

## CHAPTER 2

# THE SOLID MATERIALS OF THE EARTH'S SURFACE

### 1. INTRODUCTION

**1.1** To a great extent in this course, we will be dealing with processes that act on the solid materials at and near the Earth's surface. This chapter might better be called "the ground beneath your feet". This is the place to deal with the nature of the Earth's surface materials, which in later sections of the chapter I will be calling regolith, sediment, and soil.

**1.2** I purposely did not specify any previous knowledge of geology as a prerequisite for this course, so it is important, here in the first part of this chapter, for me to provide you with some background on Earth materials.

**1.3** We will be dealing almost exclusively with the Earth's continental surfaces. There are profound geological differences between the continents and the ocean basins, in terms of origin, age, history, and composition. Here I'll present, very briefly, some basic things about geology. (For more depth on such matters you would need to take a course like "The Earth: What It Is, How It Works", given in the Harvard Extension program in the fall semester of 2005–2006 and likely to be offered again in the not-too-distant future.)

**1.4** In a gross sense, the Earth is a layered body (Figure 2-1). To a first approximation, it consists of concentric shells: the core, the mantle, and the crust.

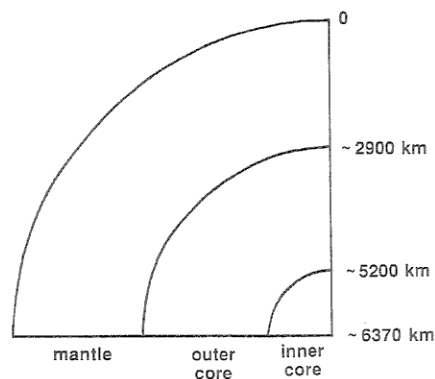


Figure 2-1: Schematic cross section through the Earth.

**The core:** The core consists mostly of iron, alloyed with a small percentage of certain other chemical elements. The outer part of the core is liquid, and the inner part is solid. There is no need here to deal with how that's known—or indeed why there is an iron core in the first place.

**The mantle:** The mantle, which constitutes the greater part of the mass of the Earth, has its lower boundary, with the core, at a depth of about 2900 km. (The radius of the Earth is about 6400 km.) The mantle consists almost entirely of solid rock, except for certain places in the uppermost mantle where magma (= melted rock) is formed at certain times. (There is a deplorably widespread and thoroughly mistaken belief that the upper mantle is everywhere molten.) Saying that the mantle is solid rock, true as that statement is, is somewhat misleading, though, because the solid rock of the mantle can flow plastically in the same way that a very viscous liquid, like honey or corn syrup, flows. You will see the same kind of behavior in glacier ice, in a later chapter. “Silly putty” is another good example of such a material. The uppermost part of the mantle, however, is sufficiently cool that it behaves not as a very stiff liquid but as a rigid solid. This situation, whereby the uppermost rigid part of the mantle (which, together with the overlying crust, is called the *lithosphere*) rides on the plastically flowing part of the mantle beneath (called the *asthenosphere*), is the basis for what is called plate tectonics (Figure 2-2).

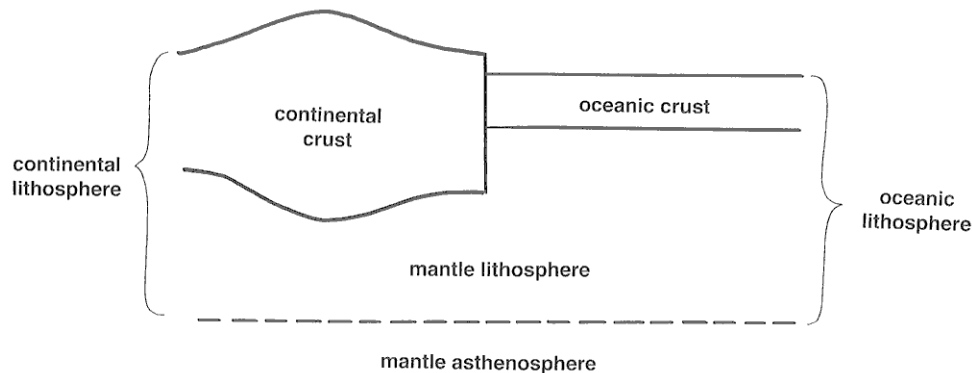


Figure 2-2. Schematic cross section through the outer part of the Earth.

**The crust:** The uppermost skin of the Earth, above the mantle, is called the crust (Figure 2-2). There are two basic kinds of crust, very different from one another in properties and origin: oceanic crust and continental crust. Oceanic crust is relatively thin, seldom more than seven to eight kilometers. Continental crust, on the other hand, is relatively thick, mostly thirty to fifty kilometers. The features of the upper surface of the continental crust, and the great variety of

physical and chemical (and biological) processes and phenomena associated with that surface, are the focus of this course.

**1.5** You probably know something about, or at least have heard about, plate tectonics. The lithosphere, the outermost part of the Earth, comprising the crust and the uppermost, rigid part of the mantle (Figure 2-2), is in the form of a number of segments, called *lithospheric plates* (or just *plates*), which are in contact with one another along what are called plate boundaries. The movements of these plates, and the consequences of those movements, especially in terms of what happens at plate boundaries, are collectively termed *plate tectonics*.

**1.6** We could get along fairly well in this course without reference to the core or the mantle, or, for that matter, even to the lower part of the continental crust. Deep subsurface processes, in the crust and in the underlying upper mantle, have a great indirect importance to this course, however: such processes (the nature of which is beyond the scope of this course!) cause *uplift and subsidence of the land surface of the continents*, both sharp and localized, in the case of active mountain belts like the western Cordillera of North America, and broad and gentle, affecting large areas of the stable continent of North America east of the Cordillera. Uplift and subsidence, in turn, are instrumental in erosion, transportation, and deposition of sedimentary materials and in shaping of the surface topography of the continent. We'll return to such matters in the final chapter.

## 2. MINERALS

**2.1** The technical definition of a mineral is elegant: *a naturally occurring crystalline solid*. That definition calls for a bit of commentary, though. By “crystalline” is meant that the atoms of the mineral are arranged in a regular three-dimensional array, called its *crystal structure*. The qualification “naturally occurring” excludes the multitude of crystalline solids that can be synthesized in the laboratory but are not found in the natural environment. Also, there are naturally occurring solids, like glass or amber, that do not have a crystal structure and therefore technically are not minerals.

**2.2** There are thousands of named minerals—but, fortunately for us in this course, only a few are common rock-forming minerals, and even fewer figure prominently in the unconsolidated material that mantles the continents.

**2.3** Most of the common minerals are of a class called *silicate minerals*. Silicate minerals have as their basic building blocks a polyatomic (five-atom) unit, called a *silica tetrahedron*, that consists of one atom of silicon, relatively small, surrounded by four atoms of oxygen, relatively large, to give the shape of a tetrahedron (Figure 2-3). The five atoms are bonded very strongly together. The reason why silicate minerals are so common is that, in terms of abundances of the

chemical elements in the crust, oxygen is the big number one and silicon is second.

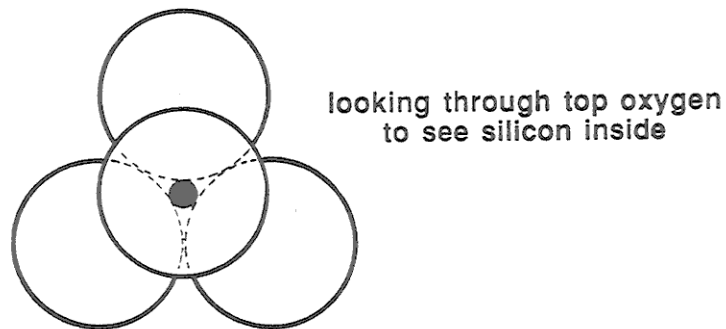


Figure 2-3. Schematic view of the silica tetrahedron, looking straight down on one of the oxygens. The small black dot represents the silica atom.

**2.3** The atoms in mineral crystals are held together by strong forces called **bonds**. An understanding of the nature of bonds necessitates some familiarity with atomic structure. All atoms consist of a positively charged nucleus around which a number of negatively charged electrons are in orbit. The fundamental nature of the electron orbits is complex, but for the purposes of understanding bonds in minerals it's sufficient to know that electron orbits exist in distinctive groups known as **shells**.

**2.4** Chemical elements like helium, neon, or argon whose atoms have just enough electrons to fill all of the shells are chemically almost entirely inert. Elements whose atoms have their outermost electron shell almost filled have a strong tendency to take on one or a few electrons to fill the shell, and in so doing they acquire a negative electric charge. Similarly, elements whose atoms have only one or a few electrons in their outermost shell lose those electrons to revert to a filled-shell configuration, and in so doing they acquire a positive electric charge. Such atoms with electric charges are called **ions**.

**2.5** According to what is known in physics as Coulomb's law, unlike electric charges attract one another and like charges repel one another. Certain combinations of positively charged and negatively charged ions can become packed together in a regular three-dimensional array in such a way that the sum of all the forces, attractive and repulsive, among all of the ions in the structure is attractive, meaning that the structure is a stable one and can exist as a mineral. Almost all minerals are of this nature, and are called **ionic crystals**; the bonds in

such a crystal are called *ionic bonds*. Ions can also be formed from two or more atoms that share their electrons in such a way that each atom effectively has a filled-shell configuration. The bonds among such atoms are called *covalent bonds*. Covalent bonds are much stronger than ionic bonds, and the covalently bonded atoms form a single ion that is available for ionic bonding with other ions.

**2.6** It turns out that in terms of electric charge the silica tetrahedron has a net negative charge. The silica tetrahedra are bonded to various positively charged atoms, mainly of the elements aluminum, iron, magnesium, calcium, potassium, and sodium. These ionic bonds are weaker than the strong covalent bonds within the silica tetrahedra. When a mineral is broken, it breaks along these ionic bonds rather than through the bonds within the silica tetrahedra.

**2.7** There is an important further complication to silicate minerals, which makes them highly diverse (and almost unique in the universe!): individual tetrahedra can be joined together by sharing one or more (two, three, or even all four) of their oxygens. The phenomenon is called *polymerization*. By polymerization, zillions of tetrahedra can be joined into chains or sheets that stretch all the way through a macroscopic mineral grain. In some minerals, all of the oxygens of the tetrahedra are shared, resulting in a complex three-dimensional network of shared tetrahedra.

**2.8** There are other important minerals besides silicate minerals. Oxide minerals are perhaps the most important kind, in the context of this course. Oxide minerals are those in which negatively charged oxygen atoms are bonded with various positively charged ions. The most important oxides are *iron oxides* and *aluminum oxides*. These oxides of iron and aluminum are generally unspectacular in their appearance, but they are voluminous and important in Earth-surface environments. Carbonate minerals are also important near-surface non-silicate minerals. In carbonates, a negatively charged carbonate ion, consisting of one carbon atom covalently bonded with three oxygen atoms, is ionically bonded with calcium, magnesium, and/or iron ions; calcite and dolomite are common carbonate minerals.

**2.9** To conclude this section, I'll list just the very most important kinds of minerals in the Earth's upper crust. It's important to know something about these minerals, because the mineralogy of solid Earth materials is of great importance in determining how those Earth materials are affected by their physical and chemical environment.

*ferromagnesian silicates* (“*femags*”): This collection of minerals (olivines, pyroxenes, amphiboles, biotite, and certain metamorphic minerals) contain iron and/or magnesium, in various proportions, as their main positive ions. They tend

to be dark in color. They are abundant in igneous and metamorphic rocks—but not in sedimentary rocks, because they tend to weather rapidly.

*micas:* Micas consist of sheets of polymerized silica tetrahedrons, each with three of their four oxygens shared, with various positive ions between the sheets. The main micas are muscovite (a potassium mica), which is common in igneous, metamorphic, and sedimentary rocks, and biotite (a ferromagnesian mineral, with iron and magnesium rather than potassium). Biotite is common in both metamorphic and igneous rocks, but it weathers rapidly, so it's uncommon in sediments and sedimentary rocks.

*feldspars:* Feldspars are the most abundant kind of mineral in the Earth's crust. They are network silicates with either potassium (these are called potassium feldspar, or just Kspar for short) or varying proportions of sodium and calcium (these are called plagioclase) in their structure. Feldspars are the major minerals of igneous rocks and are common in many metamorphic rocks. Potassium feldspar is common in sediments and sedimentary rocks, but plagioclase is not, because it weathers readily.

*quartz:* Quartz, with the simple formula  $\text{SiO}_2$ , is a network silicate mineral with a three-dimensional network of silica tetrahedra, with all of the oxygens shared with adjacent tetrahedra. It is common but not abundant in igneous rocks, is variably abundant in metamorphic rocks, and is very abundant in sediments and sedimentary rocks.

*iron oxides:* Iron oxides, of which there are several forms, are common only in sediments and sedimentary rocks. In soils, there are several poorly crystallized hydrous iron oxide minerals.

*aluminum oxides:* In poorly crystallized form, these are characteristic of soils formed by deep weathering of aluminum-bearing rocks under intense weathering conditions.

*carbonates:* The calcium carbonate minerals calcite and aragonite, along with the calcium–magnesium mineral dolomite, are the main minerals in chemically precipitated carbonate sedimentary rocks, and they are common as well in sediments in the warm and shallow oceans.

*clay minerals:* Several minerals with sheet-silicate structures are common as very fine-grained plate-shaped particles, some even smaller than a micrometer (one-thousandth of a millimeter; too small even to be seen with a powerful optical microscope). Clay minerals are the main constituent of most fine-grained muds.

### 3. ROCKS

**3.1** Rocks, in the form of what is called *bedrock*, are exposed at the Earth's surface over wide areas, and they everywhere underlie the unconsolidated surficial

materials, called *regolith*, you will learn about later in this chapter. Moreover, owing to their physical and chemical breakdown under the conditions of the Earth's surface environment they are the source of that surficial material. There's no room in this course for a systematic and detailed treatment of rocks—that's the province of an introductory course in geology—but you should know at least some general things about them.

**3.2** Rocks are naturally occurring aggregates of minerals. Again, that succinct definition is not very revealing of the nature and variety of rocks. I suppose that it verges on common knowledge that there are three kinds of rocks: *igneous*, *sedimentary*, and *metamorphic*. These three categories are for the most part distinct, although there is some fuzziness to the boundaries.

**3.3** Igneous rocks are those that form by cooling and solidification of *magma*. Magma is the term for melted rock in the Earth's interior. At certain times and in certain places in the Earth's shallow interior, down to a hundred or so kilometers, the rocks of the upper mantle or lower continental crust melt, to form magma. The melting is only partial, commonly up to fifteen to twenty percent of the rock, but the magma collects and then moves upward owing to its buoyancy (it's slightly less dense than the surrounding rock). It either becomes parked in subsurface spaces, called magma chambers, there to cool to form coarse-grained igneous rocks like granite or gabbro, or it rises all the way to surface to be ejected from volcanoes, either as flowing liquids or as already solidified particles.

**3.4** Most of the igneous rocks you are likely to encounter as bedrock outcrops on the continents are of three general kinds (although this is a great oversimplification of the wide range of igneous rocks):

- coarse-grained and light-colored rocks like *granite*, which consist mainly of feldspar with some quartz;
- fine-grained and dark-colored rocks like *basalt*, which consist mainly of ferromagnesian minerals and feldspars; and
- rocks, called *tuff*, that form by lithification of volcanic ash, the solid particles that are ejected explosively from certain kinds of volcanoes.

**3.5** Sedimentary rocks are those that form from sediments that are deposited at the Earth's surface and become lithified (turned into solid rock, by a variety of processes) as they are buried more and more deeply in the earth's crust. Sedimentary rocks vary widely in composition and origin. There are two major realms of sediments and sedimentary rocks: *siliciclastic* and *chemical*. Siliciclastic sediments and rocks are those that consist of particles of minerals or rock that are derived from weathering of bedrock on the continents. Chemical

sediments and rocks are those that are precipitated from natural surface waters, mainly in the ocean. By far the most important chemical sediments are carbonate sediments. Most carbonate sediment is produced by carbonate-secreting organisms, like clams or corals, although some is precipitated inorganically, in somewhat the same way that rock candy grows in a beaker on your windowsill.

**3.6** Most of the sedimentary rocks you are likely to encounter are of the following kinds. (Again, this is a highly simplified list, in large part because these major sedimentary rock types grade imperceptibly into one another.)

*sandstones*: Sandstones consist of sand-size particles of various minerals, mainly quartz but also potassium feldspar, as well as sand-size fragments of fine-grained rocks.

*shales*: Shales consist mostly of very fine-grained mineral particles of quartz and clay minerals.

*limestones*: Limestones consist mainly of the calcium carbonate mineral calcite. Also abundant are dolostones, which consist mainly of the carbonate mineral dolomite.

**3.7** Metamorphism refers to changes in mineral composition or rock geometry that occur in solid rocks with increasing temperature and pressure. These changes produce a rock called metamorphic rock. It is important to remember that *changes producing metamorphic rocks take place while the rock is solid*. Any part of the rock that is melted eventually cools to form an igneous rock.

**3.8** The basic idea behind changes in mineral composition during metamorphism is that each mineral has associated with it a range of temperature and pressure in which it is stable, in the sense that it remains in existence indefinitely. When a mineral is out of its stability range, it is liable to be converted into, or replaced by, one or more other minerals that are stable at those temperatures and pressures. A simple way of phrasing this is that *new minerals grow at the expense of the original minerals*. These changes in mineral composition act very slowly, because the rock remains solid and atoms can be exchanged between adjacent mineral grains only by atomic diffusion—although very small percentages of pore solutions along grain boundaries can expedite the transport of atoms.

**3.9** Deformation (change in shape) is an important factor in the development of many metamorphic rocks. For most kinds of metamorphic rocks, change in mineral composition is accompanied by strong shearing deformation. Strong shearing tends to obscure or even obliterate original geometrical features like



stratification in a sedimentary or volcanic rock. Also, most metamorphic rocks show some degree of what is called foliation: development of planar features in the rock, like layering or a tendency for the rock to split along parallel planes.

**3.10** Below are brief descriptions of the most common metamorphic rocks. As with the sedimentary rocks, there are all gradations among these types.

*slate*: Slate is produced by low-intensity metamorphism of shale. Individual crystals are too small to see, but the rock has a strong tendency to split along parallel planes, as a result of growth of new mica crystals of sheet-silicate minerals all parallel to one another.

*phyllite*: Phyllite is produced when slate undergoes further metamorphism. Phyllite typically has a silky sheen.

*schist*: Schist is formed by metamorphism of many igneous and sedimentary rocks. It has clearly visible flakes of sheet-silicate minerals (muscovite, biotite, and/or chlorite), and it tends to split along parallel planes in the same way as slate and phyllite.

*gneiss*: Gneiss consists mainly of granular minerals like feldspar, quartz, and ferromagnesian minerals. Gneiss is often prominently layered.

*quartzite*: As the name implies, quartzite consists mainly of quartz. It forms mostly by metamorphism of quartz-rich sedimentary rocks.

*marble*: Marble consists mainly of calcite. The calcite crystals tend to be large enough to see with the unaided eye, having grown larger during metamorphism without change in overall mineral composition.

## 4. BEDROCK

**4.1.1** By *bedrock* I mean *solid rock, exposed at the Earth's surface or buried at shallow depths, that is connected continuously with rock at greater depths in the crust* (Figure 2-4). A great many geologists spend most of their professional lives studying the bedrock of the Earth's continents.

**4.1.2** In some places on the continents, bedrock is exposed at the surface, in the form of what are called *outcrops* or *exposures*. There are natural exposures (to which the term *outcrop* is typically applied) and artificial exposures. Natural outcrops range in area from small, human-scale patches peeping out from the surrounding blanket of regolith, to whole mountainsides. Favorite places to find good outcrops are stream beds, hill slopes, mountain tops, and sea cliffs. These

days, highway road cuts make the best artificial exposures; in times past, railroad cuts were prime exposures. Building foundations are good but temporary.

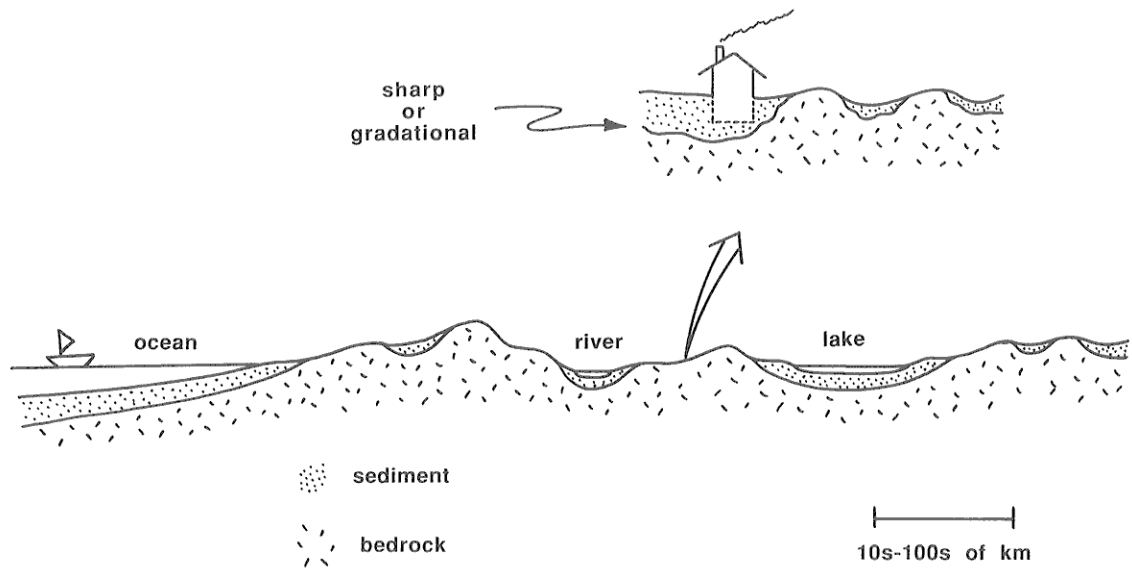


Figure 2-4. The bedrock of the continents.

**4.1.3** The percentage of continental surface where bedrock is exposed varies greatly, from nearly one hundred percent, in regions with rigorous climate and steep slopes, to zero, in regions with warm humid climates and little topographic relief. (The term *relief* is used for *the magnitude of differences in elevation of the land surface in a given region.*) The extent of exposure of bedrock depends mainly on the competition between two opposing processes: the rate of production of new regolith by weathering of exposed bedrock, and the rate of removal of existing regolith by the erosive action of running water or moving glacier ice or by downslope movement of regolith by the pull of gravity. Climate is important because the rate of production of regolith by weathering is much greater in warm and humid climates than in cold and arid climates. In some regions with a warm, humid climate and low relief, one needs to dig or drill tens or even hundreds of meters down to find fresh bedrock!

## 5. WEATHERING

### 5.1 Introduction

**5.1.1 Weathering** is the term used for the *chemical decomposition and physical disintegration of bedrock at and just below the earth's surface.*

Weathering acts upon all bedrock near the surface, although with greatly varying nature and rate depending upon a number of factors.

**5.1.2** Rock at depth tends to be brought upward toward the surface, by gradual unroofing by weathering at the surface and removal of the weathering products. It's a matter of relative motion: the bedrock at depth may be either motionless (relative to, say, the center of the earth) or moving slowly upward by uplift or downward by subsidence, but in any case *it's moving upward relative to the earth's surface if material at the surface is being stripped away*. As the rock nears the surface, it tends to be subjected to what's called *near-surface alteration*: it finds itself at much lower pressures and temperatures than at greater depth, and surface-derived waters, which tend to be oxygenated and carry dissolved CO<sub>2</sub>, have the opportunity to cause certain chemical reactions in many constituent minerals, like ferromagnesian silicates and feldspars. In a sense, you can think of such near-surface alteration as a precursor or forerunner to weathering in the stricter sense. But despite such near-surface alteration, the rock is still solidly bedrock, and nowhere near a condition that could be described as regolith.

**5.1.3** Weathering, in the usually accepted sense, acts upon rock that is close enough to the surface to be affected not only by active circulation of surface-derived waters but also by such things as temperature changes and plant growth. But in thinking about weathering you don't have to restrict yourself to the actual rock surface, upon which the sun shines!

**5.1.4** Weathering is traditionally subdivided into two sets of processes:

- *chemical weathering*, the chemical decomposition of some or all of the constituent minerals or the bedrock.
- *physical weathering*, the mechanical breakdown or disintegration of the bedrock itself into particles large and small.

This is a useful distinction, but here's a very important fact to understand: *chemical weathering and physical weathering tend to act simultaneously and reinforce one another*.

## 5.2 Chemical Weathering

**5.2.1** Chemically, *the earth's surface is a very reactive place*. To begin with, in most places and at least at certain times *the surface is an aqueous environment*, and water is not only a very effective solvent in itself but also a necessary medium for a great many chemical reactions.

**5.2.2** Owing to the high content of oxygen in the atmosphere, in most places *the earth's surface is a well oxygenated environment*, as in the shallow subsurface wherever the concentration of organic matter is not so great and the permeability so low (as is common in bogs and swamps) that the oxygen is consumed in oxidation of organic matter before it passes downward to deeper depths. (If you're unsure about the nature of oxidation, there will be some background material on it later.) Because so many of the common rock-forming minerals contain ferrous iron (that is, iron in the reduced state, and thus susceptible to oxidation to the insoluble ferric state), *oxidation reactions are an important aspect of chemical weathering*.

**5.2.3** In the same way, because the atmosphere contains a substantial percentage of carbon dioxide (CO<sub>2</sub>), *surface waters invariably contain appreciable dissolved CO<sub>2</sub>*. You know already that CO<sub>2</sub> readily dissolves in water. That dissolved CO<sub>2</sub> reacts with the water itself to form a weak acid, *carbonic acid*, according to the reaction



At equilibrium, about one percent of the dissolved CO<sub>2</sub> is in the form of carbonic acid. Most minerals are less stable in an acidic medium than in a neutral medium, so the dissolved CO<sub>2</sub> enhances the weatherability of many minerals.

**5. 2.4** Specifically, what's going on with Reaction 1? A molecule of carbon dioxide, which consists of a carbon atom tightly bound up with two oxygen atoms, combines with a molecule of water, which consists of an oxygen atom tightly bound up with two hydrogen atoms, in such a way that a single, new, larger molecule, with the chemical formula H<sub>2</sub>CO<sub>3</sub>, is produced. In the aqueous solution, this process is going on all the time, and at the same time the H<sub>2</sub>CO<sub>3</sub> molecules are being changed back into CO<sub>2</sub> and H<sub>2</sub>O molecules. Where the balance is struck depends on the imposed concentration of CO<sub>2</sub>, which in this case comes either from the atmosphere or from the chemical breakdown of organic matter. (You probably know that, by photosynthesis, plants take up water and carbon dioxide and convert it into complex carbohydrate molecules, and when the plant dies and the plant tissues are exposed to an oxygenated aqueous environment the reverse happens and the organic matter is broken back down into carbon dioxide and water.)

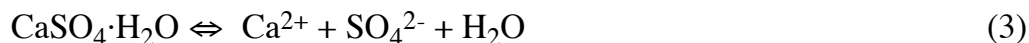
**5.2.5** Carbonic acid is not the only acid common in natural near-surface waters: in most land areas, organisms (both plants and animals) living on or in the soil (that is, the uppermost part of the layer of regolith) use carbon dioxide and

various nutrients to produce organic matter, which in its decay back to the constituent compounds tends to produce *a variety of weak organic acids*.

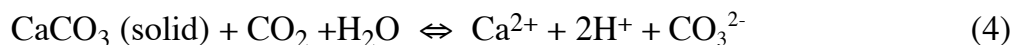
**5.2.6** In the rest of this section, several kinds of chemical reaction will show the range of weathering reactions that are important in chemical decomposition of bedrock minerals. They involve *solution, oxidation, and carbonation*, in various ways. Keep in mind that although these particular reactions are important, *they are just representative reactions*: details vary depending upon the assemblage of minerals that are being weathered, and upon their particular compositions (which, remember, vary widely because of the wide range of ionic substitutions in silicate minerals).

*solution:*

Some minerals simply *dissolve in water*. Halite (sodium chloride) and gypsum (hydrated calcium sulfate) are good examples:



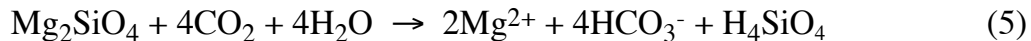
This is the simplest kind of weathering reaction to understand, but not the most important. A more important, and more complicated, kind of solution involves calcite, the constituent of limestone:



The reason this reaction is more complicated is that it involves the atmospheric carbon dioxide dissolved in the water. The  $\text{CO}_2$  has to be involved because of the  $\text{CO}_3^{2-}$  ion both in the structure of the mineral and as a product of the solution and then dissociation of  $\text{CO}_2$  in water. (Incidental note: this reaction can go in either direction, depending upon the concentration of  $\text{CO}_2$ : if  $\text{CO}_2$  is added to the system, as in weathering environments, calcite is dissolved; if it's extracted from the system, as in certain shallow warm ocean environments, calcium carbonate sediment is precipitated. Think about that the next time you're lying on a white carbonate-sand beach in the Caribbean.)

*carbonation of silicate minerals that don't contain aluminum or iron:*

Most silicate minerals that don't contain aluminum contain both Fe<sup>2+</sup> and Mg<sup>2+</sup> ions in various proportions. Rather than writing a single weathering reaction for such minerals, it's neater to look at separate reactions for those with only magnesium and those with only iron. Such minerals, although they exist, are not common, but their weathering reactions are instructive. First look at the weathering of olivine, which is a common mineral in igneous rocks, and one that's very susceptible to chemical weathering. Most olivines in igneous rocks contain some iron in addition to magnesium, but let's assume, for simplicity, that we are dealing with a pure magnesium olivine:



(I've used the single arrow to the right, rather than the double arrow, here and below because the reverse of these reactions is not actually how the minerals get built under deep Earth conditions.)

Note that "silica" (that is, SiO<sub>2</sub> in some form or other) is extracted from the original silicate mineral. In most natural waters, the dominant species of silica in solution is H<sub>4</sub>SiO<sub>4</sub>, a weak acid called *silicic acid*. It's in the form of a molecule that consists of four hydrogen atoms, four oxygen atoms, and one silicon atom, all tightly bonded together.

*carbonation of silicate minerals that don't contain aluminum but **do** contain iron:*

Now look at a ferromagnesian mineral with all Fe and no Mg. (We'll assume that the mineral is a pyroxene mineral; no details here.) In our oxygen-rich near-surface environment, it's a two-step process. The first step is like that expressed in Reaction 5 above, but the numerical values of the coefficients in front of the reactants and products are somewhat different, because we started out with a mineral with a different chemical formula. Qualitatively, the result is the same: *the iron-bearing silicate mineral is broken down in the presence of water and carbon dioxide to put iron and silica in solution.* The dissolved iron is in the ferrous ("plus-two") form, which is freely soluble in pure water. It is converted in the presence of oxygen to ferric iron oxide and silica in solution:



But that's not the end of the story, because the ferrous iron in solution reacts with oxygen dissolved in the water, to precipitate an iron oxide mineral, Fe<sub>2</sub>O<sub>3</sub>:



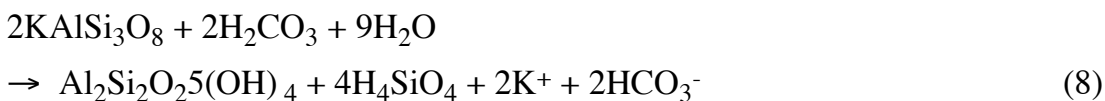
The iron in the  $\text{Fe}_2\text{O}_3$  mineral is in the ferric (“plus-three”) form, which is almost totally insoluble in pure water.

The representation on Reaction 7 is still a bit of a fake, however, because ferric oxides produced at the earth’s surface are commonly hydrous or hydrated: one could describe them as hydrous ferric oxides. There are several closely related minerals of that kind; the most common are goethite,  $\text{FeO}(\text{OH})$ , and ferrihydrite,  $5\text{Fe}_2\text{O}_3 \cdot 9\text{H}_2\text{O}$ . They all have colors that range from yellow through orange to brown; that’s why weathered surfaces on bedrock are so commonly of those colors. In a sense, ferrous-iron-bearing rocks *rust* when exposed at the earth’s surface.

If we had started with a real pyroxene, with a combination of  $\text{Fe}^{2+}$  ions,  $\text{Mg}^{2+}$  ions, and  $\text{Ca}^{2+}$  ions, the latter two would have been freed to solution while the first would have found its way into insoluble oxides; that’s because magnesium and calcium are readily soluble in water but ferric iron is almost completely insoluble.

*weathering of aluminosilicates:*

Now to the biggest show of all in chemical weathering. The example here will be what is certainly *the single most important weathering reaction on earth: hydrolysis and carbonation of feldspar*. We’ll use potassium feldspar as the example; weathering of plagioclase feldspar is similar.



Comments:

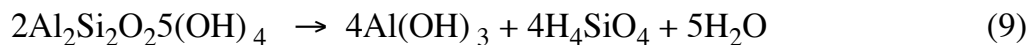
- (1) The first product on the right is the clay mineral *kaolinite*. (There will be more material on clay minerals in a later section.) Other kinds of clay minerals are possible as solid products of weathering of feldspar as well, depending upon the details of the chemical environment.
- (2) The products in solution are silicic acid, as before, and potassium ions.
- (3) If the feldspar had been plagioclase,  $\text{Na}^+$  and  $\text{Ca}^{2+}$  ions would have gone into solution rather than  $\text{K}^+$  ions.

(4) The reaction could have been rewritten slightly to start with  $\text{CO}_2$  rather than with  $\text{H}_2\text{CO}_3$ . The differences are immaterial. Those of you who are well versed in chemistry can try to rewrite the reaction by adding to it the following reaction:  $2\text{CO}_2 + 2\text{H}_2\text{O} \rightleftharpoons 2\text{H}_2\text{CO}_3$ .

In conclusion, if you don't remember anything else about chemical weathering, remember this:

*when feldspar is weathered, by reaction with water and carbon dioxide, the result is a clay mineral, positive ions ( $\text{K}^+$  and/or  $\text{Na}^+$  and/or  $\text{Ca}^{2+}$ ) in solution, and silica, mostly in the form of  $\text{H}_4\text{SiO}_4$ , in solution.*

In intense weathering situations, in warm and humid environments, weathering of aluminosilicate minerals like feldspar can go a step further, to the ultimate weathering product:



The kaolinite, which itself was the weathering product of potassium feldspar, is stripped further of all of its silica, to produce a totally insoluble aluminum hydroxide mineral and more silica in solution. This is the ultimate weathering residue! There are several hydrous aluminum oxide minerals that come out of a reaction like this; the mineral *gibbsite* is the most common. Surficial materials that are especially rich in such insoluble hydrous aluminum oxides are called *bauxite*. As you probably know, such weathering-residue mixtures of hydrous aluminum oxides are the world's supply of aluminum ore. Depending upon the composition of the original rocks, such ultimate weathering residues are commonly mixed with insoluble hydrous iron oxides as well.

**5.2.6** One final point, which will be important for our later consideration of sediments and soils: all of the solid weathering products in the reactions above—clay minerals and various oxides of aluminum and ferric iron—are commonly *of finer particle size than the minerals from which they are produced during weathering*.

**5.2.7** What are the controls on chemical weathering? Viewed on the large scale, there are four major controls on chemical weathering:



*climate:* The intensity of chemical weathering increases with precipitation and temperature. The reason it increases with precipitation is that weathering requires an aqueous medium for the reactions. The greater the percentage of time a given mass of weatherable rock is saturated with surface water, the more chemical weathering it is likely to experience. The reason that the intensity of chemical weathering increases with temperature is that, as with all chemical reactions, the rate of the reaction increases with temperature. That's basically because the vigor of the thermal agitation of the atoms and molecules increases with temperature.

*rock type:* Some kinds of rocks are more susceptible to chemical weathering than others. Rocks that contain minerals that were formed under conditions of the high temperatures and pressures of the deep crust or upper mantle are much more weatherable than rocks that contain abundant minerals like quartz, calcite, or muscovite, which are stable, or at least only slowly weatherable, under most Earth-surface conditions. For that reason, most igneous and metamorphic rocks are generally more susceptible to weathering than most sedimentary rocks.

*relief:* Regions with high relief usually also have steep slopes. Gravity-driven mass movements (more on that in a later chapter) then are effective at stripping away newly weathered material, thereby exposing fresher bedrock to continued chemical weathering. In regions with low relief, however, the layer of regolith thickens to the point where the underlying bedrock is largely sealed off from further contact with the surface environment of chemical weathering.

*vegetation:* As mentioned earlier, when plant materials decay they release carbon dioxide and weak acids into the regolith environment. These substances acidify the surficial material, thereby tending to drive the various chemical weathering reactions discussed above to the right.

**5.2.8** There's another way of looking at the controls on chemical weathering: from the standpoint of *what a small mass of weatherable material actually senses*. This is the small-scale view, in contrast to the foregoing large-scale view. It boils down, largely, to four more specific physicochemical factors:

- mineral composition of the rock
- chemical composition of the water
- rate of passage of water through the rock
- temperature

### 5.3 Physical Weathering

**5.3.1** Most rocks are strong: you need to pound on them with a hammer to break them. The strongest rocks, like tough quartzites, yield only reluctantly to even the most vigorous hammer blows. Other rocks, like soft sandstones, you can break with your own hands. (In fact, there is a continuous gradation between loose and unconsolidated sediment and strongly lithified sedimentary rock, depending in large part upon depth of burial.) Rock strength is a complicated topic, which is important for the construction industry, as well as for geoscientists who try to understand rock fracture that causes large earthquakes.

**5.3.2** Let me a bit more specific about the nature of the strength of solids like rocks. There are three ways of breaking solids: by *compression*, by *tension*, and by *shear* (Figure 2-5). In rocks, the compressive strength and the shear strength are both large, but the tensile strength is much less: generally, it's much easier to break a rock by pulling it apart than by squeezing it. (That's true of concrete, an artificial rock, as well.)

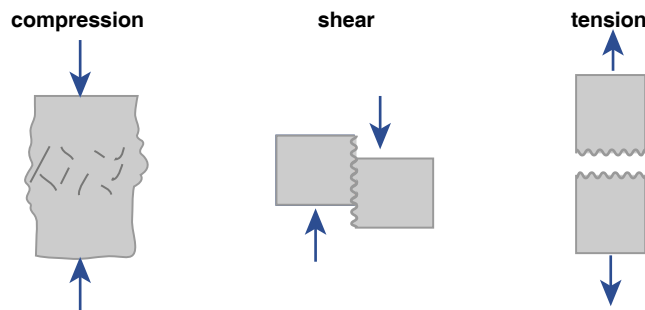


Figure by MIT OCW.

Figure 2-5. Three ways of breaking solids.

**5.3.3** The easiest way to break a rock is to *wedge open a preexisting crack*. That's partly because of the low tensile strength, but also because of an effect known as *stress concentration* at the tip of the propagating crack (for reasons that are beyond the scope of this course). A simple analogy can serve well here: it's much easier to "break" a sheet of paper in tension by starting a cut and then pulling the edges of the cut apart than by grasping the sheet at opposite edges and pulling on it uniformly. When a rock fractures in tension, either rapidly as a result of a blow or an explosion, or slowly, by a variety of natural weathering processes, the crack enlarges as the tip of the crack propagates into the non-fractured rock.

**5.3.4** Breakage of polycrystalline aggregates like rocks (or metals) is more complicated than breakage of homogeneous materials like glass. That's because

*the crystal grains of the rock are joined together at the grain boundaries*, so the strength of the grain contacts needs to be taken into consideration as well as the strength of the mineral grains themselves. When a rock is stressed, extremely fine cracks tend to develop first along the grain boundaries.

**5.3.5** Rocks at the surface tend to be broken down (disaggregated, disintegrated) by a variety of processes, whose importance varies greatly from place to place and from time to time. These physical processes (you might also call them mechanical processes) have traditionally been cited as important in physical weathering, but definitive quantitative studies of the various effects are discouragingly scarce. I'll set them out for you here without making much of an attempt to rank them in importance, except for some concluding comments at the end of this section.

***sheeting:***

Sheeting is *a style of rock fracturing that develops sets of fractures just below the bedrock surface and concordant with it*. The reason such fractures tend to form is that as rock is brought up to the surface it is *unloaded*: the weight of overlying rock is reduced. The rock tries to expand slightly (rock has small but definite compressibility), and *it can expand only upward, not sideways*. This leads to development of fractures that are roughly parallel with the rock surface. Sheetting is most common in otherwise massive and non-fractured bedrock. The sheetting joints decreases rapidly in spacing downward from the surface, usually disappearing within several meters. The next time you see a photo of the mountain slopes in Yosemite, think in terms of sheetting fractures.

***exfoliation:***

Exfoliation (Figure 2-6) is the *separation, during weathering, of successive shells, a few millimeters to some tens of centimeters thick, from massive bedrock*. The mechanical effect is somewhat like that of sheetting: as the thin near-surface layer of bedrock undergoes chemical weathering, the weathering-product minerals tend to be of lower density than the original minerals, which means larger volume, so the thin uppermost zone of the rock tries to expand. It can expand only upward, so fractures parallel to the rock surface are formed. In this situation, chemical weathering and mechanical weathering act together to disintegrate the bedrock. A related process leads to what's called ***spheroidal weathering***: the exfoliation and spalling is slightly more effective at edges and corners than on rock faces, so in an outcrop cut by sets of planar fractures in various orientations (such fractures are called ***joints***; they are ubiquitous in bedrock outcrops) *the edges and corners are rounded off* to produce a very spheroidal or "blobby" look to the outcrop. In

advanced cases the spheroids are completely freed from the outcrop, and lie around on the surface or roll downslope.

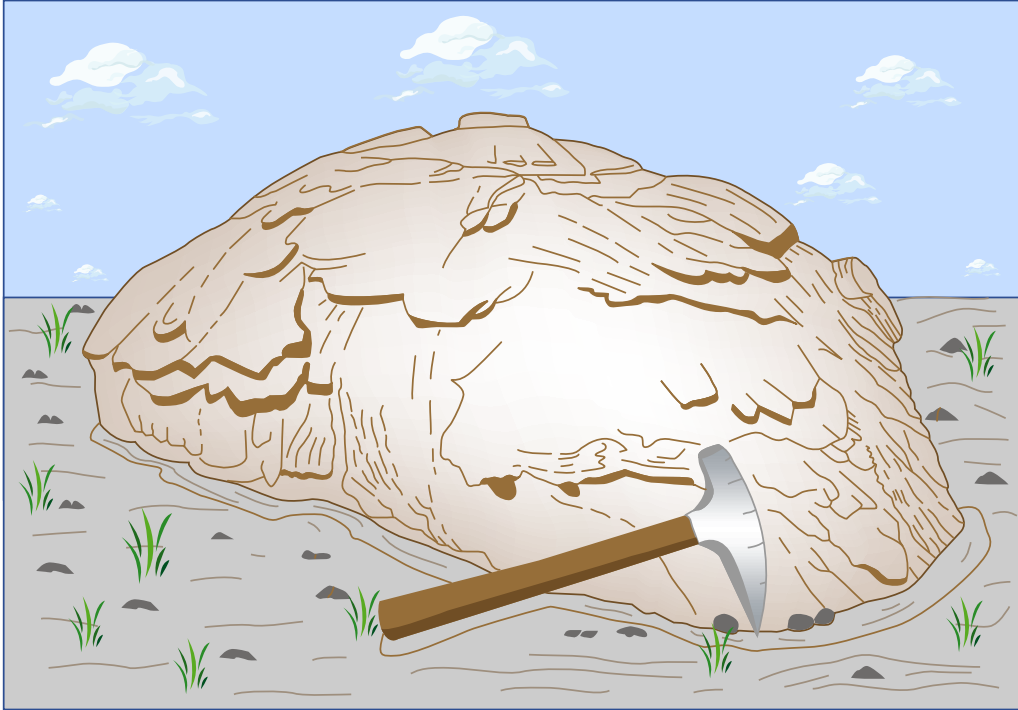


Figure by MIT OCW.

Figure 2-6. Disintegration of a surface boulder by exfoliation. (From Hunt, 1972.)

***granular disintegration:***

Granular disintegration is another process that is in part a mechanical effect of chemical weathering. Many if not most rocks consist of a mixture of minerals, some susceptible to chemical weathering and others not. As the weatherable minerals expand upon incipient chemical weathering, while the non-weatherable minerals retain their original size, intergranular forces are set up, which tend to break the bonds between mineral grains. The mineral grains fall from the rock surface, one by one, to leave an overall smoothly rounded but granular-rough surface. The mass of loose grains at the base of the outcrop is called ***grus***. This kind of weathering is most common in crystalline rocks like granite.

***thermal expansion and contraction:***

Expansion and contraction in response to changes in temperature can by itself aid in propagation of fractures in bedrock ranging in scale from intergranular fractures to widespread joints. Such a process is most effective in arid regions where the day-to-night temperature differences are great. It's unclear whether the process is effective in thoroughly dry rocks, or whether the presence of water is an important element in the process. Experiments on heating and cooling of rock samples, over a great many cycles to simulate many decades or even centuries of the natural environment, have not been successful in producing non-negligible fracturing, but maybe the conditions of the simulation have not been representative. It's well

known, however, that fire causes breakdown of rock by sudden expansion of the surface layer exposed to the extreme heating.

***frost wedging:***

Frost wedging (Figure 2-7) is important in climates where surface water is abundant and the day-to-night temperature range spans the freezing point many times through the year, as in New England in the late fall and especially the early spring. Water expands upon freezing, to the tune of about 9% in volume, an extremely large change. When water confined deeply in a thin crack in bedrock freezes, it can exert enormous side forces on the walls of the crack. The maximum possible force, which has been computed to be over 2000 kilograms per square centimeter (!), is not anywhere close to being achieved, however, because the ice does not seal the crack as strongly as the rock walls themselves, but the process is nonetheless very effective. Mountaintops with expanses of bedrock subjected to freeze–thaw cycles tend to be a jumble of frost-shattered rock; hikers in the Presidential Range in New Hampshire know that well.

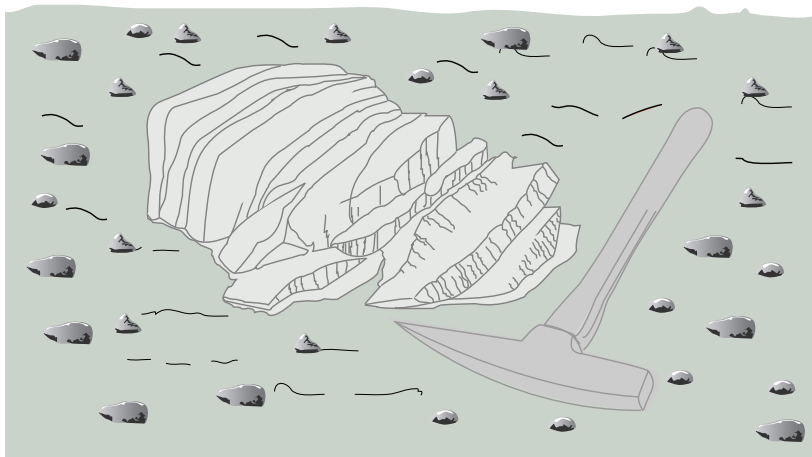


Figure by MIT OCW.

Figure 2-7. Disintegration of a surface boulder, probably by a combination of freeze-thaw, thermal expansion and contraction, and chemical weathering. (From Hunt, 1972.)

***hydrofracturing:***

It's thought by some investigators that the frost-wedging effect is more complicated than just simple expansion upon freezing. Water as thin films in very small cracks may stay liquid even at very low temperatures, for subtle thermodynamic reasons. As the freezing front progresses inward from the rock

surface, water can be forced under high pressures into microscopic cracks, causing propagation of fractures. The phenomenon is known as hydrofracturing.

***“boulder lifting”:***

It’s often said that boulders “grow” in New England fields. Frost action provides a good explanation for the effect (Figure 2-8). As the water-saturated ground freezes from the top down, freezing reaches the base of a large boulder sooner than it reaches the surrounding soil, because the rock of the boulder has a much lower heat capacity than the water-rich soil around it and is therefore cooled faster. Freezing of the water beneath the boulder lifts or pushes the boulder upward a short distance. Upon thawing of the ice beneath the boulder, the boulder does not fall as far as it was raised, because of the tendency for collapse and flowage of soil and finer rock fragments into the space beneath the boulder. Given enough cycles, the boulder makes its appearance at the surface.

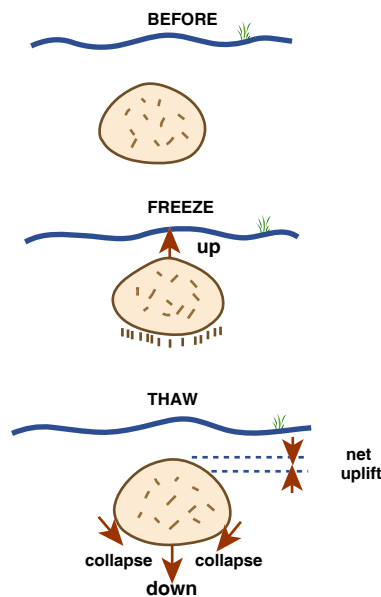


Figure by MIT OCW.

Figure 2-8. Uplift of a boulder by freezing and thawing of surrounding water-saturated regolith.

***root growth:***

Root growth can be very effective at widening preexisting cracks in bedrock near the surface. All of you have seen plants growing heroically in cracks in rock. In

many cases the plant just occupies an already large crack, but in other cases it's clear that the growing root has opened the crack.

**5.3.6** Which of these various processes are the more important, and which are the less important? I can do no more than make an educated guess. (It depends, of course, on the particular setting of bedrock and climate, in any case.) In the right circumstances, *frost wedging* is clearly of great importance, but *granular disintegration*, and perhaps *exfoliation*, are probably of more general importance. Thermal expansion and contraction alone is generally considered to be less important than the other processes, except perhaps in particularly favored environments.

#### ***5.4 Weathering Profiles and Weathering Fronts***

**5.4.1** You've probably had the experience of picking up a broken stone and noticing that the outer part is colored reddish or brownish while the inner part looks a fresh gray. What's going on is that the stone, having been derived from some outcrop as a fresh piece and then perhaps rounded at least somewhat by transport, rested somewhere at the Earth's surface for a time long enough for the outer zone to have experienced some chemical weathering. The boundary between the outer, weathered zone and the inner, non-weathered zone is called the ***weathering front***. There's commonly a very abrupt change in the degree of weathering at the weathering front. In many cases the characteristic difference in color arises from the conversion of ferrous iron in minerals of the fresh rock to ferric iron oxides in the weathered zone.

**5.4.2** The same effect tends to be present on a much larger scale when a bedrock landscape in an area of low relief is subjected to protracted weathering and the weathering products remain where they were formed. To see the effect, you need to have a vertical or steeply dipping cut down through the weathered rock, all the way down into fresh bedrock, which may be many meters or even tens of meters below the surface. Above the weathering front is a gradation from the most weathered material, at and near the land surface, down to the least weathered rock, at the weathering front. The succession of zones of degrees of weathering, from the surface down to the weathering front, is called a ***weathering profile***.

**5.4.3** Weathering profiles vary greatly in their nature, depending upon the factors listed in an earlier section (largely a matter of rock composition and climate). Figure 2-9 shows a common weathering profile, of the kind that tends to develop on deeply in-place-weathered coarse-grained igneous rocks like granite. The material just above the weathering front, called ***saprock***, consists of partly chemically weathered minerals together with yet-unweathered minerals. The

saprock preserves all of the original textures and structures of the parent rock, but its composition has been somewhat changed and its strength has been reduced. The saprock grades upward into material called *saprolite*, which is more altered chemically than the saprock below but still retains much of the textures and structures of the parent rock. You might describe such rock as “rotten rock” (and that is in fact the etymology of the term saprolite). Above the saprolite is even more strongly altered material, which has lost even more of the coherence of the parent and shows little of the original textures and structures. Above this, where the material has lost its coherence entirely and could be described as “loosed”, it is called *regolith* (see later section). The uppermost layer is the *soil* (see later section).

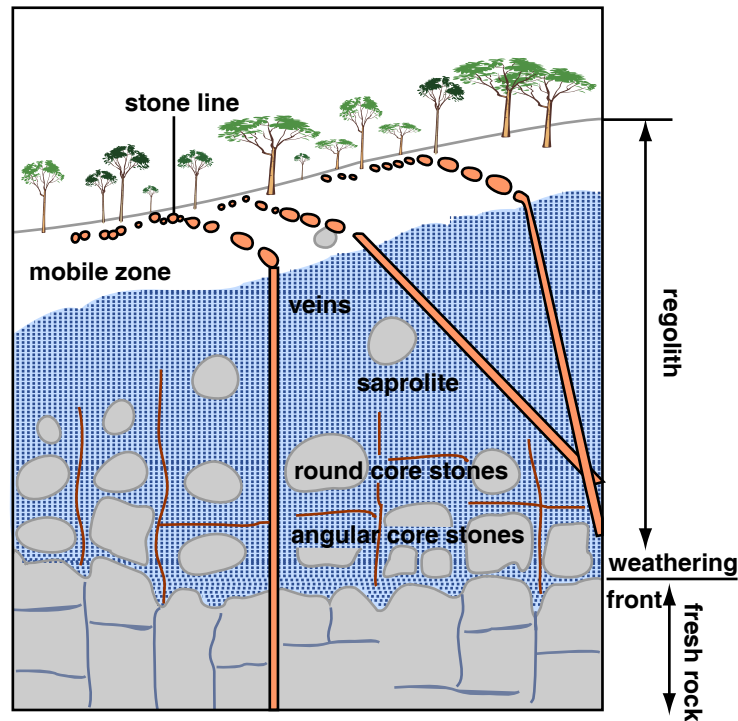


Figure by MIT OCW.

Figure 2-9. A typical weathering profile developed by deep, in-place weathering of coarse-grained granitic igneous rock. (From Taylor and Eggleton, 2001.)

**5.4.4** Note in Figure 2-9 that *the geometry of the weathering front is strongly controlled by the presence of joints in the fresh rock*. The weathering can make better progress along the joints than in the non-jointed intervening rock. In the process, there’s a tendency for non-weathered masses to be left behind as the weathering front progresses downward. These rounded, irregular masses are called *corestones*. They are similar to the weathering spheroids mentioned earlier, but they differ in being formed well below the land surface, whereas weathering spheroids develop at the rock surface as weathering proceeds at the same time that the weathering products are removed from the outcrop.



## 6. THE MINERAL PARTICLES OF THE EARTH'S SURFACE MATERIALS

**6.1** It seems a good idea, at this point, to describe the most common kinds of mineral particles in the ground under your feet—the materials that in the sections to follow are called regolith, sediment, and soil (regolith being the most inclusive term, as you will see). I think it's accurate to say that *most regolith consists mostly of mineral particles*, coarse and fine. A wide variety of kinds of minerals can be found in typical regolith, but only a very few are of any abundance.

**6.2** In terms of relatively coarse mineral particles, at the top of the list is **quartz**. Quartz is common in most of the source rocks from which surface materials are derived, and it's largely immune to chemical weathering. This is a first-order fact: *most of the mineral grains in most of the regolith you're ever going to see (with the notable exceptions of fine muds) consist of quartz*. Quartz grains are mostly equant (meaning approximately equidimensional) in “overall” shape, and usually in between subangular (that is, a bit more rounded than a classically jagged angular particle) and well rounded in “local” shape. The degree of roundness depends partly on the shape of the particle upon being freed from the parent rock by weathering and partly on the extent of mechanical wear the quartz particle has experienced, and perhaps also on some dissolution if the quartz particle is exposed to acidic environments (because the solubility of quartz in water, although extremely low in pure water, increases markedly with increasing acidity of the water).

**6.3 Potassium feldspar** is also important as coarse particles in regolith, depending upon the composition of the parent rock. Potassium feldspar is more susceptible to weathering than is quartz, but commonly some or even much of it escapes weathering before it is deposited as sediment and passes out of the reach of near-surface processes. Survival of potassium feldspar to be seen in regolith is especially common in situations where weathering is curtailed, either by a cold and rigorous climate or by rapid stripping of the weathered material from steep slopes. Commonly, potassium feldspar in regolith is partly or mostly weathered to kaolinite, and under magnification has a “punky” or “chalky” appearance.

**6.4 Rock fragments** are also common in regolith. This is especially the case for sediments. Imagine a raging river eating its way into fractured or partly weathered bedrock: it can entrain fragments of even fresh rock, large and small, and transport them to sites of eventual deposition. Glaciers can do the same thing even more effectively. Identifying large rock fragments is easy; also important, however, are sand-size fragments of various common fine-grained rocks. It takes careful microscopic work to identify such fragments.

**6.5 Muscovite** (the most common form of mica, but not the only one) as coarse particles is common in regolith as well. Micas are sheet-silicate minerals. Sheet silicates are those in which the silica tetrahedra are polymerized by sharing of three of the four oxygens to form a sheet structure, with hexagonal symmetry, and with all of the unshared oxygens pointing in the same direction. Figure 2-10 is a drawing of the sheet structure, looking down on the side with the unshared oxygens. There's more to the mica structure than that, however: the basic sheet unit consists of two such silica sheets, each with their unshared oxygens pointing inward, with aluminum ions in between the two silica sheets. (If you want more detail on mica structure, see the clay mineral "advanced topic" below.)

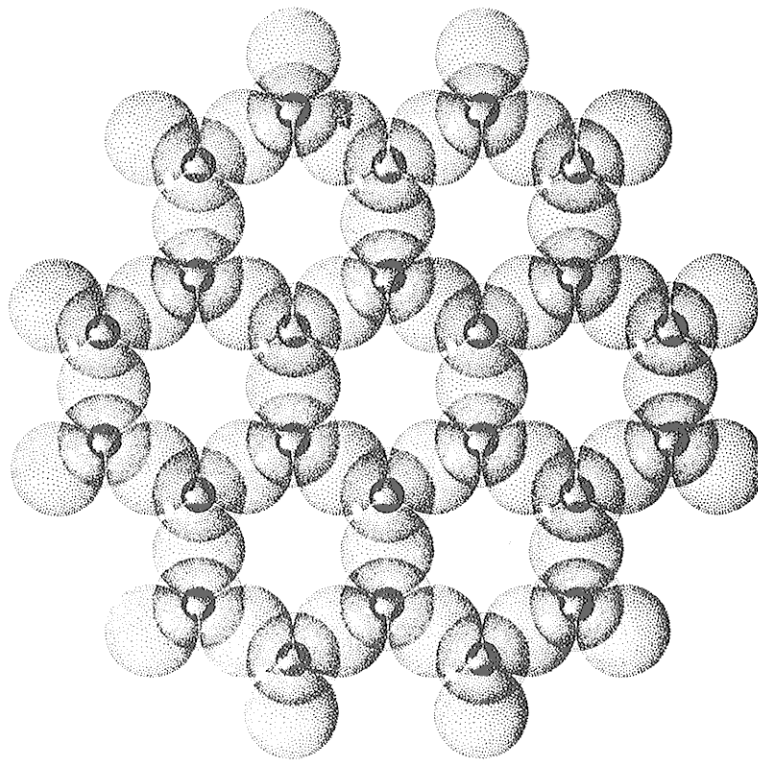


Figure 2-10 Polymerization of silica tetrahedra to form the sheet structure, a basic building block of sheet silicates.

**6.6** The mica crystal is very weak in the direction parallel to the sheets of the structure, so in regolith it's usually in the form of plates or flakes. When it's coarse, so it can be seen with the unaided eye or with a hand lens, it is easy to identify. In sediments it's usually noticeably coarser than the equant particles like quartz, because its platy shape makes it easier for flowing fluid to transport it.

**6.7** In terms of relatively fine mineral particles in regolith, at the top of the list is the group of **clay minerals**. These are a group of minerals rather than a single mineral. Clay minerals are mainly sheet-silicate minerals. Some, called illite or smectite, have the same structure as muscovite, and nearly the same composition. Owing to various substitutions of some ions for others, though, the range of compositions of illite and smectite is wide. In fact, sedimentologists who have to deal with clay minerals often just refer to *illite–smectite* (or I–S for short). Kaolinite and chlorite, with yet different sheet structures, are other common clay minerals. For more detail on the nature of clay-mineral structures, see the following optional “advanced topic”. The proportions of the various clay-mineral types in a given sample of regolith vary depending upon two major factors: *source-rock composition*, and *weathering environment*.

**6.8** An important thing for you to know about clay minerals is that *clay-mineral particles are all very small*, mostly less than a few micrometers. They range in size down to very small fractions of a micrometer, into what is called the colloidal size range. There will be more on colloids in the later section on soils, where they are especially significant.

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## ADVANCED TOPIC: CLAY MINERALS

### 1. Introduction

**1.1** The word “clay” is used, somewhat ambiguously and confusingly, in three different but largely overlapping senses:

(1) As a *mineral* term, for a group of minerals, mostly but not entirely sheet silicates, that are produced from weathering of aluminosilicate minerals (mainly feldspars). They are almost always very fine, from the sub-micrometer range up to several micrometers, and they are mostly platy in shape, although some are curvy and curled. The principal kinds are *kaolinite*, *smectites*, *illites*, and *chlorites*. (I’ve used the plural for the last three because there’s a very wide compositional range in each of them.)

(2) As a *size* term. According to the official grade scale for sediment sizes (see the later section on sediment), all particles finer than 1/256 of a millimeter, or about 4 micrometers, are termed clay.

(3) As a *material* term (“clay material”, or just “clay”), which has a sticky and tenacious consistency. It’s the stuff that potters use.

Most clay-mineral particles are of clay size; most clay-size particles are clay-mineral particles; and most clay material consists of clay-mineral particles of clay size!

**1.2** As you can imagine, *clay minerals are not easy to study*: even in the coarser size range it's difficult to see them microscopically. Standard techniques for study include electron microscopy, which allows you to see the particle shapes beautifully, and x-ray diffraction, which allows you to identify their mineralogy, after a fashion. More sophisticated instrumental techniques have been developed more recently. Clay mineralogy is worth a whole course in itself. What I'll do here is just give you a cursory account of the most important clay minerals and their occurrence.

## 2. Structure

**2.1** Two structural units are involved in clay-mineral structures:

*octahedral sheets*: two planes of closely packed oxygens and hydroxyls ( $\text{OH}^-$ ) with a plane of Al, Fe, and/or Mg between. The Al, Fe, and/or Mg are in octahedral coordination with the O and OH; that is, each of the former is equidistant from six of the latter, which are arranged as an octahedron around the former. Figure 2-11 shows a rough sketch of the arrangement. Figure 2-12 is a more detailed view of the geometry, perpendicular to the sheet. The oxygen-to-oxygen distance is  $2.94 \text{ \AA}$ , and the thickness of the sheet is  $5.05 \text{ \AA}$ . If Al is in the octahedral positions, only two-thirds of the positions are filled; if Mg and/or Fe are in the octahedral positions, all of the positions are filled.

*tetrahedral sheets*, in which silica ( $\text{SiO}_4$ ) tetrahedra are polymerized by sharing of three of the four oxygens to form a sheet structure with hexagonal symmetry. Figure 2-13 shows two diagrammatic representations. The oxygen-to-oxygen distance is  $2.55 \text{ \AA}$ , and the thickness of the sheet is  $4.65 \text{ \AA}$ .

**2.2** These two layers, or sheetlike arrangements, fit together, one on top of the other, in such a way that the vertices of the tetrahedra in the tetrahedral sheet point toward the octahedral sheet. The unshared oxygens of the tetrahedral sheet and the oxygens of the octahedral sheet are the same; they are shared for double duty. The hydroxyls that form part of the octahedral arrangement fit into the hexagonal holes of the tetrahedral sheet.

**2.3** The dimensions of the two sheets are almost but not quite right for this shared arrangement. There has to be some strain in the lattices to make them go

together to share oxygens. This seems to be accommodated by alternate cocking of the tetrahedra, to contract the tetrahedral sheet a little, and stretching and thinning of the octahedral sheet to make it a little bigger. I'll schematize this two-layer structure as shown in Figure 2-14.

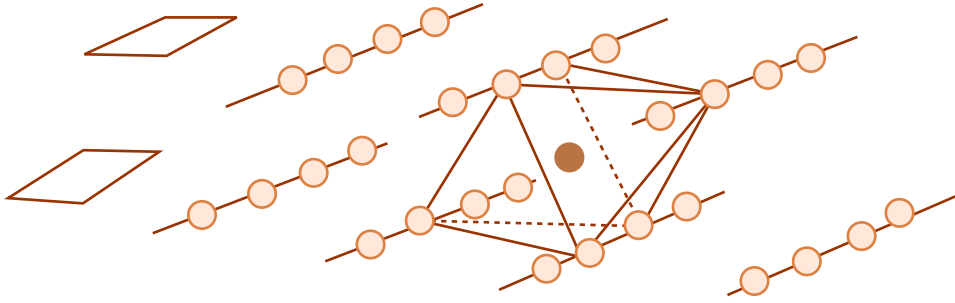


Figure by MIT OCW.

Figure 2-11. Sketch of the arrangement of atoms in the octahedral sheet of clay-mineral structures.

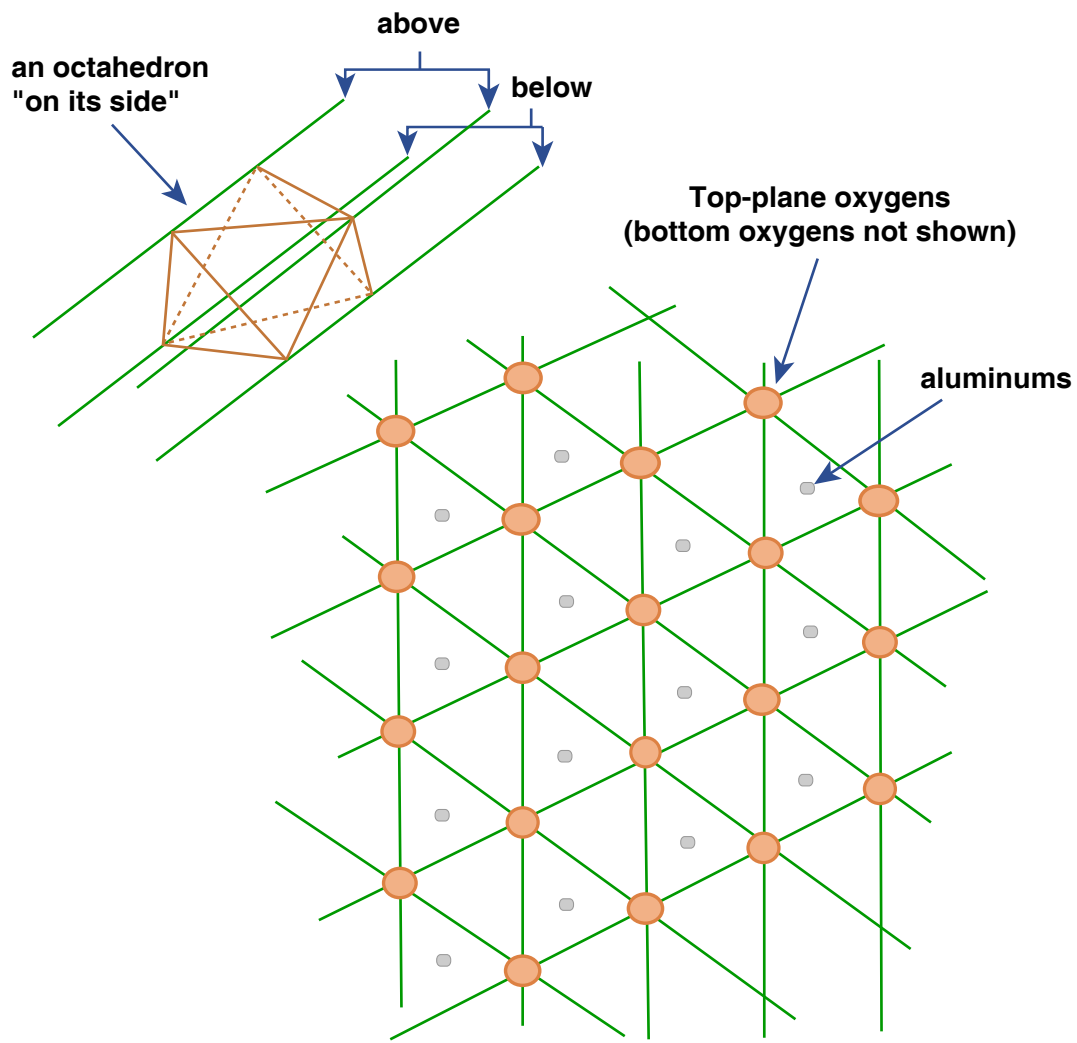


Figure by MIT OCW.

Figure 2-12. A more detailed view of the geometry of the octahedral sheet of clay-mineral structures.

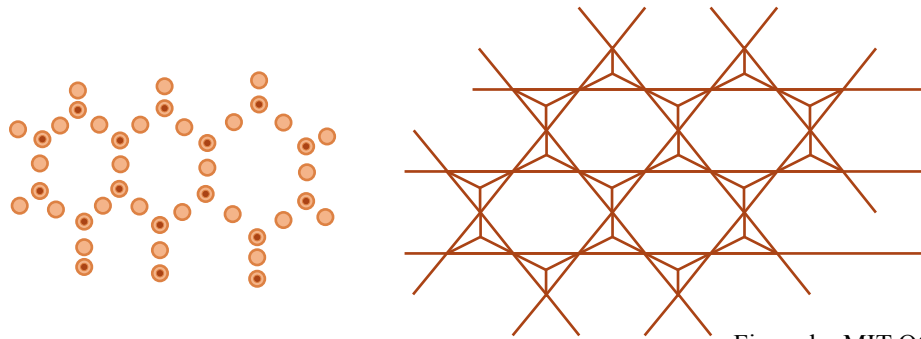


Figure by MIT OCW.

Figure 2-13. Diagrammatic representations of the arrangement of silica tetrahedra in the tetrahedral sheet of clay-mineral structures.

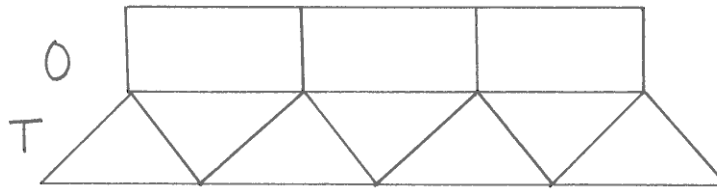


Figure 2-14. Highly diagrammatic sketch of the combination of octahedral and tetrahedral sheets in two-layer clay-mineral structures.

**2.4** It's also possible to construct a three-layer arrangement in which a tetrahedral layer is sandwiched between two tetrahedral layers. Figure 2-15 shows the highly schematic way I'll represent this three-layer structure.

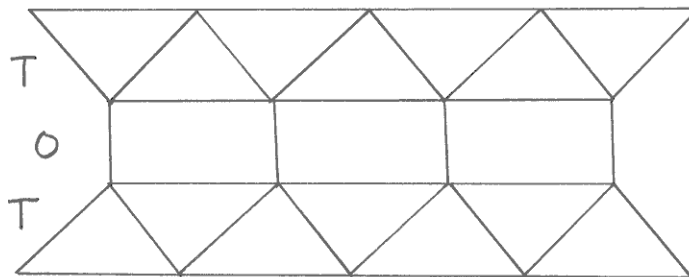


Figure 2-15. Highly diagrammatic sketch of the combination of octahedral and tetrahedral sheets in three-layer clay-mineral structures.

### 3. The Major Clay Minerals

#### ***Kaolinite:***

Kaolinite (Figure 2-16) is the main two-layer mineral. It's fairly simple chemically:  $\text{Al}_4(\text{Si}_4\text{O}_{10})(\text{OH})_8$ , with no isomorphous substitutions in either the tetrahedral sheet or the octahedral sheet; it's the purest of the clay minerals. Only two-thirds of the octahedral positions are filled, by aluminum; minerals like that are called *dioctahedral*. There's no net charge on the tetrahedral–octahedral layers; the layers are held together only by a kind of weak bonds called van der Waals bonds. The repeat distance normal to the layers is  $7.2 \text{ \AA}$ , so this group of minerals is sometimes called *seven-Ångstrom clays*. The crystals can get relatively large, because there's very little stress in the structure; kaolinite is the coarsest of the clay minerals. Kaolinite is formed mainly by weathering of feldspars; production of kaolinite is favored by acidic and warm environments.

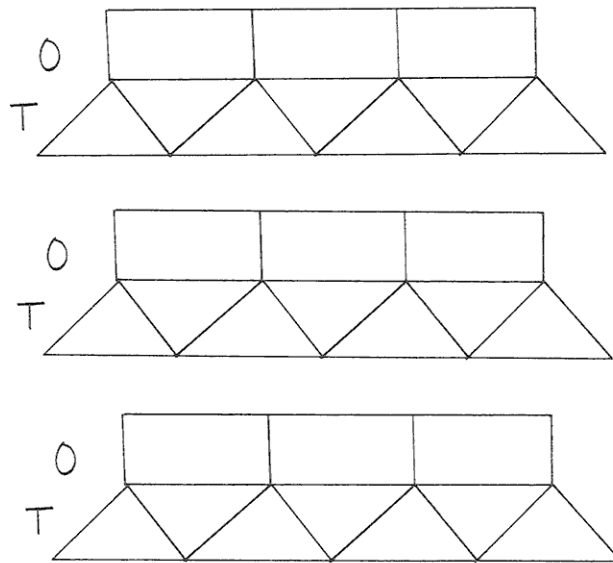


Figure 2-16. Highly simplified sketch of the structural organization of kaolinite.

#### ***Illite:***

Illite (Figure 2-17) is a hydrous muscovite-like clay mineral. It has the same three-layer structure as muscovite, but less than the one-in-four replacement of Si by Al that's characteristic of muscovite (the figure is between one-in-five and one-in-six). So there isn't the same charge deficiency as in muscovite, and not as many  $\text{K}^+$  ions in the interlayer positions.  $\text{H}_3\text{O}^+$  ions are also present in the interlayer positions. Also, substitutions in the octahedral layer are greater and



more random; there's lots of variability. Illite is never found well crystallized, only as clay-size particles. The particle size is typically less than one micrometer. This is because there's considerable stress in the layers. Illite is closely similar in structure to the common muscovite you see in sedimentary, metamorphic, and igneous rocks; it's just much finer grained. Illite is formed mainly by weathering of potassium feldspar in temperate weathering environments. With increasing intensity of weathering, kaolinite is formed instead.

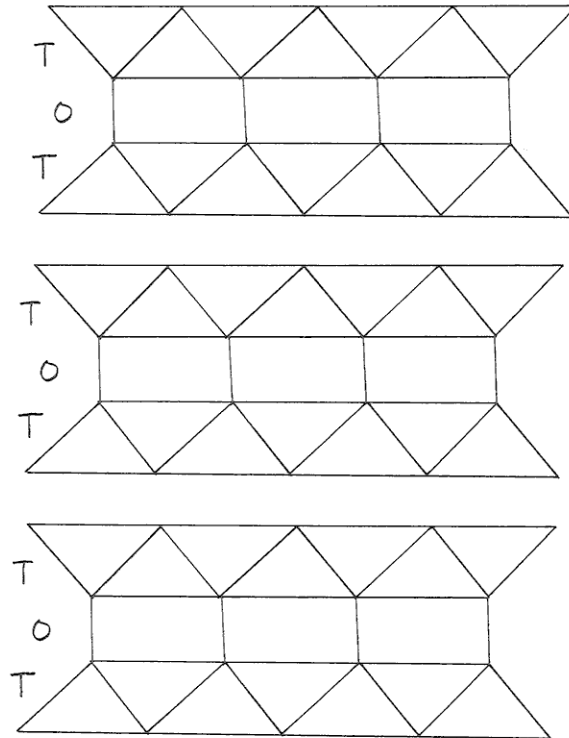


Figure 2-17. Highly simplified sketch of the structural organization of illite and smectite.

### *Smectite*

Smectite (Figure 2-17) is a clay mineral which, like illite, has the basic muscovite arrangement but which is able to expand by taking up a lot of water (or other polar molecules) in the interlayer positions. The reason is that the bonds between the TOT units are weaker, because of details of stacking of the layers, and also because the charge imbalance is located mainly in the interior octahedral layer rather than in the outer tetrahedral layers. When the mineral is collapsed the layer spacing is 9.6 Å, but when fully expanded it's almost 20 Å. Smectite has variable

percentages of large cations,  $\text{Ca}^{2+}$  and  $\text{Na}^+$ , in the interlayer positions. There's substitution of Al for Si in the tetrahedral layer, and variable proportions of  $\text{Fe}^{2+}$ ,  $\text{Mg}^{2+}$ , and  $\text{Al}^{3+}$  in the octahedral layer. Smectite is formed by the weathering of plagioclase, ferromagnesian aluminosilicates, and volcanic glass; its development is favored by alkaline environments. One important kind of smectite mineral is called *montmorillonite*, a name you are likely to encounter if you have to deal with clay minerals.

### ***Chlorite:***

Chlorite (Figure 2-18) is a kind of mixed-layer clay mineral. Its structure is an alternation of trioctahedral TOT layers, with  $\text{Mg}^{2+}$  and  $\text{Fe}^{2+}$  in the octahedral positions, and layers with the composition  $(\text{Mg}^{2+}, \text{Fe}^{2+})_3(\text{OH})_6$ , giving a repeat spacing of about 14 Å. There's a great range and diversity of ionic substitutions in each of the three ion sites, as well as differences in the geometry of stacking of the sheets. Chlorite is produced in abundance only in relatively mild weathering environments, where it survives from the source rocks, although in the form of finer particles. So it's characteristically a high-latitude product.

**6.9** Various oxides of iron and aluminum are especially important in the highly weathered soils of warm and humid regions, where (as you saw in the section on weathering) silicate minerals are broken down further to oxides. The main aluminum oxide is **gibbsite**. The formula for gibbsite,  $\text{Al}(\text{OH})_3$ , shows that it is a hydroxide mineral. Gibbsite is formed by further weathering of silicate clay minerals, in which all of the silica is stripped out of the clay minerals; see Reaction 9 in the section on chemical weathering, above. The main iron oxide mineral in soils is **goethite**,  $\text{FeOOH}$ . It's goethite that imparts the characteristic yellow-orange color to weathered iron-bearing rocks and regolith. Recall from the section of chemical weathering that ferrous iron is freed from ferrous-iron-bearing silicate minerals and then oxidized to minerals like goethite. These aluminum and iron oxides (there are others, as well) in regolith are in the form of extremely small particles, in the colloidal size range. (See the later section on soils for more on colloids.)

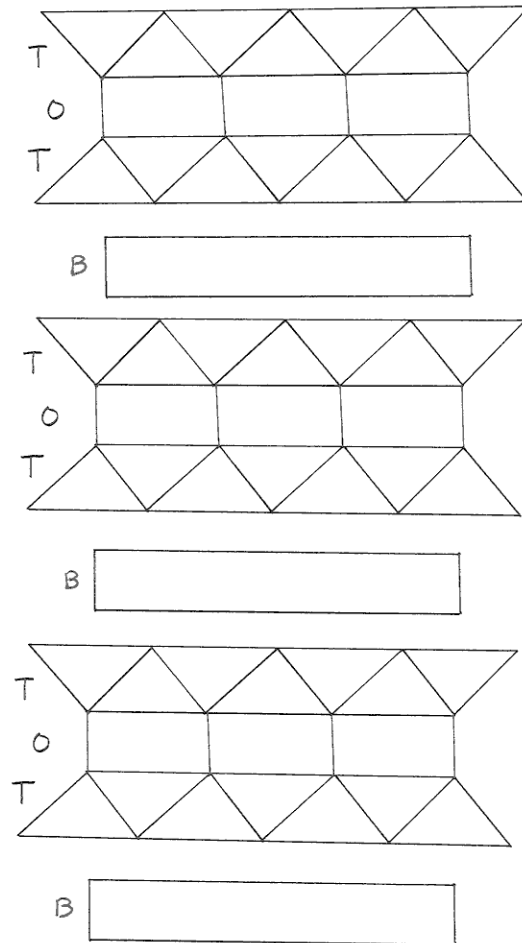


Figure 2-18. Highly simplified sketch of the structural organization of chlorite.

## 7. REGOLITH

### 7.1 *The Nature of Regolith*

**7.1.1** What is regolith? The term *regolith* is used for *the layer or mantle of fragmental and unconsolidated rock and mineral material, whether residual or transported, that rests on bedrock*. One might quibble that the definition excludes fragmental material that happens to be locally cemented by surficial processes but

is otherwise closely related to nearby unconsolidated material. (You'll be hearing more, much later in the course, about one such kind of material: caliche.) The definition is not specific about age; most regolith is young, geologically and even by human standards, but some is much older, not having been subjected to consolidation, for one reason or another, for geologically long times.

**7.1.2** For most of us geologists, continental bedrock is our bread and butter: we inspect it minutely, hammer on it, sample it, photograph it, and think about it day and night. But in most parts of the world, finding bedrock to inspect is not easy, and in many places almost impossible: it is covered by a layer of regolith. When was the last time you walked across an outcrop of bedrock? When we are out and about in the natural world, away from the ubiquitous pavement that's so characteristic of our civilization, we are mostly walking on regolith. If only for that reason alone, study of the Earth's regolith should be high on the list of desiderata for Earth scientists. A good case can be made that *the Earth's mantle of regolith is for humankind by far the most important component of the solid Earth beneath our feet*: it feeds us, in the form of soil. In fact, however, there is a tendency for regolith studies to be low on the priority list. (I think that there lurks in the minds of many geologists the notion that bedrock studies hold far more interest, and glory, than regolith studies.)

**7.1.3** Regolith originates by the complex of processes collectively termed weathering, which was the subject of the preceding section. What happens to regolith once it is produced? You can detect from the foregoing definition that regolith is either *residual* or *transported*. The terms are largely self-explanatory, are they not? In areas of low relief and with a climate conducive to deep weathering, a thick mantle of regolith, in some places many tens of meters thick, lies where it has been produced. In most places, however, the regolith we see has been moved from its place of origin, nearby or far away, by various agents of transportation.

**7.1.4** Keep in mind, however, that it's usually not a matter of an episode of production and then a later episode of transportation: generally, both production and removal are going on at the same time. The thickness of the mantle of regolith depends in great part on the relative magnitude of the rate of production and the rate of removal. In some situations, as in regions with steep slopes, high relief, and rigorous climate, regolith is stripped away as fast as it is produced; in areas with gentle slopes and climates conducive to deep weathering, the layer of regolith is very thick.

## ***7.2 Agents That Mobilize Regolith***

**7.2.1** Regolith resting on a sloping surface is pulled inexorably downward by the force of gravity (or, more accurately, by *the downslope component of the*

*force of gravity*) (Figure 2-19). Such movement of the regolith may be imperceptibly slow or at speeds in excess of a hundred meters per second. An entire later chapter will be devoted to the subject of downslope movements of near-surface material by gravity, collectively termed *mass movements* (absolutely no relation to sociopolitical mass movements).

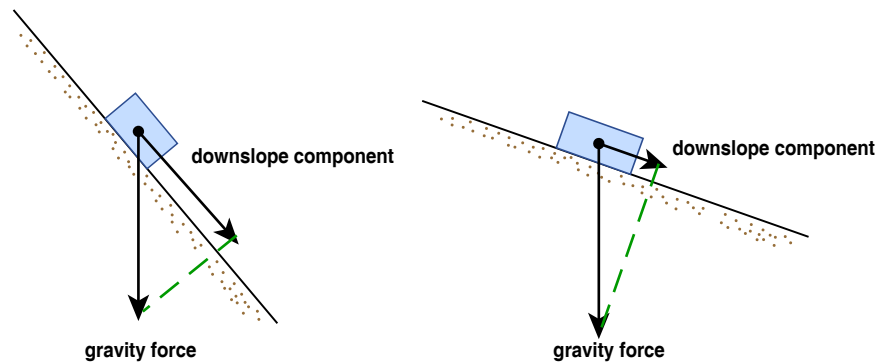


Figure by MIT OCW.

Figure 2-19. The downslope component of gravity acting on the layer of regolith. A) On a steep slope. B) On a less steep slope.

**7.2.2** The principal agent for the transport of regolith, aside from the direct pull of gravity, is *the flow of water in streams and rivers*. Of course, river flow is itself a consequence of the downslope pull of gravity—but regolith transport by flowing water involves physical processes and effects that are fundamentally different from the direct pull of gravity. The mobilization and transportation of regolith by flowing water is the subject of part of the later chapter on streams and rivers. Movement of regolith by the action of land-based glaciers is of great importance in certain Earth-surface environments as well. Wind can also be a significant transporting agent, as described in the final chapter.

### 7.3 Kinds of Regolith

**7.3.1** The main distinction in kinds of regolith is between *residual regolith* (also called *sedentary regolith*), on the one hand, and *transported regolith*, on the other hand (Table 5-1). With respect to transported regolith, several kinds are recognized, depending upon the agent of transportation. Transported regolith is collectively termed *sediment*. There will be much more material, later in this chapter as well as in later chapters, on sediment, and how it is eroded, transported, and deposited. Colluvial deposits (called *colluvium*) are those transported down steep slopes by the pull of gravity. Mainly, these are talus deposits (also called

scree deposits) on steep slopes, and avalanche and landslide deposits of various kinds. Alluvial deposits (called *alluvium*) are those transported and deposited by rivers and streams. In addition, there are glacial deposits and eolian (wind-blown) deposits.

regolith	residual	soils of various kinds; residual subsoil materials
	transported	colluvium: talus, scree, etc.
		alluvium
		glacial sediment
		eolian sediment

Table 5-1. Classification of regolith. (After Merrill, 1897.)

## 8. SEDIMENT

**8.1** When regolith is mobilized or entrained, by flowing water, or by the wind, or by moving glacier ice, it is called *sediment*, because, as the term implies (the Latin verb *sedere* means to sit), it eventually comes to rest again as a deposit. Here I will touch upon only a few aspects of sediments. You will hear much more about sediment movement, and sediment deposits, in later chapters, especially the chapters on rivers, on glaciers, and on coasts.

**8.2** One of the characteristic things about sediment is that *it tends to become fractionated, by size, during transport*. The basic reason is that a given transporting agent, like water or the wind, moving at a given speed, can move sediment particles up to a certain size but no larger—although another important effect is that the water or the wind moves finer particles faster than coarser particles. The end result of such fractionation is that the sorting of the particles in the deposit is usually much better than in the original material that was the source of the transported material. (The *sorting* of a sedimentary deposit describes the spread of particle sizes around the mean particle size: in a well sorted sediment the particle sizes cluster closely around the mean, whereas in a poorly sorted sediment the particle sizes range widely around the mean.)

**8.3** For convenience and clarity in communication, sedimentary geologists have adopted a series of terms for sediment particles over the entire range of particle sizes (Figure 2-21). Such a subdivision is called a *grade scale*. It's based on powers of two, so it's geometric (or, what comes out mathematically to be the same, logarithmic), rather than arithmetic, for reasons that are obvious but not easy to justify: for example, the difference in behavior between a 1 mm particle and a 2 mm particle is bound to be far greater than the difference in behavior between a 101 mm particle and a 102 mm particle, right? You can see from Figure 2-20 that in geoscience such things as mud, silt, sand, and gravel are not just qualitative concepts but precisely defined entities!

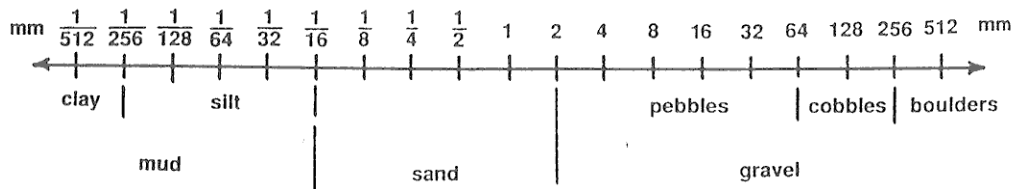


Figure 2-20. The Udden–Wentworth grade scale for sediment particle size.

**8.4** The logic behind the grade scale in Figure 2-20 is fine for well-sorted materials, but how about poorly sorted materials? Various classifications have been proposed, but the one that's in most common use (with some variations) among regolith specialists is shown in Figure 2-21. Here you see what “loam” (so commonly pronounced “loom” locally!) really is, in the technical sense.

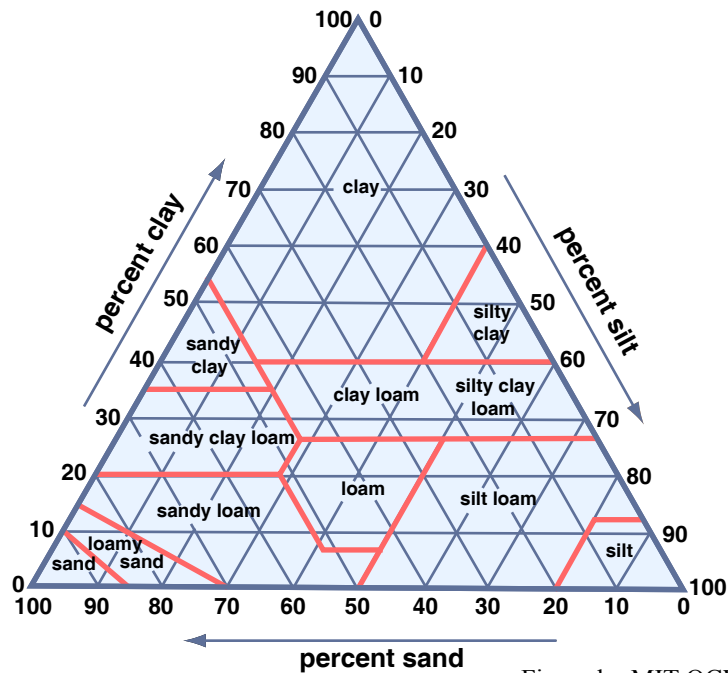


Figure by MIT OCW.

Figure 2-21. Conventional system for classification and terminology of sediment mixtures. (From FitzPatrick, 1986)

### ADVANCED TOPIC: MEASURING SEDIMENT SIZE

1. Scientists or engineers who are responsible for studying surface deposits usually have to measure or characterize, in some way, the distribution of particle sizes. For large particles, in the gravel size range, it's possible to measure the size of each piece separately, with calipers, although that's a challenging endeavor unless the particles is very well rounded, and even then the matter is not entirely straightforward (as it would be if the particles were perfect triaxial ellipsoids). For smaller particle sizes, in the silt to fine pebble range, the classic way of measuring particle size is by *sieving*.

2. Here is how particle size is measured by sieving:

- make up a stack of special sieves, from coarsest on the top to finest on the bottom, with a lid on the top and a pan on the bottom;
- put a small sample of deposit in the top sieve;



- shake the stack until the particles have found their size bin;
- weigh each of the fractions on a balance;
- compute, for each sieve size, the total weight percent finer than that size;
- plot the results on graph paper with size along the horizontal axis, logarithmically, and, along the vertical axis, weight percent finer than, from 0% to 100%, arithmetically.

An example of the result, called a *cumulative distribution curve* (common in many other applications of statistics as well), is shown in Figure 2-22.

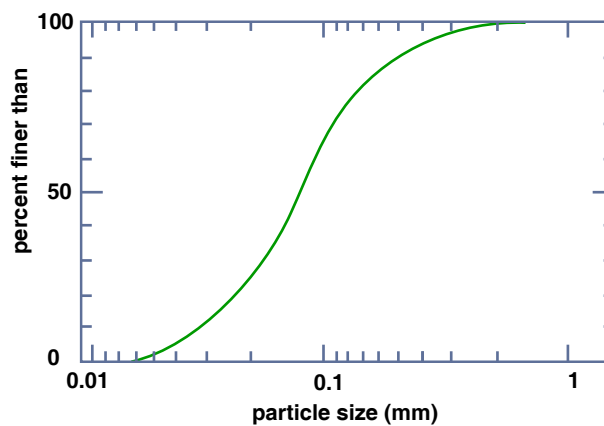


Figure by MIT OCW.

Figure 2-22. A typical cumulative distribution curve of sediment size.

**3.** The problem with sieving is that it works well only for sizes down to about 0.05 mm. For sediment finer than that, there are various techniques that involve suspension of the sample in water, settling of that suspension, and sampling during settling. The basic principle is that the coarser sizes settle faster than the finer sizes. Devices that are used to measure fine particle sizes in this way are called *settling tubes*. The governing principle is that each particle has a terminal fall speed that is a function of its size. In recent years, a variety of more sophisticated instruments have been developed to measure particle size, either optically or sonically, by sensing each particle as it passes through a thin tube in suspension. Such instruments are now obtainable “off the shelf”. There is still the tricky matter of intercalibrating between results from such instruments and traditional settling tubes.

## 9. SOILS

### 9.1 Introduction

**9.1.1** Along with oxygen and water, soils are essential to human existence. Almost all of our food comes, directly or indirectly, from crops grown in the soil. The only significant exception, of course, is food from the oceans. Hydroponically grown crops are an insignificant exception—and the infrastructure and cost involved in hydroponic agriculture are likely to doom it forever to a minor role in human food supply. All of our lumber, most of our paper products, and much of our clothing ultimately come from the soil. Our dependence on soils will continue, but the world stock of soils cannot increase. In fact, it is *decreasing*, because of soil erosion, as well as urbanization (and suburbanization). As our reserves of fossil fuels continue to shrink, energy resources from biomass (living plant material) are likely to increase.

**9.1.2** The meaning of the term “soil” varies greatly depending upon the field of endeavor in which it is used. *Agriculturalists* use the term narrowly to mean the uppermost layer of surface deposits, in which plants grow and agriculture is practiced. The *soil scientist* has a somewhat wider view: soil is the uppermost layer of surface deposits, almost always extending well below the agriculturalist’s soil, which has developed upon underlying material by the influence of various soil-forming processes acting over long times. The *geologist* has an even broader view of soils as surface materials developed by weathering processes that have broken down bedrock into regolith, and the *civil engineer* (who practices the discipline of soil mechanics, and who is responsible for the siting and design of structures with foundations) takes a similarly wide view of soil as all materials at the surface that can be excavated without blasting. In the following sections, we’ll look at the nature of soils mainly from the standpoint of the soil scientist (what is the nature of soils, how do they develop, and how are they classified?).

**9.1.3** The study of soils as natural bodies is called *pedology* (and its practitioners call themselves soil scientists, or pedologists). You can also think in terms of the *pedosphere*. (There is a tendency among science educators nowadays to think in terms of “spheres”: geosphere, hydrosphere, biosphere, atmosphere, cryosphere. These terms are useful, in a broad and loose sense, to describe zones or domains of related materials and processes. I think that the suffix “-sphere” began to be used because in a very general sense these domains are like spherical shells around the Earth. I myself think that “sphere-ology” tends to be overdone on the part of the “spherophiles”, but it remains a useful pedagogical device.)

**9.1.4** Soils can be thought about and studied on a wide range of scales: from large regions of continents, in the context of major soil types, to the distribution of soils in a local area of varied topography and substrate composition (Figure 2-23A), to the vertical profile of soil at a given point on the land surface (Figure 2-23B), to the texture and structure of the soil at a point in such a profile

(Figure 2-23C), to the details of physics, chemistry, and biology of the soil on the colloidal and atomic scale (Figure 2-23D).

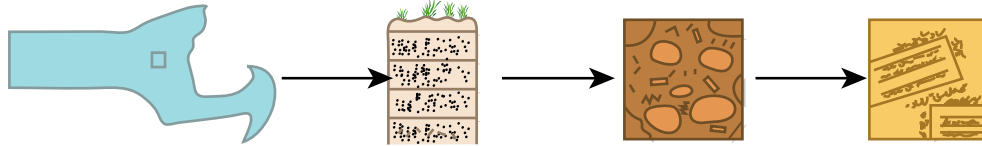


Figure by MIT OCW.

Figure 2-23: Thinking about soils on a wide variety of scales. (Inspired by Brady and Weil, 2002.)

**9.1.5** Soils have a great variety of constituents. Here is a list of the most important kinds of soil constituents:

- mineral particles
- colloids
- organic matter
- microorganisms
- soil solutions
- soil gases

**9.1.6** Most soils consist mainly of mineral particles. Such soils are often called *mineral soils*. Some soils, however, consist mostly of organic matter. In general, soils contain the same kinds of mineral particles as do sediments; go back and review the section on sediments above. Most of the soils you are likely to encounter consist mainly of quartz, plus variable percentages of potassium feldspar and rock fragments, in the coarse fraction, and mainly particles of clay minerals, iron oxides, and aluminum oxides, in the fine fraction. The remaining constituents in the list above are described in the following sections.

## 9.2 Colloids

**9.2.1** *Colloids* are mixtures of materials in which one substance exists as extremely small particles, called colloidal particles, which are dispersed in a second substance, which you can think of as the medium. In the case of colloidal

particles in a uniform liquid or gaseous medium, the term colloidal solution is commonly used. In the context of soils, we're dealing with small particles consisting of minerals or organic matter that are dispersed in the aqueous pore solutions in the soil or are aggregated together in great numbers, or cling to the larger soil particles. Colloids are important in a great many materials, both natural and manufactured.

**9.2.2** Most clay-mineral particles have a platy shape, and they have a tendency to stick together in aggregates of many particles, held together by certain short-range electrostatic forces. Such forces are small, but so are the particles—and it's important to realize that because *the ratio of surface area to volume increases with decreasing particle size*, clay-mineral particles are in a sense “almost all surface and very little volume”. The state of aggregation of the clay-mineral fraction is widely variable, and the macroscopic strength and consistency properties of the deposit vary greatly corresponding to that. Also, as you will soon see, the small size of the particles, together with the electrical charges characteristic of their surfaces, make the tiny particles chemically extremely reactive.

**9.2.3** Colloidal particles are in a kind of “no person's land” between true solutions and undoubtedly solid particles. At one extreme are ions and molecules in an aqueous solution. The smallest ions are of the order of atomic dimensions, about 0.1 nanometers (that is, about  $10^{-10}$  meters). Many molecules that consist of a large number of atoms are much larger, but they still behave like smaller ions in true solutions. At the other extreme are particles larger than a few micrometers, which is large enough to be seen with light microscopes. Such particles are large enough to settle in the surrounding fluid under the influence of gravity.

**9.2.4** Typical soils contain several important kinds of colloids. Here's a briefly annotated list of the most important:

***clay-mineral particles:***

There are several types; see the earlier background section on clay mineralogy. These types differ in their physical and chemical properties, partly because of their differing size but mainly because of their differing surface charges and their differing susceptibility to take on or release positive ions and water molecules to and from their internal surfaces between the TOT layers.

***iron and aluminum oxides:***

These are especially important in the highly weathered soils of warm and humid regions, where silicate minerals are broken down further to oxides. The main

aluminum oxide is *gibbsite*, and the main iron oxides are *goethite* and *ferrihydrate* (no details of mineralogy here).

***humus:***

These are especially abundant in the upper parts of soils, nearest the surface. Humus colloids are not crystalline structures: they are complex molecular chains and rings of carbon atoms bonded to hydrogen, oxygen, and nitrogen atoms. They are among the smallest of colloidal particles in soils. They have an enormous capacity to adsorb water onto their surfaces.

**9.2.5** The great importance of soil colloids arises from their chemical reactivity. There are two reasons for that: *surface area*, and *surface charge*. The following elaborates on those points.

***surface area:***

Chemical reactions between a solid and the fluid with which it is in contact take place at the surface of the solid, at the interface between the solid phase and the liquid phase. The rate of such a chemical reaction depends upon a number of important factors, aside from the nature of the reaction itself. Temperature is one of those important factors: in general, the higher the temperature, the faster the reaction. That's because the energy of thermal vibration of the atoms and molecules in both the solid phase and the liquid phase increases with temperature. Also important, however, is *the area of the surface at which the reaction takes place, per unit volume of the reacting system*. The extremely small size of colloidal particles means that their surface area, per unit bulk volume of material, is spectacularly large.

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**BACKGROUND: SURFACE AREA**

1. Take a nostalgia trip back to secondary-school math and think about the formulas for the volume and the surface area of some regular geometric solid. A sphere is a good example. The volume of a sphere is  $(1/6)\pi D^3$ , where  $D$  is the diameter of the sphere. In scientific parlance, *the volume goes as the cube of the diameter*. The surface area of the sphere is  $\pi D^2$ , so the surface area goes as the *square* of the diameter.

2. What's important here is not the particular formulas, or the values for volume and area of a particular sphere they give, but *how the ratio of surface area to volume varies with the diameter of the sphere*. You can easily see from the two formulas given above that the ratio of surface area to volume of any sphere is  $[\pi D^2]/[(1/6)\pi D^3]$ , or, simplifying,  $6/D$ . Ignore the factor 6 there; what's important is the  $1/D$  part. The ratio of surface area to volume goes as  $1/D$ . That means that *as the diameter decreases, the ratio of surface area to volume gets bigger*.

3. Now for a revealing example of this  $1/D$  effect. Think about a sphere 10 cm in diameter—about the size of a softball, a grapefruit, or a ball of pizza dough. It has a surface area of about 0.0314 square meters (or, in scientific notation,  $3.14 \times 10^{-2} \text{ m}^2$ ). Now compare that with a sphere 1 mm in diameter. That sphere has a surface area of only  $3.14 \times 10^{-6} \text{ m}^2$ , far smaller, by a factor of ten thousand. But suppose you took the entire volume of the 10 cm sphere and converted it into an equivalent volume of 1 mm spheres. With some further simple math, and using the formula for the volume of a sphere, you would determine that the 10 cm sphere is equivalent to  $10^6$  (that's one million) of the little ones. Now the total surface area of the little ones is  $(10^6)(3.14 \times 10^{-6}) \text{ m}^2$ , or  $3.14 \times 10^0 \text{ m}^2$ . That doesn't sound like a lot, but compare it with the surface area of the original 10 cm sphere, which was  $3.14 \times 10^{-2} \text{ m}^2$ . We see, then, that the aggregate area of the little spheres is a hundred times the area of the volume-equivalent big sphere! This is a concrete example of the  $1/D$  effect: we reduced the diameter of the sphere by a hundred (ten centimeters to one millimeter), and the surface area of the equivalent volume increased by a factor of a hundred!

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### ***surface charge:***

At any given tiny region inside a mineral particle, the electric charges associated with the ions of the structure balance out. At the surface of the particle, however, the picture is different—because the ions right at the surface are bonded only to the ions below them. That means that there is a kind of “unsatisfied” electric charge at the surface of the particle. With regard to clay-mineral particles, the surface charge is generally negative on the broad faces of the exposed TOT sheets and positive along the narrow edges of the plates. For macroscopic mineral particles, like sand grains, these surface charges are almost entirely negligible, but when the particles are extremely small their enormous area-to-volume ratio makes these surface charges very important.

**9.2.6** The combination of surface area and surface charge is a potent one. We need to deal with two different but related effects: coagulation, and ion

exchange. Figure 2-24 is a cartoon that shows the various phenomena discussed in the following paragraphs.

***coagulation:***

Clay-mineral particles, with their negative charges on faces and positive charges on edges, tend to clump together by electrostatic attraction, in an edge-to-face arrangement. (Keep in mind that like charges repel and unlike charges attract.) This tendency for coagulation is one of the important factors that make clay-rich soils sticky and tenacious.

***ion exchange:***

As you will see in a later chapter, on water, the water molecule has a positively charged “side” and a negatively charged “side”. Because of that, water molecules tend to be attracted to the charged surfaces of the colloidal particles. Picture a swarm of water molecules, perhaps up to several molecules thick, at the clay-particle surface, jiggling around and jostling one another, now and then trading places with other water molecules from the solution beyond. That layer of water molecules acts as a kind of “cushion” for the particle. Included in the layer of loosely bound water molecules are positive ions, which are attracted to the negatively charged faces of the clay particles. Also, smectite clays take up and give off water and positive ions to and from their internal surfaces. These loosely bound water molecules are the last to be extracted by plant roots as a soil dries between rains. You know that potassium is one of the most important plant nutrients. Its continued abundance in the soil is largely the result of its attraction to colloids, together with recycling in organic matter. Because of this combination of attraction to colloids and recycling by plants, under natural conditions the storehouse of available potassium, released by chemical weathering of primary minerals, is maintained at a workable level for plant growth in the face of continuing loss in solution to downward-percolating water from the surface.

## ***9.3 Organic Matter***

### ***9.3.1 Introduction***

**9.3.1.1** The uppermost parts of surface deposits—the soil of the soil scientist and especially of the agriculturalist—usually has an abundance of organic matter as well as inorganic mineral matter. It’s not easy to specify what’s meant by organic matter. It comprises a wide and complex variety of chemical compounds that a chemist would call organic compounds. Organic compounds are those that consist of carbon atoms, bonded to one another and to atoms of other elements, mainly oxygen and hydrogen but including such elements as nitrogen and sulfur as

well, in a great variety of ways. Organic compounds are both natural and laboratory-synthesized.

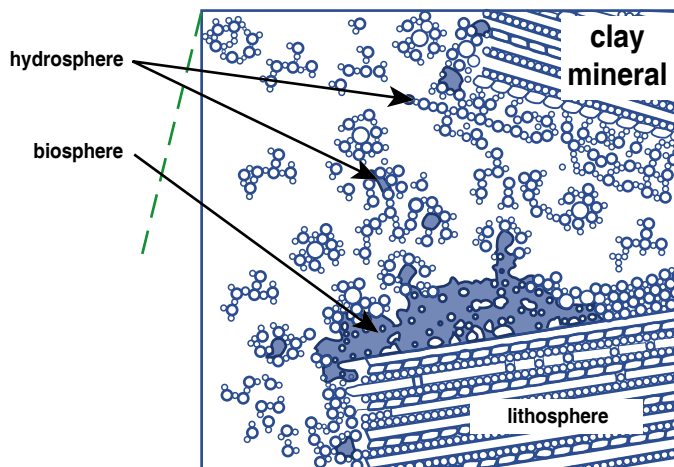


Figure by MIT OCW.

Figure 2-24. Submicroscopic view of a clay-mineral colloid particle in its aqueous environment.

**9.3.1.2** The organic matter of soils is *ultimately derived from plant growth*. (Living plants and animals are not conventionally considered to be organic matter: you have to wait until the organism dies!) Nonliving plant tissues themselves are organic matter, as are the great variety of organic compounds that result from partial decay of that plant material. The tissues of dead animals and their decay products also represent soil organic matter, although not nearly in the same abundance as plant-derived organic matter.

**9.3.1.3** A deep understanding of soil organic matter would necessitate a much greater familiarity with organic chemistry (the branch of chemistry that deals with organic compounds) than is appropriate for this course. Here we can only touch upon some of the most important considerations and kinds of materials.

### ***9.3.2 The Role of Soil Organic Matter in the Carbon Cycle***

**9.3.2.1** The element carbon is a key substance in the workings of the Earth's near-surface environment. That comes about because of its important role in living organisms, soil organic matter, the atmosphere and the oceans in the form of gaseous and dissolved carbon dioxide, and carbonate sediments. There will be a more detailed treatment on the Earth's carbon cycle in a much later chapter on



geochemical cycles. It must suffice here to point out that the soil plays a major role in that carbon cycle. It's been estimated that the Earth's soil contains almost twice as much carbon as the stock of living vegetation and the atmosphere combined.

**9.3.2.2** Plants take up carbon dioxide gas from the atmosphere and convert it, by sunlight-powered photosynthesis, into the organic compounds that constitute plant material. When the plants die, some of the dead plant material is decomposed back to carbon dioxide and water by oxidation. In part that decomposition involves non-biological chemical reactions, but in great part it is mediated by the metabolic activity of soil microorganisms. The end result, however, is the same: production of CO<sub>2</sub>. Some of that CO<sub>2</sub> remains dissolved in soil water and some is released into the atmosphere.

### ***9.3.3 Kinds of Soil Organic Matter***

**9.3.3.1** There are many kinds of organic matter. These compounds range from easily decomposable to very difficultly decomposable. A typical plant consists of easily transformed compounds like polysaccharides and lipids, stiffened with the much more resistant materials cellulose and lignin, the principal components of woody plant tissue. Lignins, of which there are various kinds, are the slowest plant tissues to decompose. Only a few kinds of organisms, mainly fungi, can break down lignins.

**9.3.3.2** The ultimate residue left behind by breakdown of plant materials is called *humus*. What's called humus comprises a number of very stable organic compounds. Humus is always dark in color, the reason being that the chemical structures in the humus absorb nearly all wavelengths of visible light. Humus is so fine grained that its particles (which in reality are large molecules) are at the fine end of the colloidal size range. These compounds, whose basic building blocks are carbon hexagonal ring structures (called aromatic compounds, in organic chemistry) with large molecular weights (that is, containing very large numbers of polymerized atoms in their structures). They are the organic materials that are the most resistant to microbial decomposition. The water-holding capacity of humus, mentioned in an earlier section, is several times greater than that of clay-mineral particles.

**9.3.3.3** Decomposition of humus is a slow process. The typical half-life of humus (that is, the time it takes to decompose half of an initial mass of the material) is measured in many decades, or even a few centuries! An important effect that slows the decomposition is complexing of the molecules of the humus with inorganic clay-mineral particles. By poorly understood processes, this association protects the humus from decay.

### **9.3.4 The Role of Organic Matter in Soils**

**9.3.4.1** With respect to plant growth, humus is an essential soil component. It's common knowledge among gardeners that plants grow better in soils rich in humus. There are various reasons for that, both direct and indirect. It's not just a matter of supply nutrients: humus is not particularly rich in nutrients (despite what many organic gardeners have been led to believe). Most of the nutrients needed by plants are inorganic ions dissolved in aqueous soil solutions, although humus does play some role in the nutrient supply for plants.

**9.3.4.2** It's the indirect effects of humus on soil properties that are the most beneficial aspects of humus. Humus can be viewed as a soil conditioner. These effects involve physical, chemical and biological properties of the soil. In terms of *physical effects*, humus reduces the stickiness and cohesion of clay-rich soils, and it promotes the formation of granular texture, so prized by agriculturalists because it enhances the tillability of the soil. In terms of *chemical effects*, humus is even more effective at adsorbing and holding onto nutrient ions than are clay-mineral particles. In terms of *biological effects*, humus is the most important substrate for the life activities of soil microorganisms; it provides most of the food for detritus-feeding soil organisms.

**9.3.4.3** If you are willing to be overwhelmed by the multitude of ways that organic matter affects soil properties, plant productivity, and environmental quality, look at Figure 2-25.

## **9.4 Soil Organisms**

**9.4.1** The last really important component of the deposit comprises soil organisms: the plants and animals that spend part or all of their lives in the soil. The soil is one of the Earth's most important ecosystems. (For a brief brush-up on ecology and ecosystems, see the background section below.) Every handful of soil contains literally billions of organisms—almost all of them microorganisms, too small to be seen with the unaided eye. The microbiology of soils is a fully developed and very important discipline in itself, and I can only touch upon it here.

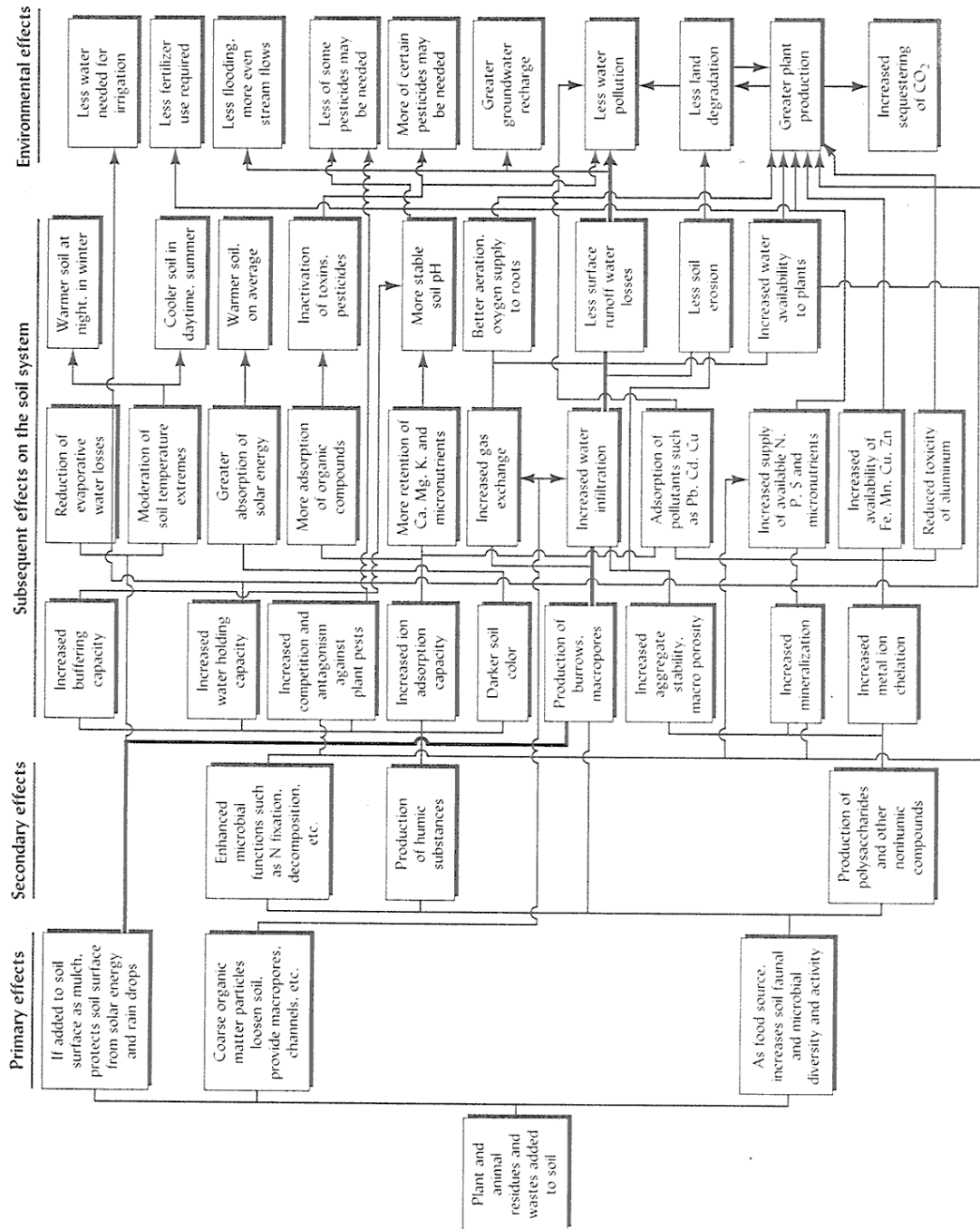


FIGURE 12.15 Some of the ways in which soil organic matter influences soil properties, plant productivity, and environmental quality. Many of the effects are indirect, the arrows indicating the cause-and-effect relationships. It can readily be seen that the influences of soil organic matter are far out of proportion to the relatively small amounts present in most soils. Many of these influences are discussed in this and other chapters in this book. The thicker line shows the sequence of effects referred to in the text in Section 12.7. (Diagram courtesy of R. Weil)

Figure 2-25. How organic matter affects soil properties, plant productivity, and environmental quality. (From Brady and Weil, 2002.)

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## BACKGROUND: ECOLOGY AND ECOSYSTEMS

1. The term **ecology**, which tends to be used too loosely in the popular media these days, is the study of *the interrelationships between organisms and their environment*. The largest entity studied in ecology is the **ecosystem**, which consists of *some chosen part of the physical and chemical environment and all of the organisms within it*. An ecosystem involves all of the physical, chemical, and biological processes and interactions that operate within the given part of the environment. An ecosystem can be as large as the entire biosphere or as small as a puddle of water at our feet.

2. What are the elements of an ecosystem? A **habitat** is *the local environment in which a given organism lives*. Ecosystems usually involve a number of different and distinctive habitats. A similar but different concept is that of an **ecological niche**, which is *the position of the organism in its habitat*, including its way of life and its role in the ecosystem. Most habitats are occupied by several species, each with its own ecological niche. Each species in an ecosystem is represented by a number (usually, but not always, large) of individual organisms, called a **population**. A population of two or more species that occupy a given habitat is called a **community**. There may be more than one community in a given ecosystem.

3. Within any ecosystem, there typically are many kinds of interactions among constituents, both living and nonliving. These interactions can be viewed in terms of flows of matter and energy through the ecosystem. Organic compounds are synthesized from the environment by **producers**, which in all but the most specialized ecosystems are photosynthesizing plants. The producers are in turn consumed by plant-eating animals, called **herbivores**. Some of the herbivores are in turn consumed by **carnivores** (or by **omnivores**, which eat both plants and animals). Other elements of the ecosystem are **parasites**, which feed on living organisms without killing them, and **scavengers**, which feed on dead organisms.

4. The sequence of species ranging from the producers, at one end, to carnivores that no other carnivores eat, at the other end, is called a **food chain**. Often the term **food web** is more appropriate, because the actual situation is more complicated than a simple linear arrangement of species. The tissues of producers, herbivores, and carnivores that are not consumed by species higher in the food chain are broken down by organisms called **decomposers**, which are usually bacteria.

5. When we think of ecosystems, we tend to think of tropical rainforests, or broad expanses of grasslands on which large mammals roam around and get photographed by ecotourists. In a real and important sense, however, *a shovelful of natural soil is an ecosystem*, and a very complex one at that.

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**9.4.2** One way of classifying soil organisms is into plants and animals. There's more to that than it might seem, however, because although we all learned, when we were small, that life is divided into an animal kingdom and a plant kingdom, most biologists these days recognize *five* kingdoms. But that's another story we need not deal with in this course. There are two more useful ways, for us, of classifying soil organisms:

- *heterotrophs vs. autotrophs*: **autotrophs** are organisms that generate their own tissues from inorganic raw material, by use of (mainly) solar energy. Plants are autotrophs. **Heterotrophs** are organisms that use organic matter produced by other organisms as food.
- soil organisms can be classified as **macroorganisms** (organisms with size ranging down to that of, say, earthworms), **mesoorganisms** (organisms ranging down to a small fraction of a millimeter), and **microorganisms** (organisms of microscopic size, smaller than about a tenth of a millimeter).

**9.4.3** The numbers of kinds of organisms per unit volume of a soil increases strongly with decreasing organism size. A typical soil might contain several species of vertebrates and several species of earthworms, along with as many as several tens of species of insects, dozens of species of mesoorganisms, but thousands of species of microorganisms. Moreover, the numbers of organisms per unit volume increases spectacularly with decreasing size: although, obviously, numbers vary widely depending upon the type and condition of the soil, numbers of macroorganisms and mesoorganisms are large but not overwhelming, whereas the numbers of microorganisms, per cubic meter, say, might be in the trillions!

**9.4.4** A whole chapter should be devoted here to the nature and significance of the life activities of soil organisms. Interesting as they are, the activities of what have been called **ecosystem engineers** (vertebrate burrowers, earthworms, ants, termites, and the like) in modifying the physical nature of the soil pale in comparison with the life activities of soil microorganisms. Microorganisms play the essential role of decomposing soil organic matter (mainly plant residues) and converting it to carbon dioxide, water, dissolved ions, and the ultimately resistant residues called humus earlier in this section.

**9.4.5** Several major kinds of microorganisms are well represented in soils. The most important are *algae*, *bacteria*, and *fungi*, and various kinds of *protozoans*. Algae, which are photosynthetic plants, are abundant only in the uppermost parts of deposits, very near the sunlit surface. Fungi and protozoans

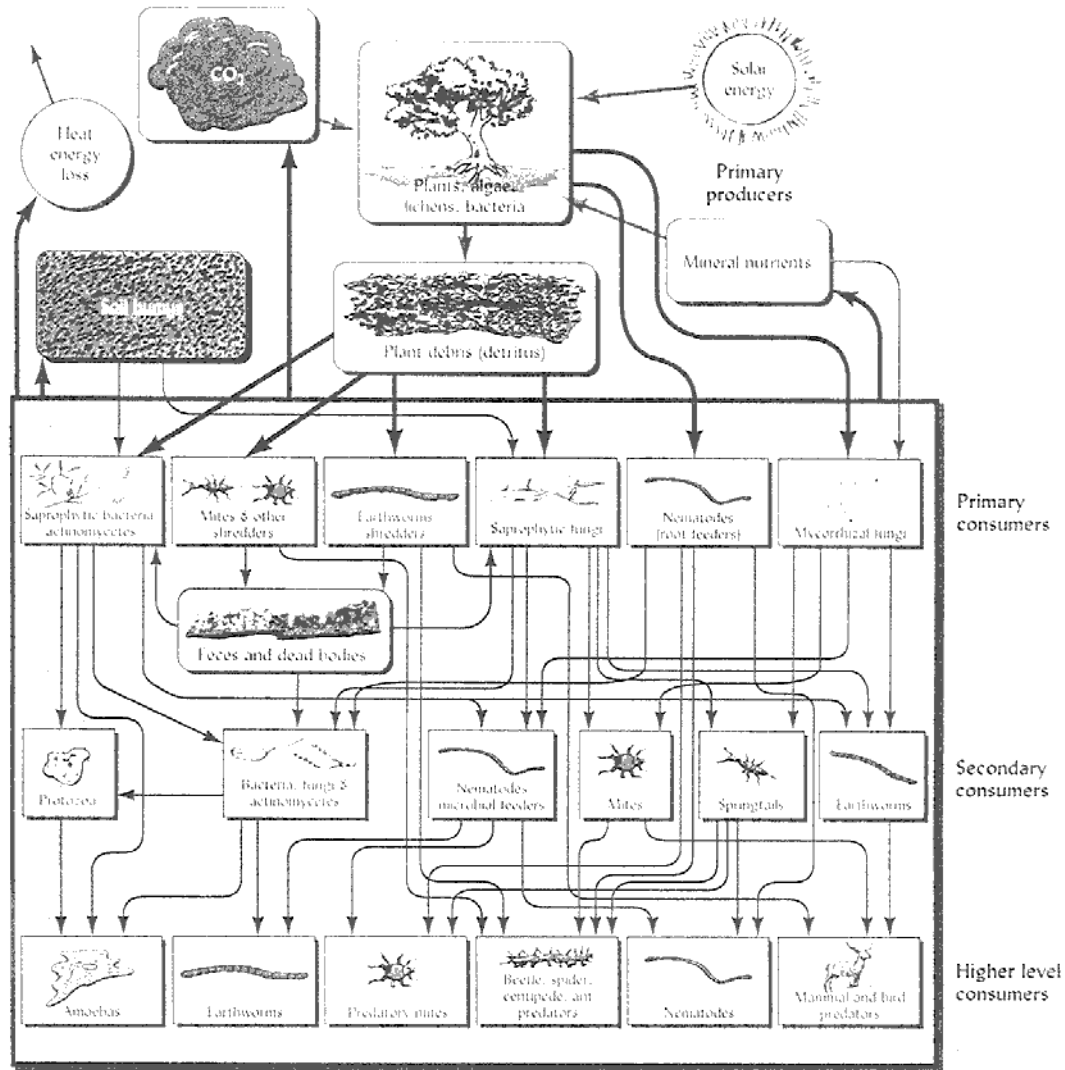


FIGURE 11.1 Greatly simplified and generalized diagram of the soil food web involved in the breakdown of plant tissue, the formation of humus, and the cycling of carbon and nutrients. The large shaded compartment represents the community of soil organisms. The rectangular boxes represent various groups of organisms; the arrows represent the transfer of carbon from one group to the next as predator eats prey. The thick arrows entering the top of the shaded compartment represent primary consumption of carbon originating from the tissue of the producers—higher plants, algae, and cyanobacteria. The rounded boxes represent other inputs that support, and outputs that result from, the soil food web. Although all groups shown play important roles in the process, some 80 to 90% of the total metabolic activity in the food web can be ascribed to the fungi and the bacteria (including actinomycetes). As a result of this metabolism, soil humus is synthesized, and carbon dioxide, heat energy, and mineral nutrients are released into the soil environment.

Figure 2-26. The soil food web. (From Brady and Weil, 2002)

are important but restricted to the shallow zones. Bacteria, on the other hand, range widely downward to great depths.

**9.4.6** Figure 2-26 shows a diagram of the soil food web. I've preserved the original caption for you. Note that it's described as "greatly simplified and generalized"!

## ***9.5 Soil Solutions and Soil Gases***

**9.5.1** All soils are porous to some extent (that is, they contain empty spaces, called *pore spaces* or *void spaces*, among the various solid constituents). It stands to reason that the pore spaces must contain either a gas or a liquid, or some of both. Whether the pore spaces are occupied by solely by liquid or whether they are occupied mainly by a gas depends upon where the region of soil lies relative to the groundwater table: below the groundwater table the soil is fully saturated with an aqueous pore solution, whereas above the groundwater table the pore spaces are occupied mainly by gas.

**9.5.2** The reason I have used the term "aqueous pore solution" rather than just "pore water" is that the water in the pore spaces contains a great variety of dissolved substances, in varying concentrations. Just after a rainstorm, as the rainwater percolates downward to replenish the water table, the concentrations of dissolved materials is very low (except for dissolved atmospheric gases), but as time goes on some of the constituents of the soil particles are leached out to go into solution. The chemistry of such soil solutions is an important part of soil science.

**9.5.3** Above the groundwater table, the pores spaces are filled mainly with gas. The composition of soil gases is not greatly different from that of the atmosphere, with one major exception: owing to continuing decay of organic matter, the concentration of carbon dioxide is typically substantially higher than in the overlying atmosphere.

**9.5.4** Even in the soil zone above the groundwater table, the soil is seldom entirely dry: some residual water clings tenaciously to the soil particles. I will defer discussion of that until the chapter on groundwater.

## ***9.6 Processes of Soil Development***

**9.6.1** The development of a soil (in the agriculturalist's and soil scientist's sense) depends upon five major (and interacting) factors, as shown in Figure 2-27. Of these, you might think that the parent material is the most important. It's generally believed, however, that climate is even more important than parent

material. The relative importance of climate and parent material depends to a great extent upon time of development: in the initial stages of soil development, parent material is certainly the more important, but, as time goes on, soils developed upon different parent materials in a given climate tend to become more and more similar. Obviously, important differences as a function of parent material must remain: if for one soil the parent material is rich in quartz, and for another soil the parent material is a limestone or a mafic igneous rock like basalt, the soils can never converge.

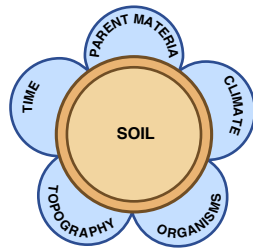


Figure by MIT OCW.

Figure 2-27. The five factors of soil development. (From FitzPatrick, 1986.)

**9.6.2** Topography is generally of secondary importance; its role lies largely in governing the downslope creep of already developed soil, which tends to “freshen” the uppermost layer of the surface deposit by exposing less weathered material beneath.

**9.6.3** The importance of organisms lies in their ability to add the all-important organic matter to the soil and to process that organic matter back into inorganic compounds, including nutrients for yet more plant growth.

**9.6.4** I might also point out that *the five factors are not entirely independent of one another*; for example, organisms depend strongly on climate, and topography depends quite strongly upon climate as well, and also upon parent material and time.

**9.6.5** This is a good place to point out that *some soils develop directly upon bedrock, but in many if not most cases, soil develops upon surface deposits, called sediments earlier in this chapter, that were derived from bedrock somewhere else and transported to the site of soil formation*. In some places, the surface deposit is grossly “out of place”. That’s true of almost all glaciated areas, New England being no exception. It’s also true of areas of deposition from streams and rivers. In such areas, soil formation is usually nowhere near the “ultimate” or “equilibrium” soil, of the kind you would find in a low-latitude, non-glaciated area of low relief, where bedrock is weathered very deeply over a very long period of geologic time and the soil is highly developed.



**9.6.6** (But just because a soil has had a long time to develop doesn't mean that it will be rich and deep: in the rain forests of the humid tropics, intense leaching removes nutrients released by chemical weathering so rapidly that the soil is thin, and has only a thin layer rich in organic matter at the surface; the lush stand of plants is in tenuous existence, recycling nutrients very efficiently. When such vegetative cover is removed, the soil rapidly becomes irreversibly barren.)

**9.6.7** A variety of physical and chemical processes act to develop soil from fresh regolith. One of the most important of these we have already dealt with: weathering of preexisting bedrock and/or further weathering of relatively fresh and unweathered surface deposits. Another important process is vertical transport of matter in either dissolved or colloidal form by upward or downward flow of soil water. Such processes come under the soil scientists' term *translocation*. Translocation also includes vertical mixing or churning by macroorganisms, most importantly earthworms and certain kinds of insects. Biological processes, most importantly *humification* of plant material, and chemical and organism-aided processes of *cycling* of nutrients and other compounds (most importantly nitrogen and carbon) are further soil-forming processes.

**9.6.8** The nature of *water flow in the vertical direction in the soil*, which is largely a function of climate (intensity and frequency of rainfall; the effect of temperature, humidity, and vegetative cover upon evaporation from the soil), determines whether matter in solution undergoes net downward transport or net upward transport in the soil. This is the aspect of climate that's most important in soil development, along with the nature and rate of growth of vegetation.

## **9.7 Soil Profiles**

**9.7.1** Soil scientists classify soil on the basis of the sequence and nature of vertical zones or layers, called *horizons*, in the soil. Each of the enormous variety of soil types has a *characteristic succession of such horizons*, each with its thickness, composition, physical properties, and transitions to overlying and underlying horizons. The particular succession of horizons is called a *soil profile*.

**9.7.2** To have a good picture of a soil profile, you need to go into your backyard and dig a trench, carefully, about a meter deep and wide enough for a good side view, with one side planar and almost vertical and shaved off carefully for good viewing with minimal smearing of soil material on the viewing face. Excavations for foundations, or even big holes for planting of trees or shrubs, can serve the purpose well. For the best and most representative results, you need to be sure to dig at a spot that hasn't been filled or excavated in recent times, by humans, either before or after the time of European settlement, and hasn't been cultivated, either. In New England, woodlands that served as pasture land rather than as tilled fields are the best places to see undisturbed soil profiles.

**9.7.3** It's dangerous for me to describe one or a few soil profiles for you, because there is such a multiplicity of important soil types (see the next section), but for concreteness, and a good start in thinking about soil horizons and classification, here's *a description of a typical New England soil* (Figure 2-28). The two conditions such a soil experiences that are the most important factors in its development are (1) *strong net downward movement of water through the profile*, and (2) *acidic waters*. In the traditional classification of soils, such a soil would be called a *podzol* soil.

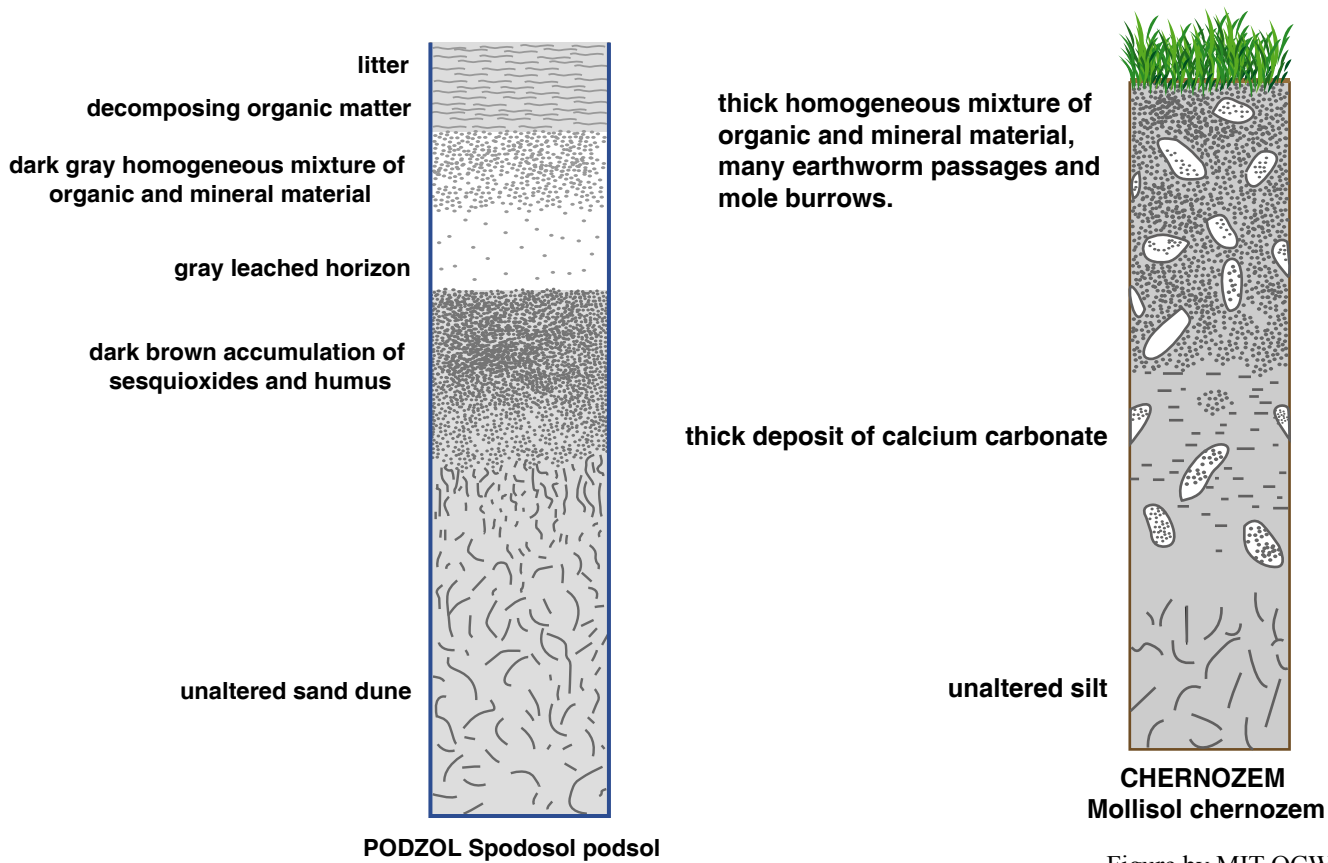


Figure by MIT OCW.

Figure by MIT OCW.

Figure 2-28 [left]. A vertical profile through a typical New England soil. (From FitzPatrick, 1986.)

Figure 2-29 [right] . A vertical profile through a typical North American mid-continent soil. (From FitzPatrick, 1986.)

**9.7.4** Right at the surface is a layer of forest litter: leaves and twigs that have fallen recently, within the last few years. That grades down into a layer of organic matter in a progressively more advanced state of humification with depth. Together these uppermost layers, called the *H horizon* if usually water-saturated or the *O horizon* if not usually water-saturated, are the part of the soil that's by far the richest in organic matter and living organisms—and most conducive to cultivation, in which case the original organic-rich layer is tilled and mixed downward to form the best medium for plant growth. Its thickness commonly ranges from just a few centimeters upward to ten to twenty centimeters. It passes rather abruptly into a layer consisting mainly of mineral grains, usually mostly quartz (if the soil is well developed and the parent material had a fairly high percentage of quartz), with a characteristic light gray color. Much, or most, or in extreme cases even all, of the other silicate minerals have been leached out of this layer by decomposition by net downward passage, from the surface, of waters rendered acidic by (1) dissolved CO<sub>2</sub> from the atmosphere and (2) organic acids produced in the uppermost organic-rich layer. This horizon is an example of what's called the *A horizon*. The A horizon passes downward, again fairly abruptly, into a layer, called the *B horizon*, with a characteristic color, usually combinations of yellow or orange or red or brown. Here is where iron leached out of the material of the A horizon is reprecipitated as hydrous iron oxides within the pore spaces among the existing mineral grains, and fine clay-mineral particles produced by weathering reactions in the A horizon have been deposited. (The general term for mobilization and removal of such things as iron and aluminum from a soil horizon is *eluviation*; their deposition in a given horizon is called *illuviation*.) The B horizon passes gradually into underlying material that's relatively little affected by soil processes; it's called the *C horizon* (but not universally considered to really be a soil horizon!).

**9.7.5** The precise nature of the physicochemical processes that mobilize, transport, and deposit iron and, especially, aluminum are not yet entirely well understood. Iron, when weathered from parent minerals, is probably complexed with organic compounds and then precipitated as insoluble oxides lower down in the soil profile where the carrier organic compounds are degraded. The aluminum presents more of a problem. The aluminum is not actually put into true solution, but neither can it make its way down the profile as large clay-mineral particles. It probably travels either as a colloid (i.e., particles so fine as to be in the range between true solution and fine mineral particles) or physically as very fine, submicron-size clay-mineral particles.

**9.7.6** Where the average water table lies far below, the B horizon tends to be diffuse; where the average water table is not far below, the B horizon tends to be more concentrated, and its colors stronger. (In areas where the average water table lies close to the surface, the B horizon tends instead to be gray from deposition of compounds containing iron in the reduced, ferrous state, as well as clay minerals.)

**9.7.7** Now for something entirely different: a capsule description of a typical soil from the mid-continent of North America (Figure 2-29) Such soils are common in a wide area including western Kansas, western Nebraska, eastern Colorado, most of the Dakotas, and on into south-central Canada. In the traditional terminology, such soil would be called *chernozem soils* (from the Russian for “black soils”), *chestnut soils*, or *brown soils*. In such areas, with a semiarid (rainfall 15–25 inches) and strongly seasonal climate and dominantly grassland vegetation (now largely converted to agricultural use, especially in the eastern parts of the area), the balance between precipitation and evaporation is such that there is little net downward movement of water in the soil, with two important consequences: (1) the soil is largely a *closed system*, in the sense that chemical constituents produced by weathering remain in the soil rather than being carried away in solution in groundwater, and (2) the pH of soil waters tends to be *alkaline* rather than acidic. In such soils, upwards of a meter thick, the upper layer (A horizon) of the soil, beneath the surface organic layer, are a deep, homogeneous, and friable mixture of organic matter and mineral particles, not strongly leached, which is kept thoroughly mixed by the activity of burrowing organisms. This layer grades downward into a lighter-colored material with less organic matter, and then passes downward into a layer (the B horizon) in which calcium carbonate has been precipitated by mobilization of calcium ions in the upper layers and precipitation at deeper levels. The carbonate is precipitated as soil moisture is drawn back up toward the surface. In areas where the parent material contains carbonate minerals, the carbonated layer is better developed, but carbonate minerals tend to be deposited in the lower part of the soil profile even in the absence of carbonate minerals in the parent material. Why? Because of the common presence of calcium ions in parent silicate minerals, and the presence of carbon dioxide in the surface waters, to supply the carbonate ion.

## **9.8 Classification of Soils**

**9.8.1** Now, on to the troublesome and confusing matter of soil classification. Early classifications, before the middle of the 1900s, relied heavily upon terms and concepts developed in the late 1800s by Russian soil scientists, who were true pioneers in the scientific study of soils. These terms are common still, but in recent decades a systematic and very detailed classification system, developed and continually refined by soil scientists at the U.S. Department of Agriculture, has come into widespread use, despite its complexity and almost absurdly non-intuitive terminology. Also, other newer classifications are in use in other parts of the world.

**9.8.2** (The “almost absurdly” in the preceding paragraph a value judgment, I know, but it’s a widely shared view. Just for laughs, I am appending, as Figure 2-31, from the thick tome on soil taxonomy published by the Soil Survey Staff of the

Department of Agriculture, the official names of the “great groups”, which are the third stage in the taxonomic hierarchy. Each of the 185 great groups is in turn subdivided into subgroups. There are nearly a thousand subgroups! This classification is a marvel of rational and synthetic classification and terminology. Also in Figure 2-30 is a typical page from the same source, giving a description of some of the subgroups of one of the great groups. Another other interesting point about this classification is that, in contrast to any other classification I know of, in any field of science, they insist upon capitalizing every soil name: Durorthidic Xerorthents, rather than durorthidic xerorthents.)

**9.8.3** I think that what would be most valuable for you at this point is to have soil classification stripped to its barest essentials. I will do that for you in two ways. First, Figure 2-31 gives the soil “orders” of the USDA classification system. Even this most boiled-down presentation needs a few words of explanation. A *pedon* (rhymes with “head on”) is the basic soil unit: the column of soil, with its various horizons, in a small representative area of the order of a square meter. The *epipedon* is just the surface horizon of the pedon; it’s usually rich in organic matter. Second, Figure 2-32, in three parts, gives (A) a simple classification of the earth’s climatic zones, and corresponding to that, (B) the major vegetation types and (C) the major soil groups. The typical New England soil, described in the preceding section, is a podzol developed in a climate with severe winters (!) and with spruce–fir forest. (Minor note: the climate of southern New England is gradational into the colder end of the “humid” box in Figure 2-32 and the box for gray brown podzolic soil in Figure 2-32. The native forests in southern New England are a mixture of conifers and deciduous hardwoods.)

Names of orders, suborders, and great groups

Order	Suborder	Great group	Order	Suborder	Great group
Alfisol	Aqualf	Albaqualf.	Histosol	Fibrif	Borofibrif.
		Duraqualf.			Cryofibrif.
		Fragiaqualf.			Luvifibrif.
	Boralf	Glossaqualf.		Medifibrif.	
		Natraqualf.		Sphagnofibrif.	
		Ochraqualf.		Tropofibrif.	
		Plinthoqualf.		Borofolist.	
		Tropaqualf.		Cryofolist.	
		Umbraqualf.		Tropofolist.	
		Cryoboralf.		Borohehemi.	
	Udalf	Eutroboralf.		Cryohemi.	
		Fragiboralf.		Luvihemi.	
		Glossoboralf.		Medihemi.	
		Natroboralf.		Sulfihemi.	
		Paleboralf.		Sulfohemi.	
		Agrudalf.		Tropohemi.	
		Ferrudalf.		Borosaprist.	
	Ustalf	Fragiudalf.		Cryosaprist.	
		Fraglossudalf.		Medisaprist.	
		Glossudalf.		Troposaprist.	
Hapludalf.		Cryandep.			
Natrudalf.		Durandep.			
Paleudalf.		Dystrandep.			
Rhodudalf.		Eutrandep.			
Tropudalf.		Hydrandep.			
Durustalf.		Placandep.			
Haplustalf.		Vitrandep.			
Xeralf	Natrustalf.	Andaquep.			
	Paleustalf.	Cryaquep.			
	Plinthustalf.	Fragiaquep.			
	Rhodustalf.	Halaquep.			
	Durixeralf.	Haplaquep.			
	Haploxeralf.	Humaquep.			
	Natrixeralf.	Placaquep.			
Aridisol	Argid	Palexeralf.	Ochrept	Cryochrept.	Tropaquep.
		Plinthoxeralf.			Durochrept.
		Rhodoxeralf.			Dystrochrept.
	Orthid	Durargid.		Eutrochrept.	
		Haplargid.		Fragiochrept.	
		Nadurargid.		Ustrochrept.	
		Natargid.		Xerochrept.	
		Paleargid.		Plaggep.	
		Calcioorthid.		Tropep.	
		Camborthid.		Dystropep.	
Durorthid.	Eutropep.				
Entisol	Aquent	Gypsiorthid.	Mollisol	Umbrept	Humitropep.
		Paleorthid.			Sombrित्रोपेप.
		Salorthid.			Ustropep.
	Arent	Cryaquent.		Cryumbrept.	
		Fluvaquent.		Fragiumbrept.	
		Haplaquent.		Haplumbrept.	
		Hydraquent.		Xerumbrept.	
		Psammaquent.		Argialboll.	
		Sulfaquent.		Natraboll.	
		Tropaquent.		Argiaquoll.	
Fluvent	Arent.	Calciquoll.			
	Cryofluvent.	Cryaquoll.			
	Torrifluvent.	Duraquoll.			
	Tropofluvent.	Haplaquoll.			
	Udifluvent.	Natraquoll.			
	Ustifluvent.	Argiboroll.			
	Xerofluvent.	Calciboroll.			
Orthent	Cryorthent.	Cryoboroll.			
	Torrorthent.	Haploboroll.			
	Troporthent.	Natroboroll.			
	Udorthent.	Paleboroll.			
	Ustorthent.				
	Xerorthent.				

(continued on next page)

	Psamments.....	Cryopsamments. Quartzpsamments. Torripsamments. Tropopsamments. Udipsamments. Ustipsamments. Xeropsamments.		Rendolls.....	Vermiborolls. Rendolls. Argidolls. Hapludolls. Paleudolls. Vermudolls. Argustolls.
		Calcustolls. Durustolls. Haplustolls. Natrustolls. Palcustolls. Vermustolls.		Ustolls.....	Fragihumods. Haplohumods. Placohumods. Trophumods. Cryorthods. Fragiorthods. Haploorthods. Placorthods. Troporthods. Albaquults. Fragiaquults. Ochraquults. Paleaquults. Plinthaquults. Tropaquults. Umbraquults. Haplohumults. Palehumults. Plinthohumults. Sombrihumults. Trophumults.
	Xerolls.....	Argixerolls. Calcixerolls. Durixerolls. Haploxerolls. Natrixerolls. Palexerolls.	Ultisols.....	Aquults.....	Umbraquults. Haplohumults. Palehumults. Plinthohumults. Sombrihumults. Trophumults. Fragiudults. Hapludults. Paleudults. Plinthudults. Rhodudults. Tropudults. Haplustults. Paleustults. Plinthustults. Rhodustults. Haploxerults. Palexerults.
Oxisols.....	Aquox.....	Gibbsiaquox. Ochraquox. Plinthaquox. Umbraquox.		Humults.....	Umbraquults. Haplohumults. Palehumults. Plinthohumults. Sombrihumults. Trophumults.
	Humox.....	Aerohumox. Gibbsihumox. Haplohumox. Sombrihumox.		Udults.....	Fragiudults. Hapludults. Paleudults. Plinthudults. Rhodudults. Tropudults. Haplustults. Paleustults. Plinthustults. Rhodustults. Haploxerults. Palexerults.
	Orthox.....	Aerorthox. Eutorthox. Gibbsiorthox. Haploorthox. Sombriorthox. Umbriorthox.		Ustults.....	Haplustults. Paleustults. Plinthustults. Rhodustults. Haploxerults. Palexerults.
	Torrox.....	Torrox.		Xerults.....	Torrerts. Torrerts. Chromuderts. Pelluderts. Chromusterts. Pellusterts. Chromoxererts. Pelloxererts.
	Ustox.....	Acrustox. Eustrustox. Sombriustox. Haplustox.		Torrerts.....	Torrerts.
		Cryaquods. Duraquods. Fragiaquods. Haplaquods. Placaquods. Sideraquods. Tropaquods.	Vertisols.....	Uderts.....	Chromuderts. Pelluderts. Chromusterts. Pellusterts. Chromoxererts. Pelloxererts.
Spodosols.....	Aquods.....	Fragiaquods. Haplaquods. Placaquods. Sideraquods. Tropaquods.		Usterts.....	Chromusterts. Pellusterts. Chromoxererts. Pelloxererts.
	Ferrods.....	Ferrods.		Xererts.....	Chromoxererts. Pelloxererts.
	Humods.....	Cryohumods.			

Figure 2-30. A) Orders, suborders, and great groups of the soil classification system developed by the Soil Survey Staff at the US Department of Agriculture. B) (next page) A random page from the system of soil taxonomy of the US Department of Agriculture. (From Soil Survey Staff, 1994)

States, but the subgroup is provided for use where needed.

**Epiaquic Tropohumults.**—Soils in this subgroup have some yellowish brown or olive brown colors in the upper horizons and have redder mottles inside the peds. The hue becomes redder with depth. The rainfall is seasonally high, and the upper horizons become saturated. The reducing conditions are commonly but not everywhere too mild to produce mottles of low chroma, but the hue in the upper horizons approaches that of Aquults. These soils are in the areas of high rainfall in Puerto Rico but are not extensive. Their slopes are mostly strong, and the soils are mostly cultivated.

**Epiaquic Orthoxic Tropohumults.**—These soils have some yellowish brown or olive brown colors in the upper horizons and have redder mottles inside the peds. The hue becomes redder with depth. The reducing conditions are commonly too mild to produce mottles of low chroma. The clay in these soils has low activity. The soils have strong slopes and are in the areas of high rainfall in Puerto Rico, but they are not extensive.

**Humoxic Tropohumults.**—These soils have a clay fraction that has low activity. They are common in the United States in Hawaii, where they formed on steep slopes or fans from basic rocks. They are largely used for pasture, but where slopes are favorable some of them are cultivated.

**Orthoxic Tropohumults.**—These soils have a clay fraction that has low activity, and they tend to be warmer than most of the Typic Tropohumults. They are not extensive in either Puerto Rico or Hawaii. Their slopes are mostly steep, and the soils are used mainly for pasture or tree crops.

**Ustic Tropohumults.**—These soils have an ustic moisture regime but are otherwise like Typic Tropohumults in defined properties. They are not known to occur in the United States, but the subgroup is provided for use if needed elsewhere.

**Ustoxic Tropohumults.**—Soils in this subgroup have an ustic moisture regime and a clay fraction that has low activity. They are not extensive in the United States.

## Udults

These are the more or less freely drained, humus-poor Ultisols (plate 9D) in humid climates in mid or low latitudes that have well-distributed rainfall. Most of them have light-colored upper horizons, commonly a grayish horizon that rests on a yellowish brown to reddish argillic horizon. A few that developed from basic rocks have a dark brown or reddish brown surface horizon that rests on a dark red or dusky red argillic horizon. Some have a fragipan or plinthite, or both, in or below the argillic horizon.

The Udults developed in sediments and on surfaces that range from late Pleistocene to Pliocene or possibly older. Many are cultivated, either with the use of soil amendments or in a system in which they are

cropped for a very few years and then are returned to forest to allow the trees to regather in their tissues the small supply of nutrients. Most of these soils have or had a forest vegetation, but some have a savanna that probably is anthropic. The moisture regime is udic. The soil temperature regime is mesic, isomesic, or warmer.

## Definition

- Udults are the Ultisols that
1. Are never saturated with water or have a redder hue or higher chroma than Aquults;
  2. Have <0.9 percent organic carbon in the upper 15 cm of the argillic horizon and have <12 kg organic carbon in a unit volume of 1 square meter to a depth of 1 m exclusive of any O horizon; and
  3. Have a udic moisture regime.

## Key to great groups

FCA. Udults that have a fragipan in or below the argillic horizon.

**Fragiudults**, p. 360

FCB. Other Udults that have plinthite that forms a continuous phase or constitutes more than half the volume in some horizon within the upper 1.25 m of the soil.

**Plinthudults**, p. 366

FCC. Other Udults that have an argillic horizon that has <10 percent weatherable minerals in the 20- to 200-micron fraction in the upper 50 cm and have a clay distribution such that the percentage of clay does not decrease from its maximum amount by more than 20 percent of that maximum within 1.5 m of the soil surface, or the layer in which the percentage of clay is less than the maximum has skeleton on ped faces or has 5 percent or more plinthite by volume.

**Paleudults**, p. 364

FCD. Other Udults that have

1. An epipedon that has a color value, moist, of less than 4 in all parts; and
2. An argillic horizon that has a color value, dry, of less than 5 and not more than 1 unit higher than the value, moist.

**Rhodudults**, p. 367

FCE. Other Udults that have an isomesic or warmer iso temperature regime.

**Tropudults**, p. 367

FCF. Other Udults.

**Hapludults**, p. 362

## Fragiudults

These are the Udults that have a fragipan in or below the argillic horizon. They were formed mainly in loamy sediments or in residuum. The fragipan normally has an upper boundary about 50 to 75 cm below the soil surface. Perched ground water is present above the pan at some period during the year, and there are commonly thick gray skeleton at the top of the pan.

The Fragiudults in the United States are principally on gentle slopes throughout the southeastern states. Their temperature regime is mesic or thermic. Their vegetation in the United States has been forest, either of conifers or of broadleaf deciduous trees. Some were considered Planosols and others Red-Yellow Podzolic soils in the 1938 classification or the 1949 revision.

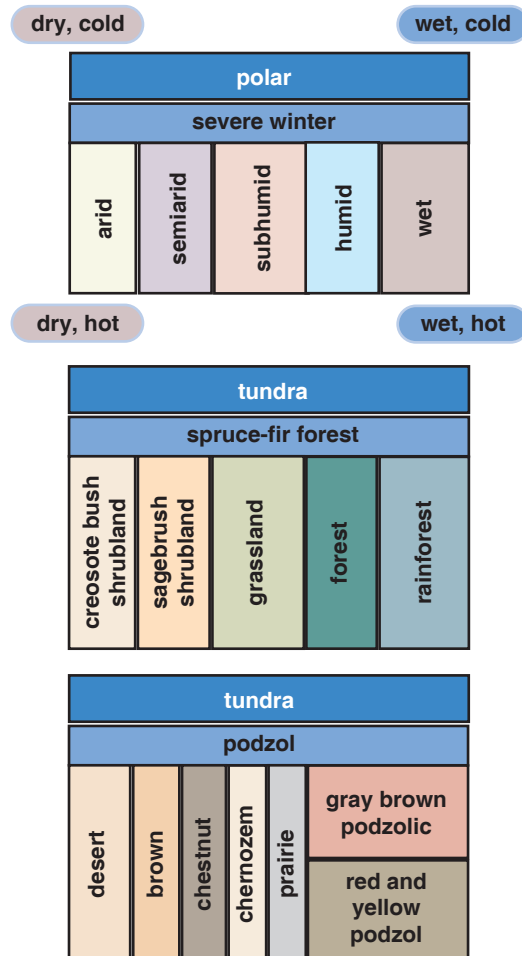


Orders of the U.S. Comprehensive Soil Classification System			
Order	Derivation	Formative Element	Description
Alfisols	Meaningless syllable	alf	Gray to brown epipedons; formed mostly in humid region areas under native deciduous forests
Aridisols	L. "aridus" dry	id	Desert or dry soils with an ochric epipedon
Entisols	Meaningless syllable	ent	Recent soils lacking profile development; found under wide variety of climatic conditions
Histosols	Gr. "histos", tissue	ist	Organic soils
Inceptisols	L. "inceptum",	ept	Young soils, more developed than Entisols; Dark (mollic) epipedons, includes some of the world's most important agricultural soils
Mollisols	L. "mollis", soft	oll	Dark, (mollic) epipedons, includes some of the world's most important agricultural soils
Oxisols	Fr. "oxide", oxide	ox	Oxic subsurface horizon from intense leaching of silica leaving Fe and Al oxides
Spodosols	Gr. "spodos", wood ash	od	Lighter colored (usually albic) horizon above spodic horizon
Ultisols	L. "ultimus", last	ult	Old, moist soils developed under warm to tropical climates, argillic horizons with low base saturation
Vertisols	L. "verto", turn	ert	High content of swelling clays which can develop deep, wide cracks when dry

**Alfisols (high-nutrient soils)**  
**Andisols (volcanic soils)**  
**Aridisols (desert soils)**  
**Entisols (new soils)**  
**Gelisols (permafrost soils)**  
**Histosols (organic soils)**  
**Millisols (prairie soils)**  
**Oxisols (tropical forest soils)**  
**Spodosols (conifer forest soils)**  
**Ultisols (low-nutrient soils)**  
**Vertisols (swelling-clay soils)**

Figure by MIT OCW.

Figure 2-31. Orders of the US Department of Agriculture soil classification system. (Modified from Winegardner, 1996.)



Highly simplified relationship of vegetation type and soil type to major climatic zones. (Top) distribution of climate types. (Middle) Distribution of vegetation types on the climate base. (Bottom) Distribution of soil groups on the climate base.

Figure by MIT OCW.

Figure 2-32. Highly simplified relationship of vegetation type and soil type to major climatic zones. (Top) distribution of climate types. (Middle) Distribution of vegetation types on the climate base. (Bottom) Distribution of soil groups on the climate base. (From Hunt, 1972.)

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