor maintenance, and photosynthesis. Since plants use water continuously, but in most places it rains only occasionally, the **water-holding capacity** of soils is essential for plant survival.

As well as moderating moisture changes in the root environment, the soil also **moderates temperature fluctuations**. Perhaps you can recall digging in the garden soil on a summer afternoon and feeling how hot the soil was at the surface and how much cooler just a few centimetres below. The insulating properties of soil protect the deeper portion of the root system from the extremes of hot and cold that often occur at the soil surface.

There are many potential sources of **phytotoxic substances** in soils. These toxins may result from human activity, or they may be produced by plant roots, by microorganisms, or by natural chemical reactions. An important role of the soils is to protect plants from toxic concentrations of such substances by ventilating gases, by decomposing or adsorbing organic toxins, or by suppressing toxin-producing organisms.

Soils supply plants with inorganic, **mineral nutrients** in the form of dissolved ions. These mineral nutrients include such metallic elements as potassium, calcium, iron and copper, as well as such non-metallic elements as nitrogen, sulphur, phosphorus, and boron. By eating plants, humans and other animals usually obtain the minerals they need (including several elements, that plants take up but do not appear to use themselves) indirectly from the soil. The plant takes these elements out of the soil solution and incorporates most of them into the thousands of different organic compounds that constitute plant tissue. A fundamental role of soils in supporting plant growth is to provide a continuing supply of dissolved mineral nutrients in amounts and relative proportions appropriate for plant growth.

Of the 92 naturally occurring chemical elements, only the 18 listed in Table 1.1 have been shown to be **essential elements** without which plants cannot grow and complete their life cycles. Essential elements used by plants in

relatively large amounts are called **macronutrients**; those used in smaller amounts are known as **micronutrients**.

Table 1.1 Elements essential for plant growth

Macronutrients: used in relatively large amounts (>0.1% of dry plant tissue)		Micronutrients: used in relatively small amounts (<0.1% of dry plant tissue)
Mostly from air and water	Mostly from soil solids	From soil solids
Carbon (CO_2)	Nitrogen (NO_3^-, NH_4^+)	Iron (Fe^{2+})
Hydrogen (H_2O)	Phosphorus ($H_2PO_4^-$, HPO_4^{2-})	Manganese (Mn^{2+})
Oxygen (O_2, H_2O)	Calcium (Ca^{2+})	Boron (HBO_3)
	Magnesium (Mg^{2+})	Zinc (Zn^{2+})
	Sulfur (SO_4^{2-})	Copper (Cu^{2+})
		Chlorine (Cl^-)
		Cobalt (Co^{2+})
		Molybdenum (MoO_4^{2-})
		Nickel (Ni^{2+})

* Many other elements are taken up from soils by plants, but are not essential for plant growth. Some of these (such as sodium, silicon, iodine, fluorine, barium, and strontium) do enhance the growth of certain plants, but do not appear to be as universally required for normal growth as are the 18 listed in this table.

In addition to the mineral nutrients just listed, plants may also use minute quantities of organic compounds from soils. However, uptake of these substances is certainly not necessary for normal plant growth. The organic metabolites, enzymes, and structural compounds making up a plant's dry matter consist mainly of carbon, hydrogen, and oxygen, which the plant obtains by photosynthesis from air and water, not from the soil. So it is true, that plants can be grown in nutrient solutions without soil (hydroponics), but then the plant-support functions of soils must be engineered into the system and maintained at a high cost of time, effort and management.

1.1.2 Regulator of Water Supplies

If we think about improving water quality, we must recognize that nearly every drop of water in our rivers, lakes, estuaries, and aquifers has either travelled

through the soil or flowed over its surface. (Of course this excludes the relatively minor quantity of precipitation that falls directly into bodies of fresh surface water).

Imagine, for example, a heavy rain falling on the hills surrounding a river. If the soil allows the rain to soak in, some of the water may be stored in the soil layers to the groundwater, eventually entering the river over a period of months or years as base flow. If the water is contaminated, as it soaks through the upper layers of soil it is purified and cleansed by soil processes that remove many impurities and kill potential disease organisms.

Contrast the preceding scenario with what would occur if the soil were so shallow or impermeable that most of the rain could not penetrate the soil, but ran off the hillsides on the soil surface, scouring surface soil and debris as it picked up speed, and entering the river rapidly and nearly all at once. The result would be a destructive flash flood of muddy water. Clearly, the nature and management of soils in a watershed will have a major influence on the purity and amount of water finding its way to aquatic systems.

1.1.3 Recycler of Raw Materials

What would a world be like without the recycling functions performed by soils? Without reuse of nutrients, plants and animals would have run out of nourishment long ago. The world would be covered with a layer, possibly hundreds of meters high, of plant and animal wastes and corpses. Obviously, recycling must be a vital process in ecosystems, whether forests, farms, or cities.

The soil system plays a pivotal role in the major geochemical cycles. Soils have the capacity to assimilate great quantities of organic waste, turning it into beneficial **humus**, converting the mineral nutrients in the wastes to forms that can be utilized by plants and animals, and returning the carbon to the atmosphere as carbon dioxide, where it again will become a part of living organisms through plant photosynthesis. Some soils can accumulate large amounts of carbon as soil organic matter, thus having a major impact on such global changes as the much-discussed *greenhouse effect*.

1.1.4 Habitat for Soil Organisms

When we speak of protecting ecosystems, most people envision a stand of old-growth forest with its abundant wildlife, or perhaps a marsh such as the surroundings of Fertő lake with its thousands of migrating birds and the famous hill of the European ground squirrel. (Hopefully, when you have visited this course, you will envision a handful of soil when someone speaks of an ecosystem)! Soil is not a mere pile of broken rock and dead debris. A handful of soil may be home to billions of organisms, belonging to thousands of species. In even this small quantity of soil, there are likely to exist preadators, prey, producers, consumers, and parasites.

1.1.5 Engineering Medium

We usually think of the soil as being firm and solid, a good base on which to build roads and all kinds of structures. Indeed, most structures rest on the soil, and many construction projects require excavation into the soil. Unfortunately, some soils are not as stable as others. Reliable construction on soils, and with soil materials, requires knowledge of the diversity of soil properties. Designs for roadbeds or building foundations that work well in one location on one type of soil may be inadequate for another location with different soils.

1.2 Soil as Environmental Interface

The importance of soil as a natural body derives in large part from its role as an interface between the worlds of rock (the lithosphere), air (the atmosphere), water (the hydrosphere), and living things (the biosphere). Environments, where all four of these worlds interact are often the most complex and productive on Earth. An estuary, where shallow waters meet the land and air, is an example of such an environment. Its productivity and ecological complexity far surpass those of a deep ocean trench, for example (where the hydrosphere is rather isolated), or the upper atmosphere (where rocks and water have little influence). The soil, or **pedosphere**, is another example of such and environment (Figure 1.1).
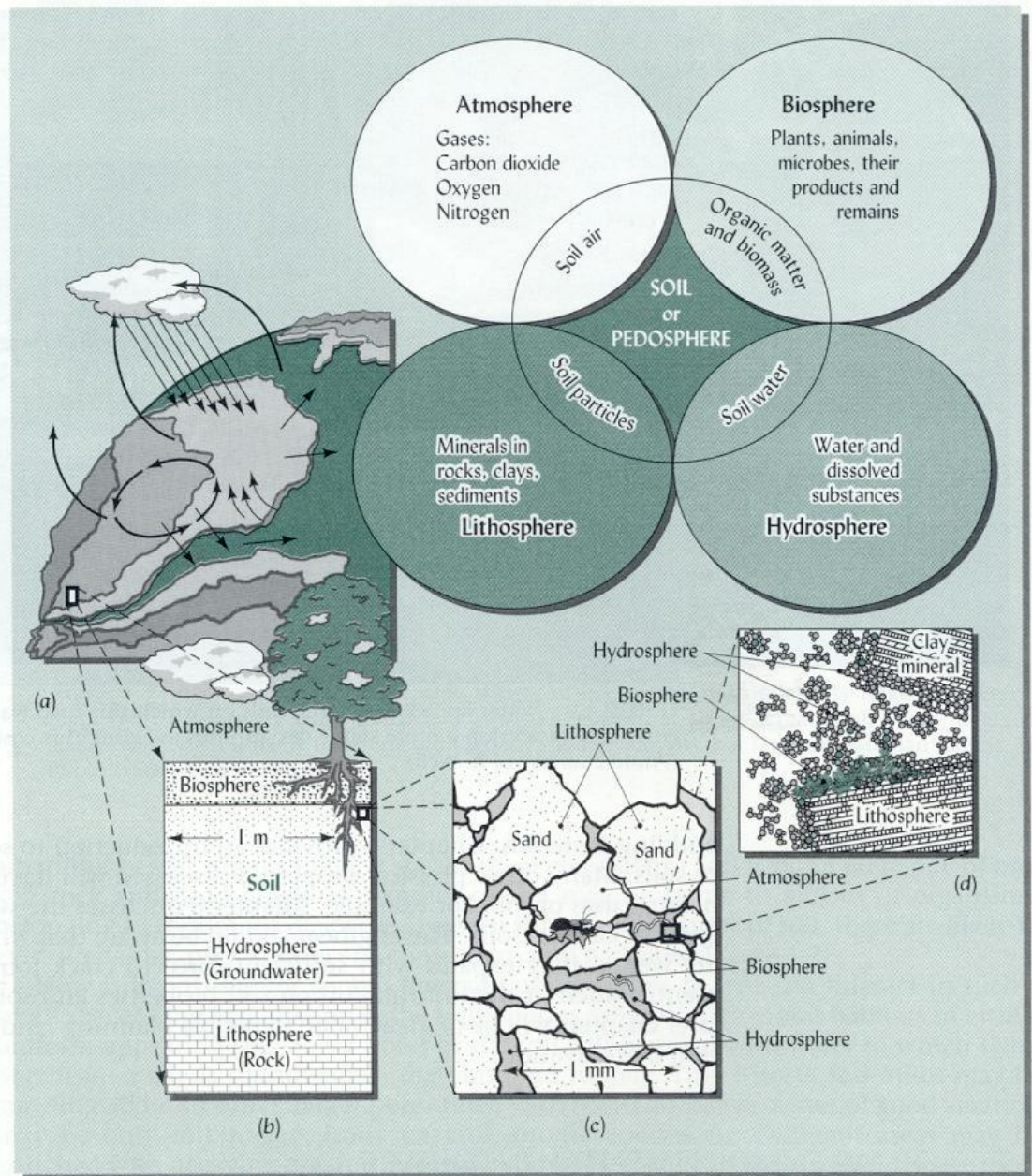
Figure 1.1 The pedosphere, where the worlds of rock (the lithosphere), air (the atmosphere), water (the hydrosphere), and life (the biosphere) all meet.

The concept of the soil as interface can be understood at many different scales: a) At the kilometer scale, soil participates in global cycles of rock weathering, atmospheric gas changes, water storage and partitioning, and the life of terrestrial ecosystems. b) At the meter scale, soil forms a transition zone between the hard rock below and the atmosphere above – a zone through which surface water and groundwater flow, and in which plants and other living organisms thrive. c) A thousand times smaller, at the millimetre scale,

mineral particles form the skeleton of the soil that defines pore space, some filled with air, and some with water, in which tiny creatures lead their lives. d) Finally, at the micro- and nanometer scales, soil minerals (lithosphere) provide charges, reactive surfaces that adsorb water and cations dissolved in water (hydrosphere), gases (atmosphere), and bacteria and complex humus macromolecules (biosphere).

1.3. Soil as a Natural Body

The soil is often said to cover the land as the peel covers an orange. However, while the peel is relatively uniform around the orange, the soil is highly variable from place to place on Earth. In fact, *the soil* is a collection of individually different soil bodies. One of these individual bodies, *a soil*, is to *the soil* as an individual tree is to the earth's vegetation.

In most places, the rock exposed at the earth's surface has crumbled and decayed to produce a layer of unconsolidated debris overlying the hard, unweathered rock. This unconsolidated layer is called the **regolith** (Figure 1.2), and varies in thickness from virtually nonexistent in some places (i.e., exposed bare rock) to tens of meters in other places. Where the underlying rock has weathered in place to the degree that it is loose enough to be dug with a spade, the term **saprolite** is used.

Through their biochemical and physical effects, living organisms such as bacteria, fungi and plant roots have altered the upper part – and in many cases, the entire depth – of the regolith. Here, at he interface between the worlds of rock, air, water, and living organisms, soil is born. Although generally hidden from everyday view, the soil and regolith can often be seen in road cuts and other excavations.

A soil is the product of both destructive and creative (synthetic) processes. Weathering of rock and microbial decay of organic residues are examples of destructive processes, whereas the formation of new minerals, such as certain clays, and of new stable organic compounds are examples of synthesis. Perhaps the most striking result of synthetic processes is the formation of contrasting layers called **soil horizons**. The development of these horizons in

the upper regolith is a unique characteristic of soil that sets it apart from the deeper regolith materials.
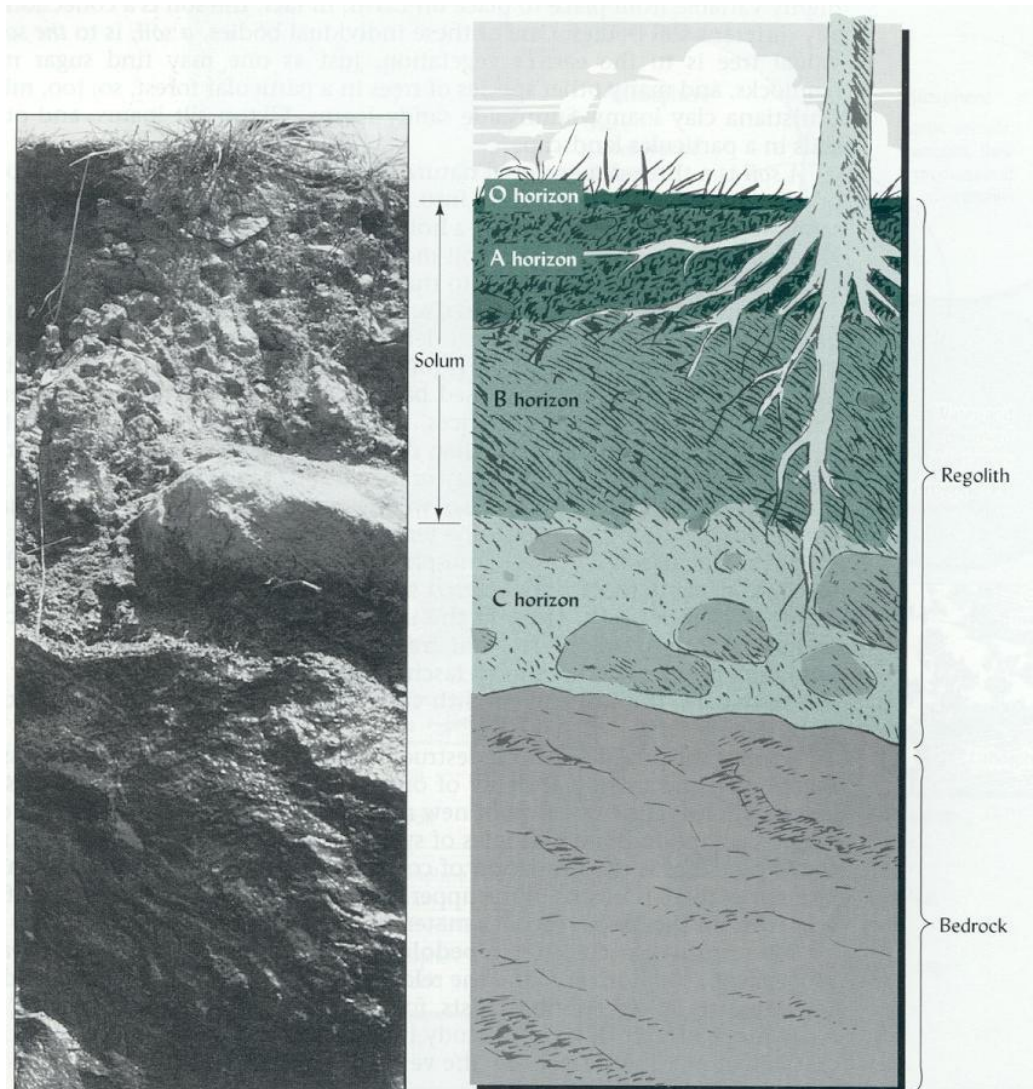

Figure 1.2 Relative positions of a regolith, its soil, and the underlying bedrock. Note, that the soil is a part of regolith, and that the A and B horizons are part of the **solum** (from the Latin word *solum*, which means soil or land). The C horizon is the part of the regolith that underlies the solum, but may be slowly changing into soil in its upper parts. Sometimes the regolith is so thin that it has been changed entirely to soil; in such a case, soil rests directly on the bedrock.

1.4 The Soil Profile and its Layers (horizons)

Soil scientists often dig a large hole, called a *soil pit*, usually several meters deep and about a meter wide, to expose soil horizons for study. The vertical section exposing a set of horizons in the wall of such a pit is termed a ***soil profile***. Road cuts and other ready-made excavations can expose soil profiles and serve as windows to the soil. In an excavation open for some time, horizons are often obscured by soil material that has been washed by rain from upper horizons to cover the exposed face of lower horizons. For this reason, horizons may be more clearly seen if a fresh face is exposed by scraping off a layer of material several centimetres thick from the pit wall.

Horizons within a soil may vary in thickness and have somewhat irregular boundaries, but generally they parallel the land surface. This alignment is expected since the differentiation of the regolith into distinct horizons is largely the result of influences, such as air, water, solar radiation, and plant material, originating at the soil-atmosphere interface. Since the regolith occurs first at the surface and works its way down, the uppermost layers have been changed the most, while the deepest layers are most similar to the original regolith, which is referred to as the soil's ***parent material***. In some cases, the regolith material has been transported long distances by wind, water or glaciers and deposited on top of dissimilar material. In such case, the regolith material found below a soil may be quite different from the upper layer of the regolith in which the soil formed.

In undisturbed ecosystems, especially forests, organic materials formed from fallen leaves and other plant and animal remains tend to accumulate on the surface. There they undergo varying degrees of physical and biochemical breakdown and transformation, so that layers of older, partially decomposed materials may underlie the freshly added debris. Together, these organic layers at the soil surface are designated the ***O horizons*** (Figure 1.3).
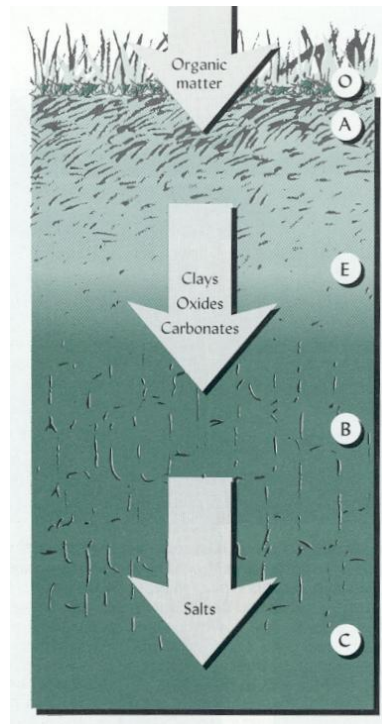
Figure 1.3 Horizons begin to differentiate as materials are added to the upper part of the profile and other materials are translocated to deeper zones. Under certain conditions, usually associated with forest vegetation and high rainfall, a leached E horizon forms between organic-matter rich A and the B horizon. If sufficient rainfall occurs, soluble salts will be carried below the soil profile, perhaps all the way to the groundwater.

Soil animals and percolating water move some of these organic materials downward to be mixed with the mineral grains of the regolith. These join the decomposing remains of plant roots to form organic materials that darken the upper mineral layers. Also, because weathering tends to be most intense nearest to the soil surface, in many soils the upper layers lose some of their clay or other weathering products by leaching to the horizons below. *A horizons* are the layers nearest the surface that are dominated by mineral particles but have been darkened by the accumulation of organic matter.

In some soils, intensely weathered and leached horizons that have not accumulated organic matter occur in the upper part of the profile, usually just below the A horizons. These horizons are designated *E horizons*.

The layers underlying the A and O horizons contain comparatively less organic matter than the horizons nearer the surface. Varying amounts of silicate clays, iron and aluminium oxides, gypsum, or calcium carbonate may accumulate in

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the underlying horizons. The accumulated materials may have been washed down from the horizons above, or they may have been formed in place through the weathering process. These underlying layers are referred to as *B horizons*. The A and B horizons together are part of the *solum* (from the Latin word *solum*, which means soil or land).

Plant roots and microorganisms often extend below the B horizon, especially in humid regions, causing chemical changes in the soil water, some biochemical weathering of the regolith, and the formation of *C horizons*. The C horizons are the least weathered part of the soil profile.

In some soil profiles, the component horizons are very distinct in color, with sharp boundaries that can be seen easily by even novice observers. In other soils, the color changes between horizons may be very gradual, and the boundaries more difficult to locate. However, color is only one of many properties by which one horizon may be distinguished from the horizon above or below it. Delineation of the horizons present in a soil profile often requires a careful examination, using all senses. In addition to seeing the colors in a profile, a soil scientist may feel, smell, and listen to the soil, as well as conduct chemical tests, in order to distinguish the horizons present.

1.5 Topsoil and Subsoil

The organically enriched A horizon at the soil surface is sometimes referred to as *top-soil*. When a soil is plowed and cultivated, the natural state of the upper 12-25 centimeters is modified. In this case, the topsoil may also be called the *plow layer*.

In cultivated soils, the majority of plant roots can be found in the topsoil. The topsoil contains a large part of the nutrient and water supplies needed by plants. The chemical properties and nutrient supply of the topsoil may be easily altered by mixing in organic and inorganic amendments, thereby making it possible to improve or maintain the soil's fertility and, to a lesser degree, its productivity. The physical structure of the topsoil, especially the part nearest the surface, is also readily affected by management operations such as tillage and application of organic materials.

The soil layers that underlie the topsoil are referred to as *subsoil*. The characteristics of the subsoil horizons can greatly influence most land uses. Much of the water needed by plants is stored in the subsoil. Many subsoils also supply important quantitites of certain plant nutrients. In some soils there is an abrupt change in properties between the topsoil and the subsoil. In other soils, the change is gradual and the upper part of the subsoil may be quite similar to the topsoil. In most soils, the properties of the topsoil are far more conducive to plant growth than those of the subsoil. That is why there is often a good correlation between the productivity of a soil and the thickness of the topsoil in a profile.

On the other hand, it is equally important not to reduce one's attention to the easily accessible "topsoil", for many soil properties are to be discovered only in the deeper layers. Plant-growth problems are often related to inhospitable conditions in the B or C horizons that restrict the penetration of roots. Similarly, the great volume of these deeper layers may control the amount of plant-available water held by a soil.

Traditionally, the lower boundary of the soil has been considered to occur at the greatest rooting depth of the natural vegetation, but soil scientist are increasingly studying layers below this in order to understand ecological processes such as groundwater pollution, parent material weathering, and geochemical cycles.

1.6 Soil: the Interface of Air, Minerals, Water and Life

The four major components of the soil are air, water, mineral matter, and organic matter. The relative proportions of these four components – which can be represented in a simple pie chart – greatly influence the behaviour and productivity of soils. Figure 1.4 shows the approximate proportions (by volume!) of the components found in a loam surface soil in good condition for plant growth.
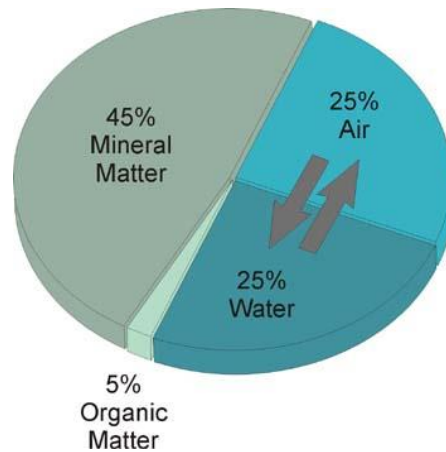
Figure 1.4 Volume composition of a loam surface soil when conditions are good for plant growth.

Although a soil may at first seem to be a solid thing, it should be noted that only about half the soil volume consists of solid material (mineral and organic); the other half consists of pore spaces filled with air or water. From the solid material, typically most is mineral matter derived from the rocks of the earth's crust. Only about 5% of the *volume* in this ideal soil consists of organic matter. However, the influence of the organic components on soil properties is often far greater than its small proportion would suggest. Since it is far less dense than mineral matter, the organic matter accounts for only about 2% of the *weight* of this soil.

The spaces between the particles of solid material are just as important to the nature of a soil as are the solids themselves. It is in these *pore spaces* that air and water circulate, roots grow, and microscopic creature live. Plant roots need both air and water. In an optimum condition for most plants, the pore space will be divided roughly equally among the two, with 25% of the soil volume consisting of water and 25% consisting of air. If there is much more water than this, the soil will be waterlogged. If much less water is present, plants will suffer from drought. The relative proportions of water and air in a soil typically fluctuate greatly as water is added or lost. Soils with much more than 50% of their volume in solids are likely to be too compacted for good plant growth. Compared to surface soil layers, subsoils tend to contain less organic matter, less total pore space, and a larger proportion of small pores (*micropores*) which tend to be filled with water rather than with air.

1.7 Mineral (inorganic) Constituents of Soils

Except in the case of organic soils, most of a soil's solid material consists of **mineral**[*] particles.[*]

The mineral particles present in soils are extremely variable in size. Excluding, for the moment, the larger rock fragments such as stones and gravel, soil particles range in size over four orders of magnitude: from 2.0 millimeters (mm) to smaller than 0.0002 mm in diameter. Individual **sand** particles are large enough (2.0 to 0.02 mm) to be seen by the naked eye and feel gritty when rubbed between the fingers. Sand particles do not adhere to one another; therefore, sands do not feel sticky. **Silt** particles are somewhat smaller (0.02 to 0.002 mm). Silt particles are too small to see without a microscope or to feel individually, so silt feels smooth but not sticky, even when wet. The smallest class of mineral particles are the clays (<0.002 mm), which adhere together to form a sticky mass when wet and hard clods when dry.

The smaller particles (<0.001 mm) of clay (and similar-sized organic particles) have **colloidal**[**] properties and can be seen only with the aid of an electron microscope.

To anticipate the effect of clay on the way a soil will behave, it is not enough to know only the amount of clay in a soil. It is also necessary to know the *kinds* of clays present. As home builders and highway engineers know all too well, certain clayey soils, such as those high in smectite clays, make very unstable material on which to build because the clays swell when the soil is wet and shrink when the soil dries. This shrink-and-swell action can easily crack foundations and cause even heavy retaining walls to collapse. These clays also become extremely sticky and difficult to work when they are wet. Other types of clays, formed under different conditions, can be very stable and easy to work with. Learning about the different types of clay minerals will help us understand many of the physical and chemical differences among soils in various parts of the world.

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[*] The word *mineral* is used in soil science in three ways: (1) as a general adjective to describe inorganic materials derived from rocks; (2) as a specific noun to refer to distinct minerals found in nature, such as quartz and feldspars; and (3) as an adjective to describe chemical elements, such as nitrogen and phosphorus, in their inorganic state in contrast to their occurrence as part of organic compounds.

1.7.1 Primary and Secondary Minerals

Minerals that have persisted with little change in composition since they were extruded in molten lava (e.g., quartz, micas, and feldspars) are known as *primary minerals*. They are prominent in the sand and silt fractions of soils. Other minerals, such as silicate clays and iron oxides, were formed by breakdown and weathering of less resistant minerals as soil formation progressed. These minerals are called *secondary minerals* and tend to dominate the clay and, in some cases, silt fractions.

The inorganic minerals in the soil are the original source of most of the chemical elements essential for plant growth. Although the bulk of these nutrients is held rigidly as components of the basic crystalline structure of the minerals, a small but important portion is in the form of charged ions on the surface of fine colloidal particles (clays and organic matter). Mechanisms of critical importance to growing plants allow plant roots to have access to these surface-held nutrient ions.

1.7.2 Soil Structure

Sand, silt, and clay particles can be thought of as the building blocks from which soil is constructed. The manner in which these building blocks are arranged together is called *soil structure*. The particles may remain relatively independent of each other, but more aggregates may take the form of roundish granules, cubelike blocks, flat plates, or other shapes.

Soil texture, which means the relative amounts of different sizes of particles, is just as important as soil structure in governing how water and air move in soils. Both structure and texture fundamentally influence the suitability of soils for the growth of plant roots.

1.8 Soil Organic Matter

Soil organic matter consists of a wide range of organic (carbonaceous) substances, including living organisms (the *soil biomass*), carbonaceous remains of organisms that once occupied the soil, and organic compounds produced by current and past metabolism in the soil. The remains of plants, animals, and microorganisms are continuously broken down in the soil and

new substances are synthesized by other microorganisms. Over time, organic matter is lost from the soil as carbon dioxide is produced by microbial respiration. Because of such loss, repeated additions of new plant and/or animal residues are necessary to maintain soil organic matter.

Under conditions, that favour plant production more than microbial decay, large quantitites of atmospheric carbon dioxide, used by plants in photosynthesis, are sequestered in the abundant plant tissues which eventually become part of the soil organic matter. Since carbon dioxide is a major cause of the greenhouse effect which is believed to be warming the earth's climate, the balance between accumulation of soil organic matter and its loss through microbial respiration has global implications. In fact, more carbon is stored in the world's soils than in the world's plant biomass and atmosphere combined.

Even so, organic matter comprises only a small fraction of the mass of a typical soil. By weight, typical well-drained mineral surface soils contain from *1 to 6%* organic matter. The organic matter content of subsoils is even smaller. However, the influence of organic matter on soil properties, and consequently on plant growth, is far greater than the low percentage would indicate.

Organic matter binds mineral particles into a *granular* soil structure that is largely responsible for the loose, easily managed condition of productive soils. Part of the soil organic matter, that is especially effective in stabilizing these granules consists of certain gluelike substances produced by various soil organisms, including plant roots.

Organic matter also increases the amount of water a soil can hold and the proportion of water available for plant growth. In addition, it is a major soil source of the plant nutrients phosphorus and sulfur, and the primary source of nitrogen for most plants. As soil organic matter decays, these nutrient elements, which are present in organic combinations, are released as soluble ions that can be taken up by plant roots. Finally, organic matter, including plant and animal residues, is the main food that supplies carbon and energy to soil organisms. Without it, biochemical activity so essential for ecosystem functioning would come to a near standstill.

Humus, usually black or brown in color, is a collection of very complex organic compounds which accumulate in soil because they are relatively resistant to

decay. Humus consists of the combined residues of organic materials which have lost their original structure following the rapid decomposition of the simpler ingredients and includes synthesized cell substance as well as by-products of microorganisms. It is not a definite substance and is in a continual state of flux, disappearing by slow decomposition, and being constantly renewed by incorporation of residual matter.

Just as clay is the colloidal fraction of soil mineral matter, so humus is the colloidal fraction of soil organic matter. Because of their charged surfaces, both humus and clay act as contact bridges between larger soil particles; thus, both play an important role in the formation of soil structure. The surface charges of humus, like those of clay, attract and hold both nutrient ions and water molecules. However, gram for gram, the capacity of humus to hold nutrients and water is far greater than that of clay. Unlike clay, humus contains certain components that can have a hormone-like stimulatory effect on plants. All in all, amounts of humus may remarkably increase the soil's capacity to promote plant growth.

1.9 Soil Water: a Dynamic Solution

Water is of vital importance in the ecological functioning of soils. The presence of water in soils is essential for the survival and growth of plants and other soil organisms.

Water is held within soil pores with varying degrees of strength depending on the amount of water present and the size of the pores. The attraction between water and the surfaces of soil particles greatly restrict the ability of water to flow as it would flow in a drinking glass.

Because soil water is never pure water, but contains hundreds of dissolved organic and inorganic substances, it may be more accurately called the *soil solution*. An important function of the soil solution is to serve as a constantly replenished, dilute nutrient solution bringing dissolved nutrient elements (e.g., calcium, potassium, nitrogen, and phosphorus) to plant roots.

When the soil moisture content is optimal for plant growth, the water in the large- and intermediate-sized pores can move about in the soil and can easily be used by plants. As the plant grows, however, its roots remove water from

the largest pores first. Soon the largest pores hold only air, and the remaining water is found only in the intermediate- and smallest-sized pores. The water in the intermediate-sized pores can still move toward plant roots and be taken up by them. However, the water in the smallest pores is so close to solid particles that it is strongly attracted to and held on the particle surfaces. This water may be so strongly held that plant roots cannot pull it away. Consequently, *not all soil water is available to plants*. Depending on the soil, one-sixth to one-half of the water may remain in the soil after plants have wilted or died for lack of moisture.

1.9.1 Soil Solution

The soil solution contains small but significant quantities of soluble inorganic compounds, some of which supply elements that are essential for plant growth. Refer to Table 1.1 for a listing of the 18 *essential elements*, along with their sources. The soil solids, particularly the fine organic and inorganic colloidal particles, release these elements to the soil solution, from which they are taken up by growing plants. Such exchanges, which are critical for higher plants, are dependent on both soil water and the fine soil solids.

An other critical property of the soil solution is its *acidity* or *alkalinity*. Many chemical and biological reactions are dependent on the levels of the H^+ and OH^- ions in the soil solution. The levels of these ion also influence the solubility, and in turn the availability, of several essential nutrient elements (including iron and manganese) to plants.

1.10 Soil Air: a Changing Mixture of Gases

Approximately half of the volume of the soil consists of pore spaces of varying sizes, which are filled with either water or air. When water enters the soil, it displaces air from some of the pores; the air content of a soil is therefore inversely related to its water content. If we think of the network of soil pores as the ventilation system of the soil, connecting airspaces to the atmosphere, we can understand that when the smaller pores are filled with water, the ventilation system becomes clogged. Because oxygen can not enter the clogged pores, nor carbon dioxide leave it, the air in the pore space

surrounded with water-filled smaller pores will soon become depleted of oxygen and enriched in carbon dioxide and water vapour by the respiration (breathing) of the plant roots and microorganisms.

Therefore, soil air differs from atmospheric air in several respects. First, the composition of soil air varies greatly from place to place in soil. Second, soil air generally has a higher moisture content than the atmosphere; the relative humidity of soil air approaches 100% unless the soil is very dry. Third, the content of carbon dioxide (CO_2) is usually much higher, and that of oxygen (O_2) lower, than contents of these gases found in the atmosphere. Carbon dioxide in soil air is often several hundred times more concentrated than the 0.038% commonly found in the atmosphere. Oxygen decreases accordingly and, in extreme cases, may be only 5 to 10%, or even less, compared to about 20% for atmospheric air.

The amount and composition of air in a soil are determined to a large degree by the water content of the soil. The air occupies those soil pores not filled with water. As the soil drains from a heavy rain or irrigation, large pores are the first to be filled with air, followed by medium-sized pores, and finally the small pores, as water is removed by evaporation and plant use.

1.11 Interaction of Four Components to Supply Plant Nutrients

As you read our discussion of each of the four major soil components, you may have noticed that the impact of one component on soil properties is seldom expressed independently from that of the others. Rather, the four components interact with each other to determine the nature of a soil. Thus, soil moisture, which directly meets the needs of plants for water, simultaneously controls much of the air and nutrient supply to the plant roots. The mineral particles, especially the finest ones, attract soil water, thus determining its movement and availability to plants. Likewise, organic matter, because of its physical binding power, influences the arrangement of the mineral particles into clusters and, in so doing, increases the number of large soil pores, thereby influencing the water and air relationships.

1.12 Soil Quality, Degradation and Resilience

Soil is a basic resource underpinning all terrestrial ecosystems. Managed carefully, soils are a *reusable* resource, but in the scale of human lifetimes they *cannot be considered a renewable resource*. In all regions of the world, human activities are destroying some soils far faster than nature can rebuild them. As mentioned before, growing numbers of people are demanding more and more from the Earth's fixed amount of land. Nearly all of the soils best suited for growing crops are already being farmed. Therefore, as each year brings million more people to feed, the amount of cropland per person continuously declines.

Finding more land on which to grow food is not easy. Most additional land brought under cultivation comes at the cost of clearing natural forests, savannas, and grasslands. Images made from the Earth from orbiting satellites show the resulting decline in land covered by forests and other natural ecosystems. If there is to be space for both people and wildlife, the best of our existing farmland soils will require improved and more intensive management. Soils completely washed away by erosion or excavated and paved over by urban sprawl are permanently lost, for all practical purposes. More often, soils are degraded in quality rather than totally destroyed.

Soil quality is a measure of the ability of a soil to carry out particular ecological functions, such as those described in this lecture. Soil quality reflects a combination of *chemical*, *physical*, and *biological* properties. Some of these properties are relatively unchangeable, inherent properties that help define a particular type of soil. Soil texture and mineral makeup are examples. Other soil properties, such as structure and organic matter content, can be significantly changed by management. These more changeable soil properties can indicate the status of a soil's quality relative to its potential.

Mismanagement of forests, farms, and rangeland causes widespread *degradation* of soil quality by erosion that removes the topsoil, little by little. Another widespread cause of soil degradation is the accumulation of salts improperly irrigated soils in arid regions. When people cultivate soils and harvest the crops without returning organic residues and mineral nutrients, the soil's supply of organic matter and nutrients becomes depleted. Such depletion is particularly widespread in sub-Saharan Africa. Contamination of

a soil with toxic substances from industrial processes or chemical spills can degrade its capacity to provide habitat for soil organisms, to grow plant that are safe to eat, or to safely recharge ground and surface waters. Degradation of soil quality by pollution is usually localized, but the environmental impacts and costs involved are very large.

While protecting soil quality must be the first priority, it is often necessary to attempt to restore the quality of soils that have already been degraded. Some soils have sufficient *resilience* to recover from minor degradation if left to revegetate on their own. In other cases, more effort is required to restore degraded soils. Organic and inorganic amendments may have to be applied, vegetation may have to be planted, physical alteration by tillage of grading may have to be made, or contaminants may have to be removed.

Study Questions:

1. As a society, is our reliance on soils likely to increase or decrease in the decades ahead? Explain.

2. Discuss how a soil, a natural body, differs from soil, a material that is used in building a roadbed?

3. What are the five main roles of soil in an ecosystem? For each of these ecological roles, suggest one way in which interactions occur with another of the five roles.

4. Think back over your activities during the past week. List as many incidents as you can in which you came into direct or indirect contact with soil.

5. List the essential nutrient elements that plants derive mainly from the soil.

6. Are all elements contained in plants essential nutrients? Explain.

7. Define these terms: soil texture, soil structure, soil pH, humus, soil profile, B horizon, soil quality, solum, and saprolite.

8. Describe four processes, that commonly lead to degradation of soil quality.

2. Formation of Soils from Parent Materials

The first astronauts to explore the moon laboured in their clumsy pressurized suits to collect samples of rocks and dust from the lunar surface. These they carried back to Earth for analysis. It turned out that moon rocks are similar in composition to those found deep in the Earth – so similar that scientists concluded that the moon itself began as a large chunk of molten Earth that broke away eons ago, when the young planet nearly melted in a stupendous collision with a Mars-sized object, leaving the Pacific Ocean as a scar. On the moon, this rock remained unchanged or crumbled into dust with the impact of meteors. On Earth, the rock at the surface, eventually coming in contact with water, air and living organisms, was transformed in to something new, into many different kinds of living soils. This lecture reveals the story of how rock and dust become "the living skin of the Earth".

2.1 Weathering of Rocks and Minerals

The influence of *weathering*, the physical and chemical breakdown of particles, is evident everywhere. It breaks up rocks and minerals, modifies or destroys their physical and chemical characteristics, and carries away the soluble products. It also synthesizes new minerals of great significance in soils. The nature of the rocks and minerals being weathered determines the rates and results of the breakdown and synthesis.

2.2 Characteristics of Rocks and Minerals

The rocks in the earth's outer surface are commonly classified as *igneous*, *sedimentary*, and *metamorphic*. Those of igneous origin are formed from molten magma and include such common rocks as *granite* and *diorite*.

Igneous rock is composed of primary minerals such as light-coloured quartz, muscovite, and feldspars and dark-colored biotite, augite, and hornblende. In general, dark-colored minerals contain iron and magnesium and are more easily weathered. Therefore, dark-colored igneous rocks such as *gabbro* and *basalt* are more easily broken down than are *granites* and other lighter-

colored igneous rocks. Worldwide, light-coloured minerals and quartz are generally more prominent than are dark-coloured minerals. The mineral grains in igneous rocks are randomly dispersed and interlocked, giving a salt-and-pepper appearance if they are coarse enough to see with the unaided eye.

Sedimentary rock consists of compacted or cemented weathering products from older, pre-existing rocks. For example, quartz sand weathered from a granite rock and washed into the ocean may settle on the ocean floor and eventually become cemented into a solid mass called *sandstone*. Later, movements of the Earth's crust may raise this sandstone above sea level, where it could become the material we are so familiar with in certain canyon walls, road cuts and soil parent materials. Similarly, clays may be compacted into *shale*. Other important sedimentary rocks are listed in Table 2.1 along with their dominant minerals.

Table 2.1 Some of the most important sedimentary and metamorphic rocks and the minerals commonly dominant in them

Dominant mineral	Type of rock	
	Sedimentary	Methamorphic
Calcite ($CaCO_3$)	Limestone	Marble
Dolomite ($CaCO_3 \cdot MgCO_3$)	Dolomite	Marble
Quartz (SiO_2)	Sandstone	Quartzite
Clays	Shale	Slate
Variable	Conglomerate[a]	Gneiss[b]
Variable		Schist[b]

[a]Small stones of various mineralogical makeup are cemented into conglomerate.
[b]The minerals present are determined by the original rock, which has been changed by metamorphism. Primary minerals present in the igneous rocks commonly dominate these rocks, although some secondary minerals are also present.

Because most of what is presently dry land was at some time in the past covered by water, sedimentary rocks are the most common type of rock encountered, covering about 75% of the Earth's land surface. The resistance

of a given sedimentary rock to weathering is determined by its particular dominant minerals and by the cementing agent.

Metamorphic rocks are those that have formed by the metamorphism or change in form of other rocks. As the Earth's continental plates shift, and sometimes collide, forces are generated that can uplift great mountain ranges. As a result, igneous and sedimentary masses are subjected to tremendous heat and pressure that compress, distort, and/or partially remelt the original rocks. Igneous rocks are commonly modified to form schist or gneiss in which light and dark minerals have been reoriented into bands. Sedimentary rocks, such as limestone and schale, may be metamorphosed to marble and slate, respectively. As this is the case for igneous and sedimentary rock, the particular minerals that dominate a given metamorphic rock influence its resistance to chemical weathering. A high degree of metamorphism may also physically weaken the rock mass, hastening its breakdown into smaller fragments.

2.3 Weathering: A General Case

Weathering is a biochemical process that involves both destruction and synthesis. The original rocks and minerals are destroyed by both *physical disintegration* and *chemical decomposition* (Figure 2.1).
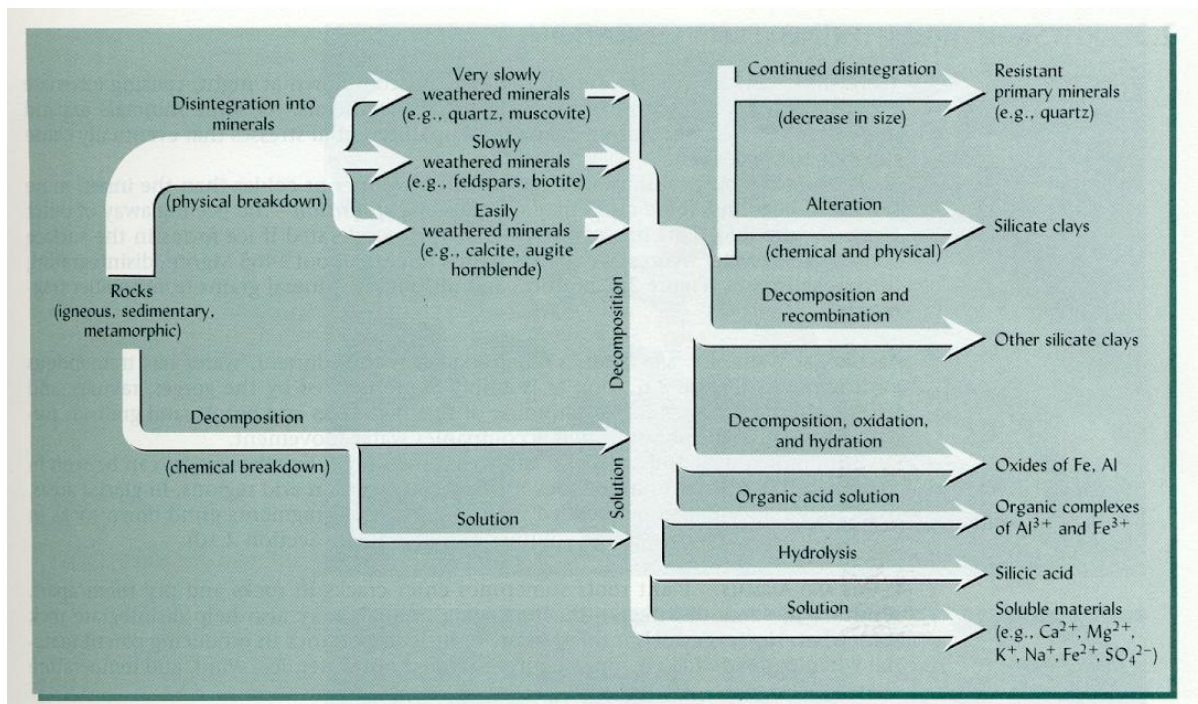
Figure 2.1 **Pathways of weathering** that occur under moderately acid conditions common in humid temperate regions. The disintegration of rocks into small individual mineral grains is a physical process, whereas decomposition, recombination, and solution are chemical processes. Alteration of minerals involves both physical and chemical processes. Note that resistant primary minerals, newly synthesized secondary minerals, and soluble materials are products of weathering. In arid regions the physical processes predominate, but in humid tropical areas decomposition and recombination are most prominent.

Without appreciably affecting their composition, physical disintegration breaks down rock into smaller rocks and eventually into sand and silt particles that are commonly made up of individual minerals. Simultaneously, the minerals decompose chemically, releasing soluble materials and synthesizing new minerals, some of which are resistant end products. New minerals form either by minor chemical alterations or by complete chemical breakdown of the original mineral and resynthesis of new minerals. During the chemical changes, particle size continues to decrease, and constituents continue to dissolve in the aqueous weathering solution. The dissolved substances may recombine into new (secondary) minerals, may leave the profile in drainage water, or may be taken up by plant roots.

Three groups of minerals that remain in well-weathered soils are shown on the right side of Figure 2.1: (1) silicate clays, (2) very resistant end

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products, including iron and aluminium oxide clays, and (3) very resistant primary minerals, such as quartz. In highly weathered soils of humid tropical regions, the oxides of iron and aluminium and certain silicate clays with low Si/Al ratios predominate because most other constituents have been broken down and removed.

2.3.1 Physical Weathering (Disintegration)

TEMPERATURE: Rocks heat up during the day and cool down at night, causing alternate expansion and contraction of their constituent minerals. As some minerals expand more than others, temperature changes set up differential stresses that eventually cause the rock to crack apart.

Because the outer surface of a rock is often warmer or colder than the inner, more protected portions, some rocks may weather by *exfoliation* – the peeling away of outer layers. This process may be sharply accelerated if ice forms in the surface cracks. When water freezes, it expands with a force of about 1465 Mg/m^2, disintegrating huge rock masses and dislodging mineral grains from smaller fragments.

ABRASION BY WATER, ICE, AND WIND: When loaded with sediment, water has tremendous cutting power, as is clearly demonstrated by the different forms of valleys around the world. The rounding of riverbed rocks and beach sand grains is further evidence of the abrasion that accompanies water movement.

Windblown dust and sand also can wear down rocks by abrasion, as can be seen in the many picturesque rounded rock formations in certain arid regions. In glacial areas, huge moving ice masses embedded with soil and rock fragments grind down rocks in their path and carry away large volumes of material.

PLANTS AND ANIMALS: Plant roots sometimes enter cracks in rocks and pry them apart, resulting in some disintegration. Burrowing animals may also help disintegrate rock somewhat. However, such influences are of little importance in producing parent material when compared to the drastic physical effects of water, ice, wind, and temperature change.

2.3.2 Biogeochemical Weathering

While physical weathering is dominant in very cold or very dry environments, chemical reactions are most intense where the climate is wet and hot. However, both types of weathering occur together, and each tends to accelerate the other. For example, physical abrasion (rubbing together) decreases the size of particles and therefore increases their surface area, making them more susceptible (vulnerable) to rapid chemical reactions.
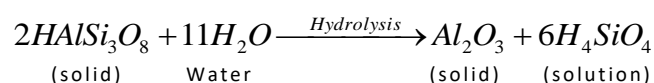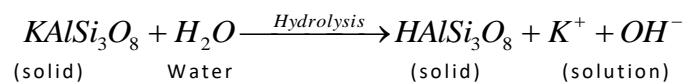
Chemical weathering is enhanced by such geological agents as the presence of water and oxygen, as well as by such biological agents as the acids produced by microbial and plant-root metabolism. That is why the term ***biogeochemical weathering*** is often used to describe the process. The various agents act in concert to convert primary minerals (e.g., feldspars and micas) to secondary mineral (e.g., clays and carbonates) and release plant nutrient elements in soluble forms. Note the importance of water in each of the **six basic types of chemical weathering reactions** discussed in the following.

HYDRATION: Intact water molecules may bind to a mineral by the process called ***hydration***.
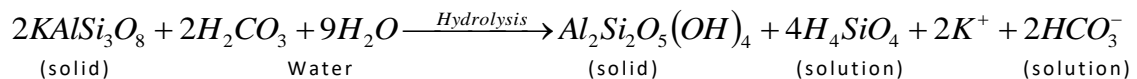
$$5Fe_2O_3 + 9H_2O \xrightarrow{\;Hydration\;} Fe_{10}O_{15} \bullet 9H_2O$$

Hematite   Water   Ferryhydrite
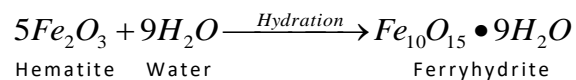
Hydrated oxides of iron and aluminium (e.g., $Al_2O_3 \cdot 3H_2O$) exemplify common products of hydration reactions.

HYDROLYSIS: In hydrolysis reactions, water molecules split into their hydrogen and hydroxyl components and the hydrogen often replaces a cation from the mineral structure. A simple example is the action of water on microcline, a potassium-containing feldspar.

$$KAlSi_3O_8 + H_2O \xrightarrow{\;Hydrolysis\;} HAlSi_3O_8 + K^+ + OH^-$$

(solid)   Water   (solid)   (solution)

$$2HAlSi_3O_8 + 11H_2O \xrightarrow{\;Hydrolysis\;} Al_2O_3 + 6H_4SiO_4$$

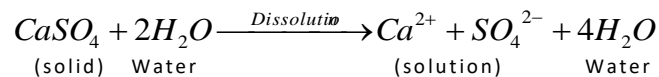(solid)   Water   (solid)   (solution)

The potassium released is soluble and is subject to adsorption by soil colloids, uptake by plants, and removal in the drainage water. Likewise, the silicic acid (H_4SiO_4) is soluble. It can be removed slowly in drainage water, or it can recombine with other compounds to form secondary minerals such as the silicate clays.
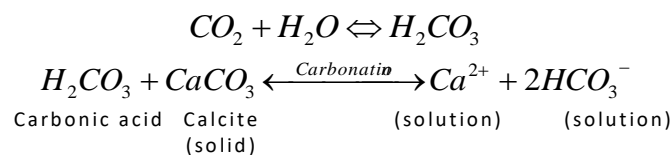
Hydrolysis is the most important process in the weathering of silicate minerals. The most common weathering reaction on earth is the hydrolysis of feldspars, producing clay minerals, e.g., K-feldspars are changed in this way to kaolinite:

$$\underset{\text{(solid)}}{2KAlSi_3O_8} + \underset{\text{Water}}{2H_2CO_3} + 9H_2O \xrightarrow{Hydrolysis} \underset{\text{(solid)}}{Al_2Si_2O_5(OH)_4} + \underset{\text{(solution)}}{4H_4SiO_4} + \underset{\text{(solution)}}{2K^+ + 2HCO_3^-}$$

DISSOLUTION: Water is capable of dissolving many minerals by hydrating the cations and anions until they become dissociated from each other and surrounded by water molecules. Gypsum dissolving in water provides an example.
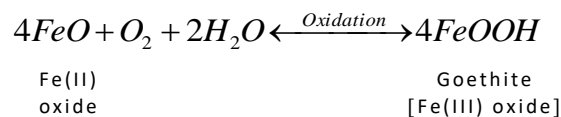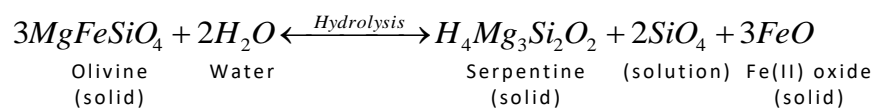
$$\underset{\text{(solid)}}{CaSO_4} + \underset{\text{Water}}{2H_2O} \xrightarrow{Dissolution} \underset{\text{(solution)}}{Ca^{2+} + SO_4^{2-}} + \underset{\text{Water}}{4H_2O}$$

CARBONATION AND OTHER ACID REACTIONS: Weathering is accelerated by the presence of acids, which increase the activity of hydrogen ions in water. For example, when carbon dioxide dissolves in water (a process enhanced by microbial and root respiration) the carbonic acid (H_2CO_3) produced hastens the chemical dissolution of calcite in limestone or marble, as illustrated when the following reactions go to the right:

$$CO_2 + H_2O \Leftrightarrow H_2CO_3$$

$$\underset{\substack{\text{Carbonic acid} \\ }}{H_2CO_3} + \underset{\substack{\text{Calcite} \\ \text{(solid)}}}{CaCO_3} \xleftarrow{Carbonation} \underset{\text{(solution)}}{Ca^{2+}} + \underset{\text{(solution)}}{2HCO_3^-}$$

Soils also contain other, stronger acids, such as nitric acid (HNO_3), sulphuric acid (H_2SO_4), and many organic acids. Hydrogen ions are also associated with soil clays. Each of these sources of acidity is available for reaction with soil minerals.
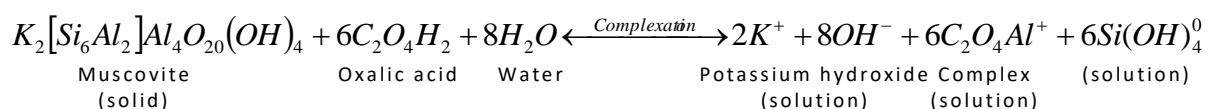
OXIDATION-REDUCTION: Minerals that contain iron, manganese, or sulphur are especially susceptible to oxidation-reduction reactions. Iron is usually laid down in primary minerals in the divalent Fe(II) (ferrous) form. When rocks containing such minerals are exposed to air and water during soil formation, the iron is easily oxidized (loses an electron) and becomes trivalent Fe(III) (ferric). If iron is oxidized from Fe(II) to Fe(III), the change in valence and ionic radius causes destabilizing adjustments in the crystal structure of the mineral.

In other cases, Fe(II) may be released from the mineral and almost simultaneously oxidized to Fe(III). For example, the hydration of olivine releases ferrous oxide, which may be oxidized immediately to ferric oxyhydroxide (goethite):

$$3MgFeSiO_4 + 2H_2O \xleftarrow{\quad Hydrolysis \quad} H_4Mg_3Si_2O_2 + 2SiO_4 + 3FeO$$

| Olivine (solid) | Water | Serpentine (solid) | (solution) | Fe(II) oxide (solid) |

$$4FeO + O_2 + 2H_2O \xleftarrow{\quad Oxidation \quad} 4FeOOH$$

| Fe(II) oxide | Goethite [Fe(III) oxide] |

The oxidation and/or removal of iron during weathering is often made visible by changes in the colors of the resulting altered minerals.

COMPLEXATION: Soil biological processes produce organic acids such as oxalic, citric, and tartaric acids, as well as the much larger fulvic and humic acid molecules. In addition to providing H$^+$ ions that help solubilise aluminium and silicon, they also form organic complexes (*chelates*) with the Al^{3+} ions held within the structure of silicate minerals. By so doing they remove the Al^{3+} from the mineral, which then is subject to further disintegration. In the following example, oxalic acid forms a soluble complex with aluminium from the mineral, muscovite. As this reaction proceeds to the right, it destroys the muscovite structure and releases dissolved ions of the plant nutrient, potassium.

$$K_2[Si_6Al_2]Al_4O_{20}(OH)_4 + 6C_2O_4H_2 + 8H_2O \xleftarrow{\quad Complexation \quad} 2K^+ + 8OH^- + 6C_2O_4Al^+ + 6Si(OH)_4^0$$

| Muscovite (solid) | Oxalic acid | Water | Potassium hydroxide (solution) | Complex (solution) | (solution) |

Had there been no living organism on Earth, the chemical weathering processes we have just outlined would probably have proceeded 1000 times more slowly, with the result that little, if any, soil would have developed on our planet

INTEGRATED WEATHERING PROCESSES: The various chemical weathering processes occur simultaneously and are interdependent. For example, hydrolysis of a given primary mineral may release ferrous iron [Fe(II)] that is quickly oxidized to the ferric [Fe(III)] form, which, in turn, is hydrated to give a hydrous oxide of iron. Hydrolysis or complexation also may release soluble cations, silicic acid, and aluminium or iron compounds. In humid environments, some of the soluble cations and silicic acid are likely to be lost from the weathering mass in drainage waters. The released substances can also be recombined to form silicate clays and other secondary silicate minerals. In this manner, the biochemical processes of weathering transform primary geologic materials into the compounds of which soils are made.

2.4 Factors Influencing Soil Formation

We understand *the soil* as a collection of *individual soils*, each with distinctive profile characteristics. This concept of soils as organized natural bodies derived initially from late-19th-century field studies by a brilliant Russian team of soil scientists led by V.V. Dokuchaev. They noted similar profile layering in soils hundreds of kilometres apart, provided that the climate and vegetation were similar at the two locations. Such observations and much careful subsequent field and laboratory research led to the recognition of **five major factors that control the formation of soils**.

1. *Parent materials* (geological or organic precursors to the soil);

2. *Climate* (primarily precipitation and temperature);

3. *Biota* (living organisms, especially native vegetation, microbes, soil animals, and human beings);

4. *Topography* (slope, aspect, and landscape position);

5. *Time* (the period of time since the parent materials became exposed to soil formation).

Soils are often defined in terms of these factors as "dynamic natural bodies having properties derived from the combined effects of climate and biotic activities, as modified by topography, acting on parent materials over periods of time".

We will now examine how each of these five factors affects the outcome of soil formation. However, as we do, we must keep in mind that these factors do not exert their influences independently. Indeed, interdependence is the rule. For example, contrasting climatic regimes are likely to be associated with contrasting types of vegetation, and perhaps differing topography and parent material, as well. Nonetheless, in certain situations one of the factors has had the dominant influence in determining differences among a set of soils. Soil scientists refer to such a set of soils as a *lithosequence*, *climosequence*, *biosequence*, *toposequence*, or *chronosequence*.

2.4.1 Parent Materials

Geological processes have brought to the earth's surface numerous parent materials in which soils form (Figure 2.2). The nature of the parent material profoundly influences soil characteristics. For example, a soil might inherit a sandy texture from a coarse-grained, quartz-rich parent material such as granite or sandstone. Soil texture, in turn, helps control the percolation of water through the soil profile, thereby affecting the translocation of fine soil particles and plant nutrients.

The chemical and mineralogical composition of parent material also influences both chemical weathering and the natural vegetation. For example, the presence of limestone in parent material will slow the development of acidity that typically occurs in humid climates. In addition, trees growing in limestone materials produce leaf litter that is relatively high in calcium. Incorporation of the calcium-rich litter into the soil further delays the process of acidification and, in humid temperate areas, the progress of soil development.

Parent material also influences the quantity and type of clay minerals present in the soil profile. First, the parent material itself may contain varying amounts and types of clay minerals, perhaps from a previous weathering cycle.

Second, the nature of the parent material greatly influences the kinds of clays that can develop as the soil evolves. In turn, the nature of the clay minerals present, markedly affects the kind of soil that develops.

2.4.1.1 Classification of Parent Materials

Inorganic parent materials can either be formed in place, as residual material weathered from rock, or they can be transported from one location and deposited at another (Figure 2.2). In wet environments (such as swamps and marshes), incomplete decomposition may allow organic parent materials to accumulate from the residues of many generations of vegetation. Although it is their chemical and physical properties that most influence soil development, parent materials are often classified with regard to the mode of placement in their current location, as seen on the right side of Figure 2.2.

Although these terms properly relate only to the placement of the parent materials, people sometimes refer to the soils that form from these deposits as *organic soils*, *glacial soils*, *alluvial soils*, and so forth. These terms are quite non-specific because parent material properties vary widely within each group, and because the effect of parent material is modified by the influence of climate, organisms, topography, and time.
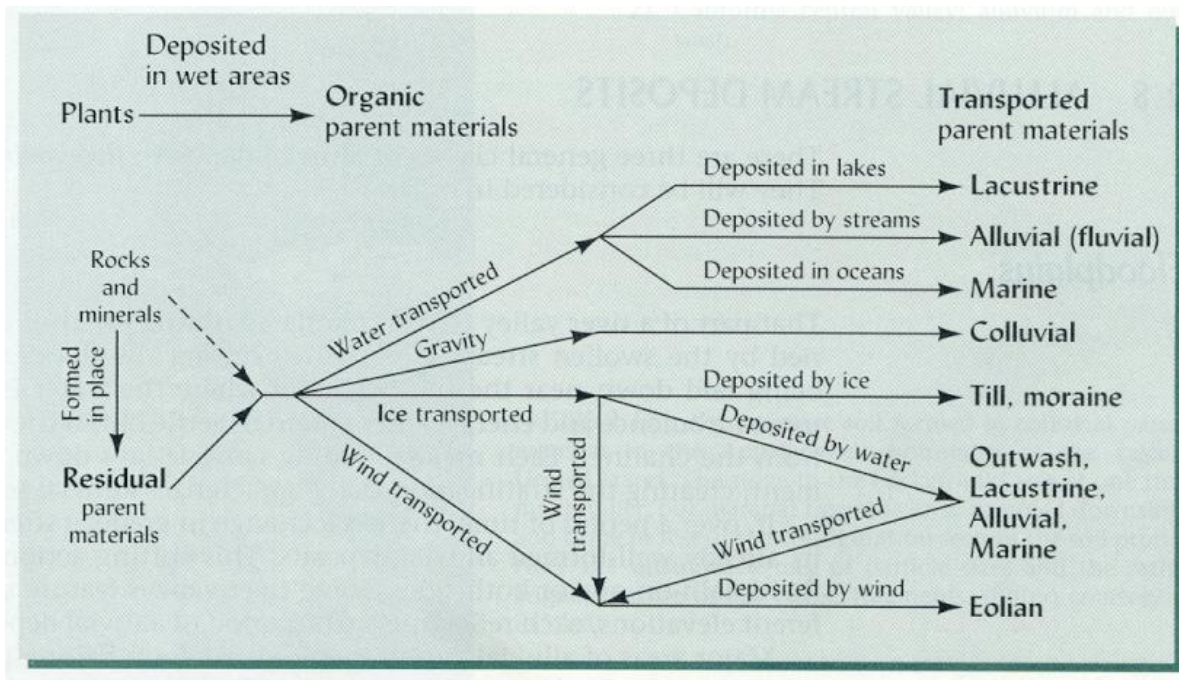
Figure 2.2 Classification of Parent Materials: How various kinds of parent materials are formed, transported, and deposited.

RESIDUAL PARENT MATERIAL: **Residual parent material** develops in place, from weathering of the underlying rock. In stable landscapes it may have experienced long and possibly intense weathering. Where the climate is warm and humid, residual parent materials are typically thoroughly leached and oxidized, and show the red and yellow colors of various oxidized iron compounds. In cooler and especially drier climates, the color and chemical composition of residual parent material tends to resemble more closely the rock from which it formed.

Residual materials are widely distributed on all continents. A great variety of soils occupy the regions covered by residual debris because of the marked differences in the nature of the rocks from which these materials evolved. The varied soils are also a reflection of wide differences in other soil-forming factors, such as climate and vegetation.

COLLUVIAL DEBRIS: Colluvial debris, or **colluvium**, is made up of poorly sorted rock fragments detached from the heights above and carried downslope, mostly by gravity, assisted in some cases by frost action. Rock fragment (talus)

slopes, cliff rock debris (detritus), and similar heterogeneous materials are good examples.

Colluvial parent materials are frequently coarse and stony because physical rather than chemical weathering has been dominant. Stones, gravel, and fine materials are interspersed (not layered), and the coarse fragments are rather angular.

ALLUVIAL STREAM DEPOSITS: There are three general classes of alluvial deposits: *floodplains*, *alluvial fans* and *deltas*.

Floodplains are the part of a river valley that is overflowed during floods. Sediment carried by the swollen stream is deposited during the flood, with the coarser materials being laid down near the river channel where the water is deeper and flowing with more turbulence and energy. Finer materials settle out in the calmer flood waters farther from the channel. Each major flooding episode lays down a distinctive layer of sediment, creating the stratification that characterizes alluvial soils.

To some degree, nutrient-rich materials lost by upland soils, are deposited on the river floodplain and delta. Soils derived from alluvial sediments generally have characteristics seen as desirable for human settlement and agriculture. These characteristics include nearly level topography, proximity to water, high fertility, and high productivity. Although many alluvial soils are well drained, others may require artificial drainage if they are to be used for upland crops or for stable building foundations.

Streams that leave a narrow valley in an upland area and suddenly descend to a much broader valley below, deposit sediment in the shape of a fan, as the water spreads out and slows down. The rushing water tends to sort the sediment particles by size, first dropping the gravel and coarse sand, then depositing the finer materials toward the bottom of the *alluvial fan*.

Alluvial fan debris is found in widely scattered areas in mountainous and hilly regions. The soils derived from this debris often prove very productive, although they may be quite coarse-textured.

Much of the finer sediment carried by streams is not deposited in the floodplain but is discharged into the lake, reservoir, or ocean into which the streams flow. Some of the suspended material settles near the mouth of the river, forming a *delta*. A delta is often a continuation of a floodplain. It is clayey in nature and is likely to be poorly drained as well.

Delta marshes are among the most extensive and biologically important of wetland habitats. Many of these habitats are today being protected or restored, but civilizations both ancient and modern have also developed important agricultural areas by creating drainage and flood-control systems on the deltas.

MARINE SEDIMENTS: Streams eventually deposit much of their sediment loads in oceans, estuaries, and gulfs. Deep marine clastic systems represent the planet's ultimate sediment sink. The coarser fragments settle out near the shore and the finer particles at a distance. Over long periods of time, these underwater sediments build up, in some cases becoming hundreds of meters thick. Changes in the relative elevations of sea and land may later raise these marine deposits above sea level, creating a coastal plain.

Marine deposits are quite variable in texture. Some are sandy, others are high in clay. Where streams have cut down through layers of marine sediments, clays, silts, and sand may be encountered side by side. All of these sediments came from the erosion of upland areas, some of which were highly weathered before the transport took place. However, marine sediments generally have been subjected to soil-forming processes for a shorter period of time than their upland counterparts. As a consequence, the properties of the soils that form are heavily influenced by those of the marine parent materials.

PARENT MATERIAL TRANSPORTED BY GLACIAL ICE AND MELTWATERS: During the *Pleistocene* epoch (about 10.000 to 1.000.000 years ago), up to 20% of the world's land surface – northern North America, northern and central Europe, and parts of northern Asia – was invaded by a succession of great ice sheets, some more than 1 km thick. Present day glaciers in polar regions and high

mountains cover about a third as much area, but are not nearly so thick as the glaciers of the Great Pleistocene Ice Age. Even so, if all present-day glaciers were to melt, the world sea level would rise by about 65 m. Some scientists predict that if the current global warming trend continues, these present-day glaciers could partially melt, causing an increase in sea level, thus flooding many coastal areas around the world.

Europe (and central Nord America) apparently sustained several distinct ice invasions over a period of 1 to 1.5 million years. These invasions were separated by long, *interglacial*, ice-free intervals of warm or even semitropical climates. We now may be enjoying the mildness of another interglacial period.

As the glacial ice pushed forward, the existing regolith with much of its mantle of soil was swept away, hills were rounded, valleys were filled, and, in some cases, the underlying rocks were severely ground and gouged. Thus, the glacier became filled with rock and all kinds of unconsolidated materials, carrying great masses of these materials as it pushed ahead. Finally, as the ice melted and the glacier retreated, a mantle of **glacial debris** or **drift** remained. This provided a new regolith and fresh parent material for soil formation.

The name **drift** is applied to all material of glacial origin, whether deposited by the ice or by associated waters. The materials deposited directly by the ice, called **glacial till**, are heterogeneous (unstratified) mixtures of debris, which vary in size from boulders to clay. Glacial till may therefore be somewhat similar in appearance to colluvial materials, except that the coarse fragments are more rounded from their grinding journey in the ice, and the deposits are often much more densely compacted because of the great weight of the overlying ice sheets. Much glacial till is deposited in irregular ridges called **moraines**.

The torrents of water streaming forth from melting glaciers carried vast loads of sediment. In valleys and on plains, these glacial waters formed an **outwash plain** from the sediment. In some places, where there was no escape for the water, the forming of ponds began. The **lacustrine deposits** formed in these glacial lakes range from coarse delta materials and beach deposits near the shore to larger areas of fine silts and clay deposited from the deeper, more still waters at the center of the lake.

PARENT MATERIALS TRANSPORTED BY WIND: Wind is capable of picking up an enormous quantity of material at one site and depositing it at another. Wind can most effectively pick up material from soil or regolith that is loose, dry, and unprotected by vegetation. Dry, barren landscapes have served, and continue to serve, as sources of parent material for soils forming as far away as the opposite side of the globe. The smaller the particles, the higher and farther the wind will carry them.

Wind-transported (*eolian*) materials important as parent material for soil formation include, from largest to smallest particle size: **dune sand**, **loess** (pronounced "luss"), and **aerosolic dust**. Windblown **volcanic ash** from erupting volcanoes is a special case that is also worthy of mention.

Dune sand: Along the beaches of the world's oceans and large lakes, and over vast barren deserts, strong winds pick up medium and fine sand grains and pile them into hills of sand called *dunes*. The dunes, ranging up to 100 m in height, may continue to slowly shift their locations in response to the prevailing winds. Because most other minerals have been broken down and carried away by the waves, beach sand usually consists mainly of quartz, which is devoid of plant nutrients and highly resistant to weathering action. Nonetheless, over time dune grasses and other pioneering vegetation may take root and soil formation may begin. Last reminders of dunes in Hungary you can find in the surroundings of Bugac.

Loess: The windblown materials called *loess* are composed primarily of silt with some very fine sand and coarse clay. They cover wide areas in the central United States, eastern Europe, Argentina, and central Asia. Loess may be blown for hundreds of kilometers. The deposits farthest from the source are thinnest and consist of the finest particles.

In the United States and in eastern Europe, the main source of loess were the great barren expanses of till and outwash left back by the retreating glaciers of the last Ice Age. During the winter months, winds picked up fine materials and moved them southward, covering the existing soils and parent materials with a blanket of loess that in some places of Hungary accumulated to as much as 30 m thick. From the coarser materials, namely sand, which stayed closer to their original place, dunes were formed.

While there were periodically warmer (interglacial) periods during the Ice Age too, in these warmer and more humid times, the plant cover was richer, so that former grasslands were replaced by forests or woody steppes. During these periods the formation of loess slowed down or came to an end, and on the formerly deposited loess soil formation began. Therefore, at the edge of the Hungarian Great Plane thick layers of loess and soil formations of different age follow upon each other.

In central and western China, loess deposits reaching 30 to 300 m in depth cover some 800.000 km^2. These materials have been windblown from the deserts of central Asia and are generally not associated directly with glaciers. These and other loess deposits tend to form silty soils or rather high fertility and potential productivity.

Aerosolic Dust: Very fine particles (about 1 to 10 μm) carried high into the air may travel fro thousands of kilometres before being deposited, usually with rainfall. These fine particles are called *aerosolic dust* because they can remain suspended in air, due to their very small size. While this dust has not blanketed the receiving landscapes as thickly as is typical for loess, it does accumulate at rates that make significant contributions to soil formation. Recent studies have shown that dust, originating in the Sahara Desert of northern Africa and transported over the Atlantic Ocean in the upper atmosphere, is the source of much of the calcium and other nutrient found in the highly leached soils of the Amazon basin in South America.

Volcanic Ash: During volcanic eruptions cinders fall in the immediate vicinity of the volcano, while fine, often glassy ash particles may blanket extensive areas downwind. Soils developed from vocanic ash are most prominent within a few dozen to hundred kilometers of volcanoes. The soils formed on volcanic ashes are uniquely light and porous, and tend to accumulate organic matter more rapidly than other soils in the area.

ORGANIC DEPOSITS: Organic material accumulates in wet places where plant growth exceeds the rate of residue decomposition. In such areas residues accumulate over the centuries from wetland plants such as pondweeds (hínár), cattails (gyékény), sedges (sás), reeds (nád), mosses (moha), shrubs, and certain trees. These residues sink into the water, where their decomposition

is limited by lack of oxygen. As a result, organic deposits often accumulate up to several meters in depth. Collectively, these organic deposits are called *peat*.

The rate of peat accumulation varies from one area to another, depending on the balance between production of plant material and its loss by decomposition. Cool climates and acidic conditions favour slow decomposition, but also slower plant production. Warm climates and alkaline conditions favour rapid losses, but also rapid plant production. Enrichment with nutrients may increase the rate of organic production more than it does the rate of decomposition, leading to very high net accumulation rates. Artificial drainage, used to remove excess water from a peat soil, lets air into the peat and drastically alters the balance between production and decomposition of organic matter. The result of such drainage is a reversal of the accumulation process and a loss or subsidence of the peat soil.

Based on the nature of the parent materials, four kinds of peat are recognized:

1. Moss peat, the remains of mosses such as sphagnum;

2. Herbaceous peat, residues of herbaceous plants such as sedges, reeds, and cattails;

3. Woody peat, from the remains of woody plants, including trees and shrubs;

4. Sedimentary peat, remains of aquatic plants (e.g., algae) and of fecal material of aquatic animals.

In the Hungarian classification of peats we speak about *moss peat* (upper peat), *plain peat* (lower peat), and of *drained peat*. Hungarian moss peats can mainly be found on soils formed on acidic parent materials, covered by moss species (e.g. *Sphagnum*), under humid climate. The formation of plain peat occurred in ponds and slow moving waters of the Hanság, Kis-Balaton, Nagyberek, Tapolcai-berek, Ecsed and Sárrét regions. In cases where a wetland area has been drained, woody and herbaceous peats tend to make very productive agricultural soils that are especially well suited for vegetable production. Note, that such use is unsustainable because, once drained, the organic deposits will decompose an disappear after a century or so; therefore, these areas might be better left in (or returned to) their natural state, in which

they can serve as important environmental buffers and natural habitats for wildlife.

While moss peats have high water-holding capacities, they tend to be quite acid. Sedimentary peat is generally undesirable as an agricultural soil.

The organic material is called **peat** or **fibric** if the residues are sufficiently intact to permit the plant fibers to be identified. If most of the material has decomposed sufficiently so that little fiber remains, the term **muck** or **sapric** is used. In mucky peats (**hemic** materials) some of the plant fibers can be recognized an some cannot.

Recognizing that the effects of parent materials on soil properties are modified by the combined influences of **climate**, **biotic activities**, **topography**, and **time**, we will now turn to these other four factors of soil formation.

2.4.2 Climate

Climate is perhaps the most influential of the four factors acting on parent material because it determines the nature and intensity of weathering that occurs over large geographic areas. The principal climatic variables influencing soil formation are *effective precipitation* and *temperature*, both of which affect the rates of chemical, physical, and biological processes.

EFFECTIVE PRECIPITATION: We have already seen that water is essential for all the major chemical weathering reactions. To be effective in soil formation, water must penetrate into the regolith. *Effective Precipitation* is the amount of precipitation that is actually added and stored in the soil. During drier periods less than 5mm of daily rainfall would not be considered effective, as this amount of precipitation would likely evaporate from the surface before soaking into the ground. Effective precipitation enters the soil and becomes available to the plant. The greater the depth of water penetration, the greater the depth of weathering and soil development. Surplus water percolating

through the soil profile transports soluble and suspended materials from the upper to the lower layers. It may also carry away soluble materials in the drainage waters. Thus, percolating water stimulates weathering reactions and helps differentiate soil horizons.

Likewise, a deficiency of water is a major factor in determining the characteristics of soils of dry regions. Soluble salts are not leached from these soils, and in some cases they build up to levels that limit plant growth. Soil profiles in arid and semiarid regions are also apt to accumulate carbonates and certain types of cracking clays.

TEMPERATURE: For every 10°C rise in temperature, the rates of biochemical reactions more than double. Temperature and moisture both influence the organic matter content of soil through their effects on the balance between plant growth and microbial decomposition. If warm temperatures and abundant water are present in the profile at the same time, the processes of weathering, leaching, and plant growth will be maximized. The very modest profile development characteristics of cold areas contrasts sharply with the deeply weathered profiles of the humid tropics.

Climate also influences the natural vegetation. Humid climates favour the growth of trees. In contrast, grasses are the dominant native vegetation in subhumid and semiarid regions, while shrubs and brush of various kinds dominate in arid areas. Thus, climate exerts its influence partly through a second soil-forming factor, the living organisms.

Considering soils with similar temperature regime, parent material, topography, and age, increasing effective annual precipitation generally leads to increasing clay and organic matter contents, greater acidity, and lower ratio of Si/Al (an indication of more highly weathered minerals). However, many places have experienced climates in past geologic epochs that were not at all similar to the climate evident today. This fact is illustrated in certain old landscapes in arid regions, where highly leached and weathered soils stand as relicts of the humid tropical climate that prevailed there many thousands of years ago.

2.4.3 Living Organisms

Organic matter accumulation, biochemical weathering, profile mixing, nutrient cycling, and aggregate stability are all enhanced by the activities of organisms in the soil. Vegetative cover reduces natural soil erosion rates, thereby slowing down the rate of mineral surface soil removal. Organic acids produced from certain types of plant leaf litter bring iron and aluminium into solution by complexation and accelerate the downward movement of these metals and their accumulation in the B horizon.

Role of Natural Vegetation:

ORGANIC MATTER ACCUMULATION: The effect of vegetation on soil formation can be seen by comparing properties of soils formed under grassland and forest vegetation near the boundary between these two ecosystems. In the grassland, much of the organic matter added to the soil is from the deep fibrous grass root systems. By contrast, tree leaves falling on the forest floor are the principal source of soil organic matter in the forest. Further differences are the frequent occurrence of fires in grassland destroying large amounts of aboveground organic matter, and the much greater acidity under many forests which inhibits the action of certain soil organisms that otherwise would mix much of the surface litter into the mineral soil. As a result, the soils under grasslands generally develop a thicker A horizon with a deeper distribution of organic matter than in comparable soils under forests, which characteristically store most of their organic matter in the forest floor (O horizons) and a thin A horizon.

The microbial community in a typical grassland soil is dominated by bacteria, while that of the forest soil is dominated by fungi. Differences in microbial action affect the aggregation of the mineral particles into stable granules and the rate of nutrient cycling. The light-coloured, leached E horizon typically found under the O or A horizon of a forested soil results from the action of organic acids generated mainly by fungi in the acidic forest litter. An E horizon is generally not found in a grassland soil.

NUTRIENT CYCLING BY TREES: The ability of natural vegetation to accelerate the release of nutrient elements from minerals by biogeochemical weathering, and

to take up these elements from the soil, strongly influences the characteristics of the soils that develop.

Soil acidity is especially affected. Differencies occur not only between grassland and forest vegetation, but also between different species of forest trees. Litter falling from coniferous tress (e.g., pines, firs, spruces, and hemlocks) will recycle only small quantities of calcium, magnesium, and potassium compared to those recycled by litter from some deciduous trees (e.g., beech, oaks, and maples) that take up and store much larger amounts of these cations. Conifer tree roots take up less Ca, Mg, and K from mineral weathering deep in the profile and allow more of these nonacidic cations to be lost by leaching. Therefore, soil acidity often develops more strongly in the surface horizons under coniferous vegetation than under most deciduous trees. Furthermore, the acidic, residous needles from conifer trees resist decomposition and discourage with distinctly separate layers of fibric (undecomposed) and sapric (highly decomposed) material. The leaves of deciduous trees are generally more readily broken down and are usually more inviting to earthworms. Thus, the deciduous trees generally form a thinner forest floor with less distinction between layers and with more litter mixed into the A horizon.

As the deciduous leaves decompose, they release large amounts of Ca, Mg, and K that displace acid cations from the adsorption sites on the clay and humus particles in the surface horizons. The more efficient uptake of cations by roots of the deciduous trees may accelerate mineral weathering and acidification in the deeper horizons. In a further chapter we will see, that nutrient cycling by plant roots can even alter the types of clay minerals in a soil.

Role of Organisms

The role of animals in soil-formation processes must not be overlooked. Large animals, such as moles, voles, gophers and prairie dogs bore into the lower soil horizons, bringing materials to the surface. Their tunnels are often open to the surface, encouraging movement of water and air into the subsurface layers. In localized areas, they enhance mixing of the lower and upper horizons by creating, and later refilling, underground tunnels. For example, dense

populations of prairie dogs may completely turn over the upper meter of soil in the course of several thousand years. Old animal burrows in the lower horizons often become filled with soil material from the overlying A horizon, creating profile features known as *crotovinas*.

EARTHWORMS AND TERMITES: Earthworms, termites, and other small animals bring about considerable soil mixing as they burrow through the soil, significantly affecting soil formation. Earthworms ingest soil particles and organic residues, enhancing the availability of plant nutrients in the material that passes through their bodies. They aerate and stir the soil and increase the stability of soil aggregates, thereby assuring ready infiltration of water. Ants and termites, as they build mounds, also transport soil materials from one horizon to another. In general, the mixing activities of animals, sometimes called *pedoturbation*, tends to undo or counteract the tendency of other soil-forming processes to accentuate the differences among soil horizons.

HUMAN INFLUENCES: Human activities also influence soil formation. For example, it is believed, that Native Americans regularly set fires to maintain several large areas of prairie grasslands in Indiana and Michigan. In contrast, nowadays the suppression of forest fires in Canada ends up in the stronger accumulation of organic matter in the forest floor, causing much more devastating fires with much higher temperatures, which seem to effect the structure of surface soils. This again leads to more frequent landslides often destroying human settlements.

Human destruction of natural vegetation (trees and grass) and subsequent tillage of the soil for crop production has in recent times abruptly modified soil formation. Likewise, irrigating an arid region soil drastically influences the soil-forming factors, as does adding fertilizer and lime to soils of low fertility. In surface mining and urbanizing areas today, bulldozers may have an effect on soils almost identical to that of the ancient glaciers; they level and mix soil horizons and set the clock of soil formation back to zero.

2.4.4 Topography

Topography relates to the configuration of the land surface and is described in terms of differences in elevation, slope, and landscape position – on other

words, the lay of the land. The topographical setting may either hasten or retard the work of climatic forces. Steep slopes generally encourage rapid soil loss by erosion and allow less rainfall to enter the soil before running off. In semiarid regions, the lower effective rainfall on steeper slopes also results in less complete vegetative cover, so there is less plant contribution to soil formation. For all of these reasons, steep slopes prevent the formation of soil from getting very far ahead of soil destruction. Therefore, soils on steep terrain tend to have rather shallow, poorly developed profiles in comparison to soils of nearby, more level sites (Figure 2.3).
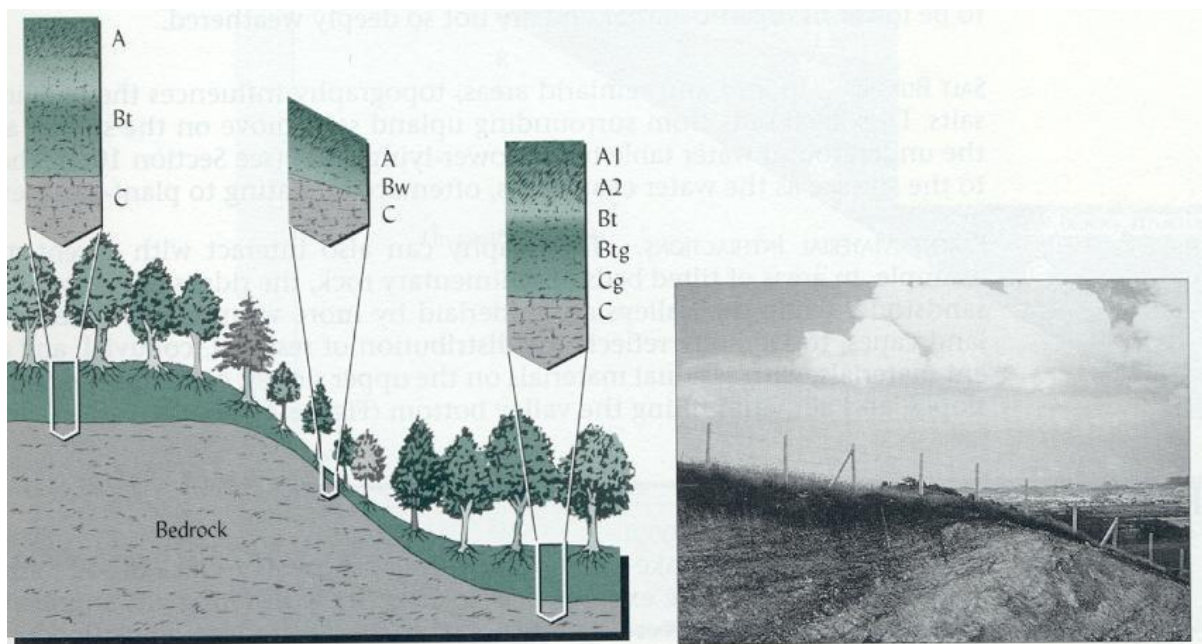
Figure 2.3 Topography influences soil properties, including soil depth. The diagram on the left shows the effect of slope on the profile characteristics and the depth of a soil on which forest trees are the natural vegetation. The photo on the right illustrates the same principle under grassland vegetation. Often a relatively small change in slope can have a great effect on soil development.

In swales and depressions where runoff water tends to concentrate, the regolith is usually more deeply weathered and soil profile development is more advanced. However, in the lowest landscape positions, water may saturate the regolith to such a degree that drainage and aeration are restricted. Here the weathering of some minerals and the decomposition of organic matter are retarded, while the loss of iron and manganese is

accelerated. In such low-lying topography, special profile features characteristic of wetland soils may develop.

Soils commonly occur together in groups, with each member of the group occupying a characteristic topographic position. When each of the soils occurs on similar parent material but in a different sequential position from the top to the bottom of a slope, the group of soils is called a *catena*.

INTERACTION WITH VEGETATION: Topography often interacts with vegetation to influence soil formation. In grassland-forestry transition zones, trees are commonly confined to the slight depressions where soil is generally wetter than in upland positions. If water stands for part or all of the year in a depression in the landscape, climate is less influential in regulating soil development. Low-lying areas may so give rise to peat bogs and, in turn, to organic soils.

SLOPE ASPECT: Topography affects the absorbance of solar energy in a given landscape. In the northern hemisphere, south-facing slopes are more perpendicular to the sun's rays and are generally warmer and thereby commonly lower in moisture than their north-facing counterparts. Consequently, soils on the south slopes tend to be lower in organic matter and are not so deeply weathered.

PARENT MATERIAL INTERACTIONS: Topography also can interact with parent material. For example, in areas of tilted beds of sedimentary rock, the ridges often consist of resistant sandstone, while the valleys are underlaid by more weatherable limestone. In many landscapes, topography reflects the distribution of residual, colluvial, and alluvial parent materials, with residual materials on the upper slopes, colluvium covering the lower slopes, and alluvium filling the valley bottom.

2.4.5 Time

Soil-forming processes take time to show their effects. The clock of soil formation starts ticking when a landslide exposes new rock to weathering environment at the surface, when a flooding river deposits a new layer of sediment on its floodplain, when a glacier melts and dumps its load of mineral

debris, or when a bulldozer cuts and fills a landscape to level a construction or mine-reclamation site.

RATES OF WEATHERING: When we speak of a "young" or a "mature" soil, we are not so much referring to the age of the soil in years, as to the degree of weathering and profile development. Time interacts with the other factors of soil formation. For example, on a level site in a warm climate, with much rain falling on permeable parent material rich in reactive minerals, weathering and soil profile differentiation will proceed far more rapidly than on a site with steep slopes and resistant parent material in a cold, dry climate.

EXAMPLES OF SOIL GENESIS OVER TIME: Figure 2.4 illustrates changes that typically take place during soil development on residual rock in warm, humid climate. During the first 100 years, lichens and mosses establish themselves on the bare exposed rock and begin to accelerate its breakdown and the accumulation of dust and organic matter. Within a few hundred years, grasses, shrubs and stunted trees have taken root in a deepening layer of disintegrated rock and soil, adding greatly to the accumulation of organic materials and to the formation of the A and C horizons. During the next 10.000 years or so, successions of forest trees establish themselves and the activities of a multitude of tiny soil organisms transform the surface plant litter into a distinct O horizon. The A horizon thickens somewhat, becomes darker in colour, and develops a stable granular structure. Soon, a bleached zone appears just below the A horizon as soluble weathering products, and iron oxides, and clays are moved down with the water and organic acids, percolating down from the litter layer. These transported materials begin to accumulate in a deeper layer, forming a B horizon. The process continues, with more silicate clay accumulating and blocky structure forming as the B horizon thickens and becomes more distinct. Eventually, the silicate clays themselves break down, some silica is leached away, and new clays containing less silica form in the B horizon. These clays often become mixed or coated with oxides of iron and aluminium, causing the B horizon to take on a reddish hue. As the entire profile continues to deepen over time, the zone of weathered, unconsolidated rock may become many meters thick.
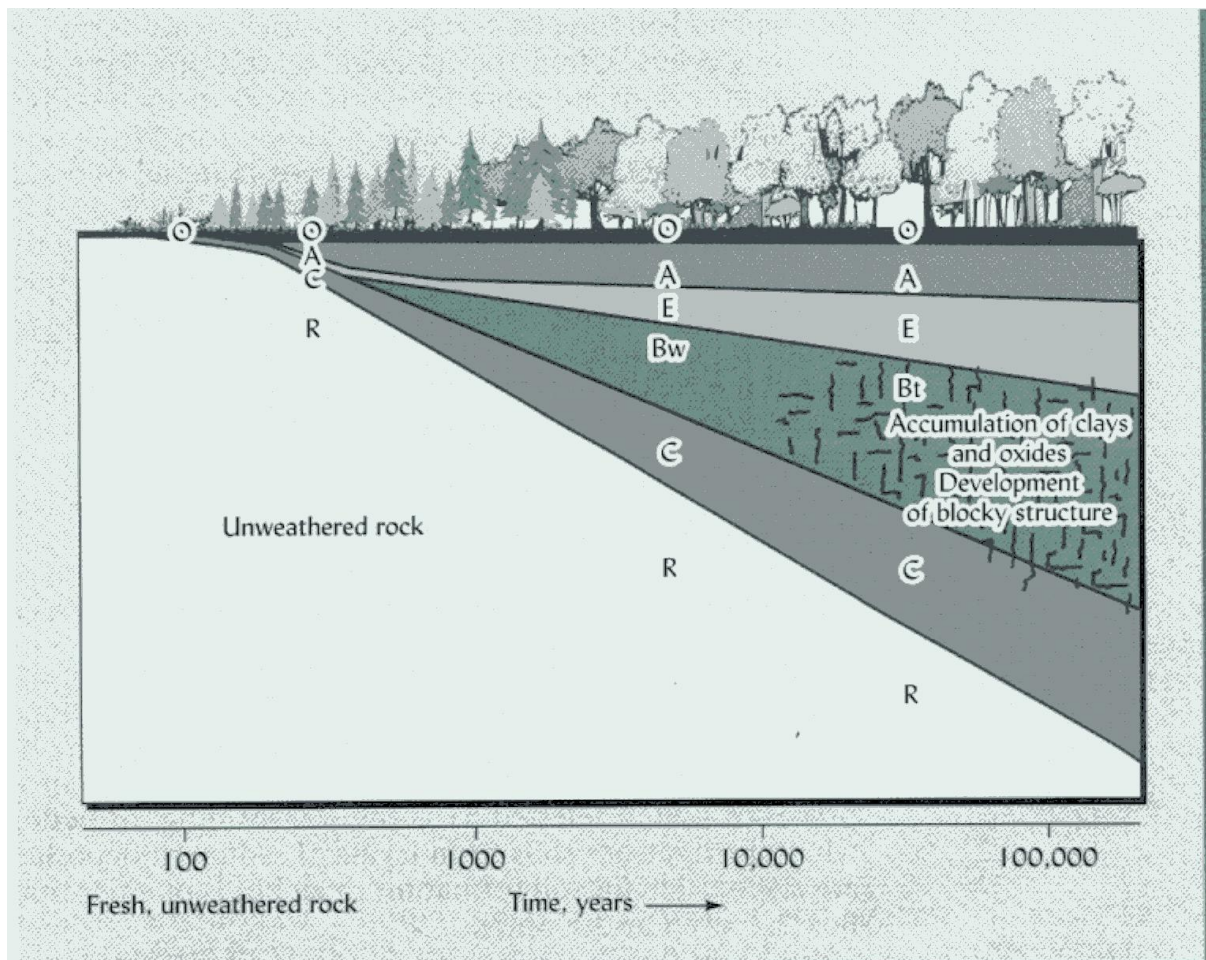
Figure 2.4 Progressive stages of soil profile development over time for a residual igneous rock, in a warm, humid climate that is conducive (favourable) to forest vegetation. The time scale increases logarithmically from left to right, covering more than 100.000 years.

CHRONOSEQUENCE: Most soil-profile features develop so slowly that it is not possible to directly measure time-related changes in their formation. Indirect methods, such as carbon dating or the presence of fossils and human artifacts, must be turned to for evidence about the time required for soil development to occur.

In a different approach to studying the effects of time on soil development, soil scientists look for a *chronosequence* – a set of soils that share a common community of organisms, climate, parent material, and topography, but differ with regard to the length of time that the materials have been subjected to weathering and soil formation. A chronosequence can sometimes be found among the soils forming on alluvial terraces of differing ages.

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The simultaneous and interdependent influences of the five factors of soil formation present many challenges to those who would like to understand just how a given soil was formed or predict what soil properties are likely to be encountered in a given environment. With this in mind, we will now focus on the processes that change parent material into a soil, processes that operate under the influence of the five major factors of soil formation we have just considered.

2.5 Four Basic Processes of Soil Formation

The accumulation of regolith from the breakdown of bedrock, or the deposition (by wind, water, ice, etc.) of unconsolidated geologic materials may precede or, more commonly, occur simultaneously with the development of the distinctive horizons of a soil profile. During the formation (*genesis*) of a soil from parent material, the regolith undergoes many profound changes. These changes are brought about by variations in the four broad soil-forming processes considered next. These four basic processes – often referred to as the soil-forming, or pedogenic processes – help define what distinguishes soils from layers of sediment deposited by geologic processes. They are responsible for soil formation in all kinds of environment.

TRANSFORMATIONS: *Transformations* occur when soil constituents are chemically or physically modified or destroyed and others are synthesized from the precursor materials. Many transformations involve weathering of primary minerals, disintegrating and altering some to form various kinds of silicate clays. As other primary minerals decompose, the decomposition products recombine into new minerals that include additional types of silicate clays and hydrous oxides of iron and aluminium. Other important transformations involve the decomposition of organic residues and the synthesis of organic acids, humus and other products. Still other transformations change the size (e.g., physical weathering to smaller particles) or arrangement (e.g., aggregation) of mineral particles.

TRANSLOCATIONS: *Translocations* involve the movement of inorganic and organic materials laterally within a horizon or vertically from one horizon up

or down to another. Water, either percolation down with gravity or rising up by capillary action, is the most common translocation agent. The materials moved within the profile include dispersed fine clay particles, dissolved salts, and dissolved organic substances. Translocations of materials by soil organisms also have a major influence on soil genesis. Important examples include incorporation of surface organic litter into the A and B horizons by certain earthworms, transport of B and C horizon material to the surface by mound-builing termites, and the widespread burrowing actions of rodents.

ADDITIONS: Inputs of materials to the developing soil profile from outside sources are considered *additions*. A very common example is the input of organic matter from fallen plant leaves and sloughed-off roots (the carbon having originated in the atmosphere). Another ubiquitous addition is dust particles falling on the surface of the soil. Still another example, common in arid regions, is the addition of salts or silica dissolved in the groundwater and deposited near or at the soil surface when the rising water evaporates.

LOSSES: Materials are lost from the soil profile by leaching to groundwater, erosion of surface materials, or other forms of removal. Evaporation and plant use cause losses of water. Leaching and drainage cause the loss of water, dissolved substances such as salts or silica weathered from parent minerals, or organic acids produced by microorganisms or plant roots. Erosion, a major loss agent, often removes the finer particles (humus, clay and silt), leaving the surface horizon relatively sandier and less rich in organic matter than before. Organic material is also lost by microbial decomposition. Grazing by animals or harvest by people can remove large amounts of both organic matter and nutrient elements. Of course, animals and people can also contribute additions, such as manure and fertilizers.

These processes of soil genesis, operating under the influence of the environmental factors discussed previously, give us a logical framework for understanding the relationships between particular soils and the landscapes and ecosystems in which they function. In analyzing these relationships for a given site, ask yourself: What are the materials being added to this soil? What transformations and translocations are taking place in this profile? What materials are being removed? And how have the climate, organisms,

topography, and parent material at this site affected these processes over time?

A SIMPLIFIED **E**XAMPLE OF **S**OIL-**F**ORMING **P**ROCESSES **I**N **A**CTION: Consider the changes that might take place as a soil develops from a thick layer of relatively uniform loess parent material in a climate conducive to grass vegetation (Figure 2.4), as often to be found in Hungary.

Although some physical weathering and leaching of carbonates and salts may be necessary to allow plants to grow in certain parent materials, soil formation really gets started when plants become established and begin to provide *additions* of litter and root residues on and in the surface layers of the parent material. The plant residues are transformed by soil organisms into humus and other new organic substances. The accumulation of humus enhances the capacity of the soil to hold water and nutrients, providing a positive feedback for accelerated plant growth and further humus buildup. Earthworms, ants, termites, and a host of smaller animals come to live in the soil and feed on the newly accumulating organic resources. In so doing, they accelerate the organic *transformations*, as well causing the *translocation* of plant residues, loosening the mineral material as they burrow into the soil.
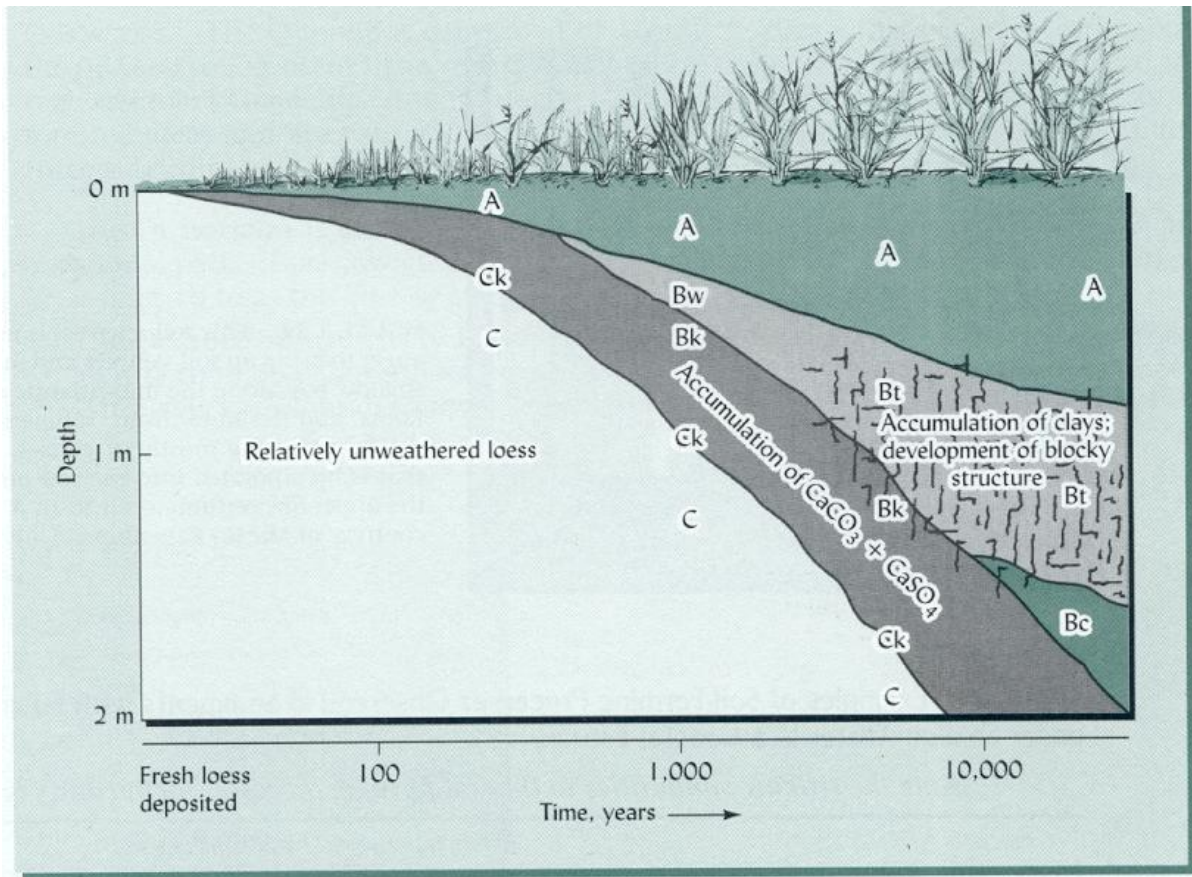
Figure 2.4 Development of a hypothetical soil in about 2 m of uniform calcareous loess deposits, where the warm subhumid climate is conducive to tall grass prairie vegetation. The time scale increases logarithmically from left to right, covering about 12.000 years. In the initial stages, rainwater, charged with organic acids from microbial respiration, dissolves carbonates from the loess and moves them downward to a zone of accumulation (Bk horizon). As this happens, plant roots take hold in the upper layer and add the organic matter necessary to create an A horizon. Ants, beetles, earthworms, and a host of smaller creatures take up residence and actively mix in surface litter and speed the release of nutrients from the minerals and plant residues. Over time the carbonate concentration zone moves deeper, the A horizon thickens, and noncalcareous B horizons develop as change in color and structure occur in the weathering loess above the zone of carbonate accumulation. Eventually, silicate clay accumulates in the B horizon (giving it the designation Bt), both by stationary weathering of primary minerals and by travelling there with water percolating from the upper horizons.

A-HORIZON DEVELOPMENT: The resulting organic-mineral mixture near the soil surface, which comes into being rather quickly, is commonly the first soil horizon developed, the A horizon. It is darker in color and its chemical and physical properties differ from those of the original parent material. Individual soil particles in this horizon commonly clump together under the influence of

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organic substances to form granules, differentiating this layer from the deeper layers and from the original parent material. On sloping land, erosion may remove some materials from the newly forming upper horizon, retarding, somewhat, the progress of horizon development.

FORMATION OF B- AND C-HORIZONS: Carbonic and other organic acids are carried by percolating waters into the soil, where they stimulate weathering reactions. The acid-charged percolating water dissolves various minerals (a *transformation*) and leaches the soluble products (a *translocation*) from upper to lower horizons, where they may precipitate. This combination of transformation and translocation create *illuviation* zones of depletion in the upper layers and zones of **accumulation** in the lower. The dissolved substances include both positively charged ions (cations; e.g., Ca^{2+}) and negatively charged ions (anions; e.g., CO_3^- and SO_4^{2-}) released from the breakdown of minerals and organic matter. In semiarid and arid regions, precipitation of these ions produces horizons enriched in calcite ($CaCO_3$) or gypsum ($CaSO_4 \cdot 2H_2O$), designated as a Bk layer in Figure 2.4.

Over time, the leached surface layer thickens, and the zone of Ca accumulation is moved deeper down to the maximum depth of water penetration. Where rainfall is great enough to cause significant drainage to the groundwater, some of the dissolved materials may be completely removed from the developing soil profile (*losses*), and the zone of accumulation may move below the reach of plant roots, or be dissipated altogether. On the other hand, deep-growing plant roots may intercept some of these soluble weathering products and return them, through leaf- and litter-fall, to the soil surface, thus retarding somewhat the processes of acid weathering and horizon differentiation.

The weathering of primary minerals into clay minerals becomes evident only long after the dissolution and movement of Ca is well underway. The newly formed clay minerals may accumulate where they are formed, or they may move downward and accumulate deeper in the profile. As clay is removed from one layer and accumulates in another, adjacent layers become more distinct from each other, and a Bt horizon (one enriched in silicate clay) is formed. When the accumulated clay in the Bt horizon periodically dries out and cracks, blocklike or prismatic units of soil structure begin to develop. As the soil matures, the various horizons within the profile generally become more numerous and more distinctly different from each other.

2.6 The Soil Profile

At each location on the land, the earth's surface has experienced a particular combination of influences from the five soil-forming factors, causing a different set of layers (horizons) to form in each part of the landscape, thus slowly giving rise to the natural bodies we call *soils*. Each soil is characterized by a given sequence of these horizons. A vertical exposure of this sequence is termed a *soil profile*. We will now consider the major horizons making up soil profiles and the terminology used to describe them.

2.6.1 The Master Horizons and Layers

Five **master horizons** are recognized and are designated using the capital letters O, A, E, B, and C. **Subordinate horizons** may occur within some master horizons and these are designated by lowercase letters following the capital master horizon letters (e.g., Bt, Ap, or Oi).

O-HORIZONS: The O group is comprised of organic horizons that generally form <u>above</u> the mineral soil or occur in an organic soil profile. They derive from dead plant and animal residues. Generally absent in grassland regions, O horizons usually occur in forested areas and are commonly referred to as the forest floor.

Often three subordinate O horizons can be distinguished.

The **Oi horizon** is an organic horizon of **fibric** materials – recognizable plant and animal parts (leaves, twigs, and needles), only slightly decomposed. The Oi horizon is referred to as the *litter* or **L layer** by foresters in some parts of Europe.

The **Oe horizon** consists of **hemic** materials – finely fragmented residues intermediately decomposed, but still with much fiber evident when rubbed between the fingers. This layer corresponds to the *fermentation* or **F layer** described by some foresters.

The **Oa horizon** consists **sapric** materials – highly decomposed, smooth, amorphous residues that do not retain much fiber or recognizable tissue

structures. This is the humified or **H layer** designated by foresters in some parts of Europe.

A-HORIZONS: The topmost mineral horizons, designated A horizons, generally contain enough partially decomposed (humified) organic matter to give the soil a color darker than that of the lower horizons. The A horizons are often coarser in texture, having lost some of the finer materials by translocation to lower horizons and by erosion.

E-HORIZONS: These are zones of maximum leaching or *eluviation* (from Latin *ex* or *e*, out, and *lavere*, to wash) of clay, iron, and aluminium oxides, which leaves a concentration of resistant minerals, such as quartz, in the sand and silt sizes. An E horizon is usually found underneath the A horizon and is generally lighter in color than either the A horizon above it or the horizon below. Such E horizons are quite common in soils developed under forests, but they rarely occur in soils developed under grassland.

B-HORIZONS: B horizons form below an O, A, or E horizon and have undergone sufficient changes during soil genesis so that the original parent material structure is no longer recognizable. In many B horizons materials have accumulated, typically by *illuviation* (from the Latin *il*, in, and *lavere*, to wash) from the horizons above. In humid regions, B horizons are the layers of maximum accumulation of materials such as iron and aluminium oxides (Bo or Bs horizons) and silicate clays (Bt horizons), some of which may have illuviated from upper horizons and some of which may have formed in place. In arid and semiarid regions, calcium carbonate or calcium sulphate may accumulate in the B horizon (giving Bk and By horizons, respectively).

C-HORIZON: The C horizon is the unconsolidated material underlying the solum (A and B horizons). It may or may not be the same as the parent material from which the solum formed. The C horizon is below the zones of greatest biological activity and has not been sufficiently altered by soil genesis to qualify as a B horizon. While loose enough to be dug with a shovel, C horizon material often retains some of the structural features of the parent rock or geologic deposits from which it formed. Its upper layers may in time become a part of the solum as weathering and erosion continue.

R LAYERS: These are consolidated rock, with little evidence of weathering.

2.6.2 Subdivisions within Master Horizons

Often distinctive layers exist within a given master horizon, and these are indicated by a numeral following the letter designation. For example, if three different combinations of structure and colors can be seen in the B horizon, then the profile may include a B1-B2-B3 sequence.

If two different geologic parent materials (e.g., loess over glacial till) are present within the soil profile, the numeral 2 is placed in front of the master horizon symbols for horizons developed in the second layer of parent material. For example, a soil would have a sequence of horizons designated O-A-B-2C if the C horizon developed in glacial till while the upper horizons developed in loess.

2.6.3 Transition Horizons

Transitional layers between the master horizons (O, A, E, B and C) may be dominated by properties of one horizon but also have prominent characteristics of another. The two applicable capital letters are used to designate the transitional horizons (e.g., AE, EB, BE and BC). The master horizon symbol that is give first designates the kind of horizon whose properties dominate the transitional horizon.

In some cases, transitional horizons have <u>distinct</u> parts with recognizable properties of two kinds of master horizons: in such a case these are indicated as above, but the two capital letters are separated by a virgule (/), such as E/B, B/E, B/C or C/R. Commonly, most of the individual parts of one of the components are surrounded by the other.

2.6.4 Subordinate Characteristics Within Master Horizons and Layers

Designations of subordinate distinctions and features within the master horizons and layers are based on profile characteristics observable in the field and are applied during the description of the soil at the site. Lower case letters are used as suffixes to designate specific kinds of master horizons and layers,

and other features. For example, three types of O horizons (Oi, Oe, and Oa) mentioned above are commonly encountered sequences of organic horizons. Other subordinate distinctions include special physical properties and the accumulation of particular materials, such as clays and salts. A list of the recognized subordinate letter designations and their meanings (by FAO) is given in Table 2.2.

Table 2.2 Lowercase letter symbols to designate subordinate distinctions within master horizons

Letter	Distinction	Letter	Distinction
A	Organic matter, highly decomposed	n	Accumulation of exchangeable sodium (Na) (pedogenetic)
B	Buried genetic soil horizon	o	Accumulation of Fe and Al (pedogenetic)
C	Concretion or nodules	p	Ploughing or other human disturbance
D	Dense layer (physically root restrictive)	q	Accumulation of pedogenetic silica
E	Organic material, intermediate decomposition	r	Strong reduction
F	Frozen soil	s	Illuvial accumulation of sesquioxides (Fe, Al) and organic matter
G	Stagnic conditions (gleying, mottling)	ss	Slickenslides (shiny clay wedges)
H	Accumulation of organic matter	t	Accumulation of silicate clays (illuvial)
I	Organic matter, slightly decomposed	v	Occurrence of plinthit (high iron, red material)
J	Jarosite accumulation (yellow sulphur mineral)	w	Development of distinctive colour or structure
Jj	Cryoturbation (frost mixing)	x	Fragipan (high bulk density, brittle)
K	Accumulation of carbonates (pedogenetic)	y	Accumulation of gypsum (pedogenetic)

| M | Cementation or induration (pedogenetic, massive) | z | Accumulation of soluble salts |

We suggest that you mark this table for future reference, and study it now to get an idea of the distinctive soil properties that can be indicated by horizon designations. In an example, a Bt horizon is a B horizon characterized by clay accumulation (t from German *Ton*, meaning clay); likewise, in a Bk horizon, carbonates (k) have accumulated pedogenetically.

It is not likely that the profile of any one soil will show all of the master horizons as indicated above. For example, in some cultivated land, serious erosion produces a **truncated profile**. As the surface soil is swept away over the years, the plough reaches deeper and deeper into the profile. Hence, the ploughed zone in many cases consists almost entirely of former B horizon material, and the C horizon is correspondingly near the surface. Comparison to a nearby non-eroded site can show how much erosion has occurred.

2.6.5 Soil Genesis in Nature

Not every contrasting layer of material found in soil profiles is a **genetic horizon** that developed as a result of the processes of soil genesis, such as those described. The parent materials from which many soils develop contained contrasting layers before soil genesis started. For example, such parent materials as glacial outwash, marine deposits, or recent alluvium may consist of various layers of fine and coarse particles laid down by separate episodes of sedimentation. Consequently, in characterizing soils, we must recognize not only the genetic horizons and properties that come into being during soil genesis, but also those layers or properties that may have been inherited from the parent material.

Study Questions?

1. What is meant by the statement, weathering *combines the processes of destruction and synthesis*? Give an example of these two processes in the weathering of a primary mineral?

2. How is water involved in the main types of chemical weathering reactions?

3. Explain the significance of the ratio of silicon to aluminium in soil minerals?

4. Give an example of how parent material may vary across large geographic regions on one hand, but may also vary within a small parcel of land, on the other.

5. Name the five main factors affecting soil formation. With regard to each of these factors of soil formation, compare a forested North Hungarian Mountains slope to the semiarid Hungarian Great Plain below.

6. How do *colluvium*, *glacial till*, and *alluvium* differ in appearance and agency of transport?

7. What is loess, and what are some of its properties as a parent material?

8. Give two specific examples for each of the four broad processes of soil formation.

9. Assuming a level area of granite rock was the parent material in both cases, describe in general terms how you would expect two soil profiles to differ, one in a warm, semiarid grassland and the other in a cool, humid pine forest.

10. For the two soils described in question 5, make a profile sketch using master horizon symbols and subordinate suffixes to show the approximately depths, sequence, and nature of the horizons you would expect to find in each soil.

11. Visualize a slope in the landscape near where you live. Discuss how specific soil properties would likely change along the toposequence of soils on this slope.

3. Soil Architecture and Physical Soil Properties

And when the crop grew, and was
harvested, No man had crumbled a hot
clod in his fingers
and let the earth sift past his
fingertips.

-JOHN STEINBECK, THE GRAPES OF WRATH

Soil physical properties profoundly influence how soils function in an ecosystem and how they can best be managed. Success or failure of both agricultural and engineering projects often hinges on the physical properties of the soil used. The occurrence and growth of many plant species are closely related to soil physical properties, as is the movement over and through soils of water and its dissolved nutrients and chemical pollutants.

Soil scientists use the color, texture, and other physical properties of soil horizons in classifying soil profiles and in making determinations about soil suitability for agricultural and environmental projects. Knowledge of the basic soil physical properties is not only of great practical value in itself, but will also help in understanding many aspects of soils considered in later chapters.

The physical properties considered in this chapter focus on soil solids and on the pore spaces between the solid particles. Together, soil texture and structure help determine the ability of the soil to hold and conduct the water and air necessary for sustaining life. These factors also determine how soils behave when used for highways and building construction and foundations, or when manipulated by tillage. In fact, through their influence on water movement through and off soils, soil physical properties also exert considerable control over the destruction of the soil itself by erosion.

3.1 SOIL COLOR

Soil colors have little effect on the behavior and use of soils. However, they do provide valuable clues to the nature of other soil properties and conditions. Because of the importance of color in soil classification and interpretation, a standard system for accurate color description has been developed using

Munsell color charts. In this system, a small piece of soil is compared to standard color chips in a soil color book. Each color chip is described by the three components of color: the **hue** (in soils, usually redness or yellowness), the **chroma** (intensity or brightness, a chroma of 0 being neutral gray), and the **value** (lightness or darkness, a value of 0 being black).

Causes of Soil Colors

Three major factors have the greatest influence on the color of a soil: (1) its organic matter content, (2) its water content, and (3) the presence and oxidation states of iron and manganese oxides. Organic matter tends to be dark brown to black in color. As a consequence, soil horizons high in organic matter are darker in color than are those that are lower in this constituent. Organic coatings tend to darken and mask the brighter colors from compounds such as iron oxides.

Water content of a soil also influences its color, moist soils generally being darker in color than dry ones. But water levels have an even more significant and indirect effect on soil color. The quantity of soil water influences the level of soil air, and more specifically of soil oxygen. Oxygen, in turn, determines the oxidation state of several elements, and especially of iron in its oxide minerals. Red or brown colors of well-drained uplands, especially in the tropics, suggest the presence of oxidized iron oxides, in contrast to the grays and blues dominant where chemically reduced conditions exist. Other minerals that sometimes give distinctive colors are manganese oxide (black), glauconite (green), and carbonates, such as calcite, that impart a whitish color to subsoils of semiarid regions.

Interpreting Soil Colors

Colors can tell us a great deal about a soil and are therefore used extensively in classifying soils. Typically A horizons are darker and B horizons are brighter in color than adjacent horizons. Color is used as a diagnostic criterion for classifying soil. For example, a mollic epipedon is so dark that both its value and its chroma are 3 or less in the Munsell color book. All soils when wet tend to be darker in color (lower value) than when dry.

The colors of iron-containing minerals tell much about the hydrologic regime or drainage status of a soil. Bright (high-chroma) colors, indicative of the oxidized state of the iron, are typical of well-drained soils through which drainage water and air move easily. Under prolonged anaerobic conditions, iron is chemically reduced to compounds that have low-chroma (gray, bluish, or gray-green) colors, a condition referred to as gley.

3.2 SOIL TEXTURE (SIZE DISTRIBUTION OF SOIL PARTICLES)

The knowledge of the proportions of different-sized particles in a soil (i.e., the **soil texture)** is critical for understanding soil behavior and management. When investigating soils on a site, the texture of various soil horizons is often the first and most important property to determine, for a soil scientist can draw many conclusions from this information. Furthermore, the texture of a soil in the field is not readily subject to change, so it is considered a basic property of a soil.

Soil Separates

Diameters of individual soil particles range over six orders of magnitude, from boulders (1 m) to submicroscopic clays ($<10^{-6}$ m). Scientists group these particles into soil **separates** according to several classification systems, as shown in Figure 3.1. The classification established by the U.S. Department of Agriculture is used in this text. The size ranges for these separates are not purely arbitrary, but reflect major changes in how the particles behave and in the physical properties they impart to soils.

Gravels, cobbles, boulders and other **coarse fragments** greater than 2 mm in diameter may affect the behavior of a soil, but they are not considered to be part of the **fine earth fraction** to which the term *soil texture* properly applies.

SAND. Sand particles are those smaller than 2 mm but larger than 0.05 mm (2 - 0.02 mm due to the Atterberg-classification, used in Hungary). They may be rounded or angular depending on the extent to which they have been worn down by abrasive processes during soil formation. The coarsest sand particles may be rock fragments containing several minerals, but most sand grains consist of a single mineral, usually quartz (SiO_2) or other primary silicate

(Figure 3.2). The dominance of quartz means that the sand separate generally has a far smaller total content of plant nutrients than do the finer separates.
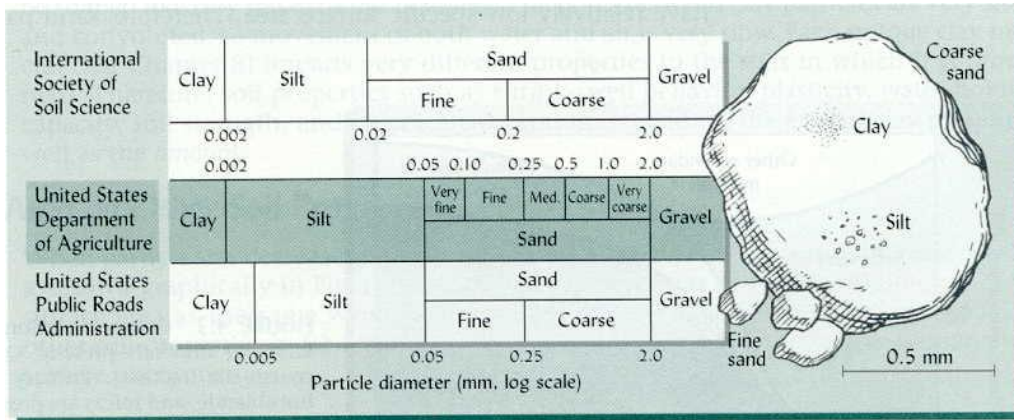

FIGURE 3.1 Classification of soil particles according to their size. The shaded scale in the center and the names on the drawings of particles follow the United States Department of Agriculture system, which is widely used throughout the world. The USDA system is also used in this book. The other two systems shown are also widely used by soil scientists and by highway construction engineers. The drawing illustrates the size of soil separates (note scale).
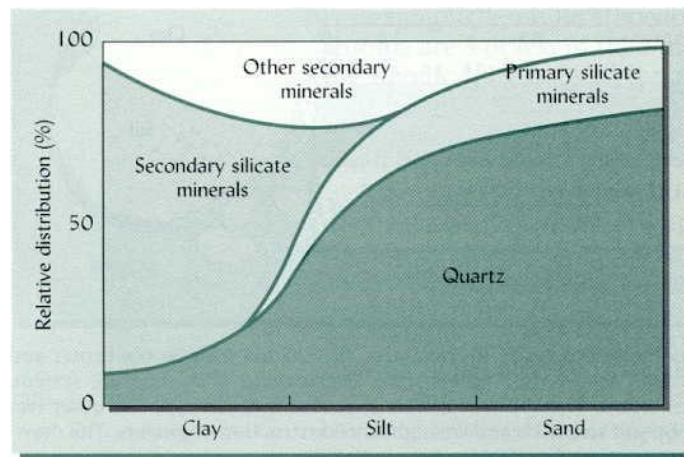
FIGURE 3.2 General relationship between particle size and kinds of minerals present. Quartz dominates the sand and coarse silt fractions. Primary silicates such as the feldspars, hornblende, and micas are present in the sands and, in decreasing amounts, in the silt fraction. Secondary silicates dominate the fine clay. Other secondary minerals, such as the oxides of iron and aluminum, are prominent in the fine silt and clay fractions.

Sand feels gritty between the fingers and the particles are generally visible to the naked eye. As sand particles are relatively large, so, too, the voids

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between them are relatively large and promote free drainage of water and entry of air into the soil. The relationship between particle size and **specific surface area** (the surface area for a given volume or mass of particles) is illustrated in Figure 3.3. Because of their large size, particles of sand have relatively low specific surface area. Therefore, sand particles can hold little water, and soils dominated by sand are prone to drought. Sand particles are considered noncohesive; that is, they do not tend to stick together in a mass.
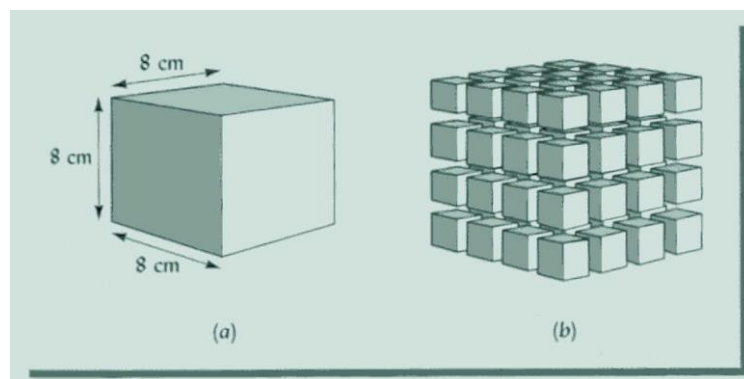
FIGURE 3.3 The relationship between the surface area of a given mass of material and the size of its particles. In the single large cube (a) each face has 64 cm² of surface area. The cube has six faces, so the cube has a total of 384 cm² surface area (6 faces x 64 cm² per face). If the same cube of material was cut into smaller cubes (b) so that each cube was only 2 cm on each side, then the same mass of material would now be present as 64 smaller cubes (4 x 4 x 4). Each face of each small cube would have 4 cm² (2 x 2 cm) of surface area, giving 24 cm² of surface area for each cube (6 faces x 4 cm² per face). The total mass would therefore have 1536 cm² (24 cm² per cube x 64 cubes) of surface area. This is four times as much surface area as the single large cube. Since clay particles are very, very small and usually platelike in shape, their surface area is thousands of times greater than that of the same mass of sand particles.

SILT. Particles smaller than 0.05 mm but larger than 0.002 mm in diameter (0.02 – 0.002 by Atterberg) are classified as silt. Individual silt particles are not visible to the unaided eye, nor do they feel gritty when rubbed between the fingers. These are essentially microsand particles, with quartz generally the dominant mineral. Where silt is composed of weatherable minerals, the smaller size of the particles allows weathering to proceed rapidly enough to release significant amounts of plant nutrients.

Although silt is composed of particles similar in shape to sand, it feels smooth or silky, like flour. The pores between silt particles are much smaller (and much more numerous) than those in sand, so silt retains more water and lets

less drain through. However, even when wet, silt itself does not exhibit much **stickiness** or plasticity (malleability). What little plasticity, cohesion (stickiness), and adsorptive capacity some silt fractions exhibit is largely due to a film of adhering clay.

CLAY. Particles smaller than 0.002 mm are classified as clay and have a very large specific surface area, giving them a tremendous capacity to adsorb water and other substances. A spoonful of clay may have a surface area the size of a football field. This large adsorptive surface causes clay particles to cohere in a hard mass after drying. When wet, clay is sticky and can be easily molded.

Clay size particles are so small that they behave as colloids-if suspended in water they do not readily settle out. Unlike most sand and silt particles, clay particles tend to be shaped like tiny flakes or flat platelets. The pores between clay particles are very small and convoluted, so movement of both water and air is very slow. Each unique clay mineral imparts very different properties to the soils in which it is prominent. Therefore, soil properties such as shrink-swell behavior, plasticity, water-holding capacity, soil strength, and chemical adsorption depend on the *kind* of clay present as well as the *amount.*

Influence of Surface Area on Other Soil Properties

When particle size decreases, specific surface area and related properties increase greatly, as shown graphically in Figure 3.4. Fine colloidal clay has about 10,000 times as much surface area as the same weight of medium-sized sand. Soil texture influences many other soil properties in far-reaching ways as a result of five fundamental surface phenomena:

1. Water is retained in soils as thin films on the surfaces of soil particles. The greater the surface area, the greater the soil's capacity for holding water.

2. Both gases and dissolved chemicals are attracted to and adsorbed by mineral particle surfaces. The greater the surface area, the greater the soil's capacity to retain nutrients and other chemicals.

3. Weathering takes place at the surface of mineral particles, releasing constituent elements into the soil solution. The greater the surface

area, the greater the rate of release of plant nutrients from weatherable minerals.

4. The surfaces of mineral particles often carry both negative and some positive electromagnetic charges so that particle surfaces and the water films between them tend to attract each other. The greater the surface area, the greater the propensity for soil particles to stick together in a coherent mass, or as discrete aggregates.

5. Microorganisms tend to grow on and colonize particle surfaces. For this and other reasons, microbial reactions in soils are greatly affected by the specific surface area.
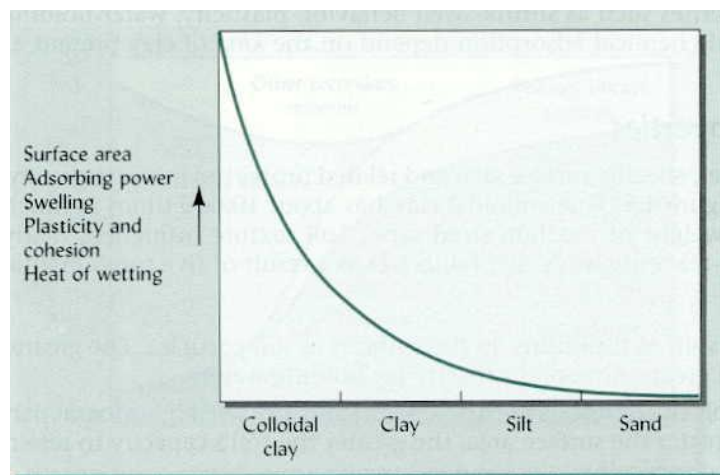
FIGURE 3.4 The finer the texture of a soil, the greater is the effective surface exposed by its particles. Note that adsorption, swelling, and the other physical properties cited follow the same general trend and that their intensities go up rapidly as the colloidal size is approached.

3.3 SOIL TEXTURE

Three broad groups of textural classes are recognized: *sandy soils, clayey soils,* and *loamy soils.* Within each group, specific **textural** class names convey an idea of the size distribution of particles and indicate the general nature of soil physical properties. The 12 textural classes named in Table 3.1 form a graduated sequence from the sands, which are coarse in texture and easy to move about, to the clays, which are very fine and difficult to handle physically.

Sands and loamy sands are dominated by the properties of sand, for the sand separate comprises at least 70% of the material by weight (less than 15% of

the material is clay). Characteristics of the clay separate are distinctly dominant in clays, sandy clays, and silty clays.

LOAMS. The loam group contains many subdivisions. An ideal **loam** may be defined as a mixture of sand, silt, and clay particles that exhibits the *properties* of those separates in about equal proportions. This definition does not mean that the three separates are present in equal *amounts* (as will be revealed by careful study of Figure 3.5). This anomaly exists because a relatively small percentage of clay is required to engender clayey properties in a soil, whereas small amounts of sand and silt have a lesser influence on how a soil behaves. Thus, the clay modifier is used in the class name of soils with as little as 20% clay, but to qualify for the modifiers *sandy* or *silt,* a soil must have at least 40 or 45% of those separates, respectively.

Most soils are some type of loam. They may possess the ideal makeup of equal proportions previously described and be classed simply as loam. However, a loam in which sand is dominant is classified as *a sandy loam.* In the same way, there may occur silt loams, silty clay loams, sandy clay loams, and clay loams.

TABLE 3.1 General Terms Used to Describe Soil Texture in Relation to the Basic Soil Textural Class Names *(U.S. Department o f Agriculture Classification System)*

Common names	General terms Texture	Basic soil textural class names
Sandy soils	Coarse	Sands Loamy sands
Loamy soils	Moderately coarse	Sandy loam Fine sandy loam[a]
	Medium	Very fine sandy loam[a] Loam Silt loam Silt
	Moderately fine	Sandy clay loam Silty clay loam Clay loam
Clayey soils	Fine	Sandy clay Silty clay Clay

[a] Although not included as class names in Figure 4.6, these soils are usually treated separately because of their fine sand content.

Alteration of Soil Textural Class

Over very long periods of time, pedologic processes such as erosion,

deposition, illuviation, and weathering can alter the textures of various soil horizons. However, management practices generally do not alter the textural class of a soil on a field scale. The texture of a given soil can be changed only by mixing it with another soil of a different textural class. For example, the incorporation of large quantities of sand to change the physical properties of a clayey soil for use in greenhouse pots or for turf would bring about such a change.

Determination of Textural Class by the " Feel" Method

Textural class determination is one of the first field skills a soil scientist should develop. Determining the textural class of a soil by its feel is of great practical value in soil survey, land classification, and any investigation in which soil texture may play a role.
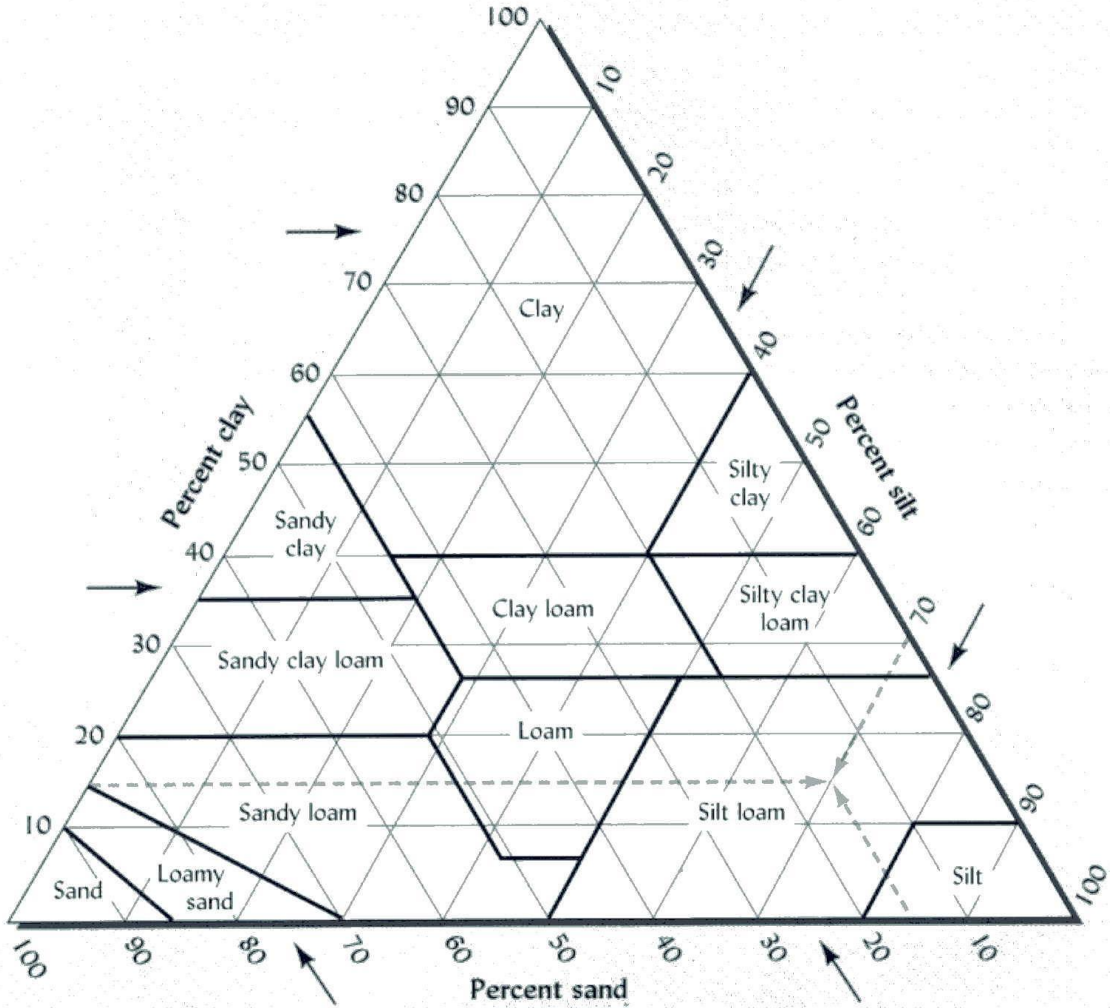
FIGURE 3.5 The major soil textural classes are defined by the percentages of sand, silt, and clay according to the heavy boundary lines shown on the textural triangle. If these percentages have been determined for a soil sample by particle size analysis, then the triangle can be used to determine the soil textural class name that applies to that soil sample. To use the graph, first find the appropriate clay percentage along the left side of the triangle, then draw a line from that location across the graph going parallel to the base of the triangle. Next find the sand percentage along the base of the triangle, then draw a line inward going parallel to the triangle side labeled "Percent silt." The small arrows indicate the proper direction in which to draw the lines. The name of the compartment in which these two lines intersect indicates the textural class of the soil sample. Percentages for any two of the three soil separates is all that is required. Because the percentages for sand, silt, and clay add up to 100%, the third percentage can easily be calculated if the other two are known. If all three percentages are used, the three lines will all intersect at the same point. Consider, as an example, a soil that has been determined to contain 15% sand, 15% clay, and 70% silt. This example is indicated by the light dashed lines that intersect in the compartment labeled "Silt loam." What is the textural class of another soil sample that has 33% sand, 33% silt, and 33% clay? The lines (not shown) for this second example would intersect in the center of the "Clay loam" compartment.

Laboratory Particle-Size Analyses

The first and sometimes most difficult step in a particle-size analysis is the complete dispersion of a soil sample in water, so even the tiniest clumps are broken down into individual, primary particles. Dispersion is usually accomplished using chemical treatments along with a high-speed blender or ultrasonic vibrator.

Separation into size groups can be accomplished by passing the suspended particles through standard sieves that are graded in size to separate the gravel and fine and coarse sand separates, permitting the silt and clay fractions to pass through. A sedimentation procedure is usually used to determine the amounts of silt and clay. The principle involved is simple. Because soil particles are more dense than water, they tend to sink, settling at a velocity that is proportional to their size. In other words, "The bigger they are, the faster they fall." The equation that describes this relationship is referred to as *Stokes' law.*

The complete expression of Stokes' law tells us the velocity V of a particle falling through a fluid is directly proportional to the gravitational force g, the difference between the density of the particle and the density of the fluid D_s - D_f) and the square of the effective particle diameter (d^2). The settling velocity is inversely proportional to the viscosity or "thickness" of the fluid η. Since velocity equals distance h divided by time t we can write Stokes' law as:

$$V = \frac{h}{t} = \frac{d^2 g \left(D_s - D_f \right)}{18\eta}$$

Where: g = gravitational force = 9.81 newtons per kilogram (9.81 N/kg)

η = viscosity of water at 20°C = 1 / 1000 newton-seconds per m² (10⁻³ Ns/m²)

D_s = density of the solid particles, *for most* soils = 2.65 x 10³ kg/m²

D_f = density of the fluid (i.e., water) = 1.0 x 10³ kg/m³

By measuring the amount of soil still in suspension after various amounts of settling time (using a pipette or a hydrometer), the percentages of each size fraction can be determined so as to identify the soil textural class and generate particle-size distribution curves such as those shown in Figure 3.6.
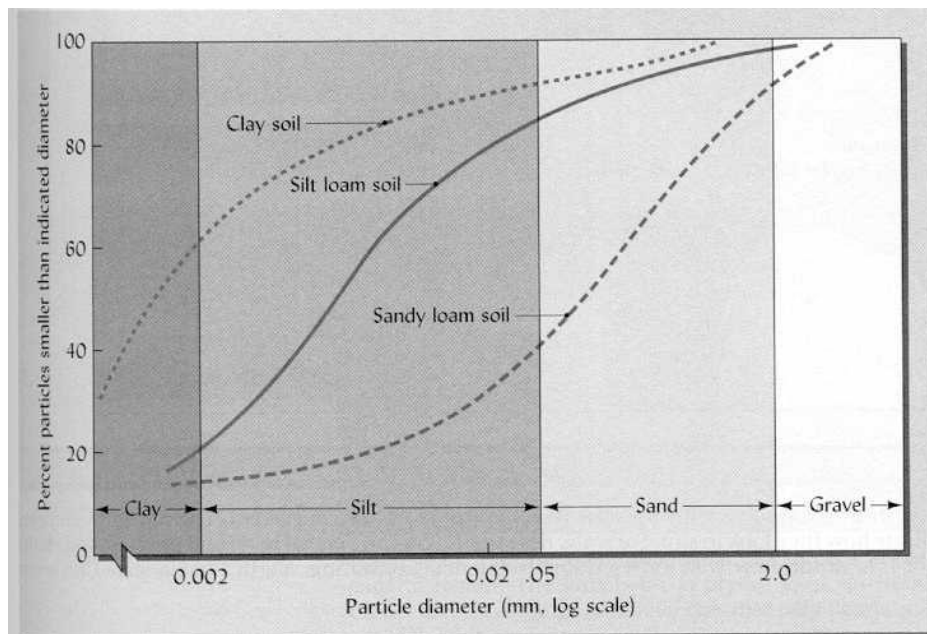
FIGURE 3.6 Particle-size distribution in three soils varying widely in their textures. Note that there is a gradual transition in the particle-size distribution in each of these soils.

Figure 3.6 presents particle-size distribution curves for soils representative of three textural classes. The fact that these curves are smooth emphasizes that there is no sharp line of demarcation in the distribution of sand, silt, and clay fractions, and suggests a gradual change of properties with change in particle

size.

It is important to note that soils are assigned to textural classes *solely* on the basis of the mineral particles of sand size and smaller; therefore, the percentages of sand, silt, and clay always add up to 100%. The amounts of stone and gravel are rated separately. Organic matter is usually removed from a soil sample by oxidation before the mechanical separation.

The relationship between such analyses and textural class names is commonly shown diagrammatically as a triangular graph (e.g., Fig. 3.5). This **textural triangle** also enables us to use laboratory particle-size analysis data to check the accuracy of field textural determinations by feel.

3.4 STRUCTURE OF MINERAL SOILS

The term **structure** relates to the arrangement of primary soil particles into groupings called **aggregates** or **peds.** The pattern of pores and peds defined by soil structure greatly influences water movement, heat transfer, aeration, and porosity in soils. Activities such as timber harvesting, grazing, tillage, trafficking, drainage, liming, and manuring impact soils largely through their effect on soil structure, especially in the surface horizons.

Types of Soil Structure

Many types of structural peds occur in soils, often within different horizons of a particular soil profile. Soil structure is characterized in terms of the shape (or *type),* size, and distinctness (or *grade)* of the peds. The four principal shapes of soil structure are *spheroidal, platy, prismlike,* and *blocklike.* These structural types (and some subtypes) are illustrated in Figure 3.7 and are described following.

SPHEROIDAL. **Granular** structure consists of spheroidal peds or **granules** that are usually separated from each other in a loosely packed arrangement (see Figure 3.7a). When the spheroidal peds are especially porous, the term **crumb** is sometimes applied. They typically range from less than 1 to greater than 10 mm in diameter.

Granular and crumb structures are characteristic of many surface soils (usually

A horizons), particularly those high in organic matter. Consequently, they are the principal types of soil structure affected by management. They are especially prominent in grassland soils and soils that have been worked by earthworms.

PLATELIKE. Platy structure, characterized by relatively thin horizontal peds or plates, may be found in both surface and subsurface horizons. In most instances, the plates have developed as a result of soil-forming processes. However, unlike other structure types, platy structure may also be inherited from soil parent materials, especially those laid down by water or ice. In some cases compaction of clayey soils by heavy machinery can create platy structure (see Figure 3.7b).

BLOCKLIKE. Blocky peds are irregular, roughly cubelike, and range from about 5 to 50 mm across. The individual blocks are not shaped independently, but are molded by the shapes of the surrounding blocks. When the edges of the blocks are sharp and the rectangular faces distinct, the subtype is designated angular blocky (see Figure 3.7c). When some rounding has occurred, the aggregates are referred to as subangular blocky (see Figure 3.7d). These types are usually found in B horizons, where they promote good drainage, aeration, and root penetration.

PRISMLIKE. **Columnar** and **prismatic** structure are characterized by vertically oriented prisms or pillarlike peds that vary in height among different soils and may have a diameter of 150 mm or more. Columnar structure (see Figure 3.7e), which has pillars with distinct, rounded tops, is especially common in subsoils high in sodium (e.g., natric horizons). When the tops of the prisms are relatively angular and flat horizontally, the structure is designated as prismatic (see Figure 3.7f). Both prismlike structures are often associated with swelling
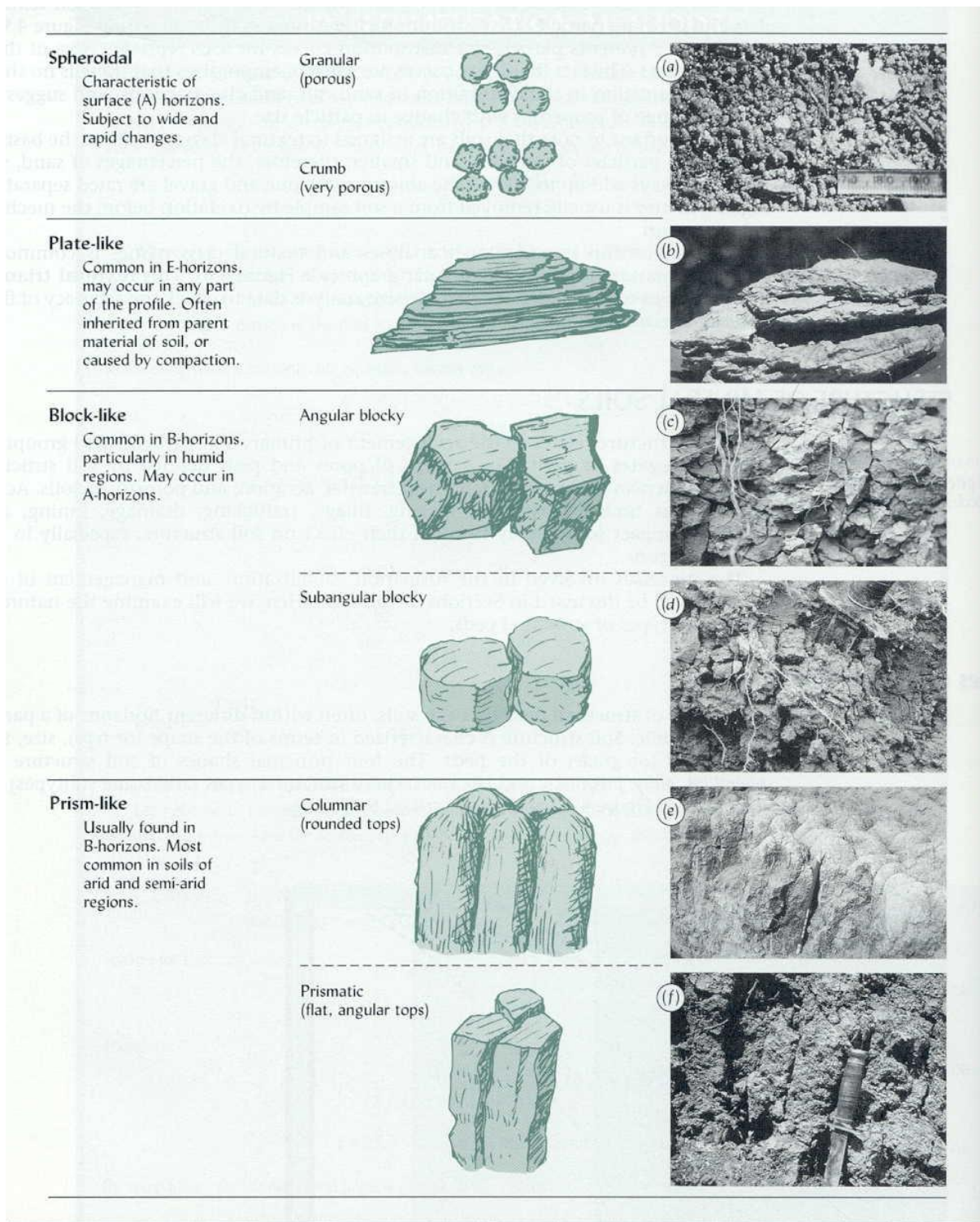
Spheroidal

Characteristic of surface (A) horizons. Subject to wide and rapid changes.

Granular (porous)

Crumb (very porous)

(a)

Plate-like

Common in E-horizons, may occur in any part of the profile. Often inherited from parent material of soil, or caused by compaction.

(b)

Block-like

Common in B-horizons, particularly in humid regions. May occur in A-horizons.

Angular blocky

(c)

Subangular blocky

(d)

Prism-like

Usually found in B-horizons. Most common in soils of arid and semi-arid regions.

Columnar (rounded tops)

(e)

Prismatic (flat, angular tops)

(f)

FIGURE 3.7 The various structure types (shapes) found in mineral soils. Their typical location is suggested. The drawings illustrate their essential features and the photos indicate how they look in situ. For scale, note the 15-cm-long pencil in *(e)* and the 3-cm-wide knife blade in *(d)* and (f). [Photo *(e)* courtesy of J. L. Arndt, now with Petersen Environmental Consulting; North Dakota State University. Others courtesy of R. Weil].

types of clay. They commonly occur in subsurface horizons in arid and semiarid

regions and, when well developed, are a very striking feature of the profile. In humid regions, prismatic structure sometimes occurs in poorly drained soils and in fragipans.

Description of Soil Structure in the Field

In describing soil structure (see Table 19.1), soil scientists note not only the *type* (shape) of the structural peds present, but also the relative *size* (fine, medium, coarse) and degree of development or distinctness of the peds *(grades* such as strong, moderate, or weak). Generally, the structure of a soil is easier to observe when the soil is relatively dry. When wet, structural peds may swell and press closer together, making the individual peds less well defined. In any case, the structural arrangement of soil particles and the pore spaces between structural peds greatly influence soil density, an aspect of soil architecture that we will now examine in detail.

3.5 SOIL DENSITY

Particle Density

Soil particle density D_p is defined as the mass per unit volume of soil *solids* (in contrast to the volume of the *soil,* which would also include spaces between particles). Thus, if 1 cubic meter (m^3) of soil solids weighs 2.6 megagrams (Mg), the particle density is 2.6 Mg/m^3 (which can also be expressed as 2.6 grams per cubic centimeter).[4]

Particle density is essentially the same as the specific gravity of a solid substance. The chemical composition and crystal structure of a mineral determines its particle density. Particle density is *not* affected by pore space, and therefore is not related to particle size or to the arrangement of particles (soil structure).

Particle densities for most mineral soils vary between the narrow limits of 2.60 to 2.75 Mg/m^3 because quartz, feldspar, micas, and the colloidal silicates that usually make up the major portion of mineral soils all have densities within this range. For general calculations concerning arable mineral surface soils (1 to 5% organic matter), a particle density of about 2.65 Mg/m^3 may be assumed

if the actual particle density is not known. This number would be reduced for soils known to be high in organic matter, which has a particle density of only 0.9 to 1.3 Mg/m^3.

Bulk Density

A second important mass measurement of soils is **bulk density** D_b, which is defined as the mass of a unit volume of dry soil. This volume includes both solids and pores. Both expressions of density use only the mass of the solids in a soil; therefore any water present is excluded from consideration.

There are several methods of determining soil bulk density by obtaining a known volume of soil, drying it to remove the water, and weighing the dry mass. A special coring instrument can obtain a sample of known volume without disturbing the natural soil structure.

Factors Affecting Bulk Density

Soils with a high proportion of pore space to solids have lower bulk densities than those that are more compact and have less pore space. Consequently, any factor that influences soil pore space will affect bulk density.

EFFECT OF SOIL TEXTURE. Fine-textured soils such as silt loams, clays, and clay loams generally have lower bulk densities than do sandy soils. This is true because the solid particles of the fine-textured soils tend to be organized in porous granules, especially if adequate organic matter is present. In these aggregated soils, pores exist both between *and* within the granules. This condition ensures high total pore space and a low bulk density. In sandy soil, however, organic matter contents generally are low, the solid particles are less likely to be aggregated, and the bulk densities are commonly higher than in the finer-textured soils. More or less similar amounts of large pores are present in both sandy and well-aggregated fine-textured soils, but that sandy soils have few of the fine, within-ped pores, and so have less total porosity.

This fact may seem counterintuitive at first because sandy soils are commonly referred to as "light" soils, while clays and clay loams are referred to as "heavy" soils. The terms heavy and light in this context refer not to the mass

per unit volume of the soils, but to the amount of effort that must be exerted to manipulate these soils with tillage implements-the sticky clays being much more difficult to till.
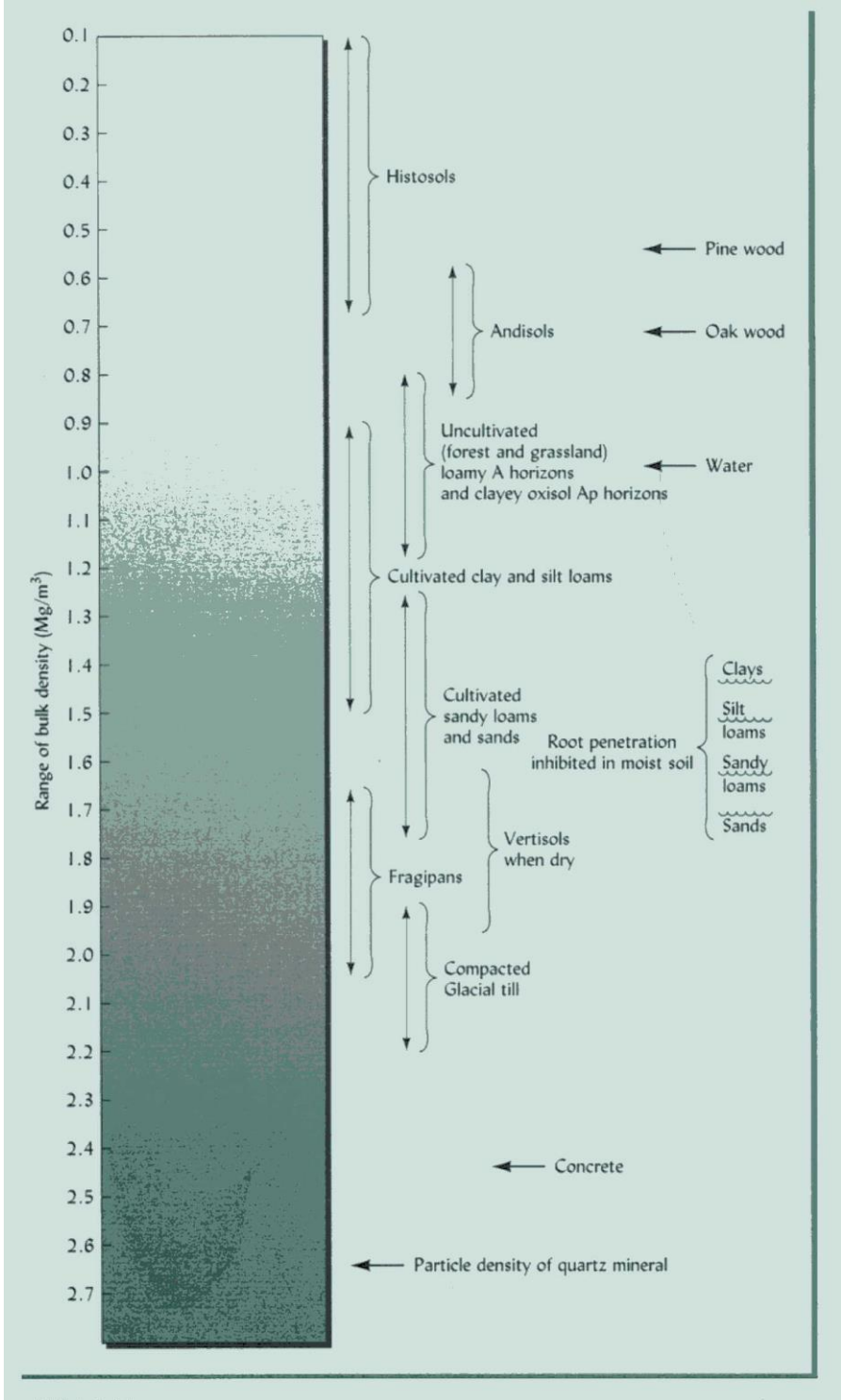

FIGURE 3.8 Bulk densities typical of a variety of soils and soil materials.

DEPTH IN SOIL PROFILE. Deeper in the soil profile, bulk densities are generally higher, probably as a result of lower organic matter contents, less aggregation, fewer roots and other soil-dwelling organisms, and compaction caused by the weight of the overlying layers. Very compact subsoils may have bulk densities of 2.0 Mg/m^3 or even greater.

Management Practices Affecting Bulk Density

Changes in bulk density for a given soil are easily measured and can alert soil managers to changes in soil quality and ecosystem function. Increases in bulk density usually indicate a poorer environment for root growth, reduced aeration, and undesirable changes in hydrologic function, such as reduced water infiltration.

FOREST LANDS. The surface horizons of most forested soils have rather low bulk densities (see Figure 3.8). Tree growth and forest ecosystem function are particularly sensitive to increases in bulk density. Conventional timber harvest generally disturbs and compacts 20 to 40% of the forest floor and is especially damaging along the skid trails where logs are dragged and at the landing decks-areas where logs are piled and loaded onto trucks. An expensive, but effective, means of moving logs while minimizing compactive degradation of forest lands is the use of cables strung between towers or hung from large balloons.

Intensive recreational and transport use of soils in forests and other areas with natural vegetation can also lead to increased bulk densities. Such effects can be seen where access roads, trails, and campsites are found. E.g., trails of herdsmen of the 19th century still show the ill effects of soil compaction. An important consequence of increased bulk density is a diminished capacity of the soil to take in water, hence increased losses by surface runoff.

In urban areas, trees planted for landscaping purposes must often contend with severely compacted soils. While it is usually not practical to modify the entire root zone of a tree, several practices can help. First, making the planting hole as large as possible will provide a zone of loose soil for early root growth. Second, a thick layer of mulch spread out to the drip line (but not too near the trunk) will enhance root growth, at least near the surface. Third, the tree roots may be given paths for expansion by digging a series of narrow trenches radiating out from the planting hole and back-filled with loose, enriched soil.

In some urban settings it may be desirable to create an "artificial soil" that includes a skeleton of coarse angular gravel to provide strength and stability, and a mixture of loam-textured topsoil and organic matter to provide nutrient- and water-holding capacities. Also, large quantities of sand and organic materials are sometimes mixed into the upper few centimeters of a fine-textured soil on which putting green turf grass is to be grown.

AGRICULTURAL LAND. Although tillage may temporarily loosen the surface soil, in the long term intense tillage increases soil bulk density because it depletes soil organic matter and weakens soil structure. The effect of cultivation can be minimized by adding crop residues or farm manure in large amounts and rotating cultivated crops with a grass sod.

In modern agriculture, heavy machines used to pull implements, apply amendments, or harvest crops can create yield-limiting soil compaction. Certain tillage implements, such as the moldboard plow and the disk harrow, compact the soil below their working depth even as they lift and loosen the soil above. Use of these implements or repeated trips over the field by heavy machinery can form **plow pans** or **traffic pans,** dense zones immediately below the plowed layer (Figure 3.9). Other tillage implements, such as the chisel plow and the spring-tooth harrow, do not press down upon the soil beneath them, and so are useful in breaking up plow pans and stirring the soil with a minimum of compaction. Large chisel-type plows can be used in subsoiling to break up dense subsoil layers, thereby permitting root penetration (Figure 3.21). However, in some soils the effects of subsoiling are quite temporary. Any tillage tends to reduce soil strength, thus making the soil less resistant to subsequent compaction.
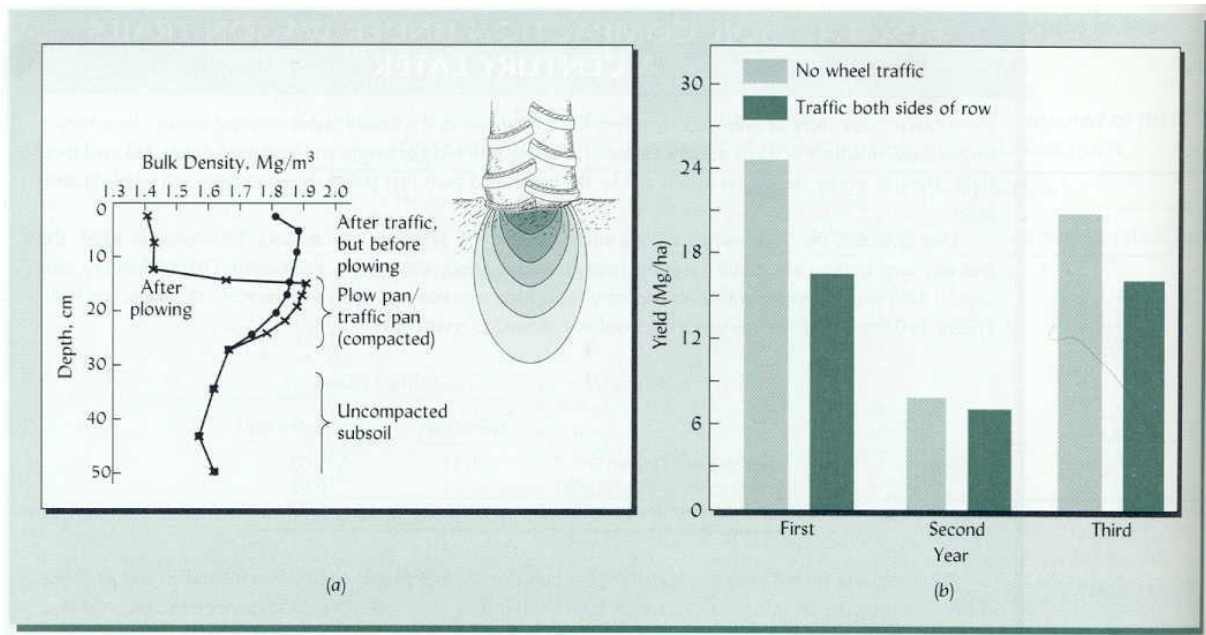
FIGURE 3.9 Tractors and other heavy equipment compact the soil to considerable depths, increasing bulk density and reducing plant growth and crop yields The effects are especia11y damaging if the soil is wet when trafficked. (a) The tires of a heavy vehicle compact a sandy loam soil to about 30 cm, creating a traffic pan. Plowing temporarily loosens the compacted surface soil (plow layer), but increases compaction just below the plowed layer, creating a combined traffic pan and plow pan. Bulk densities in excess of 1.8 Mg/m³ prevented the penetration of cotton roots in this case. (b) The yield of potatoes was reduced in two out of three years in this test on a clay loam in Minnesota. Yield reductions are often most pronounced in relatively dry years when plants have the greatest need for subsoil moisture. [Based on data from Camp and Lund (1964) and Voorhees (1984)]

Traffic is particularly damaging on wet soil. Generally, with heavier loads and on wetter soils, compactive effects are more pronounced and penetrate more deeply into the profile. To prevent compaction, which can result in yield reductions and loss if profitability, the number of tillage operations and heavy equipment trips over the field should be minimized and timed to avoid periods when the soil is wet. Unfortunately, traffic on wet agricultural soils is sometimes unavoidable in humid temperate regions in spring and fall.

Some managers attempt to reduce compaction using an opposite strategy in which special wide tires are fitted to heavy equipment so as to spread the weight over more soil surface, thus reducing the force applied per unit area.

Influence of Bulk Density on Soil Strength and Root Growth

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High bulk density may occur as a natural soil profile feature (for example, a fragipan) or it may be an indication of human-induced soil compaction. In any case, root growth is inhibited by excessively dense soils for a number of reasons, including the soil's resistance to penetration, poor aeration, slow movement of nutrients and water, and the buildup of toxic gases and root exudates.
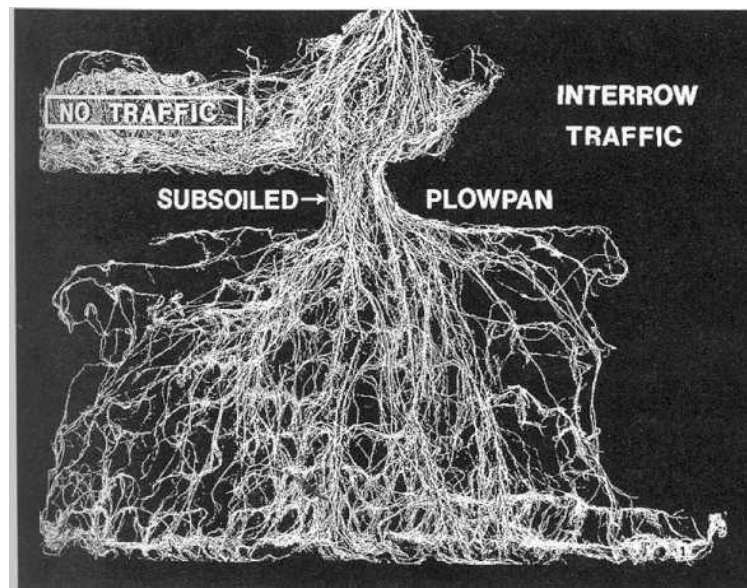

FIGURE 3.10 Root distribution of a cotton plant. On the right, interrow tractor traffic and plowing have caused a plowpan that restricts root growth. Roots are more prolific on the left where there had been no recent tractor traffic. The roots are seen to enter the subsoil through a loosened zone created by a subsoiling chisel-type implement. (Courtesy USDA National Tillage Machinery Laboratory)

Roots penetrate the soil by pushing their way into pores. If a pore is too small to accommodate the root cap, the root must push the soil particles aside and enlarge the pore. To some degree, the density per se restricts root growth, as the roots encounter fewer and smaller pores. However, root penetration is also limited by soil strength, the property of the soil that causes it to resist deformation. One way to quantify **soil strength**, is to measure the force needed to push a standard cone-tipped rod (a penetrometer) into the soil. Compaction generally increases both bulk density and soil strength. At least two factors (both related to soil strength) must be considered to determine the effect of bulk density on the ability of roots to penetrate soil.

EFFECT OF SOIL WATER CONTENT. Soil water content and bulk density both affect soil strength. Soil strength is increased when a soil is compacted to a higher bulk density, and also when finer-textured soils dry out and harden. Therefore, the effect of bulk density on root growth is most pronounced if those soils are dry, a higher bulk density being necessary to prevent root penetration when the soils are moist.

EFFECT OF SOIL TEXTURE. The more clay present in a soil, the smaller the average pore size, and the greater the resistance to penetration at a given bulk density. Therefore, if the bulk density is the same, roots more easily penetrate a moist sandy soil than a moist clayey one. The growth of roots into moist soil is generally limited by bulk densities ranging from 1.45 Mg/m^3 in clays to 1.85 Mg/m^3 in loamy sands (see Figure 3.8).

3.6 PORE SPACE OF MINERAL SOILS

One of the main reasons for measuring soil bulk density is that this value can be used to calculate pore space. For soils with the same particle density, the lower the bulk density, the higher the percent pore space **(total porosity).**

Factors Influencing Total Pore Space

In the first lesson we noted that for an "ideal" medium-textured, well-granulated surface soil in good condition for plant growth, approximately 50% of the soil volume would consist of pore space, and that the pore space would be about half-filled with air and half-filled with water. Actually, total porosity varies widely among soils for the same reasons that bulk density varies. Values range from as low as 25% in compacted subsoils to more than 60% in well-aggregated, high-organic-matter surface soils. As is the case for bulk density, management can exert a decided influence on the pore space of soils. Data from a wide range of soils show that cultivation tends to lower the total pore space compared to that of uncultivated soils. This reduction usually is associated with a decrease in organic matter content and a consequent lowering of **granulation.**

Bulk density values help us predict only *total* porosity. However, soil pores occur in a wide variety of sizes and shapes that largely determine what role

the pore can play in the soil. Pores can be grouped by size into macropores, mesopores, micropores, and so on. We will simplify our discussion at this point by referring only to **macropores** (larger than about 0.08 mm) and micropores (smaller than about 0.08 mm).

MACROPORES. The macropores characteristically allow the ready movement of air and the drainage of water. They also are large enough to accommodate plant roots and the wide range of tiny animals that inhabit the soil.

Macropores can occur as the spaces between individual sand grains in coarse-textured soils. Thus, even though a sandy soil has relatively low total porosity, the movement of air and water through such a soil is surprisingly rapid because of the dominance of the macropores.

In well-structured soils, the macropores are generally found between peds. These **interped pores** may occur as spaces between loosely packed granules or as the planar cracks between tight-fitting blocky and prismatic peds.

Macropores created by roots, earthworms, and other organisms constitute a very important type of pores termed **biopores.** These are usually tubular in shape and may be continuous for lengths of a meter or more. In some clayey soils, biopores are the principal form of macropores, greatly facilitating the growth of plant roots.

Perennial vegetation, such as forest trees and certain grasses, is particularly effective at creating channels that serve as conduits for roots, long after the death and decay of the roots that originally created them.

MICROPORES. In contrast to macropores, micropores are usually filled with water in field soils. Even when not water-filled, they are too small to permit much air movement. Water movement in micropores is slow, and much of the water retained in these pores is not available to plants. Fine-textured soils, especially those without a stable granular structure, may have a predominance of micropores, thus allowing relatively slow gas and water movement, despite the relatively large volume of total pore space. Aeration, especially in the subsoil, may be inadequate for satisfactory root development and desirable microbial activity. While the larger micropores accommodate plant root hairs and microorganisms, the smaller micropores (sometimes termed *ultramicropores* and *cryptopores)* are so tiny that their radii are measured in nanometers (10^{-9} meters), giving rise to the term *nanopores.* Such pores are

too small to permit the entrance of even the smallest bacteria or some decay-stimulating enzymes produced by the bacteria. Thus, these pores can act as hiding places for some adsorbed organic compounds (both naturally occurring and pollutants), thereby protecting them from breakdown for long periods of time, perhaps for centuries.

Clearly, the size, shape, and interconnection of soil pores, rather than their combined volume, are of greatest importance in determining soil drainage, aeration, and other such processes.

Cultivation and Pore Size

Continuous cropping, particularly of soils originally high in organic matter, often results in a reduction of macropore spaces (see Table 3.2). When native prairie lands are plowed and planted to row crops such as corn or soybeans, soil organic matter contents and total pore space are reduced. But most striking is the effect of cropping on the size of the soil pores. Such cropping drastically reduces the amount of macropore space that is so critical for ready air movement.

In recent years, conservation tillage practices, which minimize plowing and associated soil manipulations, have been widely adopted in the United States. Because of increased accumulation of organic matter near the soil surface and the development of a long-lived network of macropores (especially biopores), some conservation tillage systems lead to greater macroporosity of the surface layers.

TABLE 3.2 Effect of About 50 Years of Continuous Cropping on the Macropore and Micropore Spaces Textured Vertisol (Houston Black Clay) in Texas

Soil history	Organic matter, %	Total pore space, %	Macropore space, %	Micropore space, %	Bulk density, Mg/m³
0–15 cm depth					
Virgin prairie	5.6	58.3	32.7	25.6	1.11
Tilled 50 years	2.9	50.2	16.0	34.2	1.33
15–30 cm depth					
Virgin prairie	4.2	56.1	27.0	29.1	1.16
Tilled 50 years	2.8	50.7	14.7	36.0	1.31

Compared to the undisturbed prairie (virgin soil), the cultivated soil has far less macropore

space, but has gained some mic-ropore space as aggregates were destroyed, changing large interped pores into much smaller micropores. The loss of macropore space probably resulted from the loss of organic matter that is evident.

3.7 FORMATION AND STABILIZATION OF SOIL AGGREGATES

The formation and maintenance of a high degree of aggregation are among the most difficult tasks of soil management, yet they are also among the most important, since they are a potent means of influencing ecosystem function. The organization of surface soils into relatively large structural aggregates provides for the low bulk density and a high proportion of macropores so desirable for most soil uses.

Some aggregates readily succumb to the beating of rain and the rough-and-tumble of plowing and tilling the land. Others resist disintegration, thus making the maintenance of a suitable soil structure comparatively easy. Generally, the smaller aggregates are more stable than the larger ones, so maintaining the much-prized larger aggregates requires much care.

We will discuss practical means of managing soil structure after we consider those factors responsible for aggregate formation and those that give the aggregates stability once they are formed. Since both sets of factors are operating simultaneously, it is sometimes difficult to distinguish their relative effects on the development of stable aggregates in soils.

Hierarchical Organization of Soil Aggregates

The large aggregates (>1 mm) so desirable for most soil uses are typically composed of smaller aggregates, which in turn are composed of still smaller units, down to clusters of clay and humus less than 0.001 mm in size. You may easily demonstrate the existence of this *hierarchy of aggregation* by selecting a few of the largest aggregates in a soil and gently crushing or picking them apart to separate them into many smaller-sized aggregates.

Factors influencing Aggregate Formation and Stability in Soils

Both biological and physical-chemical (abiotic) processes are involved in the formation of soil aggregates. The physical-chemical processes tend to be most important at the smaller end of the scale, biological processes at the larger end. Also, the physical-chemical processes of aggregate formation are associated mainly with clays and, hence, tend to be of greater importance in finer-textured soils. In sandy soils that have little clay, aggregation is almost entirely dependent on biological processes.

Physical-Chemical Processes

Most important among the physical-chemical processes are (1) the mutual attraction among clay particles and (2) the swelling and shrinking of clay masses.

FLOCCULATION OF CLAYS AND THE ROLE OF ADSORBED CATIONS. Except in very sandy soils that are almost devoid of clay, aggregation begins with the **flocculation** of clay particles into microscopic clumps or *floccules* (Figure 3.11). If two clay platelets come close enough to each other, the cations compressed in a layer between them will attract the negative charges on both platelets, thus serving as bridges to hold the platelets together. This process is repeated until a small "stack" of parallel clay platelets, termed *a clay domain*, is formed. Other types of clay domains are more random in orientation, resembling a house of cards. These form when the positive charges on the edges of the clay platelets attract the negative charges on the planar surfaces (Figure 3.11). Clay floccules or domains, along with charged organic colloids (humus), form bridges that bind to each other and to fine silt particles (mainly quartz), creating the smallest size groupings in the hierarchy of soil aggregates. These domains, aided by the flocculating influence of polyvalent cations (e.g., Ca^{2+}, Fe^{2+}, and Al^{3+}) and humus, provide much of the long-term stability for the smaller (<0.03 mm) **microaggregates**. The cementing action of inorganic compounds, such as iron oxides, produces very stable small aggregates sometimes called **pseudosand** in certain clayey soils (Ultisols and Oxisols) of hot, humid regions.
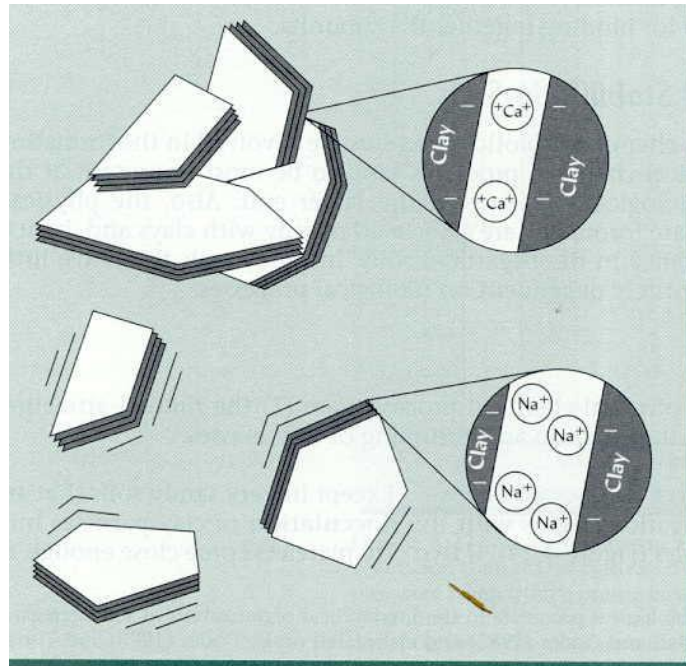
FIGURE 3.11 The role of cations in the flocculation of soil clays. The di- and trivalent cations, such as Ca^{2+} and Al^{3+} tightly adsorbed and can effectively neutralize the negative asurface charge on clay particles. These cations can also from bridges that bring clay particles close together. Monovalent ions, especially Na^+, with relatively large hydrated radii, can cause clay particles to repel each other and create a dispersed condition. Two things contribute to the dispersion: (1) the large hydrated sodium ion does not get close enough to the clay to effectively neutralize the negative charges, and (2) the single charge on sodium is not effective in forming a bridge between clay particles.

When Na^+ (rather than polyvalent cations such as Ca^{2+} or Al^{3+}) is a prominent adsorbed ion, as in some soils of arid and semiarid areas, the attractive forces are not able to overcome the natural repulsion of one negatively charged clay by another. The clay platelets cannot approach closely enough to flocculate, so remain dispersed as far apart from one another as possible. Clay in this dispersed, gel-like condition causes the soil to become almost structureless, impervious to water and air, and very undesirable from the standpoint of plant growth.

VOLUME CHANGES IN CLAYEY MATERIALS. As a soil dries out and water is withdrawn, the platelets in clay domains move closer together, causing the domains and, hence, the soil mass to shrink in volume. As a soil mass shrinks, cracks will open up along planes of weakness. Over the course of many cycles

(as occur between rain or irrigation events in the field) the network of cracks becomes more extensive and the aggregates between the cracks better defined. Plant roots also have a distinct drying effect as they take up soil moisture in their immediate vicinity. Water uptake, especially by fibrous-rooted perennial grasses, accentuates the physical aggregation processes associated with wetting and drying. This effect is but one of many examples of ways in which physical and biological soil processes interact.

Freezing and thawing cycles have a similar effect, since the formation of ice crystals is a drying process that also draws water out of clay domains. The swelling and shrinking actions that accompany freeze-thaw and wet-dry cycles in soils create fissures and pressures that alternately break apart large soil masses and compress soil particles into defined structural peds. The aggregating effects of these water and temperature cycles are most pronounced in soils with a high content of swelling-type clays, especially Vertisols, Mollisols, and some Alfisols.

Biological Processes

ACTIVITIES OF SOIL ORGANISMS. Among the biological processes of aggregation, the most prominent are (1) the burrowing and molding activities of earthworms, (2) the enmeshment of particles by sticky networks of roots and fungal hyphae, and (3) the production of organic glues by microorganisms, especially bacteria and fungi. In both cultivated and uncultivated soils, earthworms (and termites) move soil particles around, often ingesting them and forming them into pellets or casts. In some forested soils, the surface horizon consists primarily of aggregates formed as earthworm castings (see, for example, Figure 3.7a). Plant roots also move particles about as they push their way through the soil. This movement forces soil particles to come into close contact with each other, encouraging aggregation. At the same time, the channels created by plant roots and soil animals serve as macropores, breaking up large clods and helping to define larger soil structural units.

Plant roots (particularly root hairs) and fungal hyphae exude sugarlike polysaccharides and other organic compounds, forming sticky networks that bind together individual soil particles and tiny microaggregates into larger agglomerations called **macroaggregates**. The threadlike fungi that associate

with plant roots (called *mycorrhizae*) are especially effective in providing this type of relatively short-term stabilization of large aggregates, because they secrete a gooey protein called glomalin, which is very effective as a cementing agent (Figure 3.12).

Bacteria also produce polysaccharides and other organic glues as they decompose plant residues. Bacterial polysaccharides were shown intermixed with clay at a very small scale. Many of these root and microbial organic glues resist dissolution by water and so not only enhance the formation of soil aggregates but also help ensure their stability over a period of months to a few years. These processes are most notable in surface soils, where root and animal activities and organic matter accumulation are greatest.
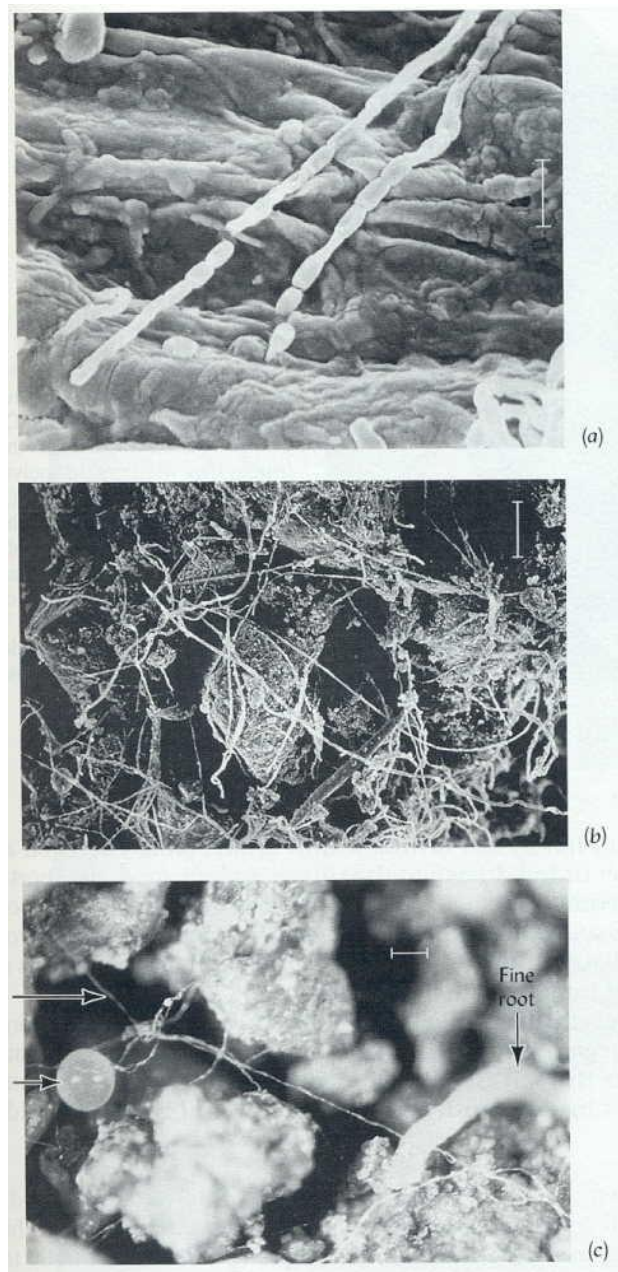
FIGURE 3.12 Fungal hyphae binding soil particles into aggregates. (a) Close-up of a hyphae growing over the surface of a mineral grain encrusted with microbial cells and debris. Bar = 10 μm. *(b)* An advanced stage of aggregation during the formation of soil from dune sands. Note the net of fungal hyphae and the encrustation of the mineral grains with organic debris. Bar = 50 μm. (c) Hyphae of the root-associated fungus from the genus *Gigaspora* interconnecting particles in a sandy loam from Oregon. Note also the fungal spore and the plant root. Bar = 320 μm.

INFLUENCE OF ORGANIC MATTER. In most temperate zone soils, organic matter is *the* major agent stimulating the formation and stabilization of granular and crumb-type aggregates (see Figure 3.14). In the first place,

organic matter provides the energy substrate that makes possible the previously mentioned activities of the fungi, bacteria, and soil animals. Second, as organic residues decompose, gels and other viscous microbial products, along with associated bacteria and fungi, encourage crumb formation. Organic exudates from plant roots also participate in this aggregating action.

Organic products of decay, such as complex polymers, chemically interact with particles of silicate clays and iron and aluminum oxides. These compounds orient the clays into packets (domains), which form bridges between individual soil particles, thereby binding them together in water-stable aggregates.

During the aggregation process, soil mineral particles (silts and fine sands) become coated and encrusted with bits of decomposed plant residue and other organic materials. At an even smaller scale, microscopic bits of decomposed residues and humus particles become encrusted with clay packets. In either case, the resulting organomineral complexes promote the formation of aggregates. Figure 3.13 shows direct evidence of the organomineral domains that bind soil particles.
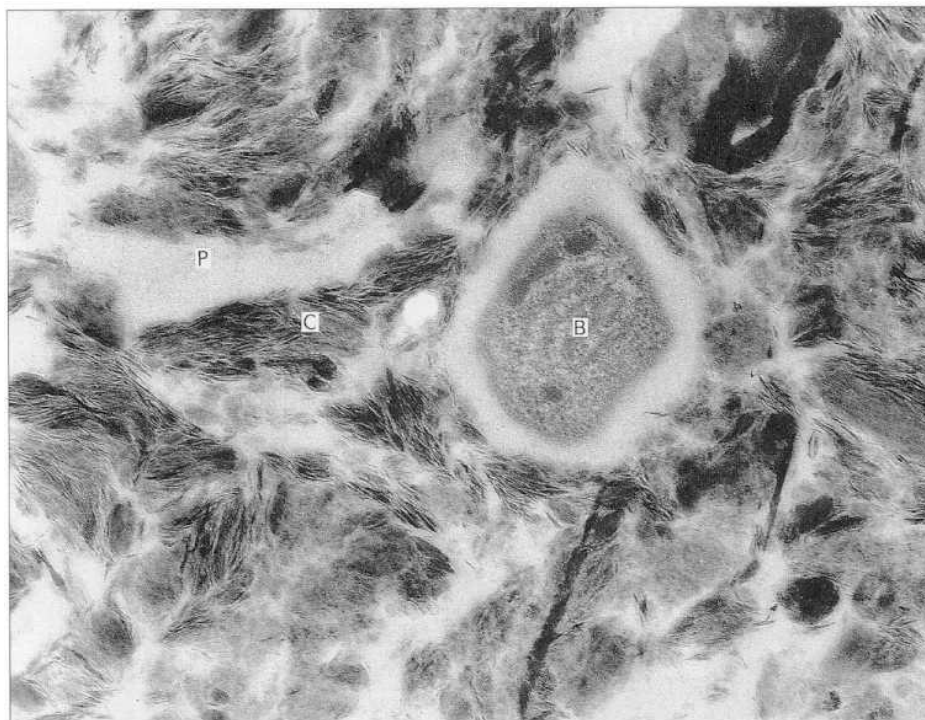
FIGURE 3.13 An ultrathin section illustrating the interaction among organic materials and silicate clays in a water-stable aggregate. The dark-colored materials (C) are groups of clay particles that are interacting with organic polysaccharides (P). A bacterial cell (B) is also surrounded by polysaccharides. Note the generally horizontal orientation of the clay particles, an orientation encouraged by the organic materials. [From Emerson, et al.
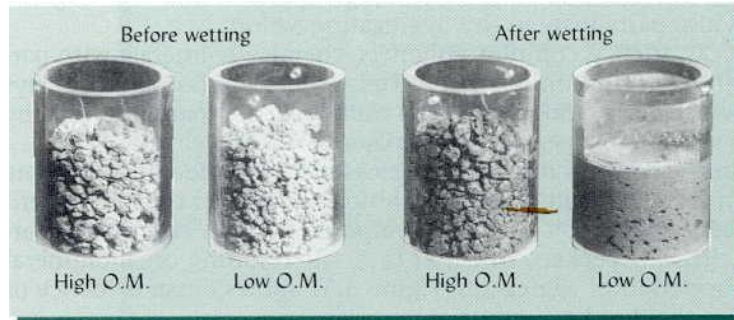
FIGURE 3.14 The aggregates of soils high in organic matter are much more stable than are those low in this constituent. The low-organic-matter soil aggregates fall apart when they are wetted; those high in organic matter maintain their stability.

INFLUENCE OF TILLAGE. Tillage can have both favorable and unfavorable effects on aggregation. If the soil is not too wet or too dry when the tillage is performed, the short-term effect of tillage is generally favorable. Tillage implements break up large clods, incorporate organic matter into the soil, kills weeds, and generally create a more favorable seedbed. Immediately after plowing, the surface soil is loosened (its cohesive strength is decreased) and total porosity is increased.

Over longer periods, however, tillage greatly hastens the oxidation of soil organic matter, thus reducing the aggregating effects of this soil component. Tillage operations, especially if carried out when the soil is wet, also tend to crush or smear stable soil aggregates, resulting in loss of macroporosity and the creation of *a puddled* condition (see Figure 3.15).
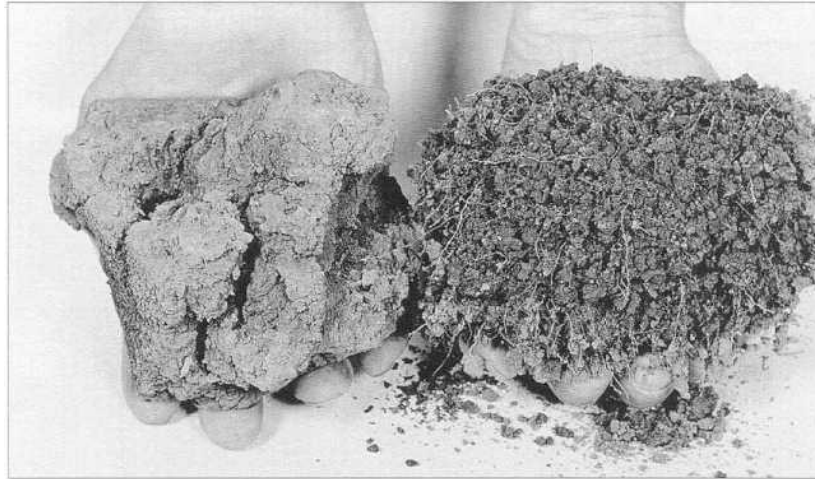
FIGURE 3.15 Puddled soil (left) and well-granulated soil (right). Plant roots and especially humus play the major role in soil granulation. Thus a sod tends to encourage development of a granular structure in the surface horizon of cultivated land. (Courtesy USDA Natural Resources Conservation Service)

INFLUENCE OF IRON/ALUMINUM OXIDES. Scientists have noted that some well-weathered soils of the tropics have more stable aggregation than soils with comparable or even higher organic matter levels in temperate regions. This is thought to be due primarily to the binding effects of the iron and aluminum sesquioxides (particularly the amorphous forms) that are plentiful in many tropical soils. Films of such compounds coat and cement soil aggregates, thereby preventing their ready breakdown when the soil is tilled.

3.8 TILLAGE AND STRUCTURAL MANAGEMENT OF SOILS

When protected under dense vegetation and undisturbed by tillage, most soils (except perhaps some sparsely vegetated soils in arid regions) possess a surface structure sufficiently stable to allow rapid infiltration of water and to prevent crusting. However, for the manager of cultivated soils, the development and maintenance of stable surface soil structure is a major challenge. Many studies have shown that aggregation and associated desirable soil properties such as water infiltration rate decline under long periods of tilled row-crop cultivation (Table 3.3).

TABLE 3.3 Effect of Period of Corn Cultivation on Soil Organic Matter, Aggregate Stability, and Water Infiltration in Five Silt Loam Inceptisols from Southwest France

Period of cultivation, years	Organic matter, %	Aggregate stability, mm MWD[b]	Infiltration	
			Of total rain, %	Prior to ponding, mm
100	0.7	.35	25	6
47	1.6	.61	34	9
32	2.6	.76	38	15
27	3.1	1.38	47	25
15	4.2	1.52	44	23

In soils with depleted organic-matter levels, aggregates easily broke down under the influence of water, forming smaller aggregates and dispersed materials that sealed the soil surface and inhibited infiltration. A level of 3% soil organic matter seems to be sufficient for good structural stability in these temperate region silt loam soils.

Tillage and Soil Tilth

Simply defined, **tilth** refers to the physical condition of the soil in relation to plant growth. Tilth depends not only on aggregate formation and stability, but also on such factors as bulk density, soil moisture content, degree of aeration, rate of water infiltration, drainage, and capillary water capacity. As might be expected, tilth often changes rapidly and markedly. For instance, the workability of fine-textured soils may be altered abruptly by a slight change in moisture.

Clayey soils are especially prone to puddling and compaction because of their high plasticity and cohesion. When puddled clayey soils dry, they usually become dense and hard. Proper timing of trafficking is more difficult for clayey than for sandy soils, because the former take much longer to dry to a suitable moisture content and may also become too dry to work easily.

Farmers in temperate regions typically find their soils too wet for tillage just prior to planting time (early spring), while farmers in tropical regions may face the opposite problem of soils too dry for easy tillage just prior to planting (end of dry season). In tropical and subtropical regions with a long dry season, soil often must be tilled in a very dry state in order to prepare the land for planting with the onset of the first rains. Tillage under such dry conditions can be very difficult and can result in hard clods if the soils contain much sticky-type silicate clay.

Soil Crusting

Falling drops of water during heavy rains or sprinkler irrigation beat apart the aggregates exposed at the soil surface. In some soils the dilution of salts by this water stimulates the dispersion of clays. Once the aggregates become dispersed, small particles and dispersed clay tend to wash into and clog the soil pores. Soon the soil surface is covered with a thin layer of fine, structureless material called a **surface seal**. The surface seal inhibits water infiltration and increases erosion losses.

As the surface seal dries, it forms a hard crust. Seedlings, if they emerge at all, can do so only through cracks in the crust. Formation of a crust soon after a crop is sown may allow so few seeds to emerge that the crop has to be replanted. In arid and semiarid regions, soil sealing and crusting can have disastrous consequences because high runoff losses leave little water available to support plant growth.

Crusting can be minimized by keeping some vegetative or mulch cover on the land to reduce the impact of raindrops. Once a crust has formed, it may be necessary to rescue a newly planted crop by breaking up the crust with light tillage (as with a rotary hoe), preferably while the soil is still moist. Improved management of soil organic matter and use of certain soil amendments can "condition" the soil and help prevent clay dispersion and crust formation.

Soil Conditioners

GYPSUM. Gypsum (calcium sulfate) is widely available in its relatively pure mined form, or as a major component of various industrial by-products. Gypsum has been shown effective in improving the physical condition of many types of soils, from some highly weathered acid soils to some low-salinity, high-sodium soils of semiarid regions (see Chapter 10). The more soluble gypsum products provide enough electrolytes (cations and anions) to promote flocculation and inhibit the dispersion of aggregates, thus preventing surface crusting. Field trials have shown that gypsum-treated soils permit greater water infiltration and are less subject to erosion than untreated soils. Similarly, gypsum can reduce the strength of hard subsurface layers, thereby allowing greater root penetration and subsequent plant uptake of water from the subsoil.

ORGANIC POLYMERS. Certain synthetic organic polymers can stabilize soil structure in much the same way as do natural organic polymers such as polysaccharides. While large applications of these polymers would be uneconomical, it has been shown that even very small amounts can effectively inhibit crust formation if applied properly. For example, polyacrylamide (PAM) is effective in stabilizing surface aggregates when applied at rates as low as 1 to 15 mg/L of irrigation water or sprayed on at rates as low as 1 to 10 kg/ha. Figure 3.16 shows the dramatic stabilizing effect of synthetic polyacrylamides used in irrigation water. A number of research reports indicate that the best results can be obtained by combining the use of PAM and gypsum products.
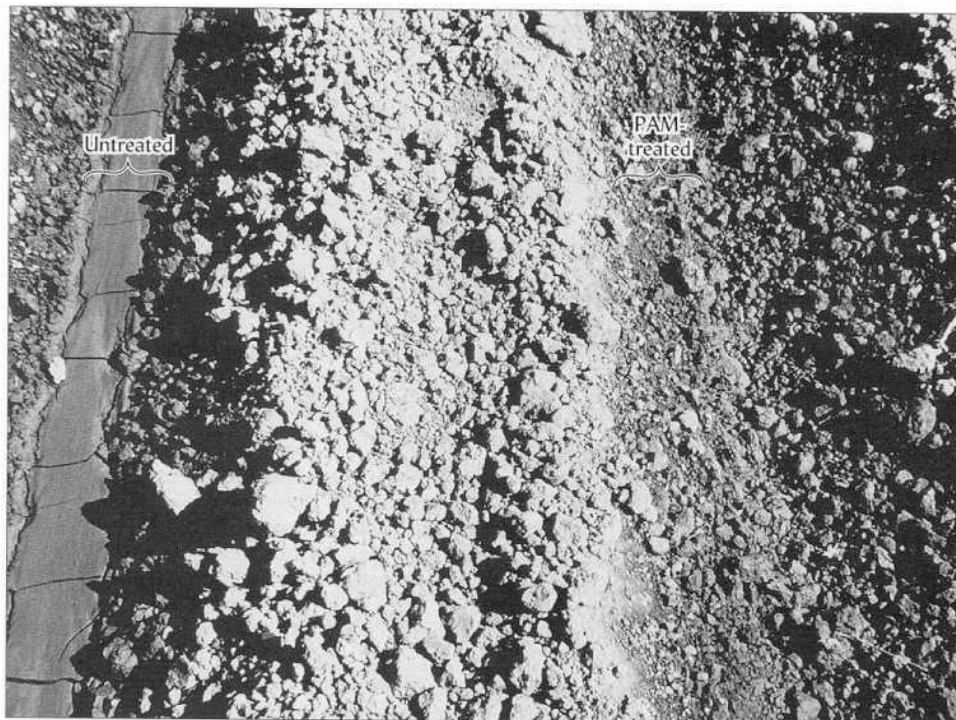
FIGURE 3.16 The remarkable stabilizing effect of a synthetic polyacrylamide is seen in the furrow on the right compared to the left, untreated, furrow. Irrigation water broke down much of the structure of the untreated soil but had no effect on the treated row. [From Mitchell (1986)]

OTHER SOIL CONDITIONERS. Several species of algae that live near the soil surface are known to produce quite effective aggregate-stabilizing compounds. Application of small quantities of commercial preparations containing such algae may bring about a significant improvement in surface soil structure. The

amount of amendment required is very small because the algae, once established in the soil, can multiply.

Various humic materials are marketed for their soil conditioning effects when incorporated at low rates (< 500 kg/ha). However, carefully conducted research at many universities has failed to show that these materials have significantly affected aggregate stability or crop yield, as claimed.

3.9 SOIL PROPERTIES RELEVANT TO ENGINEERING USES

Field rating of Soil Consistence and Consistency

CONSISTENCE. Soil **consistence** is a term used by soil scientists to describe the resistance of a soil to mechanical stresses or manipulations at various moisture contents. Soils are rated for consistence as part of describing a soil profile and for estimating suitability for traffic and tillage. This property is a composite expression of those forces of mutual attraction among soil particles, and between particles and pore water, that determine the ease with which a soil can be reshaped or ruptured. Consistence is commonly deter-mined in the field by feeling and manipulating the soil by hand. As a clod of soil is squeezed between the thumb and forefinger (or crushed underfoot, if necessary), observations are made on the amount of force needed to crush the clod and on the manner in which the soil responds to the force. The degree of cementation of the soil by such materials as silica, calcite, or iron is also considered in identifying soil consistence.

Moisture content greatly influences how a soil responds to stress; hence, moist and dry soils are given separate consistence ratings (Table 3.4). A dry clayey soil that cannot be crushed between the thumb and forefinger but can be crushed easily underfoot would be designated as *hard.* It would exhibit much greater resistance to deformation than the same soil in a wet, plastic state. Anyone who has gotten a vehicle stuck up to its axles in sticky mud has firsthand experience of these principles. While most of the terms in Table 3.4 are self-explanatory, the term **friable** requires some comment. If a clod of moist soil crumbles into aggregates when crushed with only light pressure, it is said to be friable. Friable soils are easily tilled or excavated.

The degrees of *stickiness* and *plasticity* (malleability) of soil in the wet

condition are often included in describing soil consistence (although not shown in Table 3.9).

CONSISTENCY. The term **consistency** is used in a similar manner by soil engineers to describe the degree to which a soil resists deformation when a force is applied. However, consistency is determined by the soil's resistance to *penetration* by an object, while the soil scientist's consistence describes resistance to *rupture.* Instead of crushing a clod of soil, the engineer attempts to penetrate it with either the blunt end of a pencil (some use their thumbs) or a thumbnail. For example, if the blunt end of a pencil makes onlya slight indentation, but the thumbnail penetrates easily, the soil is rated as *very firm* (Table 3.4).

TABLE 3.4 Some Field Tests and Terms Used to Describe the Consistence and Consistency of Soils

	Soil consistence[a]			Soil consistency[b]	
Dry soil	Moist to wet soil	Soil dried then submerged in water	Field rupture (crushing) test	Soil at in situ moisture	Field penetration test
Loose	Loose	Not applicable	Specimen not obtainable	Soft	Blunt end of pencil penetrates deeply with ease
Soft	Very friable	Noncemented	Crumbles under very slight force between thumb and forefinger	Medium firm	Blunt end of pencil can penetrate about 1.25 cm with moderate effort
Slightly hard	Friable	Extremely weakly cemented	Crumbles under slight force between thumb and forefinger	Firm	Blunt end of pencil can penetrate about 0.5 cm
Hard	Firm	Weakly cemented	Crushes with difficulty between thumb and forefinger	Very firm	Blunt end of pencil makes slight indentation; thumbnail easily penetrates
Very hard	Extremely firm	Moderately cemented	Cannot be crushed between thumb and forefinger, but can be crushed slowly underfoot	Hard	Blunt end of pencil makes no indentation; thumbnail barely penetrates
Extremely hard	Slightly rigid	Strongly cemented	Cannot be crushed by full body weight underfoot		

The consistency of cohesive materials is closely related to, but not exactly the same as, their consistence. Conditions, of least coherence are represented by terms at the top o f each column, those o f greater coherence near the bottom.

Field observations of both consistence and consistency provide valuable information to guide decisions about loading and manipulating soils. For construction purposes, however, soil engineers usually must make more

precise measurements of a number of related soil properties that help predict how a soil will respond to applied stress.

3.10 CONCLUSION

Physical properties exert a marked influence on the behavior of soils with regard to plant growth, hydrology, environmental management, and engineering uses. The nature and properties of the individual particles, their size distribution, and their arrangement in soils determine the total volume of nonsolid pore space, as well as the pore sizes, thereby impacting on water and air relationships.

The properties of individual particles and their proportionate distribution (soil texture) are subject to little human control in field soils. However, it is possible to exert some control over the arrangement of these particles into aggregates (soil structure) and on the stability of these aggregates. Tillage and traffic must be carefully controlled to avoid undue damage to soil tilth, especially when soils are rather wet. Generally, nature takes good care of soil structure, and humans can learn much about soil management by studying natural systems. Vigorous and diverse plant growth, generous return of organic residues, and minimal physical disturbance are attributes of natural systems worthy of emulation. Proper plant species selection, crop rotation, and management of chemical, physical, and biological factors can help ensure maintenance of soil physical quality. In recent years, these management goals have been made more practical by the advent of conservation tillage systems that minimize soil manipulations while decreasing soil erosion and water runoff.

Particle size, moisture content, and plasticity of the colloidal fraction all help determine the stability of soil in response to loading forces from traffic, tillage, or building foundations. The physical properties presented in this chapter greatly influence nearly all other soil properties and uses, as discussed throughout this book.

Study Questions?

1. If you were investigating a site for a proposed housing development, how could you use soil colors to help predict where problems might be encountered?

2. You are considering the purchase of some farmland in a region with variable soil textures. The soils on one farm are mostly sandy loams and loamy sands, while those on a second farm are mostly clay loams and clays. List the potential advantages and disadvantages of each farm as suggested by the texture of its soils.

3. Revisit your answer to question 2. Explain how soil structure in both the surface and subsurface horizons might modify your opinion of the merits of each farm.

4. Two different timber-harvest methods are being tested on adjacent forest plots with clay loam surface soils. Initially, the bulk density of the surface soil in both plots was 1.1 Mg/m^3. One year after the harvest operations, plot A soil had a bulk density of 1.48 Mg/m^3, while that in plot B was 1.29 Mg/m^3. Interpret these values with regard to the relative merits of systems A and B, and the likely effects on the soil's function in the forest ecosystem.

5. What are the textural classes of two soils, the first with 15% clay and 45% silt, and the second with 80% sand and 10% clay? (Hint: Use Figure 3.5.)

6. For the forest plot B in question 4, what was the change in percent pore space of the surface soil caused by timber harvest? Would you expect that most of this change was in the micropores or in the macropores? Explain.

7. Discuss the positive and negative impacts of tillage on soil structure. What is another physical consideration that you would have to take into account in deciding whether or not to change from a conventional to a conservation tillage system?

8. What would you, as a home gardener, consider to be the three best and three worst things that you could do with regard to managing the soil structure in your home garden?

4. Soil water: Characteristics and Behavior

Water is a vital component of every living thing. Although it is one of nature's simplest chemicals, water has unique properties that promote a wide variety of physical, chemical, and biological processes. These processes greatly influence almost every aspect of soil development and behavior, from the weathering of minerals to the decomposition of organic matter, from the growth of plants to the pollution of groundwater.

We are all familiar with water. We drink it, wash with it, swim in it, and irrigate our crops with it. But water in the soil is something quite different from water in a drinking glass. In the soil, water is intimately associated with solid particles, particularly, those that are colloidal in size. The interaction between water and soil solids changes the behavior of both.

Water causes soil particles to swell and shrink, to adhere to each other, and to form structural aggregates. Water participates in innumerable chemical reactions that release or tie up nutrients, create acidity, and wear down minerals so that their constituent elements eventually contribute to the saltiness of the oceans.

Attraction to solid surfaces restricts some of the free movement of water molecules, making it less liquid and more solidlike in its behavior. In the soil, water can flow up as well as down. Plants may wilt and die in a soil whose profile contains a million kilograms of water in a hectare. A layer of sand and gravel in a soil profile may actually inhibit drainage, causing the upper horizons to become muddy and saturated with water during much of the year. These and other soil water phenomena seem to contradict our intuition about how water ought to behave.

Soil-water interactions influence many of the ecological functions of soils and practices of soil management. These interactions determine how much rainwater runs into and through the soil and how much runs off the surface. Control of these processes in turn determines the movement of chemicals to the groundwater, and of both chemicals and eroded soil particles to streams and lakes. The interactions affect the rate of water loss through leaching and evapotranspiration, the balance between air and water in soil pores, the rate of change in soil temperature, the rate and kind of metabolism of organisms,

and the capacity of soil to store and provide water for plant growth.

4.1 STRUCTURE AND RELATED PROPERTIES OF WATER

The ability of water to influence so many soil processes is determined primarily by the structure of the water molecule. This structure also is responsible for the fact that water is a liquid, not a gas, at temperatures found on earth. Water is, with the exception of mercury, the only inorganic (not carbon-based) liquid found on Earth. Water is a simple compound, its individual molecules containing one oxygen atom and two much smaller hydrogen atoms. The elements are bonded together covalently, each hydrogen atom sharing its single electron with the oxygen.

Polarity

The arrangement of the three atoms in a water molecule is not symmetrical, as one might expect. Instead of the atoms being arranged linearly (H-O-H), the hydrogen atoms are attached to the oxygen in a V-shaped arrangement at an angle of only 105°. As shown in Figure 4.1, this results in an asymmetrical molecule with the shared electrons spending most of the time nearer to the oxygen than to the hydrogen. Consequently, the water molecule exhibits polarity; that is, the charges are not evenly distributed. Rather, the side on which the hydrogen atoms are located tends to be electropositive and the opposite side electronegative.

As a result, each water molecule does not act independently but rather is coupled with other neighboring molecules. The hydrogen (positive) end of one molecule attracts the oxygen (negative) end of another, resulting in a chainlike (polymer) grouping. Because its molecules cluster together, water has a much higher boiling point than other liquids with comparably low molecular weights (e.g., methyl alcohol).

Polarity also explains why water molecules are attracted to electrostatically charged ions and to colloidal surfaces. Cations such as W, Na^+, K^+, and Ca^{2+} become hydrated through their attraction to the oxygen (negative) end of water molecules. Likewise, negatively charged clay surfaces attract water, this time through the hydrogen (positive) end of the molecule. Polarity of water

molecules also encourages the dissolution of salts in water since the ionic components have greater attraction for water molecules than for each other.
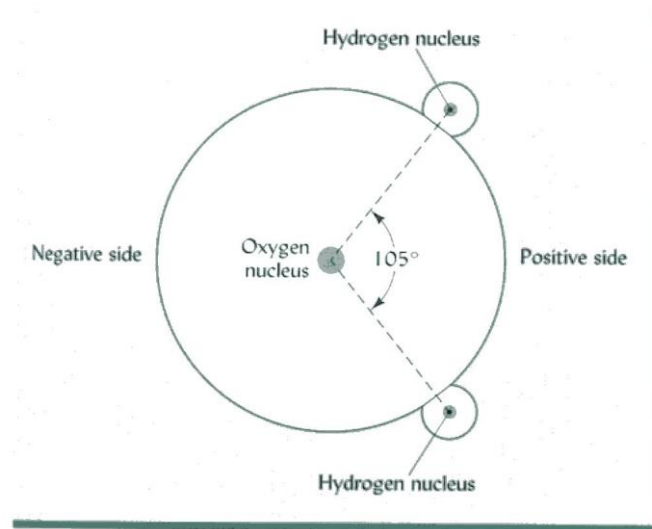

FIGURE 4.1 Two-dimensional representation of a water molecule showing a large oxygen atom and two much smaller hydrogen atoms. The HOH angle of 105° results in an asymmetrical arrangement. One side of the water molecule (that with the two hydrogens) is electropositive; the other is electronegative. This accounts for the polarity of water.

When water molecules become attracted to electrostatically charged ions or clay surfaces, they are more closely packed than in pure water. In this packed state their freedom of movement is restricted and their energy status is lower than in pure water. Thus, when ions or clay particles become hydrated, energy must be released. That released energy is evidenced as *heat of solution* when ions hydrate or as *heat of wetting* when clay particles become wet. The latter phenomenon can be demonstrated by placing some dry, fine clay in the palm of the hand and then adding a few drops of water. A slight rise in temperature can be felt.

Hydrogen Bonding

Through a phenomenon called hydrogen bonding, a hydrogen atom of one water molecule is attracted to the oxygen en of a neighboring water molecule, thereby forming a low energy bond between the two molecules. This type of bonding accounts for the polymerization of water. Hydrogen bonding also

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accounts for the relatively high boiling point, specific heat, and viscosity of water compared to the same properties of other hydrogen-containing compounds, such as H_2S, which has a higher molecular weight but no hydrogen bonding. It is also responsible for the structural rigidity of some clay crystals and for the structure of some organic compounds, such as proteins.

Cohesion versus Adhesion

Hydrogen bonding accounts for two basic forces responsible for water retention and movement in soils: the attraction of water molecules for each other (**cohesion**) and the attraction of water molecules for solid surfaces (**adhesion**). By adhesion (also called *adsorption*), some water molecules are held rigidly at the surfaces of soil solids. In turn, these tightly bound water molecules hold by cohesion other water molecules farther removed from the solid surfaces (Figure 4.2). Together, these forces of adhesion and cohesion make possible the property of plasticity possesses by clays.
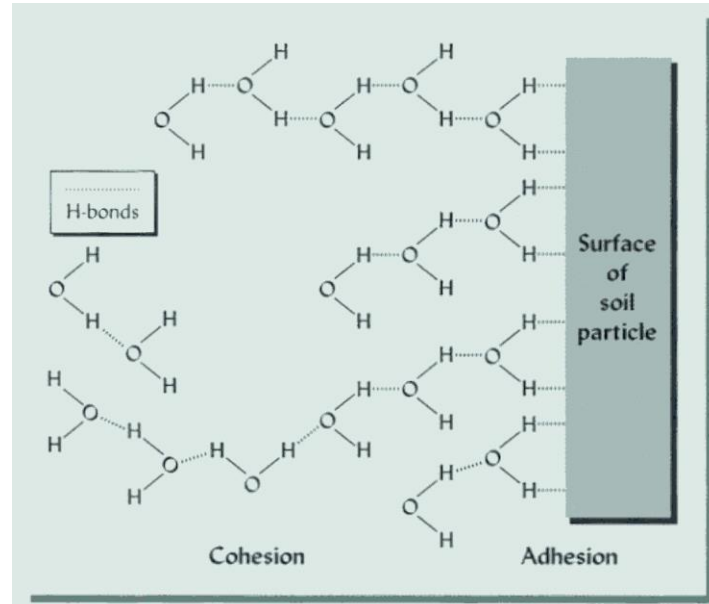
FIGURE 4.2 The forces of cohesion (between water molecules) and cohesion (between water and solid surface) in a soil-water system. The forces are largely a result of H-bonding shown as broken lines. The adhesive or adsorptive force diminishes rapidly with distance from the solid surface.

Surface Tension

Another important property of water that markedly influences its behavior in soils is that of surface tension. This property is commonly evidenced at liquid-air interfaces and results from the greater attraction of water molecules for each other (cohesion) than for the air above (Figure 4.3). The net effect is an inward force at the surface that causes water to behave as if its surface were covered with a stretched elastic membrane, and observation familiar to those who have seen insects walking on water in a pond (Figure 4.4). Because of the relatively high attraction of water molecules for each other, water has a high surface tension (72.8 newtons/mm at 20°C) compared to that of most other liquids (e.g., ethyl alcohol, 22.4 N/mm). As we shall see, surface tension is an important factor in the phenomenon of capillarity, which determines how water moves and is retained in soil.
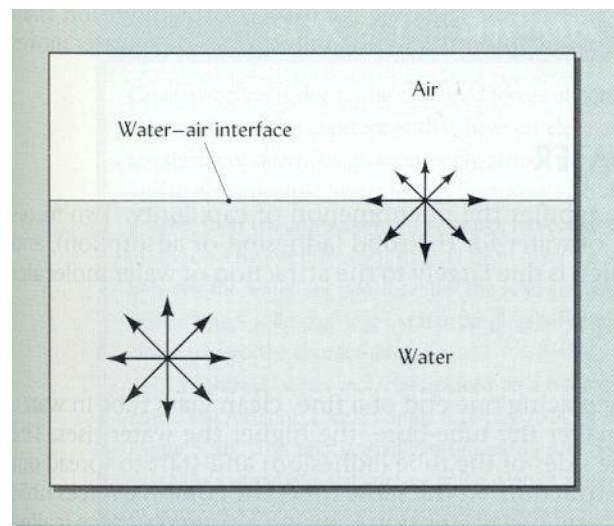
FIGURE 4.3 Comparative forces acting on water molecules at the surface and beneath the surface. Forces acting below the surface are equal in all directions since each water molecule is attracted equally by neighboring water molecules. At the surface, however, the attraction of the air for the water molecules is much less than that of water molecules for each other. Consequently, there is a net downward force on the surface molecules, and the result is something like a compressed film or membrane at the surface. This phenomenon is called *surface tension*.
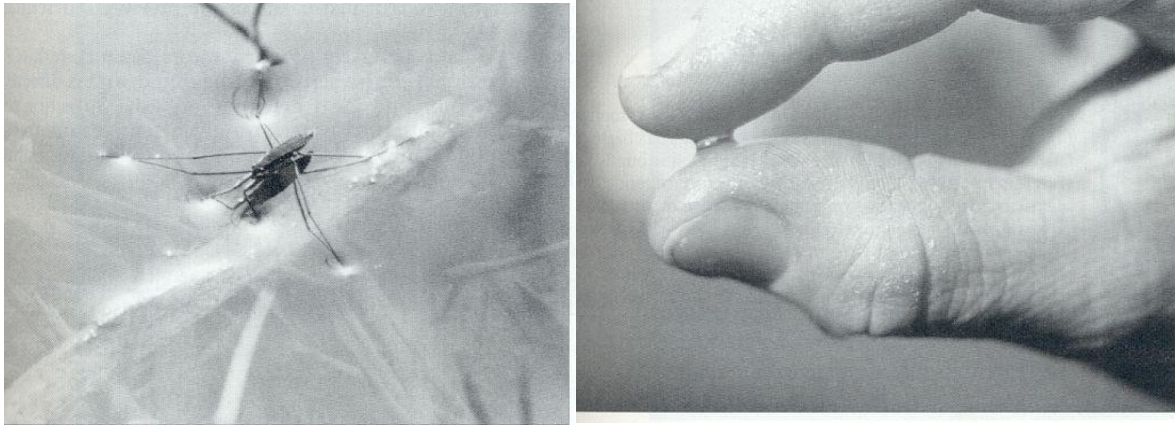
FIGURE 4.4 Everyday evidences of water's surface tension (left) as insects land on water and don not sink, and of forces of cohesion and adhesion (right) as a drop of water is held between the fingers.

4.2 CAPILLARY FUNDAMENTALS AND SOIL WATER

The movement of water up a wick typifies the phenomenon of capillarity. Two forces cause capillarity: (1) the attraction of water for the solid (adhesion or adsorption), and (2) the surface tension of water, which is due largely to the attraction of water molecules for each other (cohesion).

Capillary Mechanism

Capillarity can be demonstrated by placing one end of a fine, clean glass tube in water. The water rises in the tube; the smaller the tube bore, the higher the water rises. The water molecules are attracted to the sides of the tube (adhesion) and start to spread out along the glass in response to this attraction. At the same time, the cohesive forces hold the water molecules together and create surface tension, causing a curved surface (called a meniscus) to form at the interface between water and air in the tube. Lower pressure under the meniscus in the glass tube (P2) allows the higher pressure (P1) on the free water to push water up the tube. The process continues until the water in the tube has risen high enough that its weight just balances the pressure difference across the meniscus.

The height of rise in a capillary tube is inversely proportional to the tube radius r. Capillary rise is also inversely proportional to the density of the liquid, and is directly proportional to the liquid's surface tension and the degree of its adhesive attraction to the soil surface. By water we can calculate the height of the rise h with following equation:

$$h = \frac{0.15}{r}$$

where both h and r are expressed in centimeters. The equation tells us that the smaller the tube bore, the greater the capillary force and the higher the water rise in the tube (Figure 4.5).
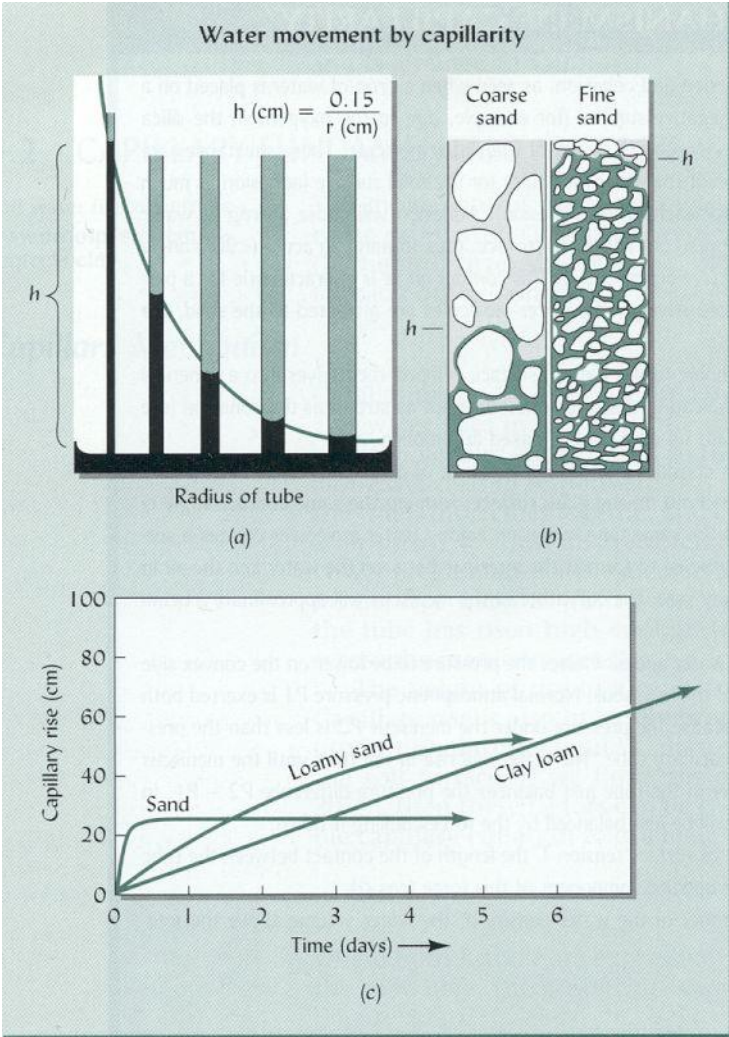
FIGURE 4.5 Upward capillarity movement of water through tubes of different bore and soils with

different pore sizes. The finer the soil texture, the greater the proportion of small-sized pores, and hence, the higher the ultimate rise of water above a free water table. However, because of the much greater frictional forces in the smaller pores, the capillarity rise is much slower in the finer-textured soil than in the sand!!!

Height and Rise in Soils

Capillary forces are at work in all moist soils. However, the rate of movement and the rise in height are less than one would expect on the basis of soil pore size alone. One reason is that soil pores are not straight, uniform openings like glass tubes. Furthermore, some soil pores are filled with air, which may be entrapped, slowing down or preventing the movement of water by capillarity (see Figure 4.5b).

Since capillary movement is determined by pore size, it is the pore-size distribution that largely determines the amount and rate of movement of capillary water in the soil. The abundance of medium- to large-sized capillary pores in sandy soils permits rapid initial capillary rise, but limits the ultimate height of rise (Figure 4.5c). Clays have a high proportion of very fine capillary pores, but frictional forces slow down the rate at which water moves through them. Consequently, in clays the capillary rise is slow initially, but in time it generally exceeds that of sands. Loams exhibit capillary properties between those of sands and clays.

Capillarity is traditionally illustrated as an upward adjustment. But movement in any direction takes place, since the attractions between soil pores and water are as effective in forming a water meniscus in horizontal pores as in vertical ones. The significance of capillarity in controlling water movement in small pores will become evident as we turn to soil water energy concepts.

4.3 SOIL WATER ENERGY CONCEPTS

The retention and movement of water in soils, its uptake and translocation in plants, and its loss to the atmosphere are all energy-related phenomena. Different kinds of energy are involved, including *potential energy* and *kinetic energy*. Kinetic energy is certainly an important factor in the rapid, turbulent flow of water in a river, but the movement of water in soil is so slow that the kinetic energy component is usually negligible. Potential energy is most

important in determining the status and movement of soil water. For the sake of simplicity, <u>in this text we will use the term energy to refer to potential energy</u>.

As we consider energy, we should keep in mind that all substances, including water, tend to move or change from a higher to a lower energy state. Therefore, if we know the pertinent energy levels at various points in a soil, we can predict the direction of water movement. It is the differences in energy levels from one contiguous site to another that influence this water movement.

Forces Affecting Potential Energy

The discussion of the structure and properties of water in the previous section suggests <u>three important forces affecting the energy level of soil water</u>. First, adhesion, or the attraction of water to the soil solids (matrix), provides a **matric force** (responsible for adsorption and capillarity) that markedly reduces the energy state of water near particle surfaces. Second, the attraction of water to ions and other solutes, resulting in **osmotic forces**, tends to reduce the energy state of water in the soil solution. Osmotic movement of pure water across a semipermeable membrane into a solution (osmosis) is evidence of the lower energy state of water in the solution. The third major force acting on soil water is **gravity**, which always pulls the water downward. The energy level of soil water at a given elevation in the profile is thus higher than that of water at some lower elevation. This difference in energy level causes water to flow downward.

Soil Water Potential

The difference in energy level of water from one site or one condition (e.g., in wet soil) to another (e.g., in dry soil) determines the direction and rate of water movement in soils and in plants. In a wet soil, most of the water is retained in large pores or thick water films around particles. Therefore, most of the water molecules in a wet soil are not very close to a particle surface and so are not held very tightly by the soil solids (the matrix). In this condition, the water molecules have considerable freedom of movement, so their energy

level is near that of water molecules in a pool of pure water outside the soil. In a drier soil, however, the water that remains is located in small pores and thin water films, and is therefore held tightly by the soil solids. Thus the water molecules in a drier soil have little freedom of movement, and their energy level is much lower than that of the water in wet soil. If wet and dry soil samples are brought in touch with each other, water will move from the wet soil (higher energy state) to the drier soil (lower energy).

Determining the absolute energy level of soil water is a difficult and sometimes impossible task. Fortunately, it is not necessary to know the absolute energy level of water to be able to predict how it will move in soils and in the environment. Relative values of soil water energy are all that is needed. Usually the energy status of soil water in a particular location in the profile is compared to that of pure water at standard pressure and temperature, unaffected by the soil and located at some reference elevation. The difference in energy levels between this pure water in the reference state and that of the soil water is termed **soil water potential** (Figure 4.6), the term potential, like the term pressure, implying a difference in energy status.

If all water potential values under consideration have a common reference point (the energy state of pure water), differences in the water potential of two soil samples in fact reflect differences in their absolute energy levels. This means that water will move from a soil zone having a high soil water potential to one having a lower soil water potential. This fact should be always kept in mind in considering the behavior of water in soils.

The soil water potential is due to several forces, each of which is a component of the **total soil water potential Ψ_t**. These components are due to differences in energy levels resulting from gravitational, matric, submerged hydrostatic, and osmotic forces and are termed **gravitational potential Ψ_g, matric potential Ψ_m, submergence potential Ψ_s**, and **osmotic potential Ψ_o**, respectively. All of these components act simultaneously to influence water behavior in soils. The general relationship of soil water potential to potential energy levels is shown in Figure 4.6 and can be expressed as:

$$\Psi_t = \Psi_g + \Psi_m + \Psi_o + \Psi_s + \ldots..$$

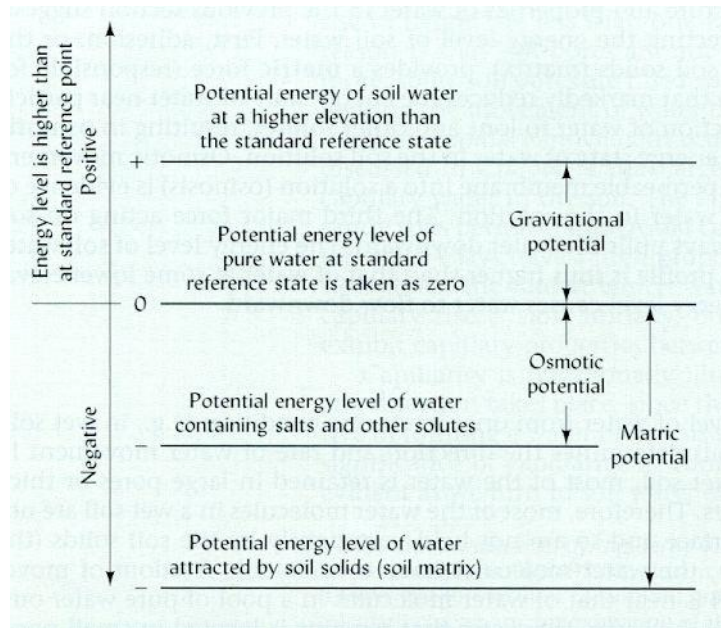where the ellipses (…..) indicate the possible contribution of additional potentials not yet mentioned.

FIGURE 4.6 Relationship between the potential energy of pure water at a standard reference state (pressure, temperature, and elevation) and that of soil water. If the soil water contains salts and other solutes, the mutual attraction between water molecules and these chemicals reduces the potential energy of the water, the degree of the reduction being termed osmotic potential. Similarly, the mutual attraction between soil solids (soil matrix) and soil water molecules also reduces the water's potential energy. In this case the reduction is called matric potential. Since both of these interactions reduce the water's potential energy level compared to that of pure water, the changes in energy level (osmotic potential and matric potential) are both considered to be negative. In contrast, differences in energy due to gravity (gravitational potential) are always positive. This is because the reference elevation of the pure, water is purposely designated at a site in the soil profile below that of the soil water. A plant root attempting to remove water from a moist soil would have to overcome alt three forces simultaneously.

Gravitational Potential

The force of gravity acts on soil water the same as it does on any other body, the attraction being toward the earth's center. The gravitational potential Vrg of soil water may be expressed mathematically as:

$$\Psi_g = g * h$$

Where *g* is the acceleration due to gravity and *h* is the height of the soil water above a reference elevation. The reference elevation is usually chosen within the soil profile or at its lower boundary to ensure that the gravitational potential of soil water above the reference point will always be positive.

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Following heavy precipitation or irrigation, gravity plays an important role in removing excess water from the upper horizons and in recharging groundwater below the soil profile.

Pressure Potential (Including Hydrostatic and Matric Potentials)

This component accounts for the effects on soil water potential of all factors other than gravity and solute levels. It most commonly includes (1) the positive hydrostatic pressure due to the weight of water in saturated soils and aquifers, and (2) the negative pressure due to the attractive forces between the water and the soil solids or the soil matrix.

The hydrostatic pressures give rise to what is often termed the **hydrostatic potential Ψ_s**, a component that is operational only for water in saturated zones below the water table. Anyone who has dived to the bottom of a swimming pool has felt hydrostatic pressure on the eardrums.

The attraction of water to solid surfaces gives rise to the **matric potential Ψ_m**, which is always negative because the water attracted by the soil matrix has an energy state lower than that of pure water. (These negative pressures are sometimes referred to as *suction* or *tension*). The *matric potential* is operational in unsaturated soil above the water table, while the *submergence potential* applies to water in saturated soil or below the water table (Figure 4.7).

While each of these pressures is significant in specific field situations, the matric potential is important in all unsaturated soils because there are omnipresent interactions between soil solids and water. The movement of soil water, the availability of water to plants, and the solutions to many civil engineering problems are determined to a considerable extent by matric potential.
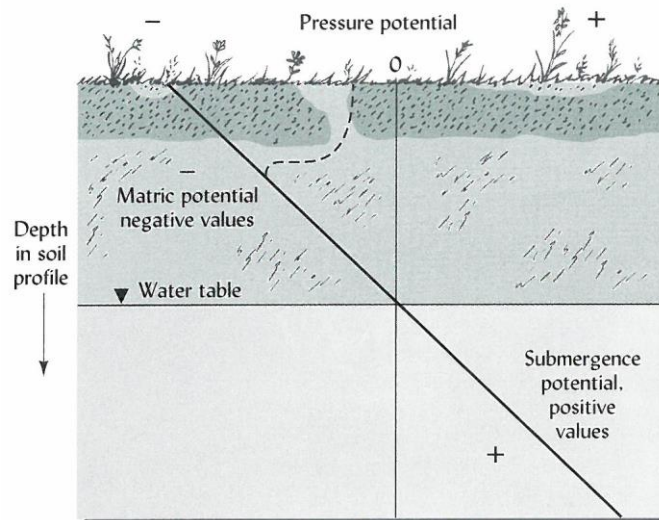
FIGURE 4.7 The matric potential and submergence potential are both pressure potentials that may contribute to total water potential. The matric potential is always negative and the submergence potential is positive. When water is in unsaturated soil above the water table (top of saturated zone) it is subject to the influence of matric potentials. Water below the water table in saturated soil is subject to submergence potentials. In the example shown here, the matric potential decreases linearly with elevation above the water table, signifying that water rising by capillary attraction up from the water table is the only source of water in this profile. Rainfall or irrigation (see dotted line) would alter or curve the straight line, but would not change the fundamental relationship described.

Matric potential Ψ_m, which results from the phenomena of adhesion (or adsorption) and of capillarity, influences soil moisture retention as well as soil water movement. Differences between the Ψ_m of two adjoining zones of a soil encourage the movement of water from moist zones (high energy state) to dry zones (low energy state) or from large pores to small pores. Although this movement may be slow, it is extremely important, especially in supplying water to plant roots.

Osmotic Potential

The **osmotic potential Ψ_o** is attributable to the presence of solutes in the soil solution. The solutes may be inorganic salts or organic compounds. Their presence reduces the potential energy of water, primarily because of the reduced freedom of movement of the water molecules that cluster around each solute ion or molecule. The greater the concentration of solutes, the more the osmotic potential is lowered. As always, water will tend to move to

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where its energy level will be lower, in this case to the zone of higher solute concentration. However, liquid water will move in response to differences in osmotic potential (the process termed **osmosis**) only if a *semipermeable membrane* exists between the zones of high and low osmotic potential, allowing water through but *preventing the movement of the solute*. If no membrane is present, the solute, rather than the water, generally moves to equalize concentrations.

Because soil zones are not generally separated by membranes, the osmotic potential Ψ_o has little effect on the mass movement of water in soils. Its major effect is on the uptake of water by plant root cells that *are* isolated from the soil solution by their semipermeable cell membranes. In soils high in soluble salts, Ψ_o may be lower (have a greater negative value) in the soil solution than in the plant root cells. This leads to constraints in the uptake of water by the plants. In very salty soil, the soil water osmotic potential may be low enough to cause cells in young seedlings to collapse (plasmolyze) as water moves from the cells to the lower osmotic potential zone in the soil.

The random movement of water molecules causes a few of them to escape a body of liquid water, enter the atmosphere, and become water vapor. Since the presence of solutes restricts the movement of water molecules, fewer water molecules escape into the air as the solute concentration of liquid water is increased. Therefore, water vapor pressure is lower in the air over salty water than in the air over pure water. By affecting water vapor pressure, Ψ_o affects the movement of water vapor in soils. The process of osmosis and the relationship between the matric and osmotic components of total soil water potential is shown in Figure 4.8.
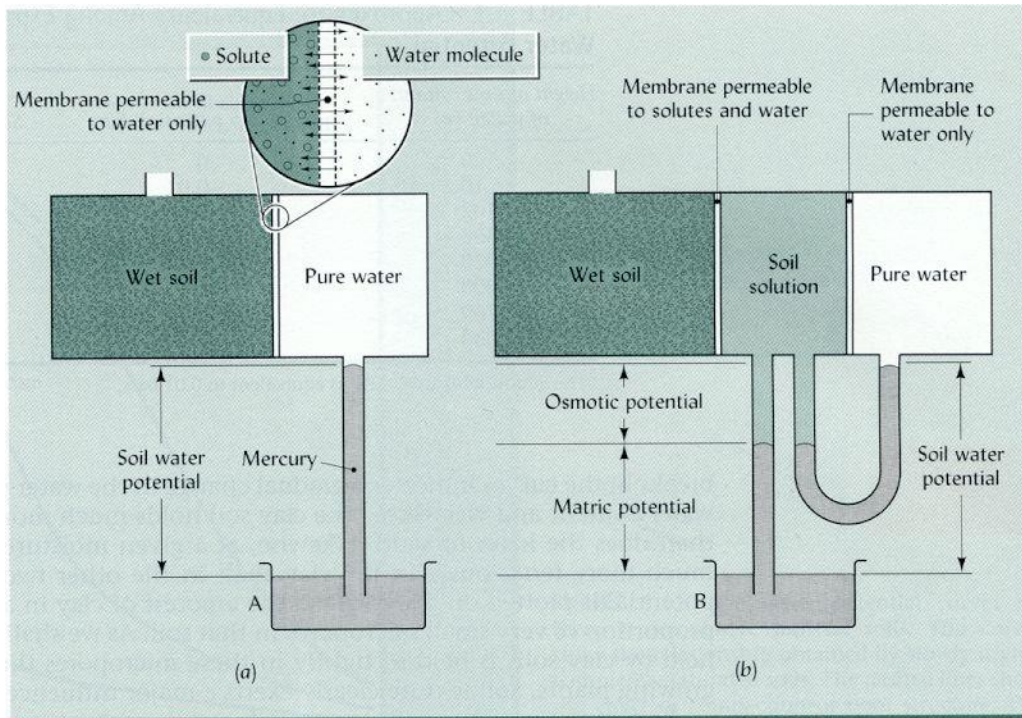
FIGURE 4.8 Relationships among osmotic, matric, and combined soil water potentials. (Left) Assume a container of soil separated from pure water by a membrane permeable only to water (see inset showing osmosis across the membrane). The pure water is connected to a vessel of mercury through a tube. Water will move into the soil in response to the matric forces attracting water to soil solids and the osmotic forces attracting water to solutes. At equilibrium the height of the mercury column above vessel A is a measure of this combined soil water potential (matric plus osmotic). (Right). Assume a second container is placed between the pure water and the soil, and this container is separated from the soil by a fine screen permeable to both solutes and water. Ions will move from the soil into this second container until the concentration of solutes in this water and in the soil water have equalized. Then the difference between the potential energies of this solution and of the pure water gives a measure of the osmotic potential. The matric potential, as measured by the column of mercury above vessel B, would then be the difference between the combined soil water potential and the osmotic component. The gravitational potential (not shown) is the same for alt compartments and does not affect the outcome since the water movement is horizontal. [Modified from Richards (1965)]

4.4 SOIL WATER CONTENT AND SOIL WATER POTENTIAL

Soil water versus Energy Curves

The relationship between soil water potential Ψ and moisture content θ of three soils of different textures is shown in Figure 4.9. Such curves are sometimes termed water release characteristic curves, or simply water characteristic curves. The absence of sharp breaks in the curves indicates a gradual change in the water potential with increased soil water content an

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vice versa. The clay soil holds much more water at a given potential than does the loam or sand. Likewise, at a given moisture content, the water is held much more tenaciously in the clay than in the other two soils. The amount of clay in a soil largely determines the proportion of very small micropores in that soil. As we shall see, about half of the water held by clay soils is held so tightly in these micropores that it cannot be removed by growing plants. Soil texture clearly exerts a major influence on soil moisture retention.

Soil structure also influences soil water content – energy relationships. A well-granulated soil has more total pore space and greater overall water-holding capacity than one with poor granulation or one that has been compacted.
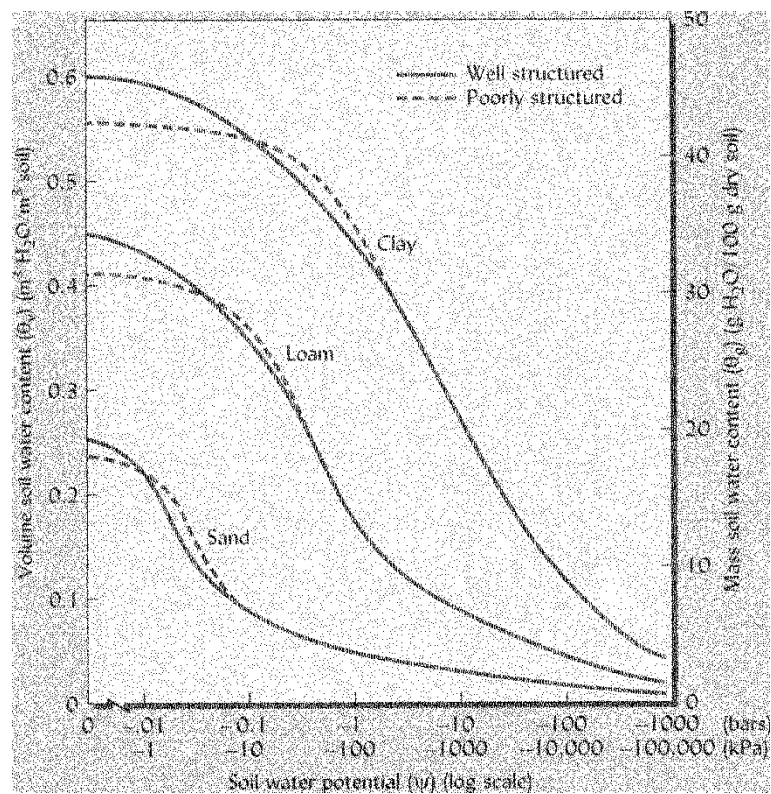

FIGURE 4.9 Soil water potential curves for three representative mineral soils. The curves show the relationship obtained by slowly drying completely saturated soils. The dashed lines show the effect of compaction or poor structure. The soil water potential Ψ (which is negative) is expressed in terms of bars (upper scale) and kilopascals (kPa) (lower scale). Note that the soil water potential is plotted on a log scale.
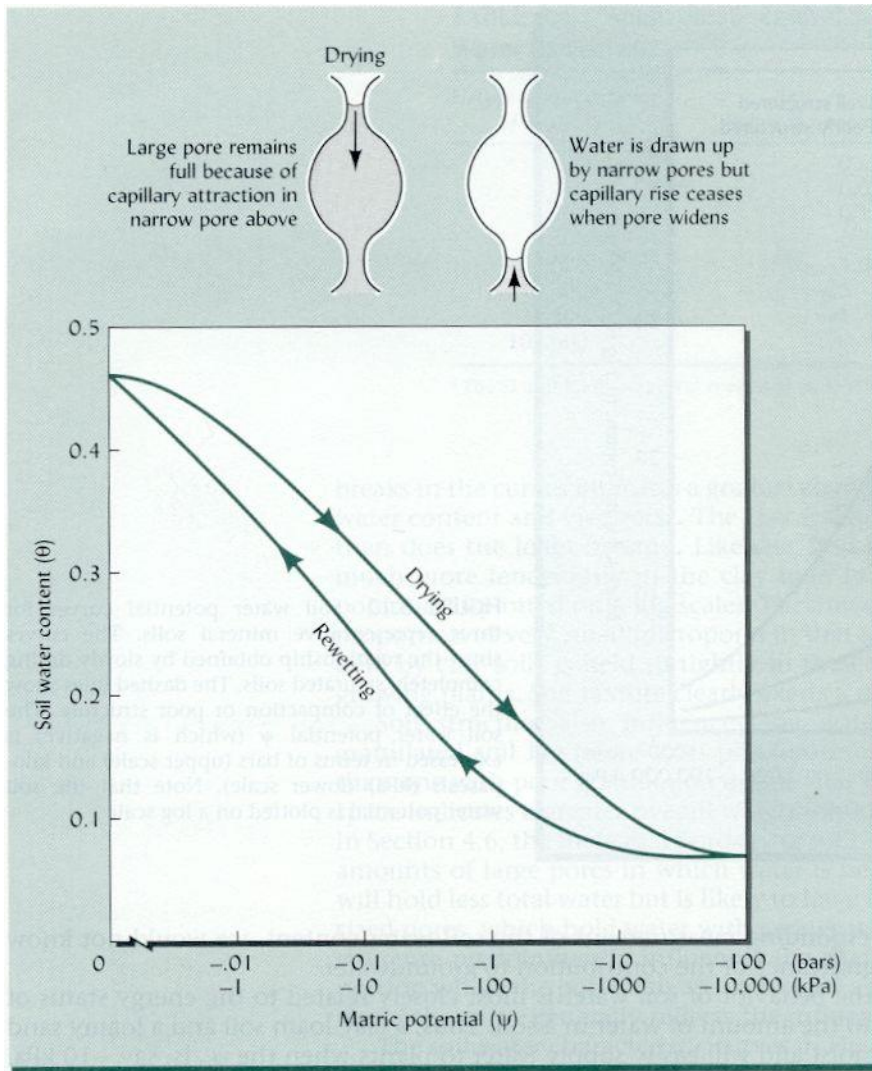
FIGURE 4.10 The relationship between soil water content and matric potential of a soil upon being dried and then rewetted. The phenomenon, known as hysteresis, is apparently due to factors such as the nonuniformity of individual soil pores, entrapped air and the swelling and shrinking that might affect soil structure. The drawings show the effect of nonuniformity of pores.

Hysteresis

The relationship between soil water content and potential, determined as a soil dries out, will differ somewhat from the relationship measured as the same soil is rewetted. This phenomena, known as **hysteresis**, is illustrated in Figure 4.10. Hysteresis is caused by a number of factors, including the nonuniformity of soil pores, swelling and shrinking of clays, etc.

Measurements of Soil Water Status

The soil water characteristic curves just discussed highlight the importance of making two general kinds of soil water measurements: the *amount* of water present (*water content*) and the *energy status* of the water (*soil water potential*). In order to understand or manage water supply and movement in soils it is essential to have information (directly measured or inferred) on *both* types of measurements.

Generally the behavior of soil water is most closely related to the energy status of the water, not to the amount of water in a soil. Thus, a clay loam soil and a loamy sand will both feel moist and will easily supply water to plants when the Ψ_m is, say, -10 kPa. However, the amount of water held by the clay loam, and thus the length of time it could supply water to plants, would be far greater at this potential than would be the case for the loamy sand.

We will mention several methods for making each of these two types of soil water measurements. Researchers, land managers, and engineers may use a combination of several of these methods to study the storage and movement of water in soil, manage irrigation systems, and predict the physical behavior of soils.

Measuring Soil Water Content

The **volumetric water content θ** is defined as the volume of water associated with a given volume (usually 1 m^3) of dry soil. A comparable expression is the **mass water content θ$_m$**, or the mass of water associated with a given mass (usually 1 kg) of dry soil. Both of these expressions have advantages for different uses. In most cases we shall use the volumetric water content θ in this text.

Because in the field we think of plant root systems as exploring a certain depth of soil, and because we express precipitation (and sometimes irrigation) as a depth of water (e.g., mm of rain), it is often convenient to express the volumetric water content as a depth ratio (depth of water per unit depth of soil). Conveniently, the numerical values for these two expressions are the same. For example, for a soil containing 0.1 m^3 of water per m^3 of soil (10% by volume) the depth ratio of water is 0.1 m of water per m of soil depth.

Some methods for measuring the soil water content are shown in **Table 4.1**. From all these methods, the *gravimetric method* is a direct measurement of soil water content, and is therefore the standard method by which all indirect methods are calibrated.

TABLE 4.1 Some Methods of Measuring Soil Water

Note that more than one method may be needed to cover the entire range of soil moisture conditions.

Method	Measures soil		Useful range, kPa	Used mainly in		Comments
	Content	Potential		Field	Lab	
1. Gravimetric	x		0 to <-10,000		x	Destructive sampling; slow (1 to 2 days) unless microwave used. The standard for
2. Resistance blocks		x	-100 to <-1,500	x		Can be automated; not sensitive near optimum plant water contents.
3. Neutron scattering	x		0 to <-1,500	x		Radiation permit needed; expensive equipment; not good in high-organic-matter soils; requires access tube.
4. Time domain reflectometry (TDR)	x		0 to <-10,000	x	x	May be automated; accurate to 1 kPa; requires wave guides; expensive instrument.
5. Tensiometer		x	0 to -85	x		Accurate to 0.1 to 1 kPa; limited range; inexpensive; can be automated; needs periodic
6. Thermocouple psychrometer		x	50 to <-10,000	x	x	Moderately expensive; wide range; accurate only to ±50 kPa.
7. Pressure membrane apparatus		x	50 to <-10,000		x	Used in conjunction with gravimetric method to construct water characteristic curve.

Measuring Soil Water Potentials

TENSIOMETERS. The tenacity with which water is held in soils is an expression of soil water potential Ψ. Field tensiometers (Figure 4.11) measure this attraction or tension. The tensiometer is basically a water-filled tube closed at the bottom with a porous ceramic cup and at the top with an airtight seal. Once placed in the soil, water in the tensiometer-moves through the porous cup into the adjacent soil until the water potential in the tensiometer is the same as the matric water potential in the soil. As the water is drawn out, a

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vacuum develops under the top seal, which can be measured by a vacuum gauge or an electronic transducer. If rain or irrigation rewets the soil, water will enter the tensiometer through the ceramic tip, reducing the vacuum or tension recorded by the gauge.

Tensiometers are useful between 0 and -85 kPa potential, a range that includes half or more of the water stored in most soils. Laboratory tensiometers called *tension plates* operate over a similar range of potentials. As the soil dries beyond -80 to -85 kPa, tensiometers fail because air is drawn in through the pores of the ceramic, relieving the vacuum. A solenoid switch can be fitted to a field tensiometer in order to automatically turn an irrigation system on and off.
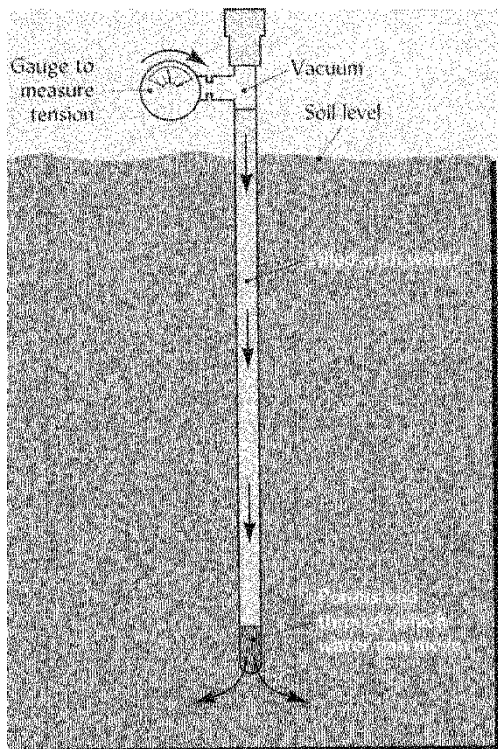

FIGURE 4.11 Tensiometer method of determining water potential in the field. Cross section showing the essential components of a tensiometer. Water moves through the porous end of the instrument in response to the pull (matric potential) of the soil. The vacuum so created is measured by a gauge that reads in kPa of tension (-kPa water potential).

4.5 THE FLOW OF LIQUID WATER IN SOIL

Three types of water movement within the soil are recognized: (1) saturated flow, (2) unsaturated flow, and (3) vapor movement. In all cases water flows in response to energy gradients, with water moving from a zone of higher to one of lower water potential. Saturated flow takes place when the soil pores are completely filled (or saturated) with water. Unsaturated flow occurs when the larger pores in the soil are filled with air, leaving only the smaller pores to hold and transmit water. Vapor movement occurs as vapor pressure differences develop in relatively dry soils.

Saturated Flow Through Soils

Under some conditions, at least part of a soil profile may be completely saturated; that is, all pores, large and small, are filled with water. The lower horizons of poorly drained soils are often saturated, as are portions of well-drained soils above stratified layers of clay. During and immediately following a heavy rain or irrigation, pores in the upper soil zones are often filled entirely with water (Figure 4.12).
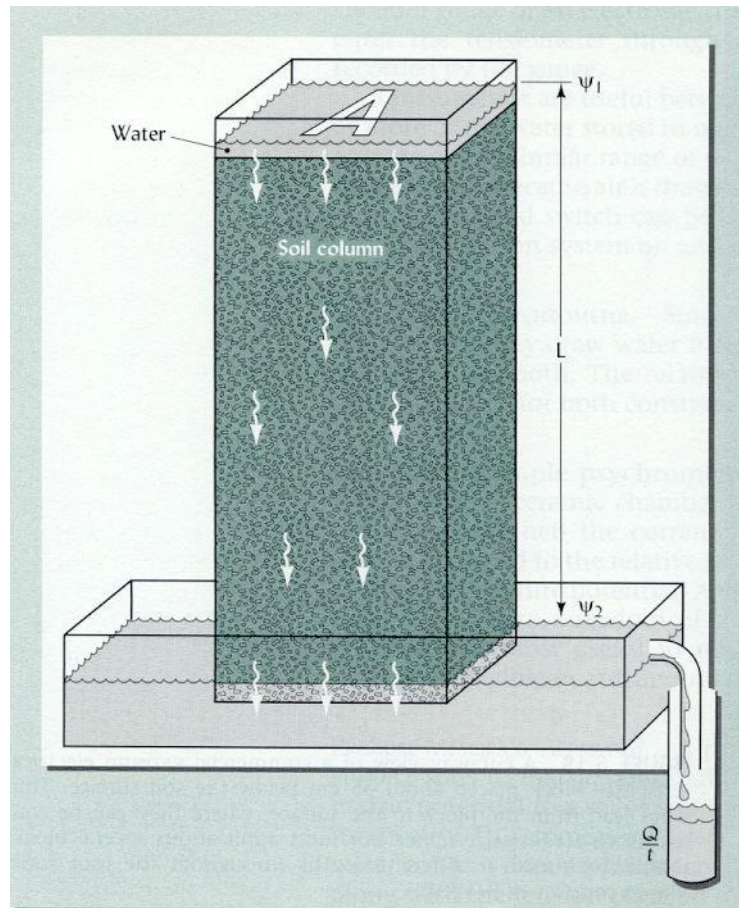
FIGURE 4.12 Saturated flow (percolation) in a column of soil with cross-sectional area A, cm². All soil pores are filled with water.

The quantity of water per unit of time Q/t that flows through a column of saturated soil can be expressed by Darcy's law, as follows:

$$\frac{Q}{t} = A \times K_{sat} \times \frac{\Delta\Psi}{L}$$

where A is the cross-sectional area of the column through which the water flows, K_{sat} is the **saturated hydraulic conductivity**, $\Delta\Psi$ is the change in water potential between the ends of the column (for example, Ψ_1 - Ψ_2), and L is the length of the column. For a given column, the rate of flow is determined by the ease with which the soil transmits water (K_{sat}) and the amount of force driving the water, namely the **water potential gradient** $\Delta\Psi/L$. For saturated flow this force may also be called the **hydraulic gradient**.

Factors Influencing the Hydraulic Conductivity of Saturated Soils

Any factor affecting the size and configuration of soil pores will influence hydraulic conductivity. The total flow rate in soil pores is proportional to the fourth power of the radius. Thus, flow through a pore 1 mm in radius (say, a small earthworm channel) is equivalent to that in 10,000 pores with a radius of 0.1 mm even though it takes only 100 pores of radius 0.1 mm to give the same cross-sectional area as a 1 mm pore. As a result, macropores (radius > 0.05 mm) account for most water movement in saturated soils. The presence of biopores, such as root channels and earthworm burrows (typically >1 mm in radius), may have a marked influence on the saturated hydraulic conductivity of different soil horizons. Because they usually have more macropore space, sandy soils generally have higher saturated conductivities than finer-textured soils. Likewise, soils with stable granular structure conduct water much more rapidly than do those with unstable structural units, which break down upon being wetted. Saturated conductivity of soils under natural vegetation is commonly much higher than where cultivated crops have been grown. Entrapped air, which is common in recently wetted soils, can slow down the movement of water and thereby reduce hydraulic conductivity.

PREFERENTIAL FLOW. Scientists have often noted more extensive pollution of groundwater from pesticides and other toxicants than would be predicted from the traditional hydraulic conductivity measurements. Apparently the toxicants are carried down rapidly by rain or irrigation water that moves through large macropores, cracks, and other such pathways, often before the bulk of the soil is thoroughly wetted. Because the water "prefers" these larger pathways, this type of flow is referred to as **preferential flow**. Mounting evidence suggests that this type of water movement greatly increases groundwater pollution.

Any process that stimulates the formation of macropores with continuity down through the profile will encourage preferential flow. Burrowing animals (e.g., worms, moles, and gophers) as well as decayed plant roots leave tubular channels through which water can flow rapidly.

Preferential flow in fine-textured soils is enhanced by the shrinkage of the clay fraction, which leaves open cracks and fissures that can extend down into

the lower subsoil horizons. In some clay soils, water from the first rains after a dry spell moves rapidly down the profile, carrying with it the more soluble toxicants.

Unsaturated Flow in Soils

Most of the time, water movement takes place when upland soils are unsaturated. Such movement occurs in a more complicated environment than that which characterizes saturated water flow. In saturated soils, essentially all the pores are filled with water, although the most rapid water movement is through the large and continuous pores. But in unsaturated soils, these macropores are filled with air, leaving only the finer pores to accommodate water movement. Also, in unsaturated soils the water content and, in turn, the tightness with which water is held (water potential) can be highly variable. This influences the rate and direction of water movement and also makes it more difficult to measure the flow of soil water.

As was the case for saturated water movement, the driving force for unsaturated water flow is differences in water potential. This time, however, the difference in the matric potential, not gravity, is the primary driving force. This **matric potential gradient** is the difference in the matric potential of the moist soil areas and nearby drier areas into which the water is moving. Movement will be from a zone of thick moisture films (high matric potential, e.g., -1 kPa) to one of thin films (lower matric potential, e.g., -100 kPa).

INFLUENCE OF TEXTURE. Figure 4.13 shows the general relationship between matric potential Ψ_m, (and, in turn, water content) and hydraulic conductivity of a sandy loam and clay soil. Note that at or near zero potential (which characterizes the saturated flow region), the hydraulic conductivity is thousands of times greater than at potentials that characterize typical unsaturated flow (-10 kPa and below).
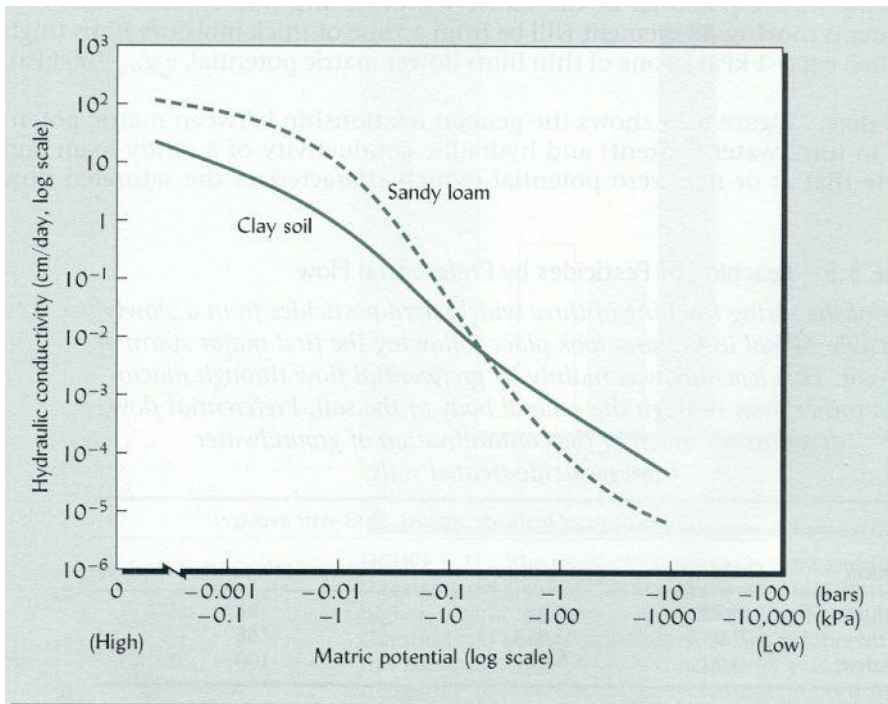
FIGURE 4.13 Generalized relationship between matric potential and hydraulic conductivity for a sandy soil and a clay soil (note log scales). Saturation flow takes place at or near zero potential, while much of the unsaturated flow occurs at a potential of -0.1 bar (-10 kPa) or below.

At high potential levels (high moisture contents), hydraulic conductivity is higher in the sand than in the clay. The opposite is true at low potential values (low moisture contents). This relationship is to be expected because the sandy soil contains mainly large pores which are water-filled when the soil water potential is high (and the soil is quite wet), but most of these have been emptied by the time the soil water potential becomes lower than about -10 kPa. The clay soil has many more micropores which are still water-filled at lower soil water potentials (drier soil conditions) and can participate in unsaturated flow.

4.6 INFILTRATION AND PERCOLATION

A special case of water movement is the entry of free water into the soil at the soil-atmosphere interface. This is a pivotal process in landscape hydrology that greatly influences the moisture regime for plants and the potential for soil degradation, chemical runoff, and down-valley flooding. The source of free water at the soil surface may be rainfall, snowmelt, or irrigation.

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Infiltration

The process by which water enters the soil pore spaces and becomes soil water is termed **infiltration**, and the rate at which water can enter the soil is termed **infiltrability** i:

$$i = \frac{Q}{A \times t}$$

where Q is the volume quantity of water (m^3) infiltrating, A is the area of the soil surface (m^2) exposed to infiltration, and t is time (s). Since m^3 appears in the numerator and m^2 in the denominator, the units of infiltration can be simplified to m/s or, more commonly, cm/h. The infiltration rate is not constant over time, but generally decreases during an irrigation or rainfall episode. If the soil is quite dry when infiltration begins, all the macropores open to the surface will be available to conduct water into the soil. In soils with expanding types of clays, the initial infiltration rate may be particularly high as water pours into the network of shrinkage cracks. However, as infiltration proceeds, many macropores fill with water and shrinkage cracks close up. The infiltration rate declines sharply at first and then tends to level off, remaining fairly constant thereafter (Figure 4.14).
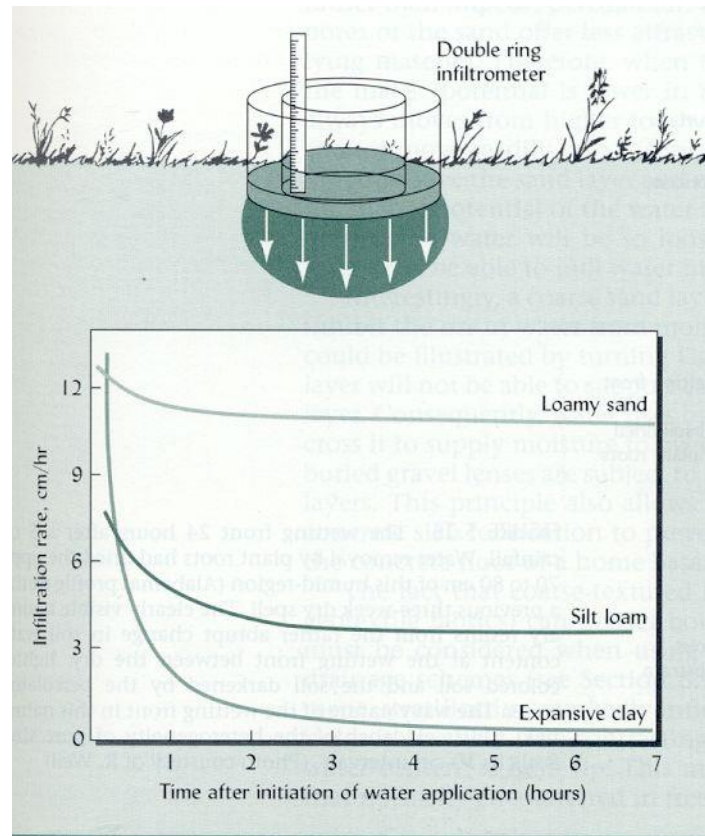
FIGURE 4.14 The potential rate of water entry into the soil, or infiltration capacity, can be measured by recording the drop in water level in a double ring infiltrometer (top). Changes in the infiltration rate of several soils during a period of water application by rainfall or irrigation are shown (bottom). Generally, water enters a dry soil rapidly at first, but its infiltration rate slows as the soil becomes saturated. The decline is least for very sandy soils with macropores that do not depend on stable structure or clay shrinkage. In contrast, a soil high in expansive clays may have a very high initial infiltration rate when large cracks are open, but a very low infiltration rate once the clays swell with water and close the cracks. Most soils fall between these extremes, exhibiting a pattern similar to that shown for the silt loam soil.

Percolation

Infiltration is a transitional phenomenon that takes place at the soil surface. Once the water has infiltrated the soil, the water moves downward into the profile by the process termed **percolation**. Both saturated and unsaturated flow are involved in percolation of water down the profile, and rate of percolation is related to the soil's hydraulic conductivity. In the case of water that has infiltrated a relatively dry soil , the progress of water movement can be observed by the darkened color of the soil as it becomes wet. There usually appears to be a sharp boundary, termed a **wetting front**, between the dry underlying soil and the soil already wetted. During an intense rain or heavy

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irrigation, water movement near the soil surface occurs mainly by saturated flow in response to gravity. At the wetting front, however, water is moving into the underlying drier soil in response to matric potential gradients as well as gravity. During a light rain, both infiltration and percolatinon may occur mainly by unsaturated flow as water is drawn by matric forces into the fine entrapped pores without accumulating at the soil surface or in the macropores.

Water movement in Stratified Soils

The fact, that at the wetting front, water is moving by unsaturated flow has important effects on how percolating water behaves when it encounters an abrupt change in pore sizes due to such layer as fragipans (strongly compacted layers) or claypans, or sand and gravel lenses.

In all cases, the effect on water percolation is similar - that is, the downward movement is impeded - even though the causal mechanism may vary. It is not surprising that percolating water should slow down markedly when it reaches a layer with finer pores, which therefore has a lower hydraulic conductivity. However, the fact that a layer of coarser pores will temporarily stop the movement of water may not be obvious.

Let's take a case, when a layer of coarse sand impedes downward movement of water in an otherwise fine-textured soil. Intuitively, one might expect the sand layer to speed, rather than impede, percolation. However, it has the opposite effect because the macropores of the sand offer less attraction for the water than do the finer pores of the overlying material. Therefore, when the unsaturated wetting front reaches the sand layer the matric potential is lower in the overlying material than in the sand. Since water always moves from higher to lower potential (to where it will be held more tightly), it cannot move readily into the sand. Eventually the downward-moving water will accumulate above the sand layer and nearly saturate the pores at the soil-sand interface (i.e., the matric potential of the water at the wetting front will fall to nearly zero). Once this occurs, the water will be so loosely held by the fine-textured soil that gravitational forces will be able to pull water into the sand layer.

Interestingly, a coarse sand layer in an otherwise fine-textured soil profile would also inhibit the rise of water from moist subsoil layers up to the surface soil. The large pores in the coarse layer will not be able to support capillary movement up from the smaller pores in a finer layer. Consequently, water rises by capillarity up to the bottom of the coarse-textured layer but cannot cross it to supply moisture to overlying layers. Thus, plants growing on some soils with buried gravel lenses are subject to drought since they are unable to exploit the lower soil layers. This principle also allows a layer of gravel to act as a capillary barrier under a concrete slab foundation to prevent water from soaking up from the soil and through the concrete floor of a home basement.

4.7 INFILTRATION AND PERCOLATION

Two types of water vapor movement occur in soils, internal and external. Internal movement takes place within the soil, that is, in the soil pores. External movement occurs at the land surface, and water vapor is lost by surface evaporation.

Water vapor moves from one point to another within the soil in response to differences in vapor pressure. Thus, water vapor will move from a moist soil where the soil air is nearly 100% saturated with water vapor (high vapor pressure) to a drier soil where the vapor pressure is somewhat lower. Also, water vapor will move from a zone of low salt content to one with a higher salt content (e.g., around a fertilizer granule). The salt lowers the vapor pressure of the water and encourages water movement from the surrounding soil.

If the temperature of one part of a uniformly moist soil is lowered, the vapor pressure will decrease and water vapor will tend to move toward this cooler part. Heating will have the opposite effect in that heating will increase the vapor pressure and the water vapor will move away from the heated area.

The actual amount of water vapor in a soil at optimum moisture for plant growth is surprisingly small, being perhaps no more than 10 liters in the upper 15 cm of a hectare of a silt loam soil. This compares with some 600,000 liters of liquid water in the same soil volume. Even though the amount of water vapor is small, its movement in soils can be of some practical significance. For example, seeds of some plants can absorb sufficient water vapor from the soil

to stimulate germination. Likewise, in dry soils, water vapor movement may be of considerable significance to drought-resistant desert plants (*xerophytes*), many of which can exist at extremely low soil water contents. For instance, at night the surface horizon of a desert soil may cool sufficiently to cause vapor movement up from deeper layers. If cooled enough, the vapor may then condense as dewdrops in the soil pores, supplying certain shallow-rooted xerophytes with water for survival.

4.8 QUALITATIVE DESCRIPTION OF SOIL WETNESS

The measurement of soil water potential and the observable behavior of soil water always depend on that portion of the soil water that is farthest from a particle surface and therefore has the highest potential. As an initially water-saturated soil dries down, both the soil as a whole and the soil water it contains undergo a series of gradual changes in physical behavior and in their relationships with plants.

To study these changes and introduce the terms commonly uses to describe varying degrees of soil wetness, we shall follow the moisture and energy status of soil during and after a heavy rain or the application of irrigation water.

Maximum Retentive Capacity

When all soil pores are filled with water from rainfall or irrigation, the soil is said to be saturated with respect to water (Figure 4.15) and at its **maximum retentive capacity**. The matric potential is close to zero, nearly the same as that of pure water. The volumetric water content is essentially the same as the total porosity. The soil will remain at maximum retentive capacity only so long as water continues to infiltrate, for the water in the largest pores (sometimes termed **gravitational water**) will percolate downward, mainly under the influence of gravitational forces. Data on maximum retentive capacities and the average depth of soils in a watershed are useful in predicting how much rainwater can be stored in the soil temporarily, thus possibly avoiding downstream floods.
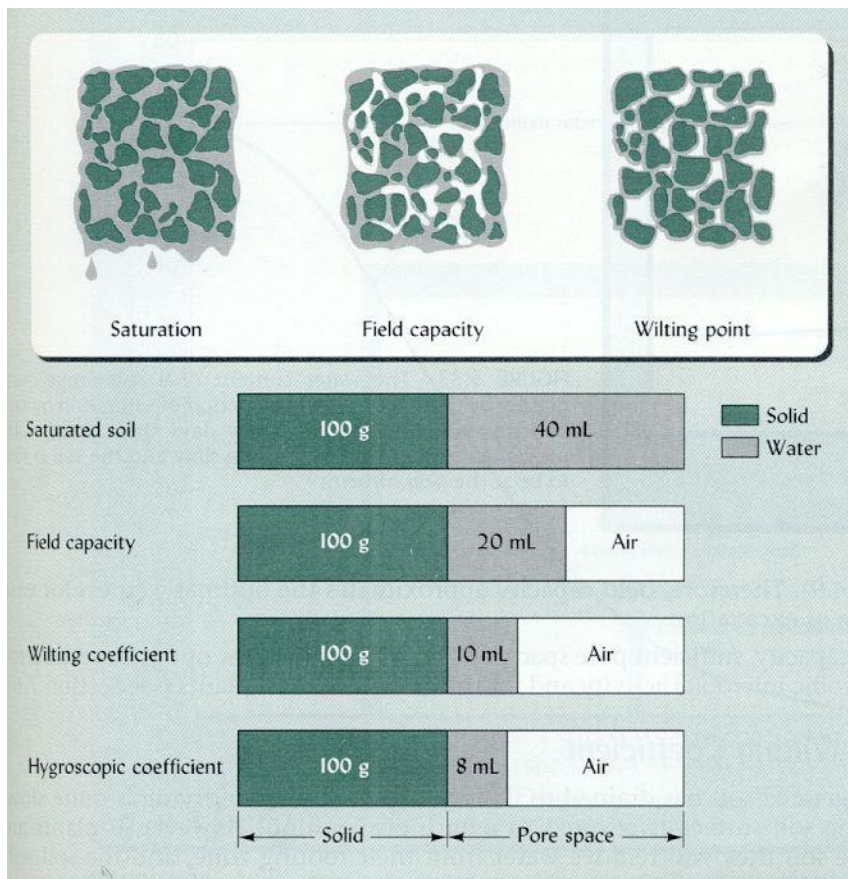
FIGURE 4.15 Volumes of water and air associated with a 100 g slice of soil solids in a well-granulated silt loam at different moisture levels. The top bar shows the situation when a representative soil is completely saturated with water. This situation will usually occur for short periods of time during a rain or when the soil is being irrigated. Water will soon drain out of the larger pores (macropores). The soil is then said to be at the field capacity. Plants will remove water from the soil quite rapidly until they begin to wilt. When permanent wilting of the plants occurs, the soil water content is said to be at the wilting coefficient. There is still considerable water in the soil, but it is held too tightly to permit its absorption by plant roots. A further reduction in water content to the hygroscopic coefficient is illustrated in the bottom bar. At this point the water is held very tightly, mostly by the soil colloids.

Field Capacity

Once the rain or irrigation has ceased, water in the largest soil pores will drain downward quite rapidly in response to the hydraulic gradient (mostly gravity). After one to three days, this rapid downward movement will become negligible as matric forces play a greater role in the movement of the remaining water. The soil then is said to be at its field capacity. In this condition, water has moved out of the macropores and air has moved in to take its place. The micropores or capillary pores are still filled with water and can supply plants with needed water. The matric potential will vary slightly from soil to soil but

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is generally in the range of -10 to -30 kPa, assuming drainage into a less-moist zone of similar porosity. Water movement will continue to take place by unsaturated flow, but the rate of movement is very slow since it now is due primarily to capillary forces, which are effective only in micropores (Figure 4.15). The water found in pores small enough to retain it against rapid gravitational drainage, but large enough to allow capillary flow in response to matric potential gradients, is sometimes termed **capillary water**.

While all soil water is affected by gravity, the term *gravitational water* refers to the portion of soil water that readily drains away between the states of maximum retentive capacity and field capacity. Most soil leaching occurs as gravitational water that drains from the larger pores before field capacity is reached. Gravitational water therefore includes much of the water that transports chemicals such as nutrient ions, pesticides, and organic contaminants into the groundwater and, ultimately, into streams and rivers.

Field capacity is a very useful term because it refers to an approximate degree of soil wetness at which several important soil properties are in transition:

1. At field capacity, a soil is holding the maximal amount of water useful to plants. Additional water, while held with low energy of retention, would be of limited use to plants because it would remain in the soil for only a short time before draining, and, while in the soil, it would occupy the larger pores, thereby reducing soil aeration. Drainage of gravitational water from the soil is generally a requisite for optimum plant growth (hydrophilic plants, such as rice or cattails, excepted).

2. At field capacity, the soil is near its lower plastic limit-that is, the soil behaves as a crumbly semisolid at water contents below field capacity, and as a plastic puty-like material that easily turns to mud at water contents above field capacity. Therefore, field capacity approximates the optimal wetness fore ease of tillage or excavation.

3. At field capacity, sufficient pore space is filled with air to allow optimal aeration for most aerobic microbial activity and for the growth of most plants.

Permanent Wilting Point or Wilting Coefficient

Once an unvegetated soil has drained to its field capacity, further drying is quite slow, especially if the soil surface is covered to reduce evaporation. However, if plants are growing in the soil they will remove water from their rooting zone, and the soil will continue to dry. The roots will remove water first from the largest water-filled pores where the water potential is relatively high. As these pores are emptied, roots will draw their water from the progressively smaller pores and thinner water films in which the matric water potential is lower and the forces attracting water to the solid surfaces are greater. Hence, it will become progressively more difficult for plants to remove water from the soil at a rate sufficient to meet their needs.

As thee soil dries, the rate of plant water removal may fail to keep up with plant needs, and plants may begin to wilt during the daytime to conserve moisture. At first the plants will regain their turgor at night when water is not being lost through the leaves and the roots can catch up with the plant's demand. Ultimately, however, the plant will remain wilted night and day when the roots cannot generate water potentials low enough to coax the remaining water from the soil. Although not yet dead, the plants are now in a permanently wilted condition and will die if water is not provided. For most plants this condition develops when the soil water potential Ψ has a value of about -1.500 kPa (-15 bars). A few plants, especially xerophytes (desert-type plants) can continue to remove water at even -1.800 or -2.000 kPa, but the amount of water available between -1.500 kPa and -2.000 kPa is very small (Figure 4.16).
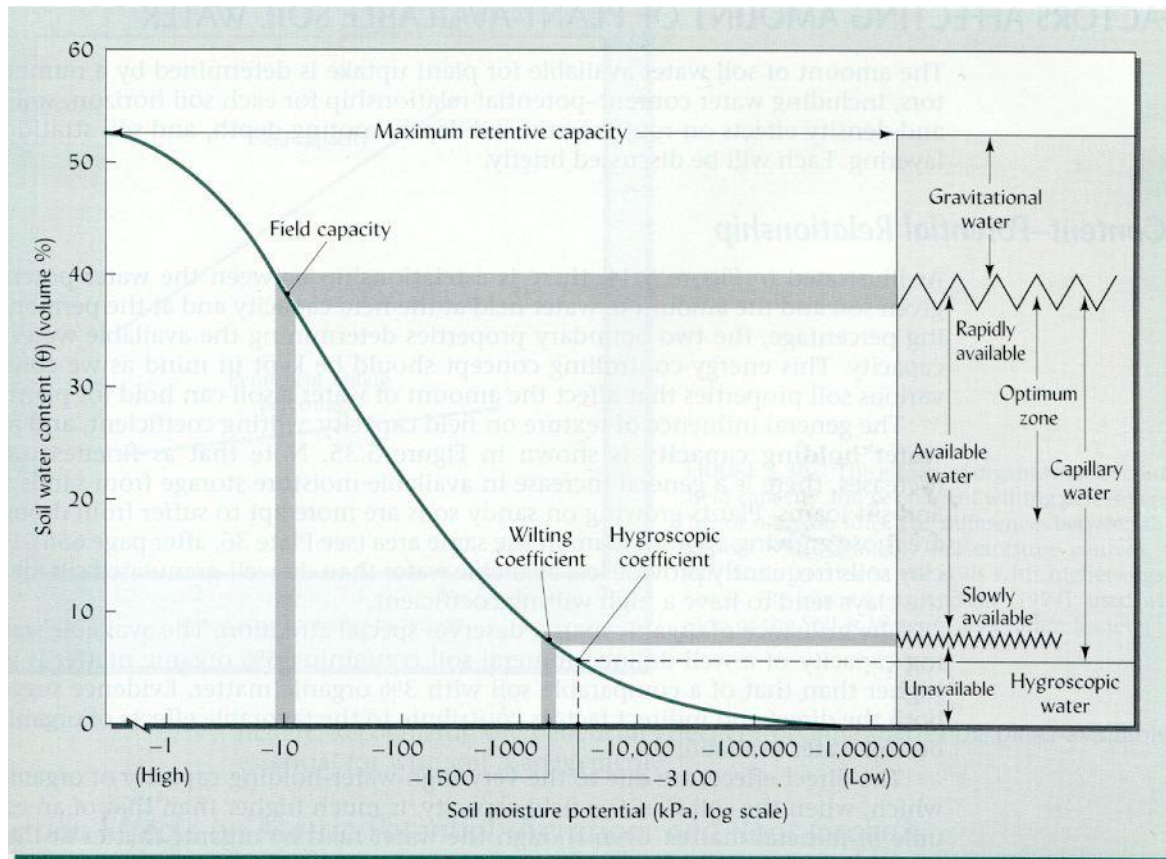
FIGURE 4.16 Water content-matric potential curve of a loam soil as related to different terms used to describe water in soils. The wavy lines in the diagram to the right suggest that measurements such as field capacity are only approximations. The gradual change in potential with soil moisture change discourages the concept of different "forms" of water in soils. At the same time, such terms as gravitational and available assist in the qualitative description of moisture utilization in soils.

The water content of the soil at this stage is called the **wilting coefficient** or the **permanent wilting percentage** and by convention is taken to be that amount of water retained by the soil when the water potential is -1.500 kPa. The soil will appear to be dusty dry, although some water remains in the smallest of the micropores and in very thin films (perhaps only 10 molecules thick) around individual soil particles (Figure 4.16). As illustrated in Figure 4.16, **plant available water** is considered to be that water retained in soils between the states of field capacity and wilting coefficient (between -10 to -30 kPa and -1.500 kPa). The amount of capillary water remaining in the soil that is unavailable to higher plants can be substantial, especially in fine-textured soils and those high in organic matter.

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4.9 FACTORS AFFECTING AMOUNT OF PLANT-AVAILABLE SOIL WATER

The amount of soil water available for plant uptake is determined by a number of factors, including water content-potential relationship for each horizon, soil strength and density effects on root growth, soil depth, rooting depth, and soil stratification or layering. Here we will only show the general relationship between soil water characteristics and soil texture (Figure 4.17).
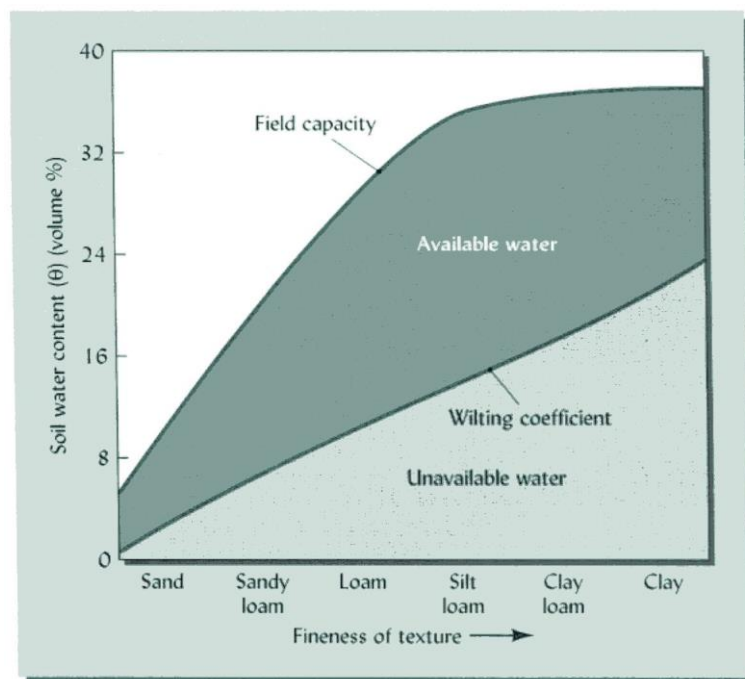

FIGURE 4.17 General relationship between soil water characteristics and soil texture. Note that the wilting coefficient increases as the texture becomes finer. The field capacity increases until we reach the silt loams, then levels off. Remember these are representative curves; individual soils would probably have values different from those shown.

Study Questions?

1. What is the role of the *reference state of water* in defining soil water potential? Describe the properties of this reference state of water.

2. Imagine a root of a cotton plant growing in the upper horizon of an irrigated soil. As the root attempts to draw water molecules from this soil, what forces (potentials) must it overcome? If this soil were compacted by a heavy vehicl, which of these forces would be most affected? Explain!

3. Using the terms adhesion, cohesion, meniscus, surface tension, atmospheric pressure, and hydrophilic surface, write a brief essay to explain why water rises up from the water table in a mineral soil.

4. Suppose you were hired to design an automatic irrigating system for a wealthy homeowner's garden. You determine that the flower beds should be kept at a water potential above -60 kPa, but not wetter than -10 kPa as the annual flowers here are sensitive to both drought and lack of good aeration. The rough turf areas, however, can do well if the soil dries to as low as -300 kPa. Your budget allows either tensiometers or electrical resistance blocks to be hooked up to electronic switching valves. Which instruments would you use and where? Explain.

5. Suppose the homeowner referred to in question 4 increased your budget and asked to use the TDR method to measure soil water contents. What additional information about the soils, not necessary for using the tensiometer, would you have to obtain to use the TDR instrument? Explain.

6. A greenhouse operator was growing ornamental woody plants in 15-cm-tall plastic containers filled with a loamy sand. He watered the containers daily with a sprinkler system. His first batch of 1000 plants yellowed and died from too much water and not enough air. As an employee of the greenhouse, you suggest that he use 30-cm-tall pots for the next batch of plants. Explain your reasoning.

7. Suppose you measured the following data for a soil:

Horizon	Bulk density, Mg/m^3	Θ_m at different water potentials, kg water/kg dry soil		
		-10kPa	-100kPa	-1.500kPa
A (0-30 cm)	1.28	.28	.20	.08
B_t (0-70 cm)	1.40	.30	.25	.15
B_x (0-120 cm)	1.95	.20	.15	.05

Estimate the total available water holding capacity (AWC) of this soil in centimeters of water.

8. A forester obtained a cylindrical core (L = 15 cm, r = 3.25 cm) of soil from a field site. She placed all the soil in a metal can with a tight-fitting lid. The empty metal can weighed 300 g and when filled with the field-

moist soil weighed 972 g. Back in the lab, she placed the can of soil, with lid removed, in an oven for several days until it ceased to lose weight. The weight of the dried can with soil (including the lid) was 870 g. Calculate both Θ_m and Θ.

9. Give four reasons why compacting a soil is likely to reduce the amount of water available to growing plants.

10.   Since even rapidly growing, finely branched root systems rarely contact more than 1 or 2% of the soil particle surfaces, how is it that the roots can utilize much more than 1 or 2% of the water held on these surfaces?"

LITERATURE: Nyle C. Brady – Ray R. Weil (2002): The Nature and Properties of Soils. Thirteenth Edition, Prentice Hall; Upper Saddle River, New Jersey (Source of all texts and figures!).