

TERRESTRIAL PLANT ECOLOGY

SECOND EDITION

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The Benjamin/Cummings Publishing Company, Inc.

Menlo Park, California • Reading, Massachusetts
Don Mills, Ontario • Wokingham, U.K. • Amsterdam • Sydney
Singapore • Tokyo • Madrid • Bogota • Santiago • San Juan

Library of Congress Cataloging-in-Publication Data

Barbour, Michael G.

Terrestrial plant ecology.

Bibliography: p.

Includes index.

1. Botany—Ecology. 2. Botany—North America—
Ecology. I. Burk, Jack H., 1942- . II. Pitts,
Wanna D., 1932- . III. Title.

QK901.B345 1987

581.5'264

86-26812

ISBN 0-8053-0541-6

CHAPTER 17

SOIL

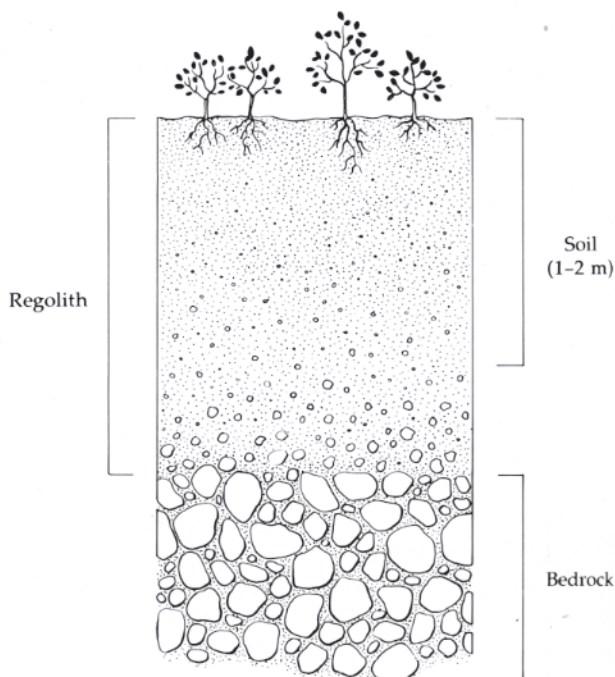
The soil is the common ground between the living and the nonliving world. We could have logically begun this book with a foundational chapter on soils. The importance of the medium for growth cannot be overstated. However, plant ecology is much like any other science in that we could begin almost anywhere and work through a cyclical progression of information with effectiveness. We can now use the knowledge gained from other chapters to support our discussion of soils as they influence plant distribution. We will consider soil development, profiles, texture, and structure. Following a brief discussion on soil physics and chemistry, we will consider the taxonomy of soils.

There are four separate components of soil: mineral grains, organic matter, water, and air. The mineral substrate provides anchorage, pore space for storage of water and air, and nutrients on an exchange basis. Organic matter refers to the plant and animal residues in various stages of decomposition, as well as to the cells, tissues, and exudates of soil organisms. Organic matter enhances the intrasystem cycling of nutrients and improves soil structure, pore space, and water storage, among other things. Soil water is the solvent medium for nutrients needed by growing plants, maintaining equilibria among the cations and anions adsorbed* on soil particles, the living plants, and the soil water (soil solution) itself. Soil air contains oxygen for cellular work, carbon dioxide that augments further mineral weathering and biotrophic processes, and atmospheric nitrogen for nitrogen-fixing soil organisms.

Soil, along with climate, is a primary agent in plant selection through evolutionary change. The term *soil* comes from the Latin word for floor, *solum*. A soil scientist, or **pedologist**, views soil as "a natural product formed from weathered rock by the action of climate and living organisms" (Thompson and Troeh 1973). The foregoing statement implies that soil is mainly composed of mineral materials, and this is true, for the most part. However, pedologists do not over-

**Adsorbed* refers to a material being attracted and held to, whereas *absorbed* refers to a material being taken in and probably held more tightly than material adsorbed.

Figure 17-1 A vertical section of soil, showing relative positions of bedrock, regolith (unconsolidated material), and soil (highly weathered portion of regolith).



look the fact that some soils are formed mainly from the weathering of accumulating organic debris (organic soils). We will use an edaphological definition of soil, considering it primarily as a medium for growing plants: Soil is a mixture of mineral and organic materials, capable of supporting plant life. Implicit in both views is the interaction of soil and life. Each one exists because of, and is formed at least in part by, the other.

The materials in the upper fraction of the earth's crust are exposed to weathering agents more intensely than are the underlying strata. Three levels can be seen in the section of an exposed hillside diagrammed in Figure 17-1, which has loose material resting on underlying rocks. The foundation is termed **bedrock**. Above that is the **regolith**, the unconsolidated mantle of weathered rock and soil material, of which the upper 1–2 meters or less of highly weathered, biochemically altered material is termed **soil**.

The Soil Cycle and Soil Development

The process of soil development may seem to be linear and formational in terms of human time, but in geologic time the process is part of a cycle. The processes that take place in the cycle of soil development include sedimentation or deposition of fragments from clay size up, lithification (stone formation), metamorphism and melting, crystallization, volcanism, and the agents of erosion and

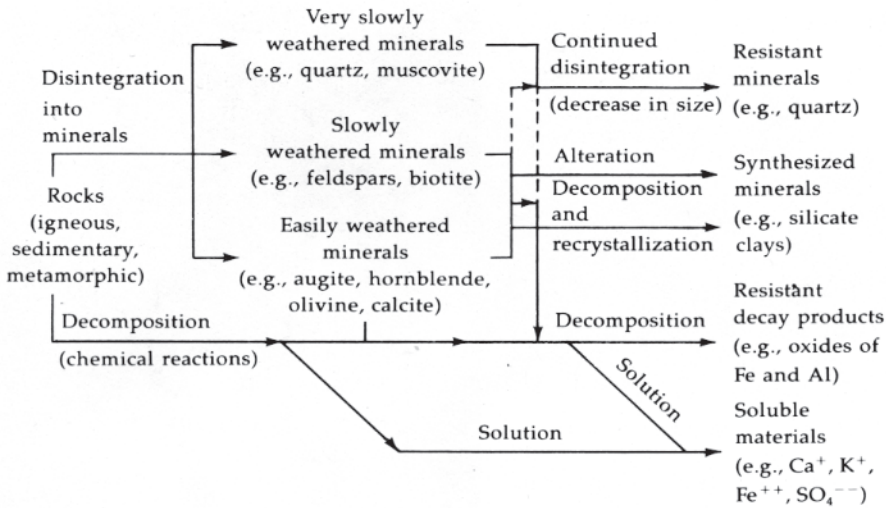


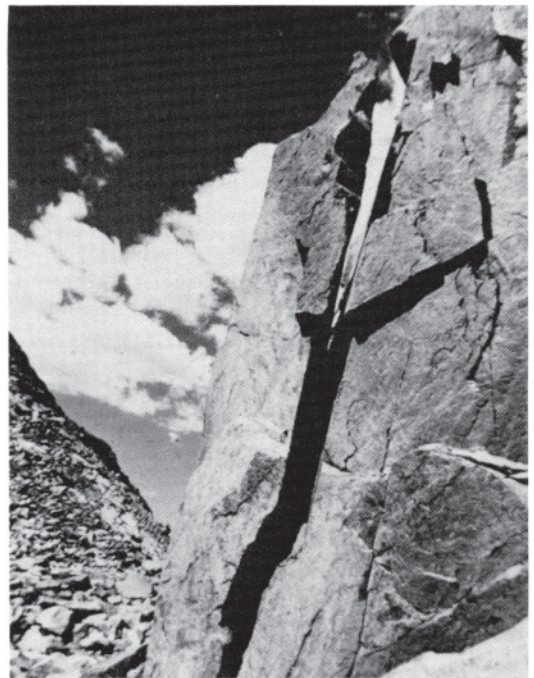
Figure 17-2 Weathering pathways under moderately acid conditions common in humid temperate regions. Major paths of weathering are indicated by the heavier arrows, minor pathways by broken lines. As one would expect, climate modifies the exact relationships. In arid regions, physical breakdown (disintegration) would dominate, and soluble ions would not be lost in large quantities. In humid regions, decomposition becomes more important, especially under tropical conditions. (Reprinted with permission of Macmillan Publishing Co., Inc. from *The Nature and Properties of Soils*, 8th Ed. by Nyle C. Brady. Copyright 1974 by Macmillan Publishing Co., Inc.)

transport. The process of forming soil may be conveniently divided into two phases. The first is rock weathering, and the second is soil formation. Whether the substrate is a granitic dome, a lava deposit, or a limestone cliff, there are certain weathering phenomena that are common to all rock substrates, and we will examine a few of them. Weathering pathways are shown in Figure 17-2. Notice that a variety of materials are produced, but that common pathways yield the same minerals, whether derived from sedimentary, igneous, or metamorphic rocks.

Mechanical Factors

Mechanical erosion, or disintegration, may be attributed to the expansion and contraction of rocks in response to thermal flux. The outer layers are heated differentially compared to the inner rock, creating stress that is relieved by cracking and spreading. The fissures and crevices thus formed, along with those caused by tectonic activity and gravitational tensions, will be filled with water. Expansion caused by the freezing of this water is impressive; normal hydrostatic pressure amounts to more than $10,000 \text{ kg cm}^{-2}$. Ice floats on liquid water, and the water at the air/liquid interface is cooled more rapidly than that within the rock, so that the ice acts as a wedge, forcing still wider cracks into the rock (Figure

Figure 17-3 Cracks in granite caused by the force exerted by the freezing of water, which seeped into minute hairline fractures.



17-3). The stresses generated by the freezing of water are also significant after initial fragmentation of rocks. The regolith, including the soil at the top, continues to be altered physically by this force.

Exfoliation, the peeling off of layers of rock, is attributed to the differential expansion just mentioned and also to the expansion allowed when surface material is removed by erosion. Igneous intrusions that crystallize far beneath the surface are, in a sense, spring-loaded. The removal of the pressure of overlying material allows exfoliation, so layers of rock are released much as the modified leaves peel away from an onion bulb. Half Dome, in Yosemite National Park in California, is an example of exfoliation.

Erosion by wind and water and deposition by these same forces are significant to rock weathering. The force exerted by plant roots must also be counted as a very important physical force, even though plants exert biological and chemical influences as well and also influence subsequent soil formation. Moving ice (or glaciers) transports tremendous amounts of material. In addition, the abrasive materials carried by the ice polish and further reduce the rocks and rock fragments. Alluvial (stream-carried), glacial, aeolian (wind-carried), and lacustrine (lake-deposited) materials form the basis for parent materials of many kinds of soils. Streams carry an enormous amount of material into lakes or into rivers that ultimately empty into the sea. The Mississippi River is estimated to discharge in excess of 660,000 kg of soil into the Gulf of Mexico each year. The size of alluvium may range up to house-sized boulders (Figure 17-4), depending on the volume and velocity of the water. By contrast, aeolian materials are fine, usually dust-like particles that abrade, polish, and further erode rocks in their passing.



Figure 17-4 Alluvium forming a fan originating at the mouth of a desert canyon. The sparseness of the vegetation allows a clear view of the sizes of material moved; some of the "particles" are huge boulders. (Courtesy of Dr. John S. Shelton.)

Chemical Factors

Chemical processes of weathering (decomposition) are active almost as soon as disintegration has begun. Included under chemical processes are a number of rather simple reactions. Oxidation, especially of iron-carrying rocks, is a very obvious form of chemical weathering. Carbonation, produced by the action of a weak concentration of hydrogen ions in percolating waters and attack by other acids including organic acids, causes decomposition. Some of the organic acids may, however, initiate processes leading to recrystallization of clays and are thus synthetic in effect. The most significant chemical weathering is exerted by water, in one of three ways: solution, the solvent action of water upon the mineral substrate; hydrolysis, the cleavage of a mineral by water; and hydration, the rigid attachment of H^+ and OH^- ions to the material being decomposed. The products of some of these processes, such as oxidation and hydration, usually have a larger volume than the initial minerals, and, if not displaced, contribute to the mechanical forces involved in exfoliation. These processes are all influenced by climate and are very effective in tropical regions.

Soil Formation

Soil formation (S) is characterized by Jenny (1941) as dependent on a set of independent variables as follows: climate (cl), organisms (o), topography (r), rock type or parent material (p), and time (t). Jenny's classic equation is as follows:

$$S = f^{(cl, o, r, p, t \dots)} \quad (\text{Equation 17-1})$$

He defines organisms as the biotic potential of a site, not just the existing organisms, which are indeed interdependent or dependent factors. The biotic potential includes consideration of soil organisms, litter, vegetation type, etc. The soil macrobiota and microbiota interact with the growth medium, altering its capacity

to support life. The amount of litter deposited on the soil is determined by the productivity of the community, and the rate of decay or removal of that litter is determined by climate, by the nature of the litter, and by fire frequency. Also, the incorporation of litter-derived materials into the soil, the maintenance of soil porosity and permeability, and ultimately the soil fertility, all rely at least in part on the activities of soil organisms. Vegetation and climate determine what kind of soil is formed. Podzol soils develop beneath coniferous forests in cool, moist climates, and they have a prominent, ashy gray upper horizon. Chernozem ("black earth") soils develop under grasslands in temperate regions. As previously noted, a warm, moist environment hastens both disintegration and decomposition. The concepts relating to plant community succession developed in Chapter 11 also apply to soil succession.

Thus, parent materials interacting with soil biota and climate produce a certain soil with characteristic features. Another very important factor in soil formation is time—time for large mineral grains to be disintegrated, decomposed, and changed to secondary minerals; time for finer and finer mineral grains to accumulate; time for organics and humus to become a significant fraction of the total soil; but, also, time for finer grains and soluble anions and cations to be leached away (Chapter 13).

The composition of parent material largely determines the chemical make-up of its derived soils. For instance, acid soils derive from (a) aluminum-rich bauxites, (b) silica-rich substrates such as sands, slates, and diatomaceous earth, (c) mine tailings of lead and zinc deposits, (d) sulfide-rich, hydrothermally altered volcanic rocks and ash, and (e) estuarine deposits. Soils with poor fertility may develop from ultrabasic (ultramafic) rock types rich in silicates of iron and magnesium. For example, serpentine develops from igneous rocks such as peridotite and dunite, which contain silicates of iron and magnesium. Serpentine soils are high in magnesium and low in calcium (Table 17-1). They may also be high in nickel and chromium and deficient in nitrogen and phosphorous; thus, they are not very fertile.

Table 17-1 Comparison of certain soil nutrients in serpentine and nonserpentine soils.

Nutrient	Nonserpentine: Serpentine
Replaceable Ca (meq 100 g ⁻¹)	7.4
Replaceable Mg (meq 100 g ⁻¹)	0.5
Percent N (total, all forms)	14.0
Percent P (total, all forms)	65.0
K (meq 100 g ⁻¹)	3.8
Ni (meq 100 g ⁻¹)	0.1
Cr (meq 100 g ⁻¹)	0.01

Soil may form from transported material or *in situ* parent material (residual soil). Topography influences the stability of a soil body in a variety of ways, including its effect on rainfall patterns and the steepness of slopes. Resistant parent materials may result in steep slopes; parent rocks or materials that erode easily produce wide valleys, rounded hills, and often deep soils. Steep slopes may continually expose new surfaces to weathering and soil formation, as alluvium or talus (material moved by gravity) is moved downslope; resident soils, if present at all, are young and shallow. Gentle slopes restrict transport and often enhance development of deep, strongly developed soils.

Soil Profiles

Soil Profile Development

Soils, especially deep soils of temperate latitudes, develop very characteristic layers or horizons, more or less differentiated and discernible from each other if a pit or trench is dug into the soil. A section encompassing all the horizons of the soil and extending into its parent material is called a **soil profile** (Figure 17-5). The rock weathering that we discussed earlier is mainly a physical process.

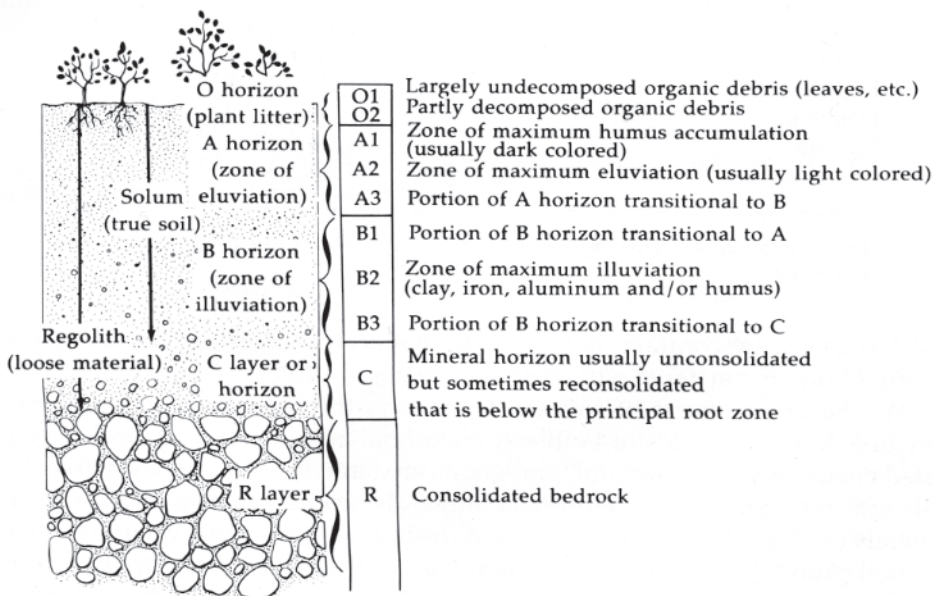


Figure 17-5 A diagrammatic representation of a soil profile (vertical section of the soil), showing soil horizon nomenclature and symbols for mineral soils. Additional subdivisions of the O horizon, not shown, are used in organic soil. A and B horizons taken together are referred to as the **solum** or true soil. (Reprinted by permission of McGraw-Hill Book Co. from *Soils and Soil Fertility* by Thompson and Troeh 1973).

Table 17-2 The more important original and secondary minerals found in soils. The original minerals are also found abundantly in igneous and metamorphic rocks. Secondary minerals are commonly found in sedimentary rocks. (Reprinted with permission of Macmillan Publishing Co., Inc. from *The Nature and Properties of Soils*, by Nyle C. Brady. Copyright 1974 by Macmillan Publishing Co., Inc.)

Name	Formula
Original minerals	
Quartz	SiO ₂
Microcline } Orthoclase }	KAlSi ₃ O ₈
Na plagioclase	NaAlSi ₃ O ₈
Ca plagioclase	CaAl ₂ Si ₂ O ₈
Muscovite	KAl ₃ Si ₃ O ₁₀ (OH) ₂
Biotite	KAl(Mg, Fe) ₃ Si ₃ O ₁₀ (OH) ₂
Hornblende*	Ca ₂ Al ₂ Mg ₂ Fe ₃ , Si ₆ O ₂₂ (OH) ₂
Augite*	Ca ₂ (Al, Fe) ₄ (Mg, Fe) ₄ Si ₆ O ₂₄
Secondary minerals	
Calcite	CaCO ₃
Dolomite	CaMg(CO ₃) ₂
Gypsum	CaSO ₄ · 2H ₂ O
Apatite	Ca ₅ (PO ₄) ₃ · (Cl, F)
Limonite	Fe ₂ O ₃ · 3H ₂ O
Hematite	Fe ₂ O ₃
Gibbsite	Al ₂ O ₃ · 3H ₂ O
Clay minerals	Al silicates

*These are approximate formulae only because these minerals are so variable in their composition.

Soil formation, by contrast, is largely biochemical weathering and the gradual mixing of organic material with inorganic material.

As the upper layers of soil are exposed to the roots of plants, to animal activities, to the activities of both soil microflora and microfauna, and to accelerated chemical weathering, mineral grains are altered. Primary or original minerals are transformed into **secondary minerals**. Some primary and secondary minerals are shown in Table 17-2. Recall from Figure 17-2 that quartz is a very resistant primary mineral. Primary minerals are found in abundance in igneous and metamorphic rocks, and secondary minerals are often found in sedimentary rocks, as we would expect.

The presence of secondary materials and abundant organics and changes due to interactions between soil and soil biota result in differentiation of soil

horizons. Two of the processes are significant enough that they provide names for the principal (master) horizons of mineral soils in which they occur. The A horizon includes the surface soil in which surface or near-surface organic matter accumulates. It is also referred to as the **zone of eluviation**, due to the movement of materials downward from the A horizon, mostly by leaching. The B horizon includes the subsurface part of the soil, or subsoil, which may be a **zone of illuviation**, where the materials lost from A are accumulated. The illuvial materials may include iron and aluminum compounds, clays, and humus. Humus is the dark brown or blackish stable fraction left over after most of the organic residues have been decomposed.* Alternatively, the B horizon may be a **zone of alteration**, where little material accumulates from A but significant changes in structure or secondary mineral formation take place.

Soil horizon nomenclature is shown in Figure 17-5. The O (organic) horizon and its subdivisions may be largely or totally lacking in certain desert soils but may be the only horizons existing in organic soils. The R layer underlies many, but not all, upland or residual soils, where it may displace the C horizon. A young soil may have A and C horizons, but no B horizon. This lack of a B horizon results when there has been no translocation or eluviation of clay from the A horizon downward; such a profile is characteristic of the chernozem or "black earth" profiles in steppe regions.

Soils can be identified in the field based on the differentiation of the horizons, the texture, the color of the principal materials, and even which horizons (or portions) are lacking. A field sheet for recording soil characteristics is shown in Figure 17-6, and Figure 17-7 is a photograph of the soil described in the field sheet. Note the site characteristics that are recorded: elevation, slope, aspect, erosion, and the like. This soil is an Inceptisol, identified in part by its lack of a high clay content in the B horizon. There is no C horizon, and the lower B horizon rests directly on the R horizon, or bedrock.

Zonal, Azonal, and Intrazonal Soils

Although we will discuss the classification of soils in detail in a later section, it is appropriate to mention here the three orders in the 1938 system of soil classification (Baldwin et al. 1938). These include zonal soils, whose characteristics and profiles are mostly influenced by climate and vegetation (e.g., desert soils, lateritic (red tropical) soils); intrazonal soils, which exhibit unique soil profiles due to unusual parent material or topography (e.g., saline soils, bog); and azonal soils, which have little or no profile development (e.g., recent alluvial soils, rapidly eroding soils). The appearance of a soil profile largely determines the fit of a soil into a soil classification scheme.

*The formation of humus is a complex topic that we will discuss in a later section.

FIELD SHEET FOR RECORDING SOIL CHARACTERISTICS

No. 8

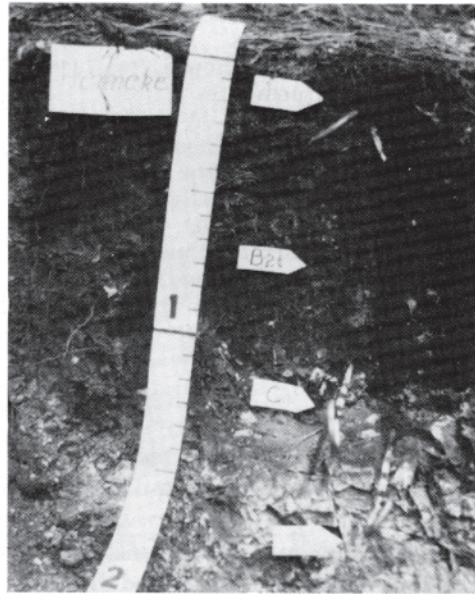
Soil Type Henneke variant
Location 9 km S of Coloma NE 1/4 SW 1/4 Sec 1 T 10 N, R 9 E MDBM
Geographical Landscape Upland
Elevation 326 m Slope 20% Aspect S-SW Erosion slight-moderate
Groundwater deep Drainage mod. well Alkali none
Mode of Formation Primary serpentinitic Parent Material Serpentine
Climate MAT 14°C: Jan 4.5°C, Jul 23°C MAP 88 cm FFD 223 @ 0°C
Natural Cover Chamise, Digger Pine, Ceanothus Soil Region VIII
Profile Group VII - II Higher Categories Loamy-skeletal, serpentinitic,
Genetically Related Soil Series Del Piedra (an alfisol); thermic, Typic Xerochrept
Dukabella (higher, timbered with Jeffrey pine)

PROFILE SKETCH	COLOR	TEXTURE	STRUC-TURE	CONSISTENCE	REACTION	MISC: Roots, Pores, Clay films, Concretions
<div style="border: 1px solid black; padding: 2px;">A11 0-13 cm <small>as</small></div>	7.5YR 4/4	gl	1 fgr	so, vfr; so, po	7.3	3vf, 3f, 2m, 2co rts 3vf, 3f, 2m, 1co pores
<div style="border: 1px solid black; padding: 2px;">A12 13-20 cm <small>aw</small></div>	5YR 4/6	gl	2 mgr	so, vfr; so, po		2vf, 2f, 1m, 1co rts 2m, 1co, 3vf, 1f pores
<div style="border: 1px solid black; padding: 2px;">B1 20-28 cm <small>cw</small></div>	5YR 4/6	gl	1c abk → 2cgr	so, vfr; ss, ps		1vf, 1f, 2m, 2co rts 2vf, 2f, 2m pores
<div style="border: 1px solid black; padding: 2px;">B21 28-51 cm <small>cw</small></div>	5YR 3/4		1m abk → 2cgr	so, fr, ss, ps		1f, 1m, 1co rts 2fi, t; 2vf i pores
<div style="border: 1px solid black; padding: 2px;">B22 51-68 cm <small>ci</small></div>	5YR 4/4		1m sbk → 2cgr	sh, fr, ss, ps		1f, 1m, 2co rts 2f, 2vf pores
<div style="border: 1px solid black; padding: 2px;">R</div>					✓	

Natural Land Division E 8-3m
Soil Rating (Storie index) 11 Soil Grade 5(6)

Figure 17-6 Field sheet for recording soil characteristics. a s = abrupt smooth; a w = abrupt wavy; c w = clear wavy; c i = clear irregular; 7.5 YR = Yellow-Red hue and 4/4 = value and chroma (in Munsell notation); g l = gravelly loam; f = fine; m = medium; c = coarse; gr = granular; abk = angular blocky; sbk = subangular blocky; so = soft; ss = slightly sticky; sh = slightly hard; vfr = very friable; po = non-plastic; ps = slightly plastic; fr = friable; vf = very fine; co = coarse; rts = roots; t = tubular pores; i = interstitial pores. For structure, 1 = weak and 2 = moderate. For roots and pores, 1 = few, 2 = common, and 3 = many.

Figure 17-7 Soil profile of Henneke variant. The soil is classified as loamy-skeletal, serpentinitic, thermic, Typic Xerochrept, an Inceptisol derived from serpentinitic materials.



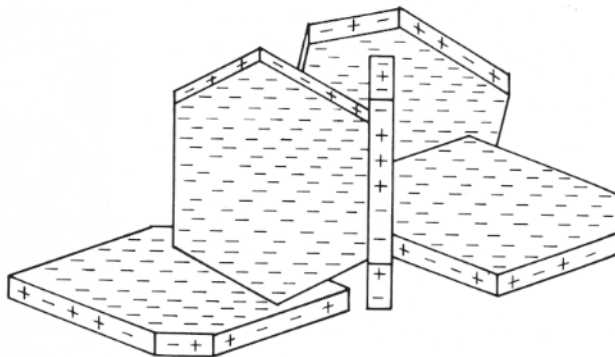
Physical Properties of Soil

Soil Texture

Soil texture is defined by the percent by weight of sand, silt, and clay. That is, only mineral components are considered in determining soil texture, even though soil is also composed of organic matter, water, and air, with organic materials contributing up to 5% or more of the total. The United States Department of Agriculture classification of soil particles places the upper limit for silt at 0.05 mm in diameter and the lower limit at 0.002 mm. Particles greater than 0.05 mm in diameter are considered sand, and particles less than 0.002 mm in diameter are considered clay (Brady 1974). Although almost any soil contains pebbles and cobbles that are larger than the coarsest sand (2 mm diameter), such coarse fragments play but a minor role in soil function unless the bulk of the soil is composed of these fragments. In the field, a good estimate of texture can be made by dampening and handling a small amount of the soil. The particles that feel grainy and gritty are sand; those that are silky, like moist talcum powder, are silt. If the sample is sticky and if a self-supporting, flexible ribbon can be extruded between the thumb and finger, there is a high percentage of clay. If it is not sticky or only slightly sticky, and no ribbon can be extruded, there is a low percentage of clay. Different soil textures can be recognized by various combinations of the foregoing.

In general, the structural or skeletal support of the soil is provided by the sand fraction, or at least by the largest soil particles, and the finer particles help

Figure 17-8 An idealized group of clay particles representing the **card house** effect in soil structure, in which positive charges on edge positions are attracted by negative charges on the broad surfaces. The negative charges provide exchange sites for cations and are thus a storage facility. (Reprinted by permission of McGraw-Hill Book Co. from *Soils and Soil Fertility* by Thompson and Troeh 1973.)



to store nutrients and to bind particles together into aggregates. Size classes are assigned arbitrarily, recognizing that minerals (and the building blocks of minerals) form a continuum of sizes, ranging from single tetrahedra* to large and very complex mineral crystals. Material that is without identifiable structure in the fine clay fraction is called **allophane**.

If mineral grains are not spherical, then their diameter is taken to be an average of their maximum and minimum dimensions. Sand grains tend to range from blocky and irregular to subround in shape, and silt particles are similar but smaller. Clay particles are very unlike sand grains; their plate-like shape is reminiscent of mica (Figure 17-8).

A more precise determination of soil texture can be made in the lab. Particles larger than 2 mm are sieved out of a given sample, which is then treated to oxidize and remove the organic matter. This is usually done with hydrogen peroxide, which also has the effect of rupturing the binding of clays by organic matter. The sample is then dried and weighed. The remaining materials are carefully flushed and shaken through a series or nest of sieves that allow particles 1 mm, 0.5 mm, 0.25 mm, 0.1 mm, and 0.05 mm to pass through. The total weight of sand and its distribution can be determined by this process. The fraction passing through the sieves contains silt and clay together, and these can be separated by using a hydrometer or pipette method of analysis.[†] Such methods, which are based on the rate of particle settling, allow the separation of the silt and clay fractions.

Once the percentage of weight of separate components has been determined, the use of a soil texture triangle (Figure 17-9) facilitates textural class determination. Assume, for instance, that our soil contains 30% clay, 40% silt, and 30% sand. This soil would be called *clay loam*. Make note of two features: (a) The designation is an area, not just a point on the triangle. (b) There are four

*A *tetrahedron* is a three-dimensional figure with four triangular sides and an oxygen ion at each of the four points.

†For a more complete discussion, consult Thompson and Troeh 1973.

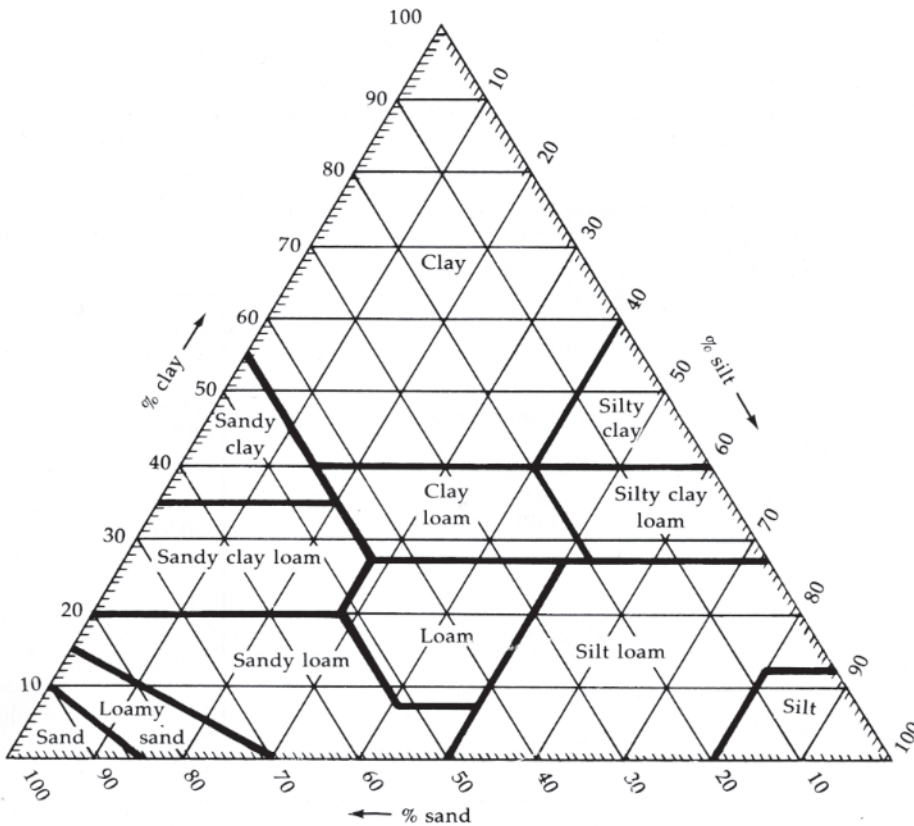


Figure 17-9 Relationship between the particle size distribution of a soil and its class name. The points corresponding to the percentages of silt and clay present in the soil are located on the silt and clay lines, respectively. Lines are then projected inward from % silt parallel to the clay side of the triangle and inward from % clay parallel to the sand side. Note that the third particle size (in this case, sand) is defined by the two projected lines. The name of the compartment in which the lines intersect is the texture class of the soil in question. (Reprinted by permission of McGraw-Hill Book Co. from *Soils and Soil Fertility* by Thompson and Troeh 1973.)

terms commonly applied to soil texture: clay, silt, sand, and loam. *Loam* designates a mixture of sand, silt, and clay that exhibits the properties of each fraction about equally. Thus, our sample exhibits the clay fraction more strongly than does loam, but is still a loamy soil, hence *clay loam*. Table 17-3 lists the general terms used by the United States Department of Agriculture to describe soil texture.

To understand the function of these various soil fractions, we must comprehend two more physical properties of soil: porosity and permeability. *Porosity* refers to the total space that is not occupied by soil particles. *Permeability* deals with the ease of passage through soil of gases, liquids, and plant roots. In gen-

Table 17-3 General terms used by the U.S. Department of Agriculture to describe soil texture in relation to the basic soil textural class names. (Reprinted with permission of Macmillan Publishing Co., Inc. from *The Nature and Properties of Soils*, by Nyle C. Brady. Copyright 1974 by Macmillan Publishing Co., Inc.)

General terms		Basic soil textural class names
Common names	Texture	
Sandy soils	Coarse	{ Sand Loamy sand
	{ Moderately coarse	{ Sandy loam Fine sandy loam
		{ Very fine sandy loam
Loamy soils	{ Medium	Loam
		{ Silt loam Silt
	{ Moderately fine	{ Clay loam Sandy clay loam Silty clay loam
Clayey soils	Fine	{ Sandy clay Silty clay Clay

eral, finer-textured soils have greater porosity, with a pore space of perhaps 40–60% of the total, and sandy soils have lesser porosity, with perhaps 35–50% pore space.

Water moves into and drains out of a sandy soil with much greater ease than for a finer-textured soil, which means that sandy soils are more permeable than fine-textured soils. This is because water has strong adhesion and cohesion capacity. Water molecules form a film (**hygroscopic water**) about each clay particle and easily bridge the gap between particles; the cohesion between water molecules is sufficient to maintain the bridge. Such soils are nutritionally rich but can be a problem agriculturally, due to low permeability and poor aeration. Hygroscopic water also forms a tight film about sand grains, but the surface-to-volume ratio is very low, there are fewer grains than in clays, and the large pore size does not allow efficient cohesion of water molecules, which therefore drain away under the influence of gravity. (See Chapter 18 for a full discussion of water potential and soil moisture.)

Clay particles and bits of organic matter within the soil are referred to as **micelles**. Micelles are negatively charged on their surfaces. The capacity of micelles to provide the primary soil nutrient storage results from (a) the negative charges on the micelles and (b) their very large surface-to-volume ratio. The hypothetical

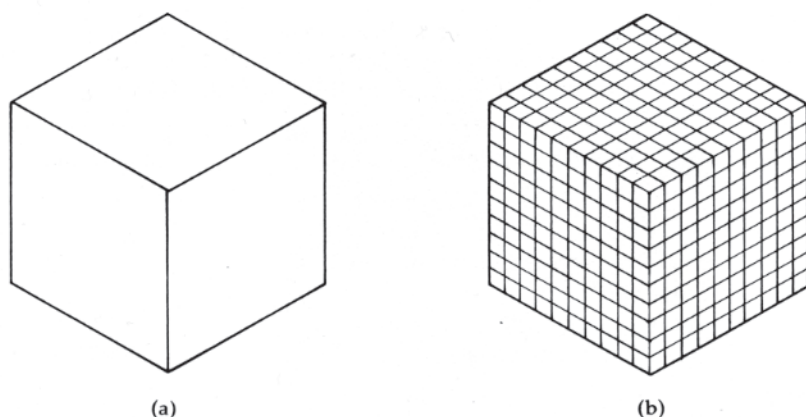


Figure 17-10 (a) A block with sides 1 mm in length. (b) The same block sliced vertically and horizontally into blocks with sides 0.1 mm in length. (Reprinted by permission of McGraw-Hill Book Co. from *Soils and Soil Fertility* by Thompson and Troeh 1973.)

dicing of a mineral grain (Figure 17-10) allows us to envision the difference in the surface-to-volume ratio of a sand grain with sides 1 mm in length and the same volume of clay particles with sides 0.001 mm in length. The sand grain has a volume of 1 mm^3 and a surface area of 6 mm^2 . If we slice this into bits with sides 0.1 mm in length, there will be 1000 bits, each one with $6 \times 10^{-2} \text{ mm}^2$ surface area, or a total of 60 mm^2 , which is an increase of ten times more than the surface area of the parent sand grain. Slice again, so that each soil particle has sides 0.01 mm in length, and the surface area increases to 600 mm^2 . Slice again, so that clay particles have sides 0.001 in length, and surface area now equals 6000 mm^2 , which is 1000 times that of the parent grain.

In summary, the sand fraction provides skeletal (structural) support, aeration and permeability for the soil. The clay fraction provides water holding and nutrient storage capacities (see also the section on soil chemistry). Silt also provides water holding capacity and, through weathering, provides plant nutrients and material to form clays, to replace losses from leaching.



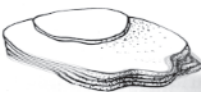


Soil Structure

Primary soil particles are arranged into secondary units, called **peds**.^{*} The nature of the arrangement or **aggregation** of the peds is called *soil structure*. Soil aeration is greatly enhanced by good soil structure, as is the movement of water and plant roots through the soil.

Aggregate stability, and the shapes and sizes of aggregates, are the most important characteristics of soil structure. **Stability** means the capacity of the

^{*}A ped is a unit of soil structure such as an aggregate, crumb, prism, block, or granule, formed by natural process.

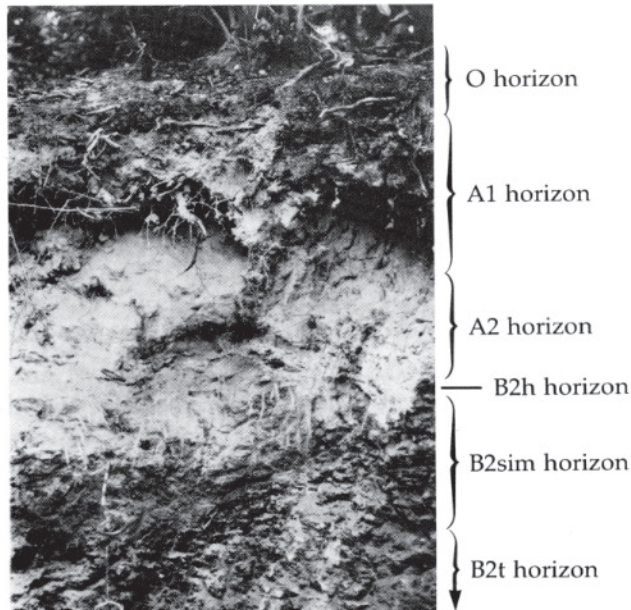
Table 17-4 Classification, description, and illustration of various types of soil structure. (Reprinted by permission of McGraw-Hill Book Co. from *Soils and Soil Fertility* by Thompson and Troeh 1973 (modified).)

Classification	Description	Illustration
Structureless		
Single grain	Each soil particle independent of all others.	
Massive	Entire soil mass clings together, no lines of weakness.	
Structured		
Granular	Primary soil particles grouped into roughly circular peds, such that there is space between them, as they do not fit tightly together. Enhances permeability.	 Granular
Platy	Peds with horizontal dimensions greater than vertical. Common at  with lateral movement of water.	 Platy
Blocky	Peds approximately equal in vertical and horizontal dimensions, but fit well together, unlike granular peds. May be angular or subangular.	 Angular blocky Subangular blocky
Prismatic	Peds taller than wide, common in B horizons of well-developed soils. <i>May be columnar due to eluviation (old age or high sodium content will produce columnar peds of prismatic peds).</i>	 Prismatic
Structure destroyed		
Puddled	Soils disturbed when wet, puddled or run together; structure is destroyed, pores collapse.	

peds to retain their structure—that is, to absorb water and not disintegrate. There are various agents that bind these aggregates. Fungal mycelia, microbial exudates, hydrogen and calcium ions, and certain clays promote stability; sodium ions and expansion and contraction of certain clays promote instability.

The size and shape of peds are peculiarly significant to the success of plants and other soil-dwelling organisms. Table 17-4 presents illustrations and descrip-

Figure 17-11 Soil profile of Blacklock sand, a podzol soil in the Pygmy Forest, Mendocino County, California. The horizon boundaries are wavy rather than smooth. Note the highly bleached A2 horizon, beneath which is a B2h horizon. Though not apparent here, it is darkly stained with humus. The B2 SIM is a silica cemented hardpan, underlain by the B2t claypan. The entire profile encompasses only about one meter in depth.



tions of several common soil aggregates and the classification of soil structures that possess them. Notice that granular soils are particularly suitable for plant growth, because both water and air can circulate between the peds. The presence of good structure—granular peds, high degree of aggregate stability—is especially desirable in the topsoil. If the texture of the B horizon is sandy, then structure is of little consequence, because drainage is assured. However, clay soils in the B horizon can create a “claypan” or hardpan, which has a deleterious effect on plant life. An excellent example of this is the Blacklock podzol soil (Typic Sideraquod) of the Pygmy Forest in Mendocino County, California. Jenny et al. (1969) note with some amusement the lack of agreement among naturalists as to the cause and effects:

naturalists indulged in an apparent *circulus vitiosus*. On field trips the professors of botany would tell their students that the podzol soil is the cause of the unusual assortment of plant species, whereas the visiting professors of pedology (soil science) would attribute—in the light of classical podzol theory—the soil horizon features to the acid-producing vegetation. While it is true that a species individual responds to its soil niche, it is also true that it modifies that niche, which, in turn, reacts upon the individual.

This Blacklock soil is one of the most acid soils known, with a pH of 2.8–3.9 and a characteristic bleached white surface horizon underlain by an iron-cemented hardpan (Figure 17-11). Leachate from the acid conifer litter cannot drain through the hardpan but continues to remove nutrients from the already depauperate surface soil.

Soil Chemistry

One of the most important aspects of soil chemistry is the state of and the relationship between chemical elements in soil systems, as soluble ions in the soil solution, as adsorbed ions on charged particles, and as constituents of mineral and organic particles. Soil chemistry deals primarily with reactive materials, which are the chemicals that are of significance to living things. Reactive materials tend toward equilibria in a changeable environment. There is virtually no correlation between the absolute and relative amounts of an element present in the soil and the significance of that element to life. Oxygen (49.5%), silicon (25.8%), and aluminum (7.5%) are the most abundant, the second most abundant, and the third most abundant elements, respectively, in the earth's crust. Yet, they are relatively inaccessible due to the bound nature of their chemical compounds. By contrast, the most important elements in soil chemistry (Table 17-5) make up less than 16% of the earth's crust materials. Soil pH, cation exchange capacity, and the nature of the interface between the solid phase and the liquid phase are all important aspects of soil chemistry.

Soil pH

Typical soils exhibit pH values that range from 4 to 8, although certain soils may have values higher or lower. For instance, saline soils commonly exhibit pH values ranging from 7.3 to 8.5. In southern Death Valley, California, saline soils support a saltbush scrub of desert holly (*Atriplex hymenelytra*), Parry saltbush (*A. parryi*), and honeysweet tidestromia (*Tidestromia oblongifolia*). The distinctly alkaline soils of the broad desert valleys of the Great Basin region of western North

Table 17-5 Elements important in soil chemistry and their chemical symbols and principal ions. (Reprinted by permission of McGraw-Hill Book Co. from *Soils and Soil Fertility* by Thompson and Troeh 1973.)

Element	Principal ions	Element	Principal ions
Aluminum	Al^{+++}	Manganese	Mn^{++} , MnO_4^-
Boron	$\text{B}_4\text{O}_7^{--}$	Molybdenum	MoO_4^{--}
Calcium	Ca^{++}	Nitrogen	NH_4^+ , NO_2^- , NO_3^-
Carbon	CO_3^{--} , HCO_3^-	Oxygen	With other elements
Chlorine	Cl^-	Phosphorus	H_2PO_4^- , HPO_4^{--}
Cobalt	Co^{++}	Potassium	K^+
Copper	Cu^{++}	Sodium	Na^+
Hydrogen	H^+ , OH^-	Sulfur	SO_4^{--}
Iron	Fe^{++} , Fe^{+++}	Zinc	Zn^{++}
Magnesium	Mg^{++}		

America support a shadscale scrub vegetation dominated by shadscale saltbush (*A. confertifolia*) and bud sagebrush (*Artemisia spinescens*). A pH of less than 6.6 is considered an acid soil, and vernal pools have pH values of perhaps 6.4–6.85 (a vernal pool is an ephemeral body of water present in spring). Very ancient soils, especially those covered by coniferous forest in humid regions, have very low pH values. The pygmy cypress (*Cupressus pygmaea*) grows in the Blacklock podzol soil of the Mendocino Forest, which we noted has an extremely acid soil (pH of 2.8–3.9). The highly weathered Blacklock podzol soils are half a million years old. They consist primarily of very resistant quartz grains and contain very small amounts of nutrients for plant growth. In general, leaching tends to remove basic cations and thus lower the pH of the soil. Strong podzols with a pH less than 4.0 develop in as little as 10,000 years in glacial materials.

The influence of plants on the soil pH is very complex. Nutrients (both cations and anions) are brought up from subsurface soil by plants and then deposited with litter on the surface. Grasses tend to use more bases, and grass cover may act to keep the soil pH from dropping. Coniferous litter, on the other hand, provides acid leachate, which effectively lowers the pH. Fewer cations are held on exchange sites, and more cations are released by acid weathering. Figure 17-12 shows the changing availability of plant nutrients in response to changing pH values.

Agricultural soils may develop a low pH due to fertilizer applications that add nitrogen and sulfur. Lime is often applied to acid soils to make them more basic. Very alkaline soils, on the other hand, may be treated with gypsum to lower pH. In most soils, the pH cannot be changed very much because of very effective soil pH buffering.

Although the direct effects of soil pH on plant growth are very limited, the indirect effects are numerous and significant. Toxicity of certain metals, such as aluminum and manganese (which are more soluble at lower pH), would be a direct effect of pH value. The most important indirect effect of soil pH is its influence on nutrient availability (Figure 17-12). Rates of weathering, availability of nutrients such as nitrogen, phosphorus, and sulfur, and the leachability of nutrients such as potassium are all influenced by soil pH. The availability of nutrients with low solubilities, such as phosphorus, is dramatically influenced by pH changes. Calcium phosphates are less soluble as the pH climbs, and iron and aluminum phosphates are less soluble as the pH drops. A near-neutral pH, between 6.5 and 7.5, is best for phosphorus availability—indeed for the availability of most nutrients needed for plant growth.

Humus Formation

The action of microorganisms on organic materials eventually converts them into humus. The organic materials become very finely divided and nearly black in color. By virtue of their tiny size, these particles can coat soil particles to the point that the soil appears black when it contains, for example, as little as 5% organic matter. Root exudates, earthworm excreta, and the like may provide a significant contribution to this nonliving, finely divided organic matter. However,

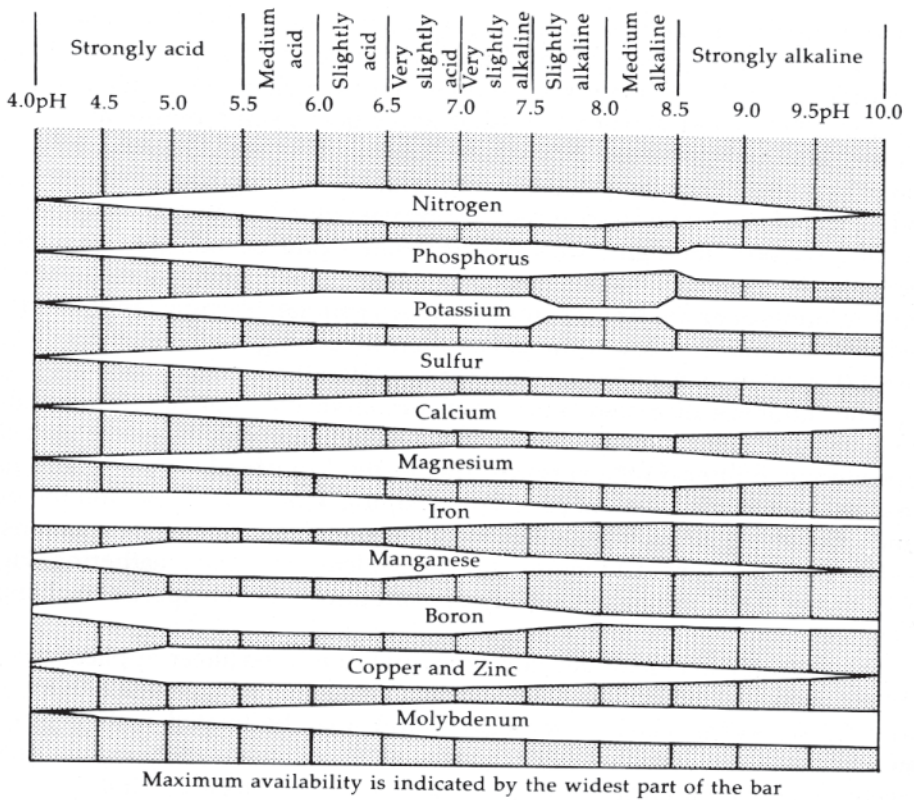


Figure 17-12 A schematic illustration of the relationship between pH and plant nutrient availability. (Reprinted by permission of McGraw-Hill Book Co. from *Soils and Soil Fertility* by Thompson and Troeh 1973.)

probably 90% of humus is attributable to microorganisms, microbial tissue, and microbial activities on litter.

Humus, an intermediate product of decomposition, may decompose at an annual turnover rate of less than 3% in temperate regions. It comprises not only the resistant fraction of the organic matter but also the microbes that have accomplished the decomposition of the more decomposable fraction. In a climax community or even a stable seral one, the amount of new humus formed each year may approximate that which is released by decomposition. The significance of slowly decomposing humus is that it provides a secure supply of continually available nutrients for plant growth.

Although the composition of humus varies, 10–15% of its weight is made up of easily identifiable organic compounds: polysaccharides such as cellulose and its decomposition products; polyphenols such as tannins; proteins and their decomposition products; and a host of hydrocarbons, organic acids, alcohols, esters, and aldehydes. The rest of the humus is not so easily identified but is characterized as humic materials, many of which contain reactive groups—car-

boxyl, amine, phenolic hydroxyl groups—combined with chains and rings, forming rather large and complex molecules.

Cation Exchange Capacity

Cation exchange capacity (CEC) is a measure of the number of negatively charged sites on soil particles that attract **exchangeable cations**, that is, positively charged ions that can be replaced by other such ions in the soil solution. Three factors strongly influence the cation exchange capacity of a soil: its clay content, the kinds of clay minerals or amorphous colloids (allophane) it contains, and the humus content. We have already commented on the significance of the surface-to-volume ratio to mineral storage. The negative charge, which provides for nutrient storage on clay mineral particles, is a result of, for instance, substitution of ions within clay mineral structures or ionization of hydrogen ions from OH groups, which also produces negatively charged surface sites.

Tropical Rainforest Soils

Let us consider the interaction of vegetation, soil texture, rainfall regime, temperature regime, and soil microorganisms, in reference to cation exchange in a particular ecosystem, the tropical rainforest. Soils of the tropical rainforest (TRF) are usually highly weathered and **laterized**—there is little or no horizon development in the reddish-brown loam, which is comprised of aluminum and ferric sesquioxides (Al_2O_3 , Fe_2O_3). Lateritic soils (Oxisols) are very acidic and nutrient poor. The litter layer is extremely thin because mobilization of nutrients occurs at a very rapid pace in the warm, moist rainforest. Even wood is quickly recycled by subterranean termites.

The luxuriant growth of vegetation ($52.5 \text{ t ha}^{-1} \text{ yr}^{-1}$ gross productivity) belies the extremely poor nutrient content of most TRF soils. Walter (1979) states that the entire nutrient reserve required by the forest is contained in the phytomass.

What mechanism allows for litter recycling, so that mobile nutrients are not leached away by the inevitable rains? Work by Went and Stark (1968) showed that feeding roots of TRF trees are directly connected to litter. These rootlets exploit the hyphae of mycorrhizal fungi to obtain nutrients. As noted in Chapter 7, these are mutualistic fungi, interacting with the roots and increasing their host's ability to extract nutrients from the forest litter. Recall that many fungi are saprobic, obtaining nutrients by absorption. Due to this highly adaptive relationship, there is little chance that free, mobile ions released from litter biomass will be lost through leaching.

As long as the rainforest is not disturbed, the lack of a soil storage compartment is unimportant. The nearly complete recycling of materials allows for stability for hundreds or even thousands of years. Yet deforestation through burning, clearing of the land for cultivation, or logging, removes the only significant storage facility, the phytomass, and quickly allows the mobilized ions to move out of the system. The luxuriant growth of the virgin forest is never (in human terms) renewed.

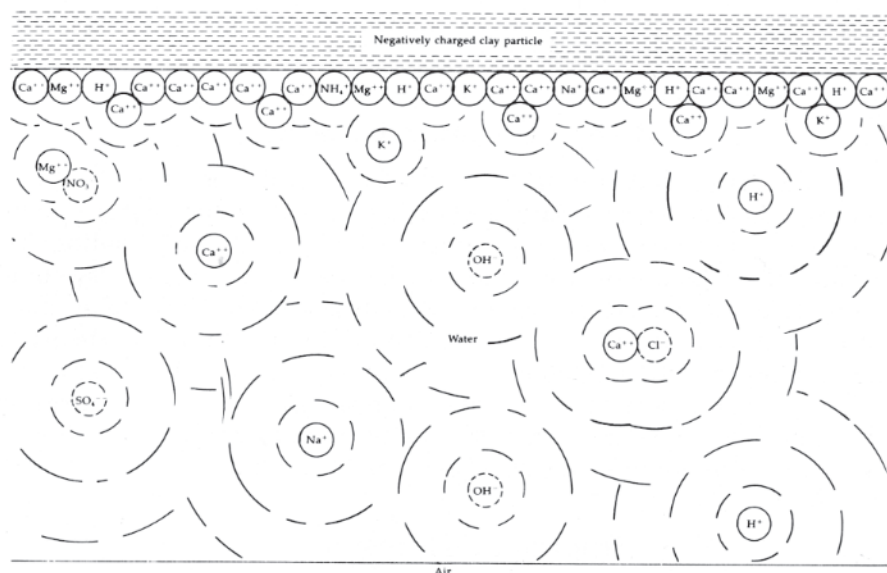


Figure 17-13 An illustration of how ions are distributed in the soil solution near a negatively charged clay particle. (Reprinted by permission of McGraw-Hill Book Co. from *Soils and Soil Fertility* by Thompson and Troeh 1973.)

Soil Solution

There is no way that soil water can be withdrawn and studied as a separate entity; it must be understood *in situ*, in relationship to its function as the liquid medium that renders the soil a habitable environment. The readily available ions are present in the soil solution and maintain an equilibrium with adsorbed ions. These, in turn, maintain an equilibrium with adsorbed mineral ions (Figure 17-13). Plant removal of nutrients from the soil solution lowers the concentration of those nutrients in the rhizosphere; ions move toward the rhizosphere down a chemical gradient. This shift, due to solubility constants of the ions involved, enhances the release of adsorbed ions, and so forth.

For example, plant uptake of phosphorous—roots removing P from the soil solution—establishes a concentration gradient in the soil solution such that the movement of P is toward the root. Under optimum conditions, daily uptake of P in grassland may be as much as 50 times the quantity of P in the solution pool, the pool itself being replenished from the labile pool (that is, from adsorbed ions). When conditions of temperature and water are exactly right, the labile pool has the potential of replacing the solution pool 250 times daily. As we experience time, there is neither gain nor loss of phosphorus in grasslands, and the cycle itself is probably governed as much by biotic factors as by abiotic factors.

With continued uptake, the concentration of P in the cytoplasm of root cells may exceed concentrations by up to two orders of magnitude in ambient soil

solution. It is well known that mycorrhizae greatly enhance uptake of P. This enhancement is attributed to the tremendously enlarged contact zone of root and soil, mediated through hyphal fungi. A generalized phosphorus cycle is depicted in Figure 13-3. Phosphorus participates in a sedimentary cycle, which means that local deficits may occur.

Soil Taxonomy

The naming of soils allows one to impart a great deal of information simply by invoking the name of a certain soil. A name is applied to an individual kind of soil that has a definable existence on the landscape. A unit of soil, large enough such that its profile and horizons can be studied and its properties defined, is called a **pedon**. This is the smallest practical volume that can be properly referred to as a soil. A pedon may range from 1 m² to 10 m² in area, depending on the continuity or cyclic discontinuity on the soil's horizons within short linear distances (1 to 7 m). Pedons of soils with continuous horizons have the smallest area (1 m²). Groups of similar adjacent pedons, termed **polypedons**, form the basis for soil survey and soil classification.

For names to be used, a classification system must be scrupulously followed and names must be assigned on the basis of certain specified properties. Characteristics that are used to define a soil are termed **differentiating characteristics** or **differentiae**.

Soil scientists strive for a classification scheme that reflects a natural relatedness; the scheme of naming is called **soil taxonomy**. *Classification* is the broader term and may include schemes of ordering for specific practical purposes, whereas taxonomy is concerned with relatedness among groups. Nevertheless, any taxonomic system is an arbitrary creation of its author, reflecting certain biases and devised to fill a particular need.

History

The term *soil* is very old, and humans have been concerned with soil as a medium for plant growth for millenia. In the 1870s, a Russian school led by Dokuchaiev developed and introduced a revolutionary new concept of soil and a system of naming soils (Soil Survey Staff 1975).

Soils were . . . independent natural bodies, each with a unique morphology resulting from a unique combination of climate, living matter, earthy parent materials, relief, and age of land form. The morphology . . . reflected the combined effects of the particular set of genetic factors responsible for its development.

The Russian soil taxonomy was adopted first in Europe, then in this country early in this century. As more soils were described and other information became available, the weaknesses of the Russian system became increasingly apparent. In 1938, a system was developed in the United States and published under the title "Classification of Soils on the Basis of Their Characteristics" (Baldwin et al.

1938). Then, starting in the 1950s, a taxonomic scheme was developed, modified, improved, and presented as a series of approximations, each one more refined and more workable than the last. Six of these approximations were tested by the Soil Survey Staff of the U.S. Department of Agriculture, various U.S. land grant universities, and interested European individuals and agencies. The seventh approximation was published and issued on a limited basis in 1960 at the Seventh International Congress of Soil Science at Madison, Wisconsin, for wider testing and evaluation (Soil Survey Staff 1960). It was generally referred to as the Seventh Approximation or "the brown book." In 1965, a modified form was officially adopted for soil classification in this country by the National Cooperative Soil Survey. With additional modification, it was finally published for unrestricted use as a comprehensive system in 1975 by the Soil Survey Staff of the U.S. Department of Agriculture as Agricultural Handbook 436, *Soil Taxonomy—A Basic System of Soil Classification for Making and Interpreting Soil Surveys*. It is this system that we will discuss here. (For current information regarding placement of various soils into this taxonomic scheme, see *Classification of Soil Series*, 1986 USDA Soil Conservation Service publication.)

To be useful, a soil taxonomy should have the following attributes.

1. The definition of each taxon should have the same meaning for each user.
2. The taxonomy should be a multicategoric system, with many taxa in the lower categories.
3. The taxa should refer to real soils, known to occupy geographic areas.
4. Differentiae should be soil properties that can be discerned in the field, inferred from other properties that are observable in the field, or taken from the combined data of soil science and other disciplines.
5. The taxonomy should be modifiable with a minimum of perturbation to the system.
6. The differentiae should maintain pristine soil and manipulated soil that is its equivalent in the same taxon whenever possible.
7. The taxonomy should provide taxa for all the soils of a landscape or at least be capable of such provision (Soil Survey Staff 1975).

Differentiating Properties of Soils

The properties of a polypedon should serve as differentiae. Soil color and soil horizons have long been used for differentiae. Properties of soil that interact with other properties should be carefully considered when differentiae are selected to define soil taxa.

There are six defined diagnostic surface horizons or **epipedons**. An epipedon is a horizon, or horizons, forming in the upper part of the soil profile. We will name these but not describe them, due to limitations of space. A subsurface horizon (often B, but perhaps part of A) is also of diagnostic value, and the names, descriptions, and genesis or mode of formation of these horizons are also important in soil classification. The diagnostic epipedons are mollic, anthropic,

umbric, histic, plaggen, and ochric. The diagnostic subsurface horizons include argillic, agric, natric, spodic, placic, cambic, oxic, and duripan, among others.

Other properties used in defining a soil include an abrupt textural change, mineralogical composition, amorphous material dominating the exchange complex, microrelief, lithic contact, organic soil materials, particle-size classes, and so on.

The Structure of Soil Taxonomy

The USDA soil taxonomy system recognizes a hierarchy of categories. Each category includes a set of taxa. The series is the unit of classification. One of the strengths of the system lies in its capacity for internal modification, as new information is gathered, without disruption of the rest of the system. The hierarchy is as follows:

<i>Categories</i>	<i>Example of taxa within categories</i>
Order	Entisol
Suborder	Orthent
Great Group	Cryorthent
Subgroup	Typic Cryorthent
Family	Loamy-skeletal, carbonatic Typic Cryorthent
Series	Swift Creek

Most of the formative elements in the names of soil orders (Table 17-6) are derived from Latin or Greek words appropriate to that soil order, and the soil order name is formed with the suffix *-sol*, from the Latin *solum* for soil. The

Table 17-6 Formative elements in the names of soil orders. (From Soil Survey Staff 1975. Courtesy of the U.S. Dept. of Agriculture.)

Name of order	Formative element in name of order	Derivation of formative element	Pronunciation of formative element
Alfisol	Alf	Meaningless syllable	Pedalfer
Aridisol	Id	<i>L. aridus</i> , dry	Arid
Entisol	Ent	Meaningless syllable	Recent
Histosol	Ist	<i>Gr. histos</i> , tissue	Histology
Inceptisol	Ept	<i>L. inceptum</i> , beginning	Inception
Mollisol	Oll	<i>L. mollis</i> , soft	Mollify
Oxisol	Ox	<i>E. oxide</i> , oxide	Oxide
Spodosol	Od	<i>Gr. spodos</i> , wood ash	Odd
Ultisol	Ult	<i>L. ultimus</i> , last	Ultimate
Vertisol	Ert	<i>L. verito</i> , turn	Invert

formative elements *alf* and *ent* are not from Greek or Latin, but are meaningless syllables. Alfisols are variable, often found in continually wet, cold, or seasonally dry climates. Entisols are often young soils with little or no horizon development. The derivations of the other formative elements are listed in Table 17-6.

The names of suborders are a combination of the formative element of the parent order as the suffix with a prefix that suggests the diagnostic properties of the soil. For example, *alb* (from *albus*, white) with *oll* (from mollisol, soft) names a suborder of the Mollisol order with a white horizon—*Alboll*. Similarly, a great group name consists of the name of the suborder coupled with a prefix containing one or two formative elements to refer to definitive properties. For instance, *natralboll* indicates a Mollisol with an *albic* (white eluvial) horizon and a *natric* horizon (an illuvial subsoil with excess exchangeable sodium).

The great group name is combined with one or more modifiers to generate the subgroup name. *Typic Natralboll* would denote the subgroup thought to typify the great group. Families are named with polynomials consisting of the subgroup name modified by adjectives in a specified order that are names of classes describing the soil's particle size mineralogy, reaction, and temperature classes, among other factors. Series names are place names, usually taken from the region where the soil was first described. In the field, profiles that do not match the descriptions of officially recognized series may be designated as **variants** of the existing, closely related series. With time and additional field information, a variant may be defined as an official soil series with its own unique name.

Summary

The components of soil are mineral grains, organic matter, water, and air. They provide, respectively, plant anchorage and nutrients, intrasystem cycling, solvent medium, and oxygen and nitrogen. The upper layers of soil weather and change in response to abiotic and biotic factors. The first phase of soil formation is rock weathering, and the second is biochemical weathering.

Soil formation relies on climate, organisms, topography, parent material, and time, as does soil profile development, which encompasses development of soil horizons. Soils can be classed as zonal, azonal, and intrazonal.

Soil texture refers to the content of the soil, by weight, of the sand, silt, and clay fractions. Skeletal support and permeability are provided by the largest particles; clay provides water and nutrient storage; silt aids in water storage and weathers to produce additional nutrients and clay.

Micelles (clay particles and bits of organic matter) provide the primary storage for nutrients in the soil. The nature of the arrangements of peds and aggregate stability are the most significant characteristics of soil structure.

Equilibrium among ions in the soil solution, adsorbed ions, and absorbed ions are concerns of soil chemistry. Most soils exhibit a pH range from about 4 to 8. Ancient soils may be very acid. Climate, vegetation, parent material, and relief determine soil pH, which in turn affects vegetation and rates of weathering.

Soil microorganisms convert organic matter to humus, a finely divided, nearly black fraction that decomposes at a stable equilibrium rate in climax communities. Cation exchange capacity is influenced by the presence and the kinds of clay minerals, allophane, and humus. Clays with expanding lattice provide more efficient cation exchange than do nonexpanding clays. Some tropical soils have little or no horizon development, and are ancient, nutrient poor soils with little exchange capacity. The phytomass associated with these soils provides the bulk of their nutrient storage.

Soils are classified by their differentiating characteristics. Soil taxonomy is a classification scheme designed to reflect a natural relatedness among soil groups. To be effective, a taxonomy must have definitions for each taxon that have the same meaning for each user. The system currently used in the United States was developed by the Soil Survey Staff of the U.S. Department of Agriculture, Soil Conservation Service. It relies on the properties of the polypedon for differentiating characteristics and has a very complex but flexible multicategoric system. Names for the soil taxa are formed with syllables selected from terms that reflect the nature of the soil.