PART II

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Source to Sink

(Chapters 3 to 8)

Landscapes are shaped by both subtle and dramatic geomorphic processes that move mass from uplands to lowlands and eventually to the coast and offshore to the deep ocean basins. In the next six chapters, we examine the processes that move earth materials and generate landforms. Chapter 3 considers how rocks weather to produce soil and sediment. In Chapter 4, we trace the movement of water at and near Earth's surface. Chapter 5 introduces the processes active on hillslopes and the landforms that these processes produce. Flow through channels is covered in Chapter 6; drainage basins, the fundamental unit of landscape analysis, are the topic of Chapter 7. In Chapter 8, we consider coastal and marine environments, the depositional archives that store material removed from the land surface. After reading these six chapters, you should understand how material moves across Earth's surface from source to sink and how surface processes drive this cycle and shape our planet's landscapes.

The Amazon is Earth's largest river system. As recently as 15 million years ago, the river flowed westward, until the rise of the Andes forced a new eastward flow path to the Atlantic. Today the material eroded from the Andes is transported east to the river mouth and the long-term sink of the Atlantic Ocean. This image shows a digital elevation model of the Amazon River Basin from Shuttle Radar Topography Mission (SRTM) data. By assigning color and brightness to the elevation values (from light gray at sea level to cool pink at the highest elevations), the image shows both the local terrain detail and the major regional landscape elements.

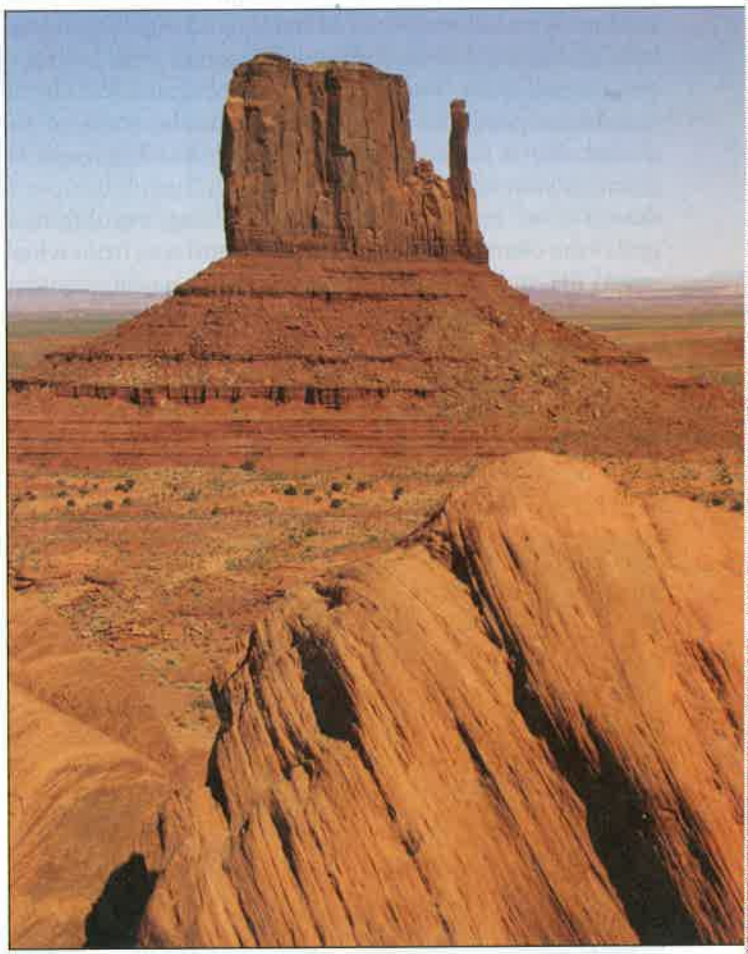
Weathering and Soils

3

Introduction

A thin layer of mechanically broken and chemically altered rock mixed with living and dead organic material defines the transition from Earth's rocky interior to its gaseous atmosphere. This layer is produced by **weathering**, the chemical or physical alteration of **primary minerals** (unaltered by weathering) that make up the **parent material**, rock or sediment. From a geomorphologist's perspective, this thin layer plays a key role in defining the shapes and processes of Earth's surface because erosion is much more effective at altering landscapes once solid rock breaks down into transportable material—a process involving both physical changes and chemical transformations that create **secondary minerals** and **dissolved ions.**

Three types of weathering are generally recognized and closely related: **physical weathering, chemical weathering,** and **biological weathering.** The greater surface area that results from physically breaking rocks into smaller fragments accelerates chemical weathering. Changes in mass or volume accompanying chemical decomposition can promote physical weathering. Even though different environmental conditions tend to favor physical versus chemical weathering, the processes are complementary and often strongly influence each other. Biological activity catalyzes both physical and chemical weathering. The weathering effects of biological activity depend on the active species, substrate, and weathering processes in question. Tree roots, for example, slowly pry rocks apart as they grow [**Photograph 3.1**], and the organic acids they


M. Miller

Rock weathering can produce spectacular landscapes. In Monument Valley, Arizona, differential rock weathering and erosion of sandstone formations result in stair-stepped topography and distinctive towers of rock. Weathering etches less resistant layers so that bedding becomes visible. The pattern of jointing and overall rock strength varies with rock type. Debris shed from rock faces forms angle-of-repose slopes.

IN THIS CHAPTER

secrete, as well as the carbon dioxide (CO_2) they respire into the soil, increase rates of chemical weathering.

Physical, chemical, and biological processes all act as weathering agents, but their rates and relative importance vary dramatically among different landscapes around the world. Weathering processes create and influence the physical and chemical properties of the layer of **regolith** (unconsolidated and weathered material covering fresh bedrock) that provides the foundation for terrestrial life. Not all regolith is produced in situ—for example, some of the glacial debris covering New England was imported by glaciers from Canada and some regolith on hillslopes is derived from upslope. Such mobile hillslope regolith integrates the characteristics of the rocks upslope, from which it was derived. Regolith properties—such as clay content, the speed with which water moves through it, and its grain size—directly influence the distribution of geomorphological processes and the resulting landforms.

Biological activity affects both the rate and style of physical and chemical weathering. Organisms speed up mineral weathering by mechanically breaking down rocks and by catalyzing chemical reactions and causing changes in environmental conditions (such as soil pH). The growth of plant roots helps disintegrate rocks by gradually prying apart cracks, fractures, and other openings. Burrowing animals, like gophers, ants, and termites, excavate rock fragments and mix them into overlying soils [**Photograph 3.2**]. The process of biologically mediated mixing, called **bioturbation**, disrupts original structures or fabric in the **parent material.**

Biological activity also increases the potential for chemical weathering by increasing surface area and creating and enlarging pathways for subsurface water flow. Because oxygenated water is reactive, the more water that moves through a soil, the more weathering takes place. Decaying organic matter and respiration of soil microorganisms and plant roots indirectly influence chemical weathering by increasing the concentration of CO_2 in soil gases, thereby promoting acidification of soilwater and enhancing chemical weathering in the fractures and pores through which soilwater flows. Plant roots release organic acids that enter soilwater and either attack fresh mineral surfaces directly or exchange hydrogen ions for nutrient cations that plant roots absorb as they take up water. Decaying organic matter also releases humic acids that facilitate further weathering. Bacteria can mediate weathering reactions, and lichen colonization is the first step in weathering of many rocky surfaces. In short, more life leads to more rapid rock breakdown and thus more weathering.

Weathering and erosion do not necessarily progress at the same pace. Where weathering outpaces erosion, landscapes develop thick mantles of **saprolite,** chemically altered but in-place rock that has lost mass and strength but not volume during weathering. Saprolite can be hundreds of meters deep in flat-lying, slowly eroding areas of tropical Africa and South America. It can erode like loose sand, yet still retain structures or features of the original parent material, such as igneous dikes, calcite veins, or sedimentary bedding planes [**Photograph 3.3**]. Saprolite density can be as low as half that of intact rock due to mass lost by solution.

D. Thompson

PHOTOGRAPH 3.1 Biological Activity and Weathering. Plant roots can pry apart rock, such as this small tree wedging open a glacial erratic in Ledyard, Connecticut.

D. Eldridge

PHOTOGRAPH 3.2 Bioturbation. Funnel ants live in the semi-arid woodlands of eastern Australia and make volcano-shaped ant hills that dominate the forest floor. Their burrowing activity turns over the entire soil profile down to a depth of 30 cm every 200 years or so.

Not everyone defines soil similarly. Some consider soil to be any unconsolidated material at Earth's surface. Others consider soil to be only the material that has been affected by pedogenesis. In this book, we define soil broadly as the unconsolidated mineral and organic material on the surface of the Earth that serves as a natural medium for the growth of land plants and animals and that differs from the material from which it is derived in its physical, chemical, biological, and morphological properties. Where it is present, soil forms the upper part of the regolith.

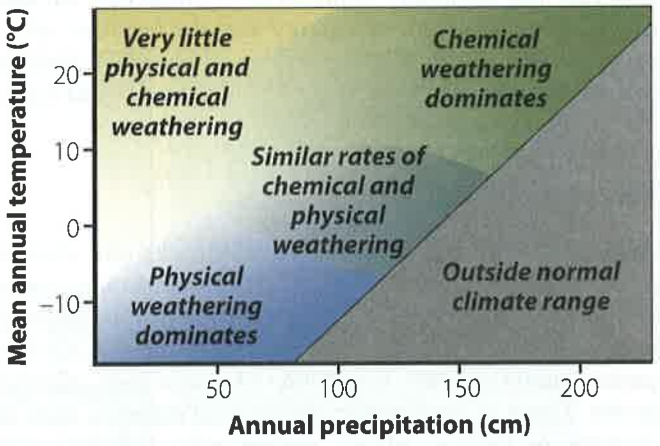
Soil forms on Earth's outer surface and provides the substrate in which plants root, terrestrial life derives sustenance, and erosion acts to shape many landscapes. Soils not only harbor and sustain life, they are themselves partly composed of organic matter. Plants and animals both depend on soils and, in turn, influence the rate of **regolith production** (rock weathering) and the rate of soil development, or **pedogenesis**—the result of physical, chemical, and biological processes that alter the appearance and properties of the parent material upon which the soil is developed. In this sense, soil development may be considered a top-down process through which the properties of surficial materials change in response to the influence of environmental factors (like climate and vegetation), whereas soil production may be viewed as a bottom-up process through which rocks break down into surficial materials.

The presence, amount, and phase of water (i.e., vapor, liquid, or ice) strongly influence patterns and rates of both chemical and physical weathering. It is not surprising, then, that global patterns in the style and intensity of weathering generally track regional climate. Differences in temperature and precipitation control both the magnitude and relative importance of physical and chemical weathering processes [**Figure 3.1**]. Chemical weathering is most important in regions with high year-round temperatures and abundant precipitation, like the equatorial regions, and least important in cold, dry areas, like the poles. Higher temperatures promote much faster reactions because of the nonlinear dependence of chemical reaction rates on temperature; thus, temperature extremes can control chemical weathering rates. A short time spent at high temperature can produce as much chemical weathering as a long time at the average temperature. Relative to other environments, physical weathering is thought to be least active in dry environments and most active in regions where temperatures repeatedly cycle through freezing for part of the year. Consequently, tropical regions are generally dominated by chemical weathering and high latitudes by physical weathering.

Topography also greatly influences weathering. Mechanical weathering dominates in high mountains with steep slope gradients that create stresses in rock, extensive fracturing from tectonic forces, and periodic freezing temperatures. Low-relief and especially low-elevation environments generally favor chemical weathering. Rates of physical and chemical weathering influence one another because high rates of mechanical breakdown promote high rates of chemical decomposition by exposing

R. Marschik

PHOTOGRAPH 3.3 Saprolite. Saprolite developed on nickel-rich weathered serpentinite in Greece. Veins of calcite stand above the weathered rock surface in the lower left.



Environmental controls on **physical weathering** and **chemical weathering** vary depending on the temperature and the amount of moisture. Chemical weathering dominates in hot, wet climates, and physical weathering dominates in cold, dry climates, where chemical weathering proceeds slowly. Chemical weathering tends to decrease with increasing elevation and at higher latitude, whereas physical weathering tends to increase at higher elevation and higher latitude.

FIGURE 3.1 Physical and Chemical Weathering. The intensity and thus importance of physical and chemical weathering vary with climate (precipitation and temperature) and degree of prior weathering.

greater surface area to chemical attack and vice versa (through reduced material strength). Consequently, the highest total rates of rock weathering typically occur in environments conducive to both physical and chemical weathering, such as steep terrain in tropical regions.

Styles and rates of weathering vary among rock types because of differences in chemical (mineralogical) characteristics and physical attributes (strength and fracturing). In particular, the ease with which water is able to penetrate rocks strongly influences weathering rates. As discussed earlier, mineral composition is a primary factor in chemical weathering. Some rock types, such as limestone and marble, are particularly vulnerable to chemical weathering in humid climates because they can be dissolved by weakly acidic soilwater or rainwater. In contrast, these rock types can be quite erosion resistant in arid regions where water is not sufficiently abundant to carry away dissolved ions. Compared to limestone, quartz is far more resistant to both chemical and physical weathering, and rocks like quartz sandstone and quartzite weather more slowly than most other rocks, regardless of environmental conditions.

This chapter reviews the processes that act to weather rocks, sediment, and minerals at and near Earth's surface and introduces the physical, chemical, and biological processes that produce soils. We explore the dominant controls on the transformation of primary rock-forming minerals into secondary minerals and consider global and lithologic controls on weathering and the resulting influences on landforms. We also address how soil scientists recognize diagnostic characteristics and classify soil types.

Physical Weathering

Physical processes mechanically break rocks into smaller pieces (**disaggregation**). During physical weathering, rocks and rock-forming minerals break apart without changes in composition. In some cases, physical weathering reduces parent material into fragments of individual mineral grains. Zones of weakness in the original material, such as cleavage or bedding planes, metamorphic foliation, and even mineral boundaries, often determine the size and shape of rock fragments. Cracks form where stresses imparted by expansion, contraction, or shearing exceed the strength of rocks or minerals. The most pronounced physical weathering in bedrock occurs where there is a strong directional contrast in pressure. This is the case at and near the land surface where overburden is minimal and where open fractures are more abundant than at depth, where confining pressures are greater and more uniform.

Important general mechanisms of physical weathering include release of confining pressure that allows rock to expand and fracture, thermal expansion from **insolation** (heating by the sun) or forest fires, and cyclic expansion and contraction from freeze–thaw action in cold environments. Disintegration as a result of wetting and drying is important in rocks with minerals susceptible to shrinking and swelling. Expansion by growth of salt crystals can crack rocks in arid and coastal environments.

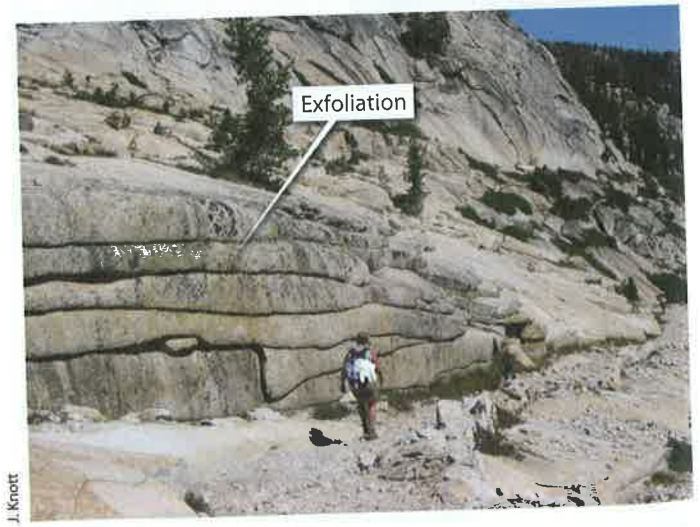
Exfoliation

Exfoliation describes the processes by which exposed outcrops of rock break into sheets oriented parallel to the land surface. These **exfoliation sheets** are typically 0.5 to 10 m thick and result in onionlike fracture patterns [**Photograph 3.4**]. Because exfoliation sheets are more readily eroded than underlying unfractured rock, slope-parallel fractures tend to mimic topographic forms as subsequent sheets form and erode.

The process of exfoliation commonly produces dome-shaped surfaces, like Yosemite Valley's famous granitic Half Dome. Exfoliation sheets are most apparent on bare rock surfaces, but slope-parallel fractures also develop within soil-mantled slopes, creating networks of discontinuities that can greatly influence near-surface groundwater flow paths and slope stability. Exfoliation is typically more common and better developed in igneous and metamorphic rocks that formed deep (>10 km) within Earth's crust than in sedimentary and volcanic rocks formed at shallower depths.

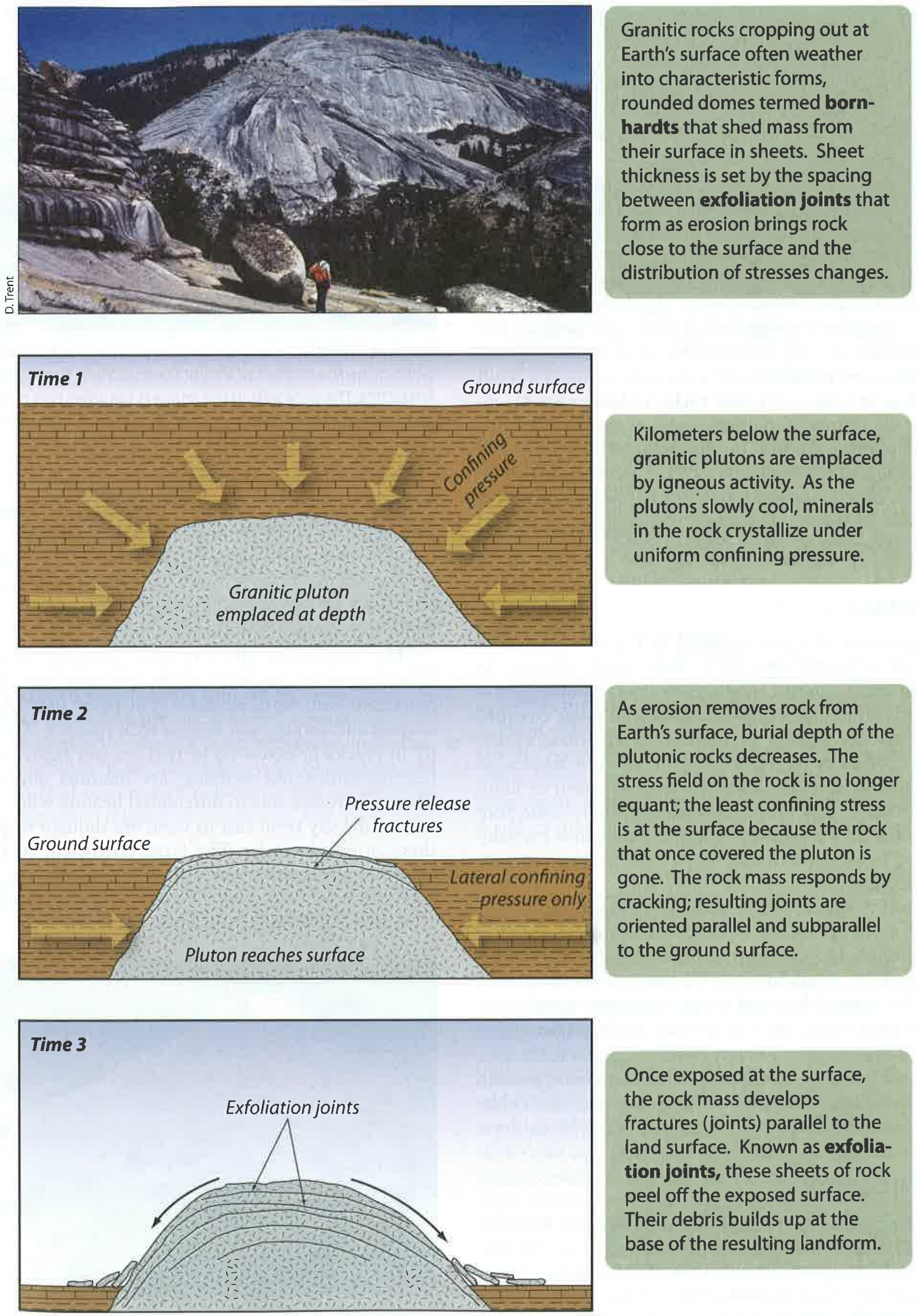
Exfoliation occurs as erosion brings rock closer to Earth's surface, increasing the contrast between the stresses locked in during crystallization or metamorphism and the stress imposed by adjacent rock [**Figure 3.2**]. When the difference becomes greater than the strength of the rock, the rock cracks. Even in sedimentary rocks, upward expansion from unloading separates bedding planes, allowing outward expansion where bedding tilt matches the topographic slope. Fracturing due to stress release is an important mechanism that allows water, oxygen, and plant roots to penetrate into a rock mass.

Rock masses subject to tectonic stresses or rocks that have shrunk upon cooling develop patterns of **joints** (fractures along which no significant movement has occurred) that provide planes of weakness along which weathering and erosion



J. Knott

PHOTOGRAPH 3.4 Exfoliation. Exfoliation (sheet jointing) in granite in Yosemite National Park, California, north of May Lake.



Granitic rocks cropping out at Earth's surface often weather into characteristic forms, rounded domes termed **bornhardts** that shed mass from their surface in sheets. Sheet thickness is set by the spacing between **exfoliation joints** that form as erosion brings rock close to the surface and the distribution of stresses changes.

Kilometers below the surface, granitic plutons are emplaced by igneous activity. As the plutons slowly cool, minerals in the rock crystallize under uniform confining pressure.

As erosion removes rock from Earth's surface, burial depth of the plutonic rocks decreases. The stress field on the rock is no longer equant; the least confining stress is at the surface because the rock that once covered the pluton is gone. The rock mass responds by cracking; resulting joints are oriented parallel and subparallel to the ground surface.

Once exposed at the surface, the rock mass develops fractures (joints) parallel to the land surface. Known as **exfoliation joints,** these sheets of rock peel off the exposed surface. Their debris builds up at the base of the resulting landform.

FIGURE 3.2 Exfoliation Sheets. Exfoliation occurs when the confining pressure on a rock mass is not equal on all sides. As erosion brings rock closer to the surface, the pressure from above is reduced and joints open.

can penetrate and preferentially remove material. Because the tensile (extensional) strength of rock is typically many times less than its compressive and shear strength, jointing is often well developed even in relatively strong, erosion-resistant rocks. Parallel sets of extensional joints develop orthogonally to the direction of maximum stress as a result of either crustal extension or cooling of igneous rock.

Joint systems can have strong topographic expression in arid and semi-arid landscapes where regolith is thin or absent and bedrock dominates hillslope morphology. Joints provide avenues for water and plant roots that focus erosion and facilitate infiltration and groundwater flow into rock that, in turn, promote more aggressive weathering. Erosion along intersecting sets of joints can produce isolated columns of rock separated by weathered-out joints. Joints also form parallel to the strike of bedding as a result of bending or folding of brittle rocks. Lithology greatly influences the degree of jointing, with joints typically being better developed in more brittle rocks such as sandstone and granite and less well-developed in more ductile rocks like shale. In addition, joint patterns developed in igneous rocks tend to be less linear and more irregularly spaced than those developed in sedimentary rocks.

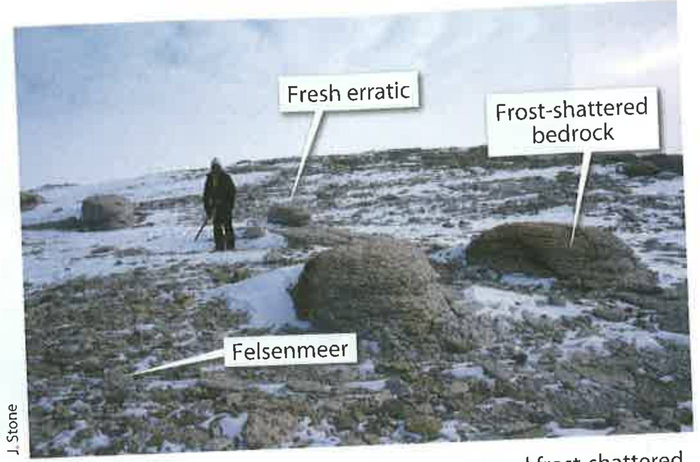
Freeze–Thaw

The expansion of water confined in fractures and pore spaces as it freezes into ice is particularly effective at breaking rocks apart. Consequently, **frost shattering** is a primary weathering process in alpine and polar environments that are subject to frequent freeze–thaw cycles. Water expands by almost 10 percent when it freezes, so rock generally will not shatter unless 90 percent or more of the available pore space is saturated. If only some pore space is filled by water, ice simply expands into partially saturated voids without generating high pressures on the surrounding rock.

Hydrofracturing, a process by which freezing of water proceeds from the outside of the rock inward, forces water into the tiny ends of fractures, producing an effect like a hydraulic jack. In addition, as ice lenses form, water and vapor flow toward them and freeze, imparting enough force to cause growing ice crystals to crack rocks. Many alpine slopes in environments subject to frost shattering are covered by a blanket of rock blocks called **felsenmeer** (German for "rock sea") [Photograph 3.5], for example, the rubble-covered summit of Mount Washington in New Hampshire.

Thermal Expansion

Rocks and minerals expand when subjected to heat, but rock's low **thermal conductivity** generally prevents the zone of heating from penetrating more than a few centimeters into rocks over the course of a forest fire or a day of exposure to intense sunlight. Extreme temperature differences between a rock's surface and its cool interior can produce large differential stresses that exceed rock strength and result in **spalling** (breaking away) of a thin outer layer that can be up to several centimeters thick [Photograph 3.6]. Such differences are particularly acute during range and forest fires and can result in extensive loss of mass from rock surfaces over time. Mass loss from rock surfaces has implications for surface exposure dating techniques examined in Chapter 2.



J. Stone

PHOTOGRAPH 3.5 Felsenmeer. Felsenmeer and frost-shattered bedrock, on the summit of Mount Darling, Marie Byrd Land, Antarctica. The geologist in the image is between two much less weathered glacial erratics deposited on the surface by glacial ice frozen to its weathered bed.

The weathering effect of daily temperature fluctuations due to heating by sunlight has long been debated, but recent field work provides convincing data that daily heating and cooling can lead to rock fracture. The majority of cracks in desert rocks that are not related to rock heterogeneities, like bedding, are oriented north–south. Thermal stresses, due to differential heating when the sun crosses the sky from east to west, are thought to produce these oriented cracks. The large **coefficient of thermal expansion** of certain minerals, such as calcite, can cause a rock to break apart along mineral boundaries as individual grains expand and contract differently with small changes in temperature. The weathering product of this type of granular disintegration, known as **grus** from the German for grit, fine gravel or debris, consists of loose, unconsolidated, individual mineral grains. Grus is most commonly formed by the disintegration of coarse-grained intrusive igneous rocks, such as granite.

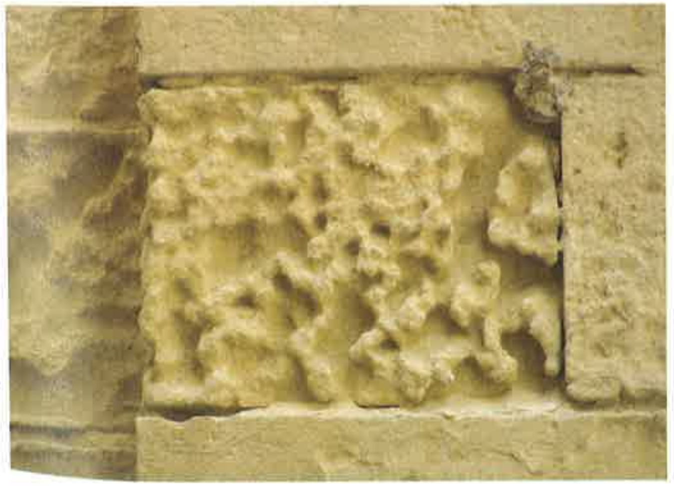
D. Thompson

PHOTOGRAPH 3.6 Fire Spalling. Fire-spalled granitic boulder in Pingree Park, Colorado, with accumulation of thin chips of rock around the base of the boulder.

Wetting and Drying

The addition and removal of water from minerals, processes known as **hydration** and **dehydration,** can cause swelling or contraction capable of fracturing rocks or disaggregating them into individual mineral grains. The volume increase in the anhydrite to gypsum transformation, in particular, has been ruinous to archaeological monuments. Most clay minerals shrink when they dry and swell when they absorb water. Some can expand to twice their original volume, and certain particularly susceptible clays, like smectite (including montmorillonite), swell severalfold when they get wet. **Expansive soils,** containing minerals with such shrink-swell behavior, cause substantial engineering problems and extensive damage (e.g., cracked foundations) in the southern United States outside of the glacial limits and in the western plains where such soils are common. Swelling of smectite-rich soils in Montana, Wyoming, and Colorado can make travel on dirt roads nearly impossible after rainstorms, because sticky mud bogs down even vehicles with four-wheel drive.

In some rocks, like granite, physical expansion of biotite micas during weathering can pry apart grain boundaries and produce grus. Similarly, cycles of wetting and drying can lead to repeated expansion and contraction that disaggregate micaceous sandstone and turn hard, erosion-resistant rock into a loose pile of sand in just a few years or decades. In rock types susceptible to this process, exposure to seasonal cycles of wetting and drying leads to rapid rates of bedrock erosion through the annual formation of a loose crust or outer covering of highly erodible material easily removed by subsequent heavy rainfalls or high river flows.



B. Smith et al.

PHOTOGRAPH 3.7 Limestone Weathering. Honeycomb weathering of limestone blocks in Malta results in the formation of gypsum at the surface. The weathering pattern on the rock surface here is controlled by the fossil pattern of bioturbation from marine animal burrows in the original sediment. The width of the image is approximately 70 cm.

G. Mustoe

PHOTOGRAPH 3.8 Salt Residue. White salt residue on honeycomb weathering in Capitol Reef National Park, Utah. The ruler is 15 cm long.

Hydration and expansion of salts such as gypsum or halite (table salt) within pore spaces also cause spalling and rock disintegration [Photograph 3.7]. Salts can expand severalfold when hydrated. Repeated wetting and drying, as well as the growth of salt crystals due to evaporation of fluids in near-surface fractures and pore spaces, can gradually pry rocks apart in arid climates and in coastal settings where spray covers rocks in salt [Photograph 3.8]. Salt weathering may be an important mechanism of rock disintegration in the dry desert landscapes on Mars.

Chemical Weathering

Chemical weathering involves breaking of chemical bonds—metallic, ionic, and covalent. The corresponding principal weathering processes are electron exchange (oxidation-reduction), solution (ionization), and ion exchange (as in acid attack). Because many rocks are dominated by a mix of ionic and covalent bonds, solution and acid attack are major weathering processes. Hydration and dehydration are also important weathering mechanisms for certain rock types and in certain environments. Chemical weathering is essential for the biosphere; vegetation needs calcium, magnesium, potassium, and phosphorus to thrive. Nutrients derived from minerals are cycled through ecological systems because of the slow pace of weathering or the depleted nutrient status of surficial materials (especially in elements such as phosphorus, which, if not present in sufficient concentrations, can limit plant growth). Such cycling is of particular importance in areas with deeply weathered, nutrient-depleted soils.

Chemical weathering occurs because the minerals in rocks form in equilibrium with deep-Earth conditions—high pressure, high temperature, and low oxygenation—quite different from those at Earth's surface. When these minerals are exposed to the relatively cool, wet, and oxygen-rich surface conditions, they are vulnerable to chemical decomposition and transformation. Chemical weathering involves reactions that change primary rock-forming minerals into **secondary minerals**, such as clays. In the process, some elements are lost in solution to surface water and groundwater.

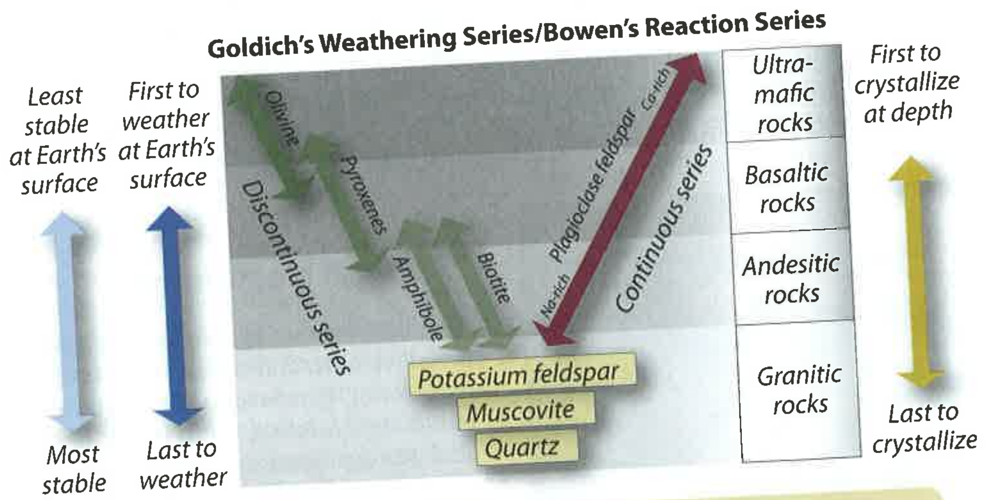
The primary weathering agent is rainwater that percolates into the ground and promotes chemical weathering because it contains dissolved ions gained from the atmosphere and from the soils through which it moves. The bipolar water molecule can be a potent solvent, given time. The metabolism of soil microorganisms and decay of organic matter enhance weathering as they add organic acids to water moving through soils. Root respiration and microbial oxidation enrich the soil atmosphere in CO_2, which makes carbonic acid when dissolved in water, thus lowering pH and increasing the aggressiveness of soilwaters. Sulfuric and nitric-acid weathering are important in some areas.

Processes of chemical weathering include **solution, oxidation, reduction, hydrolysis, ion exchange,** and the formation of new, secondary minerals like clays and hydrous oxides, which are more stable near Earth's surface than are their parent minerals. The end results of chemical weathering depend on a variety of interacting factors, including the composition and texture of the parent material and the chemical, physical, and biochemical processes acting in a particular environment. The mobility and stability of the secondary minerals and solutions produced depend on environmental conditions such as pH, redox potential, and temperature.

Mineral Stability

The general susceptibility of rock-forming minerals to weathering is the inverse of the sequence in which they form deep within the Earth. Minerals that formed at the highest temperatures and pressures are farthest from equilibrium at surface conditions and are therefore most susceptible to weathering when exposed to the elements. Among the common silicate rock-forming minerals, olivine and pyroxene are most susceptible to weathering, followed in order of decreasing susceptibility to breakdown by amphibole, biotite, muscovite, and quartz [**Figure 3.3**]. This progression, known as **Goldich's Weathering Series**, is the opposite of the order of crystallization as magma cools, familiar to geologists as **Bowen's Reaction Series**.

Under similar environmental conditions, rocks composed of more **mafic** iron- and magnesium-rich minerals (olivine, pyroxene, amphibole, and biotite) will weather faster than those composed of muscovite and quartz. But it is not always as simple as this. Feldspars have a range of compositions and thus a range of weathering susceptibilities. Due to the relative weakness of ionic bonds, silicates with complicated mineral structures break down more readily than do those with simpler covalent-bond structures like quartz (SiO_2) or zircon ($ZrSiO_4$), a very stable mineral even though it has a very high melting temperature. Both crystal complexity and formation conditions are central to mineral stability.



Minerals that cool at the highest temperature (such as olivines and calcium-feldspars) are least stable in relatively cold, moist surface environments. Some minerals, such as the feldspars, form continuous solid-solution series with a variety of compositions having different percentages of sodium and calcium. Other minerals, such as amphiboles and biotites, have discontinuous series with minerals having different and discrete compositions dependent on the temperature of crystallization.

FIGURE 3.3 Goldich's Weathering Series. Minerals formed at high temperatures tend to be less stable at and near Earth's surface than minerals formed at low temperatures. This relationship, known as Goldich's Weathering Series, is the inverse of Bowen's Reaction Series, used to describe the order in which minerals crystallize from rock.

The mobility of cations in rock-forming minerals varies greatly and influences the relative ease and order in which weathering strips cations from rocks and secondary minerals, with the sequence from most to least mobile proceeding as Ca^{2+}, Na^{+}, $Mg^{2+} > K^{+} > Fe^{2+} > Si^{4+} > Fe^{3+} > Al^{3+}$. The most mobile cations (Ca^{2+}, Na^{+}, Mg^{2+}) are readily stripped from mineral surfaces, tend to remain in solution, and are the first to be lost from rocks as they weather. The least mobile cations (Si^{4+}, Fe^{3+}, and Al^{3+}) are relatively insoluble and become concentrated in residual soils over time as weathering strips away more mobile elements.

Oxidation and Reduction

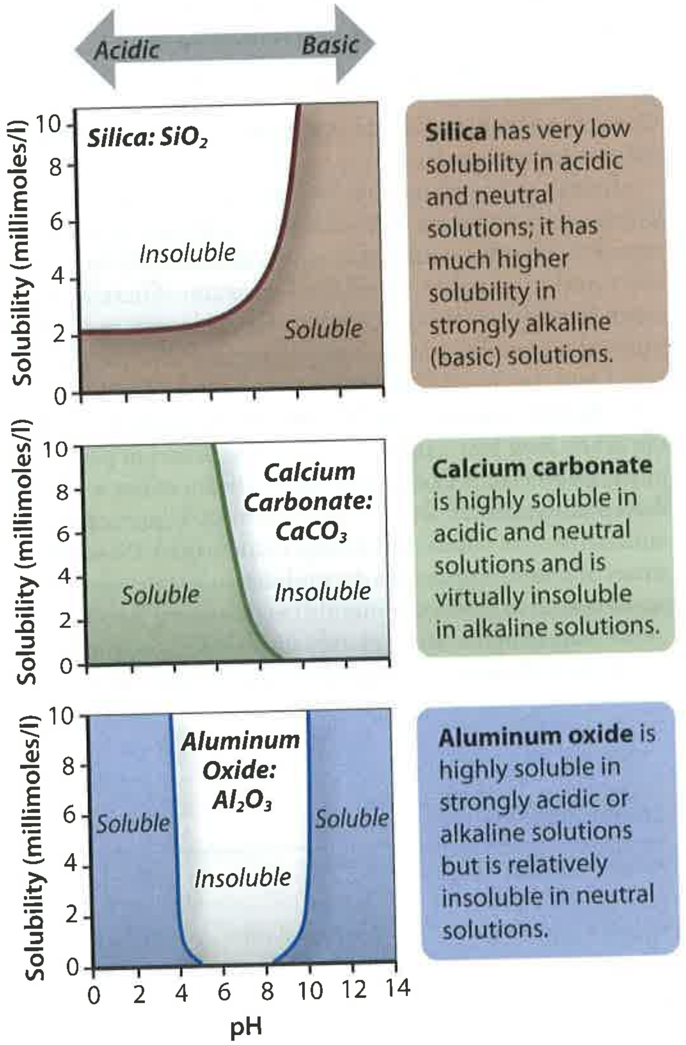
In oxidation, an element such as iron loses an electron to a receptor, often an oxygen ion—for example, when iron rusts. Conversely, reduction is defined as gaining an electron. Free oxygen is rare at crustal depths where rocks form, but abundant at Earth's surface. Certain rocks and rock-forming minerals oxidize when they are exposed to well-oxygenated soilwater, directly to the atmosphere, or to gases in soil pores. Reducing conditions in oxygen-poor waters with lots of organic matter, found for example in swamps and peat bogs with high seasonal water tables, generally prevent oxidation, retard organic decay, and slow weathering. When soils alternate between saturated and unsaturated conditions, a blotchy color pattern known as **mottling** develops—with gray colors due to reduced iron alongside reddish colors due to oxidation.

Oxidizing potential is expressed in terms of **redox potential (Eh)**, the availability of free oxygen, which is greatly influenced by the amount of dissolved organic matter in pore fluids. Soils typically have Eh high enough to oxidize most common elements, but iron, manganese, and sulfur are especially prone to rapid oxidation and typically occur as red, black, and yellow coatings in soils. Redox potential exerts a substantial influence on ion mobility, and oxidation is often the first form of weathering to alter freshly exposed rock surfaces. Most oxidized elements form hydroxides and are thus relatively insoluble and immobile, whereas most reduced elements are far more soluble and thus mobile.

Over time, rinds of oxidized material form on the surfaces of outcrops, boulders, and cobbles (see Photograph 2.6). When rocks containing common iron-bearing carbonates, sulfides, and silicates (such as olivine and biotite) oxidize, they become susceptible to additional physical weathering. Oxidation produces relatively insoluble ferric oxides like hematite (Fe_2O_3) and oxyhydroxides like goethite ($FeO(OH)$), which color soils and weathered rock various shades of reddish or yellowish brown. By the same token, oxygen-starved conditions (such as those under stagnant water rich in decomposing organic matter) reduce iron and manganese, which allows them to be dissolved and leached if the water drains or is flushed from the soil. Portions of the soil that are frequently subject to reduction processes are recognizable by their gray-blue colors.

Solution

The flow rate and acidity of pore water are two of the most important factors influencing the amount of dissolution from soil, sediment, or rock in a given weathering environment. In particular, pH strongly affects the solubility of most elements [**Figure 3.4**]. Rainwater is slightly acidic ($pH = 5.7$) from dissolved atmospheric CO_2 (which forms carbonic acid, H_2CO_3), and chemical weathering and biologic processes (respiration and decomposition) further alter the pH of water moving through soil and regolith



Silica has very low solubility in acidic and neutral solutions; it has much higher solubility in strongly alkaline (basic) solutions.

Calcium carbonate is highly soluble in acidic and neutral solutions and is virtually insoluble in alkaline solutions.

Aluminum oxide is highly soluble in strongly acidic or alkaline solutions but is relatively insoluble in neutral solutions.

FIGURE 3.4 Solubility Diagrams. Solubility relationships with pH are different for each mineral. The three geomorphically important minerals shown here have very different solubility-pH relationships.

(**Table 3.1**). In weathering zones with active groundwater circulation, freshwater comes in contact with parent material, and weathering continues as leaching removes dissolved constituents. Slowly circulating pore waters retard dissolution as the amount of dissolved ions in solution approaches an equilibrium or saturated concentration.

Natural soilwater tends to be slightly acidic, due to dissolution of soil carbon dioxide in water to produce carbonic acid (H_2CO_3). Carbonic acid is not a strong acid, but it is extremely abundant because it forms through the carbonation reaction wherever water encounters CO_2:

$$H_2O + CO_2 \longleftrightarrow H_2CO_3 \qquad \text{eq. 3.1}$$

Decay of organic matter together with respiration of soil invertebrates, bacteria, and root systems can elevate CO_2 concentrations in soil pores so that they are 10 to 100 times greater than atmospheric concentrations. Thus, carbonation is a particularly important factor in heavily vegetated areas. Cold temperatures also favor formation of carbonic acid in soilwater because the solubility of CO_2 is inversely proportional to temperature, as is true of most gases.

In solution, carbonic acid partially disassociates into hydrogen (H^+) and bicarbonate ions (HCO_3^-):

$$H_2CO_3 \longleftrightarrow H^+ + HCO_3^- \qquad \text{eq. 3.2}$$

Consequently, bicarbonate is the most common anion in natural groundwater.

Atoms exposed on the mineral surfaces of rock and soil particles tend to have surfaces with net positive or negative charges and react with dissociated hydrogen (H^+) and hydroxide (OH^-) ions in water. These interactions break bonds, effectively disassociating individual mineral molecules and causing exchanges that release cations from the mineral surface into solution. Mineral structures become unstable and vulnerable to further weathering when they lose cations, so initial weathering promotes more weathering. **Congruent dissolution** occurs when all the constituents of an individual molecule are separated and remain in solution. During **incongruent dissolution**, some of the released ions recombine to create new compounds and secondary minerals (as discussed later).

Dissolved material may remain in solution and move along with flowing groundwater, may reprecipitate elsewhere, or it may enter streams and rivers and eventually reach the ocean. Ocean salinity results from the long-term delivery of dissolved material in stream water. Most common elements are soluble to some degree in both rainwater and soilwater. Consequently, water circulation promotes solution by introducing fresh water that removes dissolved ions from mineral surfaces.

TABLE 3.1

Common pH Values

Normal rainfall	5.7 (mostly due to carbonic acid)
Acid rain	< 5.6
Normal soil (moist climate)	≈ 4.6
Vinegar	2.4
Coca-Cola	≈ 2.3

The dissolution of calcite ($CaCO_3$, calcium carbonate) is a particularly important chemical weathering reaction that occurs in the presence of carbon dioxide dissolved in water (carbonic acid, see equation 3.1) and introduces bicarbonate ions (HCO_3^-) into solution. The resulting reaction is expressed as

$$CaCO_3 + H_2CO_3 \longleftrightarrow Ca^{2+} + 2(HCO_3^-) \qquad \text{eq. 3.3}$$

The carbonate dissolution reaction is reversible. An increase in CO_2 concentration within soil gases, a decrease in pH, or dilution will drive the reaction to the right (as written above); carbonate dissolution will increase, and the bicarbonate concentration in groundwater will go up. This effect helps percolating water erode fractures and form extensive cave systems typical of regions underlain by carbonate rocks (limestone or dolomite). Conversely, decreased CO_2 concentration, increased pH, or evaporation will drive the reaction to the left and favor precipitation of calcium carbonate ($CaCO_3$). It is this reaction that deposits stalagmites and stalactites in caves, as well as calcite in desert soils.

Mineral dissolution can affect near-surface process rates by increasing pore space and thus increasing the volume of percolating water, soil acids, oxygen, and bacteria moving into and through the regolith. Calcite and salts are readily dissolved in water, so carbonate rocks and evaporites are particularly susceptible to dissolution, especially in regions with abundant precipitation. In contrast, quartz and most other rock-forming silicate minerals are not very soluble at typical Earth surface conditions, leading to slow rates of dissolution in most environments.

Iron and aluminum oxides are virtually insoluble under oxygenated soilwater conditions, so these compounds are typically left behind while more soluble, mobile material is depleted. Consequently, the abundance of iron and aluminum oxides increases over time as rocks and sediment are exposed to weathering. The red soils of unglaciated, stable continental regions are an example of how these oxides can accumulate over time.

Hydrolysis

Hydrolysis is a chemical reaction in which water molecules (H_2O) are split into protons (H^+) and hydroxide anions (OH^-) that react with primary rock-forming minerals to form new compounds (secondary minerals). Hydrolysis is an important chemical weathering process that acts to break rocks apart and transform silicate minerals (the most widespread minerals in Earth's crust) into weathering products. The process is critical in generating regolith and making clay minerals that are important for soil fertility. In hydrolysis reactions, mineral cations are released into solution and replaced by hydrogen (H^+), producing a new mineral. This process results in incongruent dissolution, the irreversible transformation of aluminosilicate minerals, like feldspars and micas, into various clay minerals or oxides.

When carbonic acid dissociates to form an "acid" or protons, the resulting weathering of aluminosilicate minerals consumes CO_2, drawing down atmospheric CO_2 levels and thus helping to cool global climate through the general reaction:

$$\text{aluminosilicate} + H_2CO_3 + H_2O \rightarrow$$
$$\text{clay mineral} + \text{cations} + HCO_3^- + H_4SiO_4 \qquad \text{eq. 3.4}$$

Earth's long-term climate is thus mediated by organic matter burial (which sequesters carbon in geologic materials), and silicate weathering, which consumes CO_2 (producing bicarbonate). Glaciations and the anthropogenic contribution to atmospheric CO_2 are short-term perturbations of these long-term geologic controls on global atmospheric composition and thus climate.

Once secondary minerals (clays) are formed, further weathering can strip additional cations and can convert secondary aluminosilicates into other, more cation-depleted clays or oxides. Each step in the weathering of clay minerals strips additional cations, sequentially reducing the complexity of mineral structures.

Clay Formation

Clay minerals are both a product and a player in processes of hydrolysis and hydration. Unlike most primary minerals, with the exception of quartz, secondary minerals such as clays and hydrous oxides are chemically stable under Earth surface conditions. They become a major constituent in soils because their relative stability and immobility leaves them as common in situ by-products of weathering.

Most clay minerals are layer silicates composed of sheets of alumina octahedra (an atom of aluminum bonded to six atoms of oxygen) or silica tetrahedra (an atom of silica bonded to four atoms of oxygen) [**Figure 3.5**]. Aluminum (and, rarely, other elements) may substitute for silica in the tetrahedral layers. Iron and magnesium commonly substitute for aluminum in the octahedral layers. These sheets generally are bonded together in either a 1:1 structure (T-O), in which each layer of silica tetrahedra (T) is paired with a layer of alumina octahedra (O), or in a 2:1 structure (T-O-T), in which each octahedral layer is sandwiched between two tetrahedral layers. These building blocks are themselves interlayered and bound together by shared ions between the sheets. Layer architecture (1:1 versus 2:1) and ionic substitution within and between the sheets determine the physical properties of different clay minerals. For example, substitution within the layers makes the difference between illite and smectite clays.

Adjacent layers in **kaolinite**, a clay mineral with a 1:1 structure, are held together by hydrogen bonds that are strong enough to prevent cations or water from entering the spaces between the sheets. Because it has few exchangeable cations held between its layers, kaolinite does not swell much when wetted, and it has low plasticity (and thus little capacity to be molded). Clays with 2:1 layer structures exhibit much more variability in the chemical composition of their octahedral sheets (typically due to substitution of Fe^{2+} and Mg^{2+} for Al^{3+}) and in the abundance and type of ions present between layers. **Smectite** clays, like **montmorillonite**, have weak bonds between the silica layers, which allow water and ions to penetrate the crystal structure readily. Also known as swelling clays, smectites expand readily upon wetting and are a main component of expansive soils. **Illite**, another common clay mineral in soils, has a strongly bonded 2:1 structure. The cations between its layers are tightly held, so it has less swelling potential than smectite.

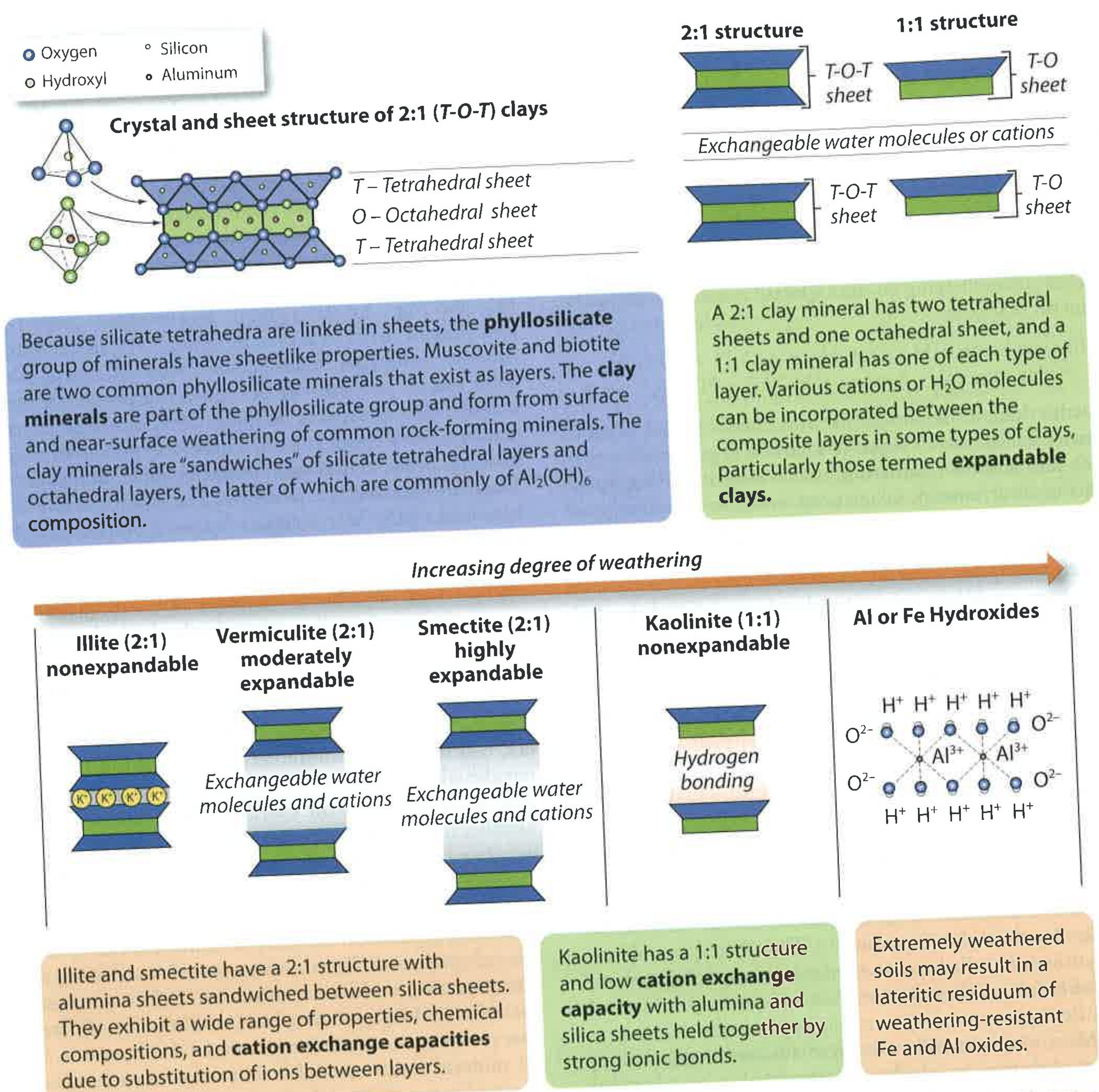
Weathering of secondary minerals involves stripping off layers of the silicate structure. The modification of muscovite (mica) to illite (clay), both of which consist of T-O-T "sandwiches," involves removal of the interlayer cations. Extreme weathering conditions can go beyond leaching of the intermediary cations and remove one of the two T layers, leaving a T-O sequence of silicate layers, resulting in kaolinite clay. Stripping the remaining tetrahedral layer leaves a basic octahedral layer of aluminum hydroxide, gibbsite ($Al(OH)_3$). In general, smectites weather to illites and ultimately become kaolinite. Deeply weathered soils generally have high concentrations of kaolinite and oxides.

The particular minerals produced by weathering processes depend on the parent materials and environmental conditions. But the sequence of weathering reactions, starting with the conversion of the rock mineral potassium feldspar (orthoclase) to illite clay, illustrates how progressive weathering changes clay minerals from 2:1 mineral structure to 1:1 mineral structure and then eventually to oxides. Orthoclase incongruently reacts with carbonic acid to produce the 2:1 clay mineral illite:

$$\begin{array}{ccc} 3KAlSi_3O_8 + 2H_2CO_3 & \rightarrow & KAl_3Si_3O_{10}(OH)_2 + 2K^+ \\ + 12H_2O & & + 2HCO_3^- + 6H_4SiO_4 \\ \textit{Orthoclase (feldspar)} & & \textit{Illite (clay)} \end{array} \qquad \text{eq. 3.5}$$

Further intensive weathering of illite produces the 1:1 clay mineral kaolinite, which consists of just hydrogen, aluminum, silica, and oxygen:

$$\begin{array}{ccc} 2KAl_3Si_3O_{10}(OH)_2 & \rightarrow & 3Al_2Si_2O_5(OH)_4 + 2K^+ \\ + 2H_2CO_3 + 3H_2O & & + 2HCO_3^- \\ \textit{Illite (clay)} & & \textit{Kaolinite (clay)} \end{array} \qquad \text{eq. 3.6}$$



Because silicate tetrahedra are linked in sheets, the **phyllosilicate** group of minerals have sheetlike properties. Muscovite and biotite are two common phyllosilicate minerals that exist as layers. The **clay minerals** are part of the phyllosilicate group and form from surface and near-surface weathering of common rock-forming minerals. The clay minerals are "sandwiches" of silicate tetrahedral layers and octahedral layers, the latter of which are commonly of $Al_2(OH)_6$ composition.

A 2:1 clay mineral has two tetrahedral sheets and one octahedral sheet, and a 1:1 clay mineral has one of each type of layer. Various cations or H_2O molecules can be incorporated between the composite layers in some types of clays, particularly those termed **expandable clays.**

Illite and smectite have a 2:1 structure with alumina sheets sandwiched between silica sheets. They exhibit a wide range of properties, chemical compositions, and **cation exchange capacities** due to substitution of ions between layers.

Kaolinite has a 1:1 structure and low **cation exchange capacity** with alumina and silica sheets held together by strong ionic bonds.

Extremely weathered soils may result in a lateritic residuum of weathering-resistant Fe and Al oxides.

FIGURE 3.5 Clay Weathering Sequence. Increasing degrees of weathering result in different, indicative types of secondary clay minerals as interlayer cations are removed.

Weathering that strips the silica from kaolinite leaves a residue of aluminum oxide, or gibbsite ($Al(OH)_3$):

$$\underset{\textit{Kaolinite (clay)}}{Al_2Si_2O_5(OH)_4} + 5H_2O \rightarrow \underset{\textit{Gibbsite (oxide)}}{2Al(OH)_3} + 2H_4SiO_4 \quad \text{eq. 3.7}$$

Secondary minerals, like kaolinite and gibbsite, can also form directly from primary minerals, depending on the parent material and specific environmental conditions.

Hydration

Hydration describes the process by which minerals combine with water or hydroxide ions (OH^-) to form hydrated compounds. Hydration is another way that primary minerals are converted to secondary minerals. Common hydration reactions include the conversion of anhydrite ($CaSO_4$) to gypsum ($CaSO_4 \cdot 2H_2O$), and the formation of relatively insoluble iron and aluminum

hydrous oxides, such as limonite ($FeO(OH) \cdot nH_2O$) in regions of intense tropical weathering such as the Amazon basin.

Chelation

Chelation is the process through which relatively immobile metal ions, such as iron and aluminum in soils, are mobilized by soluble organic compounds that form ring structures around metal ions. Chelation is facilitated by organic acids (particularly fulvic acid) produced by the breakdown of leaf litter, soil organic matter, and lichens. Iron and aluminum mobilized by chelating agents may be carried along in solution with soilwater flow until concentration changes or microbial actions break down the chelating agent, causing the metal to reprecipitate. Conifer needle decay in cool, moist environments is a common source of chelating agents, leading to the stripping of iron and other elements from the upper portions of forest soils.

Cation Exchange

An important outcome of chemical weathering is the ability of secondary minerals and organic matter to exchange cations with soilwater, thereby making vital nutrients (Ca^{2+}, Mg^{2+}, K^+) available to plants. Clays and organic compounds vary in their ability to adsorb and release cations, a property called **cation-exchange capacity.** Rates of ion exchange are controlled by soil cation-exchange capacity as well as by the ionic composition and pH of soilwater. Strongly acidic (low pH) pore fluids allow H^+ to substitute for and replace metal cations. As hydrogen ions exchange places with nutrient cations held on a clay surface, the number of potentially exchangeable cations decreases.

The degree to which the exchange sites are occupied by exchangeable cations other than H^+ and Al^{3+} is called **base saturation.** Repeated cation exchange progressively lowers base saturation in clay minerals by removing cations from between clay sheets. A clay with a high cation-exchange capacity but a low base saturation has had its balancing cations stripped out and replaced by hydrogen ions through substantial chemical weathering. Such clays are typically found in tropical regions with high temperature and rainfall, such as Hawaii. The progressive loss of exchangeable cations, which are needed for plant growth, reduces soil fertility. Older, more intensively weathered soils have low base saturation and are thus proportionately less fertile.

Soils

To a geomorphologist, soil is a layered residue of decomposed rock and organic matter left by weathering over an extended period of time. Soils have mineralogical, biological, and morphological characteristics that are distinct from those of its parent material. Other disciplines have their own definitions of soil—engineers generally consider soil to be any loose material above bedrock (if it can be dug, it is soil), and agronomists often consider soil anything in which one can grow a plant.

A tremendous variety of soils has resulted from the wide array of parent materials, climates, vegetation, and weathering environments around the world. Due to differences in soil-forming processes, patterns in the distribution of soil types mirror different climate zones and geologic settings. Soils also reflect differences in landscape history, and in some cases their characteristics can be used to infer past climate changes, periods of landscape stability, or the nature of ancient environments. Soils are generally unconsolidated, but calcite ($CaCO_3$), silica (SiO_2), and iron oxide (Fe_2O_3) cements that develop in some soils add strength and allow formation of distinct erosion-resistant landforms.

Soil-Development Processes

Soil development occurs by a variety of processes through the addition, loss, transformation, and translocation (movement) of material within a soil profile. These processes include the accumulation of organic matter at the ground surface and its subsequent integration down into the soil, decomposition of primary minerals into secondary minerals, the leaching of particular organic and mineral constituents from the parent material, and the movement of material leached or washed from upper soil layers deeper into the soil. Soil profiles lose material by either erosion or dissolution. Erosion by surface water flow and wind removes material from the top of the soil profile, and dissolution removes mineral material from within the profile.

New material may be added to a soil profile from sources below, within, or above the ground surface. Dustfall adds material to the top of a soil profile. The breakdown of bedrock adds mineral matter to soil from below. Organic matter is added to soil profiles from above when leaves and other organic debris fall to the ground and the remains of once-living plants and animals become mixed into the soil, or roots die within the soil. In addition, floods can deposit fluvial sediment on floodplains, wind can add dust to the top of a soil profile, and volcanic eruptions contribute ash to soils.

Soil scientists refer to the movement of material from one part of a soil profile to another as translocation. Dissolved matter may move within the soil or leave the soil in groundwater, a process known as **leaching.** Water percolating down through the upper soil may also physically entrain clays, silt, and colloids and redeposit them deeper in the soil. Precipitation infiltrating into a soil dissolves material from the upper layers. Soilwater flow carries the dissolved material downward and may deposit it in lower layers as water evaporates and solute concentration increases or as redox, pH, or gas contents change. The process of transporting clays and mobile ions out of a layer of soil is called **eluviation;** washing of material into a layer of soil is called **illuviation.**

Soils can be residual, colluvial, or cumulic. **Residual soils** develop in place on stable Earth materials including weathered rock and alluvial deposits, for example, river terraces. **Colluvial soils** move downslope and thus are typically well stirred (also known as turbated) by both biological and physical processes. **Cumulic soils** accumulate slowly enough, or episodically enough, that material added at the top of the soil profile weathers to some degree before being buried. Examples of such inputs include wind-blown dust or volcanic ash and flood-deposited silts on floodplains. Repeated, episodic deposition results in stacked soils, with multiple buried A and B horizons evident in soil pits.

Factors Affecting Soil Development

The factors that control soil development are climate, organisms (biological activity), topography (including slope orientation and steepness), parent material, and time. Climate and time are the predominant factors in soil development at a regional scale, but geologic factors greatly influence local soil characteristics. Other than time, soil-development factors are to some degree interdependent. Climate, for example, influences vegetation, and topography and parent material influence one another. Still, considering these factors independently helps to organize thinking about soil development.

Temperature and moisture conditions strongly influence rates and styles of weathering and thus soil development, because weathering reactions proceed faster at higher temperatures and where more water is available. In hot, wet tropical regions, intense weathering creates thick soils depleted of most of their original constituents, leaving behind relatively stable weathering products like kaolinite clay and aluminum and iron oxides and hydroxides. In contrast, soils formed slowly in cold, dry polar regions tend to be relatively shallow (thin) and have only minimal chemical alteration. In mid-latitude temperate regions that have moderate temperatures, precipitation, and evaporation, the resulting soils typically have intermediate depths and development intensities. In arid regions, low precipitation rates and high evaporation rates cause development of carbonate- or salt-enriched soils because dissolved constituents (usually derived from weathered, wind-blown dust) percolate into the soil, reprecipitate, and accumulate over time.

Climatically driven differences in soil are most pronounced on level, stable slopes where soil **residence time** (the time between a typical mineral grain's initial weathering from the underlying rock and erosion from the slope) is sufficiently long to reveal the influence of climate.

Organisms, including microbes, plants, and animals, affect the composition and behavior of soil. Vegetation influences soil composition directly when the growth of roots and tree-throws break up bedrock and mix it into the soil [**Photograph 3.9**]. The type and amount of vegetation growing in a soil influences erosion resistance and the ratio of surface water runoff to infiltration after precipitation events. Root size, density, and depth can greatly influence soil development through the location of organic inputs, whether at the surface or at depth. Vegetation also helps keep soil in place and shades the soil, helping it retain moisture and increasing rates of chemical transformation. Animals contribute to soil development and mixing through burrowing activity. In tropical regions, ants and termites can build great mounds that move soil vertically. In more temperate regions, burrowing mammals, such as gophers, churn the regolith and mix weathered rock fragments into the soil. Microbes, fungi, and other decomposers are central to the breakdown of organic matter and the cycling and recycling of key nutrients derived from rock weathering.


D. Montgomery

PHOTOGRAPH 3.9 Tree-Throw. Tree-throw stirs soils, breaking up underlying rock and moving it downslope. This tree-throw has ripped up rock held within the tree's roots.

Topography greatly influences rates and styles of soil development through its control on rates of soil erosion and patterns of soil moisture. Gentler slopes may foster greater plant growth and accelerated soil development because they retain soil moisture better than steeper slopes. In upland regions, soils are better developed on gentle slopes, while shallow soils and even bare rock characterize steep slopes. Steeper slopes also promote faster rates of downslope soil movement, leading to higher soil turnover rates and therefore younger, less well-developed soils.

Slope aspect (the direction a slope faces) can affect microclimate and local soil moisture. In some landscapes, different types and densities of vegetation and soils develop on north-facing versus south-facing slopes because systematic differences in the amount of direct sunlight translate into differences in soil moisture. North-facing slopes in the Northern Hemisphere have lower evaporation rates, retain snow cover longer in the spring, and tend to hold soil moisture longer into the summer growing season. Such differences may result in local differences in soil thickness, moisture, pH, and organic matter content that may influence vegetation patterns, such as the presence of grasses versus trees.

Rocks or unconsolidated sediments provide the raw parent material for soil, and the mineralogy, porosity, and permeability (e.g., fracturing) of the parent material influence the style of weathering and types of secondary minerals that form from weathering. In particular, the structure, hydraulic conductivity, and fracturing of rock and sediments greatly influence the movement of soilwater and groundwater and thus the progress of weathering into parent material.

Differences in parent material generally exert the greatest influence during initial soil development and determine clay mineralogy, chemistry, and certain physical characteristics of the soils that first develop from them. For example, in the Piedmont of the eastern United States, smectite clays in soils, developed from calcium- and magnesium-rich rocks like gabbro and dolomite, tend to have thicker, more organic-rich topsoil and redder, clay-rich subsoil than well-drained soils developed on granite; the reason is that expansive clays seal up soil structures and impede soil drainage. In tropical regions like Brazil, deeply weathered residual soils developed on silica-rich granite tend to consist mainly of kaolinite and quartz (SiO_2), whereas those formed from silica-poor basalt typically consist of aluminum oxides, like gibbsite ($Al(OH)_3$). Young desert soils are strongly influenced by parent material, particularly the contrast between soils developing in carbonates (limestone) versus aluminosilicates (like granite or basalt). As these desert soils age, the high influx of dust in many arid regions over time blurs the parent material–related differences.

Soils develop over time as pedogenic (soil-forming) processes mechanically and chemically transform parent material. Young surfaces have weakly developed soils, which are thin and show little evidence for pedogenesis. Old surfaces typically have more strongly developed, thicker soils with distinct evidence for pedogenesis, including color and soil structure. On relatively flat-lying surfaces, where erosion and deposition rates are low, soils generally remain in place and chemically evolve over long periods of time (tens of thousands to hundreds of thousands of years). Given enough time, a soil will progressively mature as weathering proceeds until it approaches equilibrium between the rate at which weathering products are removed and the rate at which they are produced. The timescale to reach an equilibrium soil thickness varies greatly. In general, rates of weathering and ion loss from soils tend to decrease over time because the supply of fresh minerals declines as pedogenesis progresses. Soil development can be interrupted and equilibrium conditions reset by a changing regional climate or by environmental changes like glaciation that scrape away some or all of the soil.

Most soils on steep slopes are poorly developed because high rates of erosion and downslope transport move soil before pedogenesis can alter parent material significantly. Another way to think about this is to consider **soil residence time**. In sloping, actively eroding landscapes where regolith production and soil erosion are in equilibrium, soil residence time is equal to the ratio of the soil thickness to the rate of erosion. For example, one meter of soil developed on a hillside that is eroding at an average rate of 0.1 mm/yr has a residence time of 10,000 years, whereas soil 0.1 m thick developed on a slope that is eroding at 1 mm/yr has an average residence time of just a century. In general, soils with longer residence times will be better developed than those with shorter residence times.

Processes and Rates of Soil Production

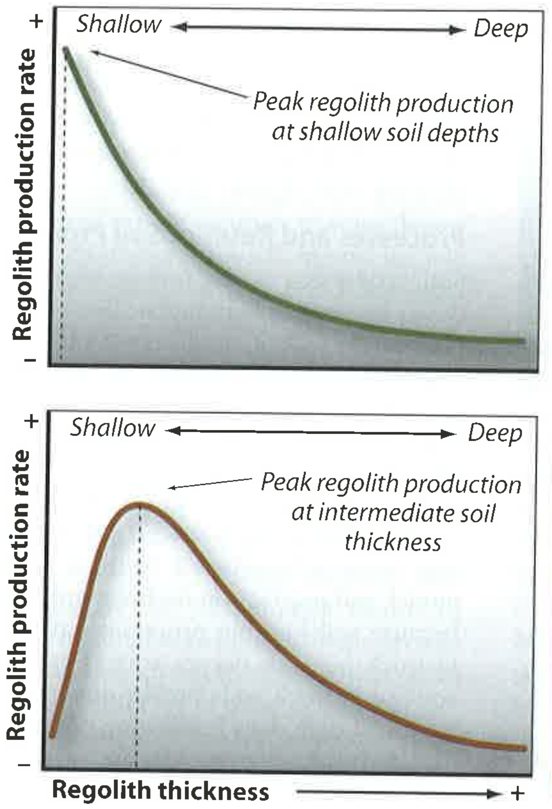
Soil production is the rate at which bedrock is broken down into erodible material. Because soil erosion occurs everywhere, albeit at different rates, soil can be maintained only if it is replaced by soil production. Soil production is critically important for maintaining soil and soil fertility in the form of fresh minerals and exchangeable cations.

Geomorphologists have proposed two general models for the relationship between soil depth and soil production from unweathered bedrock [**Figure 3.6**]. In one model, soil production declines with increasing soil depth because soil-forming processes have less access to fresh bedrock beneath deeper soils. For example, burrowing activity in thick soils breaks down less rock to produce fresh soil than does burrowing in thin soils where organisms are more likely to entrain weathered rock fragments in their digging. Likewise, once soils become deeper than the average depth of tree rooting, the roots of falling trees simply rework soil rather than ripping up bedrock. In addition, chemically percolating water becomes less aggressive as it runs out of reactive oxygen and CO_2. For soils produced by such processes, soil production rates are highest in shallow soils and decline with increasing soil thickness. In addition, development of a thick regolith can impede weathering and soil production by keeping the base of the soil profile cooler than the surficial soil where daily temperatures may vary greatly. Thus, given time, soil thickness will approach a steady state that reflects local environmental conditions in which soil production and soil erosion are matched.

In the other model, soils promote rock weathering because they hold more water and support more vegetation than bare rock outcrops. This model predicts that soil production rates are highest at some intermediate soil thickness but lower in both thin soils that hold less water and support less vegetation and in thick soils that act to protect the underlying bedrock. In landscapes that have this relationship between soil production and thickness, soil that thins to the point that soil production declines substantially will erode to expose bare rock. In this case, bedrock weathers and erodes slower than surrounding terrain, giving rise to isolated, high-standing bedrock outcrops.

Soil Profiles

Soils exhibit tremendous variability, but almost all soils have an upper zone of leaching that is composed of inorganic primary minerals mixed with organic matter in



In many landscapes, the rate of **regolith production** from bedrock weathering exhibits an exponential decrease with increasing regolith depth. In this case, the maximum rate of regolith production occurs under little to no cover because thicker regolith cover reduces the ability of regolith-producing processes to reach fresh bedrock parent material.

In landscapes where the rate of regolith production is more strongly controlled by regolith water content, the regolith production function may exhibit a maximum at an intermediate cover thickness. Regolith production rates drop where regolith is thin because the regolith/rock interface there is often dry.

FIGURE 3.6 Soil Production Functions. The rate at which soil is produced from weathered rock is a function of soil depth. Field observations and data suggest two somewhat different relationships that reflect soil-forming processes and subsurface water content.

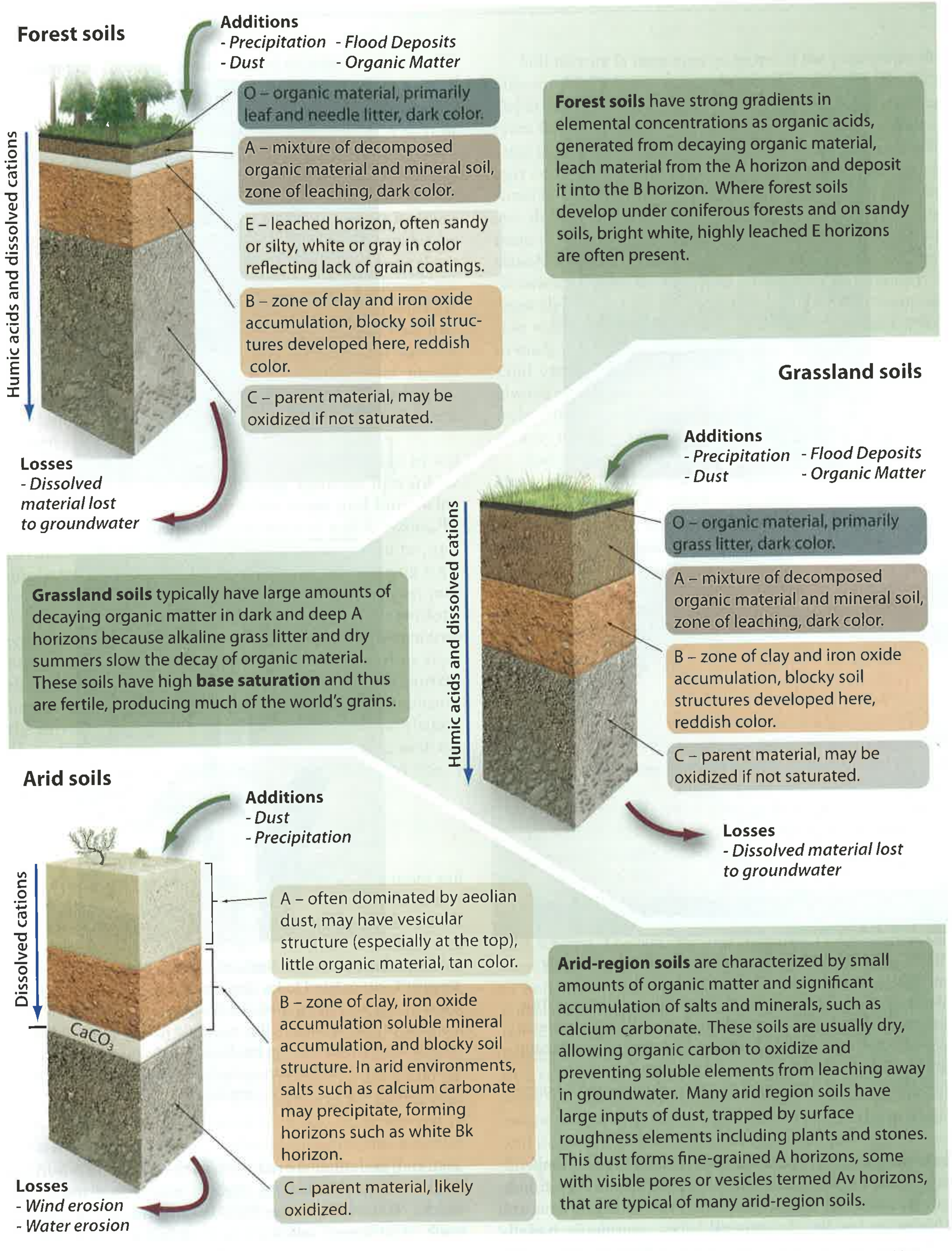
temperate climates and silt-sized dust in dry climates. Most soils also have a lower zone of accumulation, the nature of which is determined by climate. In temperate and tropical climates, clays and iron oxides dominate the accumulation zone. In dry climates, calcium carbonate ($CaCO_3$) and salts dominate. The upper zone, where losses dominate except in very dusty regions, is called the **zone of eluviation;** the lower zone where particles accumulate and chemical compounds precipitate out of solution is called the **zone of illuviation.**

As soils develop, they become differentiated into distinct **soil horizons.** These horizons are pedogenically created zones in the soil profile with unique characteristics typically found at different depths. Horizons have compositions that reflect the action and interaction of different soil-forming processes. Letters are used to identify master horizons within soil profiles reflecting major differences in physical and chemical characteristics and soil-forming processes. Soil profiles typically consist of certain associations and combinations of soil horizons that are typical of common climatic zones with distinctive environmental conditions [Figure 3.7; Photograph 3.10].

The layer of decomposing organic matter at the ground surface is known as the **O horizon.** Organic matter like leaves, twigs, pine needles, fallen logs, and animal remains dominate the volume of solids in an O horizon; inorganic minerals make up the remainder. In some soil profiles, the O horizon is a thick layer that acts as organic mulch in various stages of decomposition, for example, in wetlands. In other profiles, such as those in deserts, the O horizon is thin or missing altogether.

The next layer down, the **A horizon,** is a mixture of decomposed organic matter and mineral grains. In humid regions, the A horizon is typically dark brown to black in color because of high concentration of decomposed organic matter. Humic acids and high soil CO_2 concentrations promote decomposition of fresh mineral grains within the A horizon, and percolating water leaches material downward in the soil profile. Many A horizons have a relatively loose, friable (crumbly) texture because of soil aggregates, relatively high organic matter concentration, and pervasive disruption by plant roots. In soils developed in dry climates, the A horizons are often light tan in color and have accumulated dust rather than organic matter. Desert A horizons often have visible pores, termed vesicles. In cool, moist, acidic soils, extreme leaching removes organic matter and iron oxides, and a nutrient-poor, bleached gray **E horizon** develops below either the O or A horizon.

The **B horizon** lies below the O, A, or E horizon, and consists predominately of inorganic material in which leached elements and minerals have accumulated after



Forest soils have strong gradients in elemental concentrations as organic acids, generated from decaying organic material, leach material from the A horizon and deposit it into the B horizon. Where forest soils develop under coniferous forests and on sandy soils, bright white, highly leached E horizons are often present.

Grassland soils typically have large amounts of decaying organic matter in dark and deep A horizons because alkaline grass litter and dry summers slow the decay of organic material. These soils have high **base saturation** and thus are fertile, producing much of the world's grains.

Arid-region soils are characterized by small amounts of organic matter and significant accumulation of salts and minerals, such as calcium carbonate. These soils are usually dry, allowing organic carbon to oxidize and preventing soluble elements from leaching away in groundwater. Many arid region soils have large inputs of dust, trapped by surface roughness elements including plants and stones. This dust forms fine-grained A horizons, some with visible pores or vesicles termed Av horizons, that are typical of many arid-region soils.

FIGURE 3.7 Soil Profiles. Soil profiles and the soil horizons that typify them reflect climate and vegetation. Surface, atmospheric, and biospheric processes add mass to soil profiles. Mass is lost through groundwater transport of soluble materials and by wind- and water-induced erosion.

PHOTOGRAPH 3.10 Soil Variability. Soils are extremely variable in appearance. (a) Forest soil profile, Cape Cod, Massachusetts, formed in sandy glacial fluvial deposits. This soil has surface layers of organic and mineral sandy material (O and A horizon), a white E horizon of uncoated quartz sand grains, and a reddish B horizon. (b) Calcium carbonate collecting in a light-colored K horizon below the red B horizon in this aridisol, Escalante, Utah. (c) Red soil of Martha's Vineyard, interpreted to be developed on an older interglacial deposit, has well-developed O/A horizon over reddened B horizon. (d) Soil section in the hyperarid Paran region of the Negev Desert, Israel. At the top there is a desert pavement, below is the Av soil horizon, a clast-free horizon, the result of ~1.8 million years of dust accumulation.

percolating down from the overlying horizons. Depending on the environment, these accumulations may consist of clay minerals (Bt), carbonates (Bk), salts, or iron and aluminum oxides. In general, clay content is usually higher in B horizons than in the overlying or underlying horizons. B horizons have textures that are distinct from overlying soil horizons as well as from underlying parent material, and often are redder than the overlying A horizon because they contain less organic matter and more iron oxides. B horizons retain little to no evidence of original rock or sediment structure such as bedding or foliation. In many soils, clay films are deposited by water seeping down into the soil. These films coat grains or aggregates and line fractures and voids within the B horizon. In arid and hyperarid regions, evaporation of pore water concentrates calcium carbonate ($CaCO_3$) or increases soil CO_2, causing precipitation and development of a petrocalcic horizon, a thick erosion-resistant layer of calcium carbonate (K horizon).

The **C horizon** consists of unweathered parent material and weathered material that retains some original rock or depositional structure, including sedimentary bedding planes, depositional fabrics, and core stones surrounded by altered weathering rinds. Note that the parent material for a soil can be lithified rock or unconsolidated sediment like dune sand or river gravel. Thoroughly weathered saprolite is generally noted as a Cr horizon and consists of rock material (r) that has weathered in place. Deeply weathered soils and saprolite extending down hundreds of meters into weathered rock are common in the flat, hot, wet tropical landscapes typical of the cratons of Africa and South America.

Biological activity can disrupt the development of soil horizons and mix soil by moving material upward or downward. Bioturbation by plant roots and burrowing animals acts to disrupt development of soil horizons. In particular, the burrowing activity of worms, termites, and gophers may thoroughly mix soil profiles, dragging rock fragments to the surface and organic matter down into the soil. Significant material mixing and net downslope transport also occur when a tree falls and the material pulled up by roots gradually settles back into or near the resulting hole. The persistence of strong soil horizonation in the presence of episodic disturbance indicates the pervasive nature of soil-forming processes, bioturbation rates that are slower than pedogenic processes, and the confinement of the most rapid bioturbation to the A horizon.

Soil Classification

Soils are distinguishable by differences in one or more soil properties that reflect the degree of soil development and provide both indications of process and measures of time and relative landform stability. The most important soil properties observable in the field—horizonation, color, texture, and structure—reflect a soil's clay and organic matter content, its development, and its ability to hold moisture. For the most part, soils are classified based on such field observable characteristics.

Soil color indicates both mineral composition and organic matter content. Soils with high organic matter content are typically dark brown to black. Soils are red to yellow-brown in color in oxidizing environments and blue to black or gray in reducing environments such as wetlands. Light colors, like white, gray, and beige are generally associated with accumulations of calcite ($CaCO_3$) or leached horizons consisting mostly of residual silica (SiO_2). In the field, soil scientists use a book of color charts resembling those used at paint stores to classify soil colors based on a palette of standardized colors.

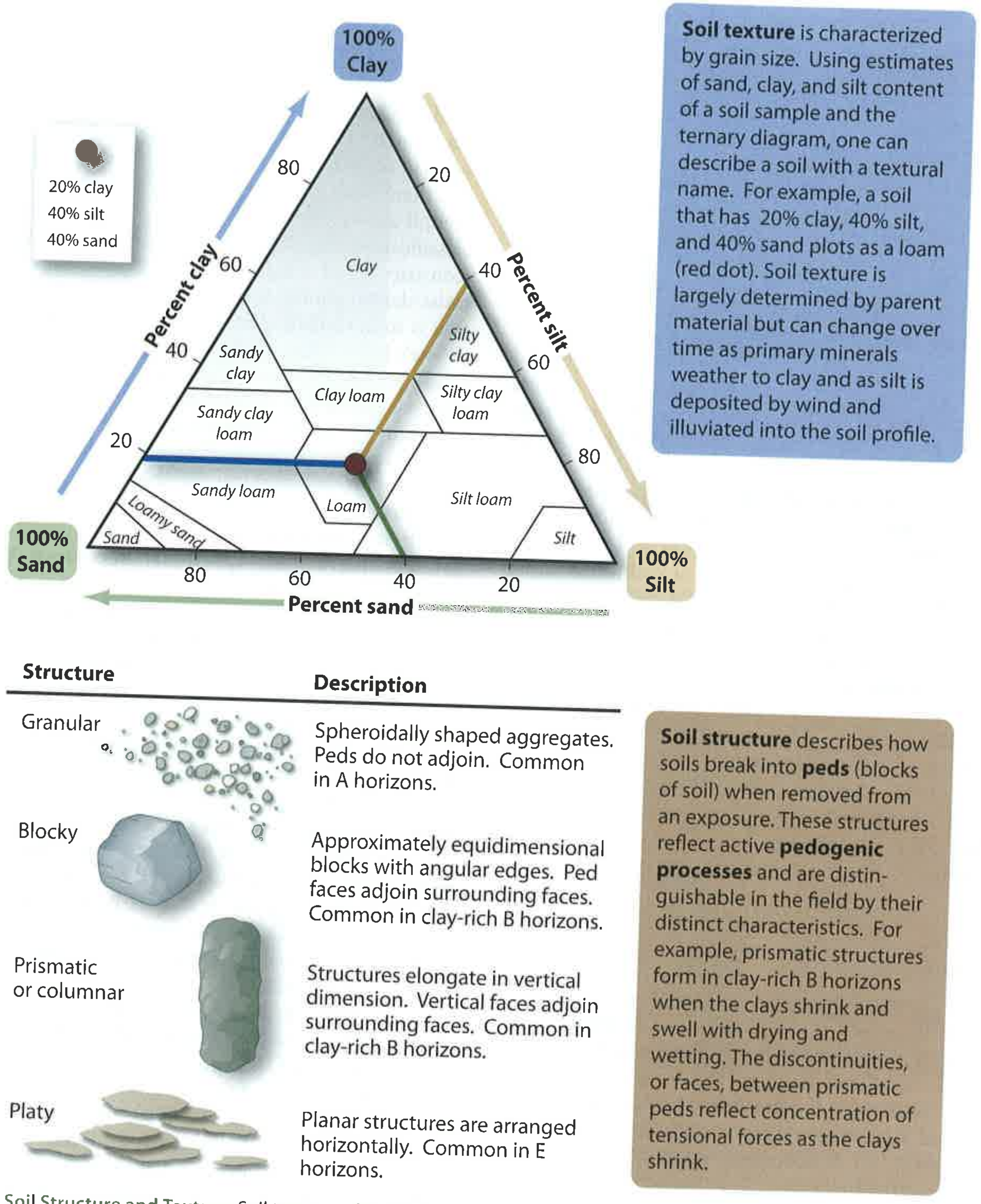
Soil texture is measured in terms of the proportions of different particle sizes in the soil [Figure 3.8]. The U.S. Department of Agriculture defines 12 standard soil textures based on the relative proportions of clay, silt, and sand (e.g., clay loam, silt loam, silty clay). Although soils may contain larger clasts (like gravel), soil properties measured in the field generally are described based on particles less than 2 mm in diameter (i.e., sand and smaller) and then, if appropriate, the modifier gravelly is added to the classification. Soil texture greatly influences the **cohesion** (consistency) of a soil as well as its **plasticity** (ability to resist deformation). A simple field test for soil composition is to moisten a sample of soil and try to make a ball in one's hand. Clay can be rolled into a "worm"; silt and sand cannot. Another simple field test—one that is not always advisable to employ—is that medium to coarse silt feels gritty between one's teeth, while clay has a smooth, creamy texture. Sand grains are visible to the naked eye.

Soil structure describes the shapes in which soil particles cluster together. Most A horizons have granular or crumb structures in which individual clumps, or aggregates, are more or less spheroidal. In contrast, B horizons often have structures, called **peds**, that are shaped like blocks, prisms, or plates with rounded or angular edges. Peds can range from pea size to fist size and reflect the deposition of clay enhanced by wetting and drying along gaps or discontinuities in the soil structure.

Soil organic matter includes undecomposed material like leaves, branches, and bones (**litter**) as well as amorphous decomposed material (**humus**). Soil microorganisms that decompose litter into humus become active at temperatures slightly above freezing (generally >5°C), and rates of microbial decomposition increase as temperatures rise. At temperatures above about 25°C, little humus accumulates because it decomposes as fast as it is produced. Consequently, soil organic matter content tends to be greatest in mid-latitude regions with mean annual temperatures between 5° and 25°C.

Soil taxonomy is one way in which scientists classify soils. **Soil orders** are the highest level of this soil classification scheme. The U.S. Natural Resource Conservation Service (formerly the Soil Conservation Service) uses a system of soil taxonomy that recognizes 12 soil orders, distinguished by diagnostic horizons and characteristics that reflect different environments, processes, and soil residence time on the landscape [Figure 3.9]. Soils can be further subdivided into 64 common suborders and then into more specific soil types, but the major soil orders are sufficient to characterize soil characteristics, parent material, and degree of weathering at a regional scale.

Young, incompletely developed soils are either **entisols** that have a faint A horizon, or **inceptisols** with a weak B horizon. These soils do not have enough horizon development to qualify as one of the more developed soil orders. Soils formed in seasonally submerged or frozen environments have distinctive characteristics. **Histosols** are organic-rich soils that form in wetlands where reducing

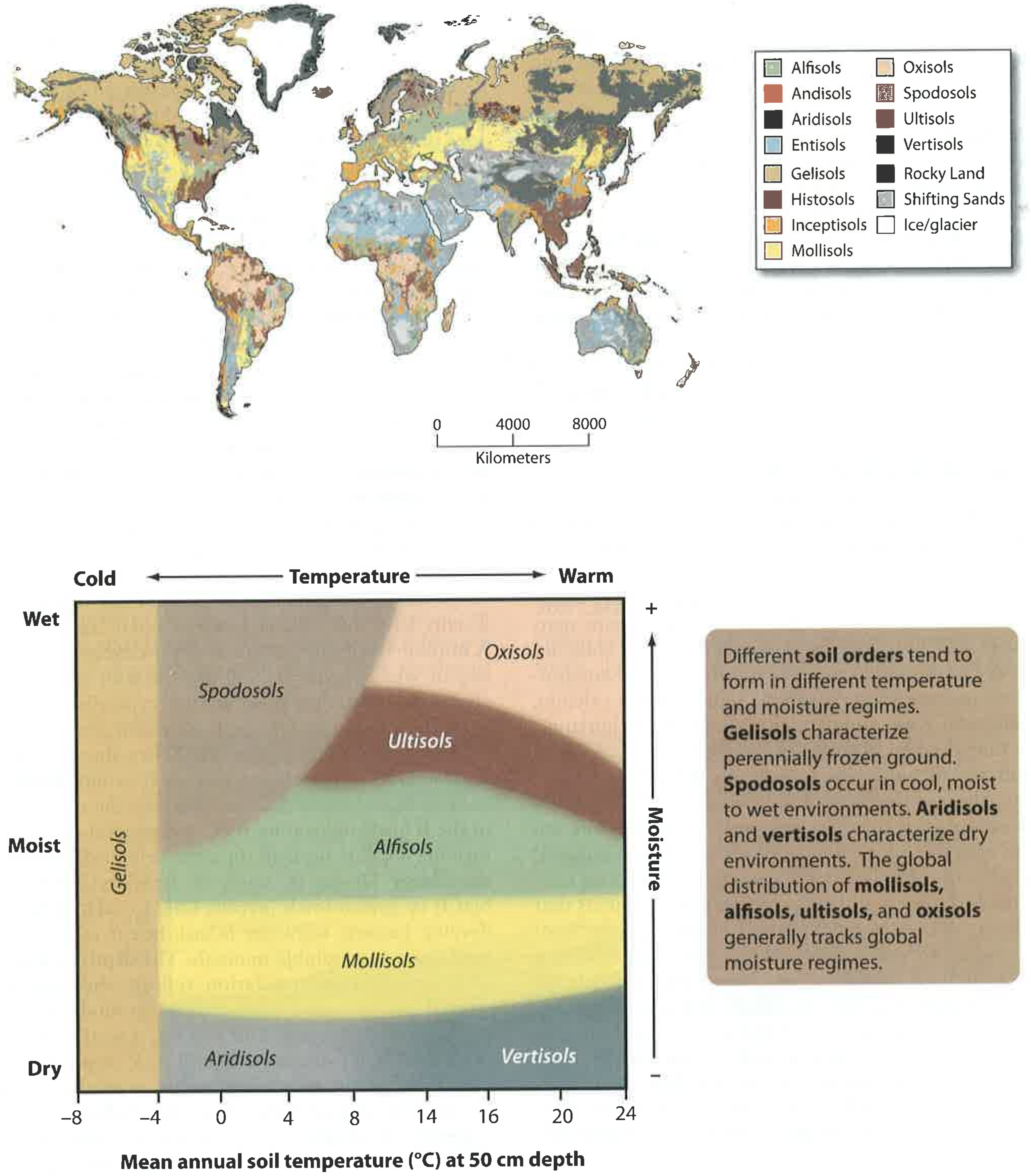


Soil texture is characterized by grain size. Using estimates of sand, clay, and silt content of a soil sample and the ternary diagram, one can describe a soil with a textural name. For example, a soil that has 20% clay, 40% silt, and 40% sand plots as a loam (red dot). Soil texture is largely determined by parent material but can change over time as primary minerals weather to clay and as silt is deposited by wind and illuviated into the soil profile.

Structure	Description
Granular	Spheroidally shaped aggregates. Peds do not adjoin. Common in A horizons.
Blocky	Approximately equidimensional blocks with angular edges. Ped faces adjoin surrounding faces. Common in clay-rich B horizons.
Prismatic or columnar	Structures elongate in vertical dimension. Vertical faces adjoin surrounding faces. Common in clay-rich B horizons.
Platy	Planar structures are arranged horizontally. Common in E horizons.

Soil structure describes how soils break into **peds** (blocks of soil) when removed from an exposure. These structures reflect active **pedogenic processes** and are distinguishable in the field by their distinct characteristics. For example, prismatic structures form in clay-rich B horizons when the clays shrink and swell with drying and wetting. The discontinuities, or faces, between prismatic peds reflect concentration of tensional forces as the clays shrink.

FIGURE 3.8 Soil Structure and Texture. Soil texture reflects the relative percentages of sand, silt, and clay in the soil. Pedogenic processes, soil materials, and water interact to create soil structure, which can be recognized when examining soil profiles in the field.



Different **soil orders** tend to form in different temperature and moisture regimes. **Gelisols** characterize perennially frozen ground. **Spodosols** occur in cool, moist to wet environments. **Aridisols** and **vertisols** characterize dry environments. The global distribution of **mollisols, alfisols, ultisols,** and **oxisols** generally tracks global moisture regimes.

FIGURE 3.9 Soil Order Classification. The distribution of soil orders, the most general level of soil classification, is controlled by climate, predominantly temperature and precipitation. [Adapted from Birkeland (1999).]

conditions restrict rates of organic decomposition. Histosols are identified by their black color, high organic matter content (more than 25 percent), thick organic-rich O and A horizons, and lack of evidence for oxidation. **Gelisols** form in polar and subpolar environments where the ground at some depth stays frozen all year. These cold climate soils include, by definition, a layer of permanently frozen soil (**permafrost**) within 2 meters of the land surface. They typically show little evidence of chemical weathering and have only an A horizon developed over permafrost. Gelisols are often structurally disrupted and mixed by seasonal freeze–thaw processes called **cryoturbation**, which complicates foundation engineering in periglacial environments.

Several types of soil reflect distinctive parent materials. **Andisols** are formed from volcanic parent material. Many andisols contain amorphous colloids from weathering of glassy volcanic ash and have high cation-exchange capacities. **Vertisols** are soils that contain high concentrations of swelling clay (smectite) and exhibit significant shrink-swell behavior; thus, vertisols do not have distinctive horizons because of the physical mixing. They exhibit large desiccation cracks when dry and tend to become quite sticky when moist. The other six soil orders are associated with particular climate or environmental settings and form under certain combinations of temperature and relative wetness.

Mollisols typically develop in temperate regions with grassland vegetation, like the prairies of the Great Plains of North America. Mollisols have well-developed, black, organic-rich A horizons that are often more than a meter thick and B horizons that are enriched in clay. Grassland soils accumulate abundant organic matter because of the dense and deep root structures of most grasses. In many grasslands, the soils are too cold for decomposition in the winter and are too dry for much decomposition in the summer. Such environments permit incomplete decay of organic material, producing organic colloids that leach down into the soil, darkening both A and B horizons. The abundant organic matter and clays with high base saturation in mollisols generally make fertile agricultural soils. The fertility of the grassland soils of the U.S. Midwest is provided by A horizons so thick and rich that they remain productive even after more than a century of agriculturally induced erosion.

Surface litter produced by forests, particularly coniferous forests, tends to be more acidic than organic matter produced by grasses. Consequently, forest soils are generally more acidic and have A horizons that are thinner and more leached of soluble ions (notably calcium, sodium, and magnesium) than grassland soil horizons. Soils that develop beneath temperate forests are called **alfisols** and **ultisols.** Alfisols and ultisols have thin A horizons. Alfisols form under deciduous forests where decaying leaves provide organic acids, and as such are common in the eastern United States. Alfisols are less weathered than ultisols; alfisols have higher base saturation. Ultisols tend to form in old and highly weathered areas that are warm and wet and were not glaciated (e.g., the Piedmont region of southeastern North America), reflecting more prolonged or intense weathering under deciduous or coniferous forests. In both alfisols and ultisols, chelating agents (organic acids from leaf and needle decomposition) leach iron, aluminum, and organic matter from the A horizon and deposit these materials in the B horizon. A distinctive, light-colored, leached E horizon between the A and B horizon characterizes many forest soils (see Photograph 3.10a).

Spodosols are forest soils that form under cooler temperature regimes and a range of moisture conditions in environments characterized by coniferous or boreal forests. Spodosols exhibit severe leaching due to acidic conditions under coniferous forests resulting from organic acids produced by decaying needles. This results in translocation (elluviation) of clays, organic matter, iron, and aluminum from the A and E horizons. Thus, the B horizon in spodosols is enriched in iron and organic matter, typically forming a hardpan horizon (**fragipan**) within the soil. Spodosols are most common in parent materials with high permeability, such as sands on a coastal plain.

Oxisols are the most deeply weathered of all soil orders and are characterized by highly oxidized soil horizons that form in hot, wet (tropical) forest environments. Abundant rain falling through a surficial organic layer leads to acidified water that promotes intensive leaching, which strips nutrients from the soil. Oxisols generally have distinctive red coloration and little organic matter accumulation because plant and animal remains decay rapidly in such conditions. Red-colored oxisols have an abundance of iron oxides (hematite and goethite) and tend to be depleted in exchangeable nutrient cations. The most heavily weathered tropical soils (**laterites**) consist of little more than residual iron and aluminum oxides such as gibbsite and kaolinite. The components, once heated and dried by the sun, harden into bricklike peds. Organic matter and nutrients in oxisols are contained in the thin O or A horizon or in the standing vegetation, posing impressive challenges to sustaining agricultural productivity when these surficial horizons or the vegetation are removed by erosion.

Aridisols are soils that develop in dry regions. They usually have low organic matter content and a minimal A horizon overlying a B horizon, the development intensity of which varies by soil age. In semi-arid and arid climates, infiltrating precipitation typically evaporates from the ground surface or from within the soil profile before reaching the water table. When this occurs, ions and particles in percolating rainwater or mobilized from the A horizon precipitate and deposit calcium carbonate in the B horizon, coating rock and mineral surfaces and forming a calcic horizon. In extremely arid settings like the Negev Desert in southern Israel, salt-rich (salic, NaCl) or gypsum-rich (gypsic, $CaSO_4 \cdot 2H_2O$) horizons develop because soilwater is insufficient to leach away even these very soluble minerals. The depth to this zone of evaporative accumulation reflects the amount of annual precipitation and shallow groundwater flowpaths. In arid and semi-arid regions, if water containing silica dissolved from fine dust and rock evaporates, silica can reprecipitate. Increasing aridity generally results in less well-developed A horizons and shallower zones of evaporite accumulation. Given enough time, however, red, clay-rich B horizons can develop in some of Earth's driest deserts.

Dustfall is important in the development of many soils. In particular, dust-influenced desert soils typically have a stone-free, silt-rich horizon just below the surface. This accumulation of fine, wind-delivered sediment is known as an Av horizon because of the presence of small vesicles (air pockets) that form when the horizon wets from rain and then dries. Sometimes, Av horizons are capped by a layer of coarse clasts known as a desert pavement (see Photograph 3.10d).

Soils and Landscapes

Soil properties and orders vary among climate zones and physiographic regions [**Figure 3.10**]. Soil orders and the degree of soil development, particularly soil thickness and organic matter content, generally track latitudinal patterns in temperature and precipitation. Deeply weathered oxisols are typical of the equatorial tropics. Aridisols are typical of mid-latitude deserts. Organic-rich mollisols and forest soils are typical of temperate latitudes. However, soils also differ within single landscapes because of variations in environmental factors, rates of erosion, and landscape history.

Soil Development over Time

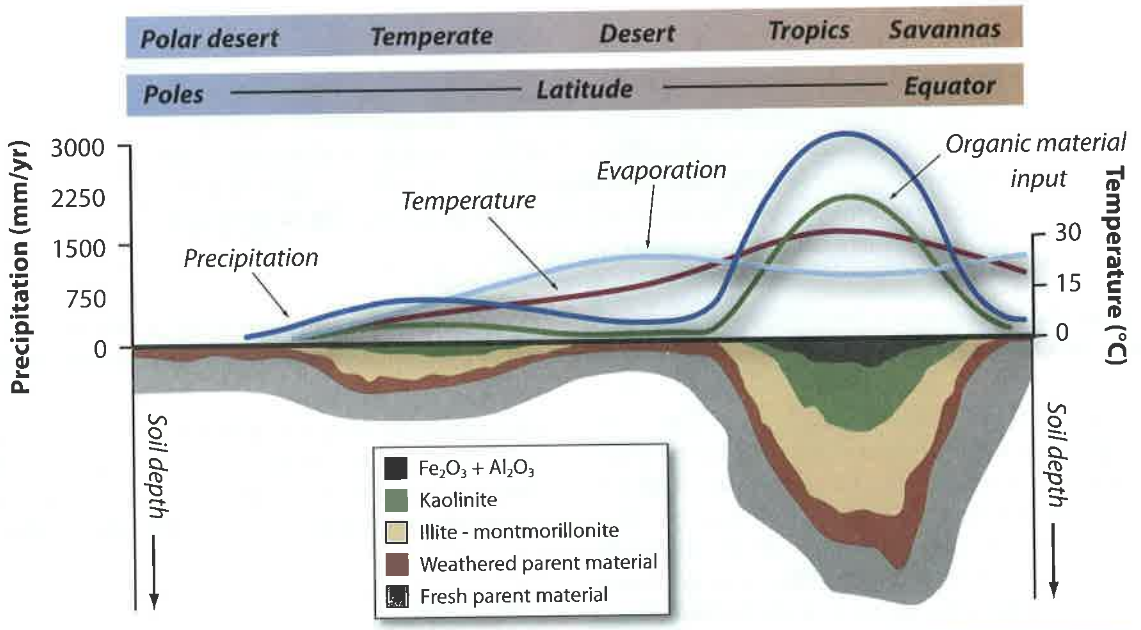
As soils develop over time, they gradually lose the physical and chemical characteristics of their parent materials and increasingly take on characteristics that reflect soil-forming processes and the four other factors affecting soil formation, especially climate and vegetation. Consequently, differences in parent material are best expressed in poorly developed entisols and inceptisols. Soil properties like the amount of organic matter, the degree and depth of oxidation, the removal of cations or minerals, development of clay minerals, and clay or $CaCO_3$ content in the B horizon change as the soil weathers and develops. Thus, the degree of weathering can serve as a proxy for the stage of development and relative age of a soil.

The parts of a soil profile develop at different rates. As a soil ages, the A horizon generally achieves steady state or equilibrium first, followed by the B horizon, and then such features as carbonate accumulation and oxidation. Different soil orders take different times to reach equilibrium. Mature spodosols and alfisols are capable of forming in millennia. Mature ultisols may take tens of thousands of years to form. Mature oxisols may require hundreds of thousands of years to form.

A **chronosequence** is a series of soils of different ages that formed on the same parent material under similar conditions of vegetation, topography, and climate—for example, a sequence of soils formed on alluvial terraces at different heights above an incising river [**Figure 3.11**]. Chronosequences of soils that have been dated by radiometric or other means are used to evaluate rates of change in soil properties. Calibrated soil development rates can then be used to infer the ages of landforms from soil properties or to estimate the amount of time that a soil was exposed at the surface before being buried.

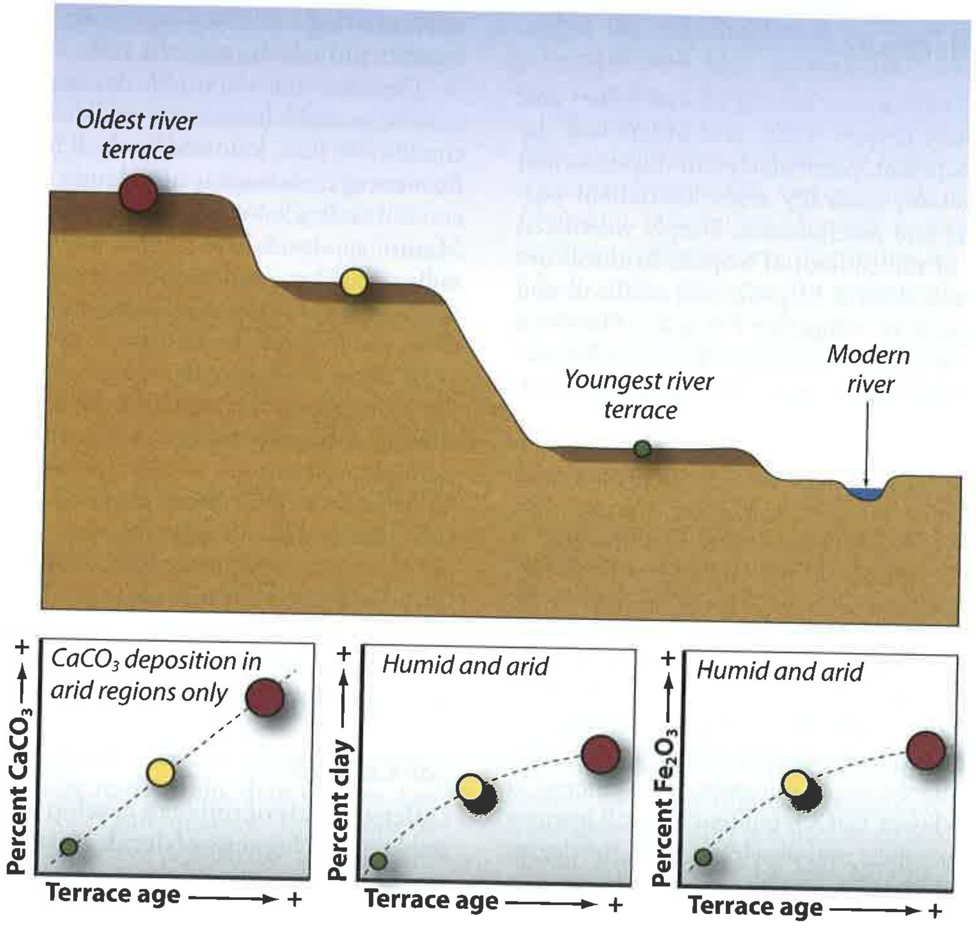
Soil Catenas

Different kinds of soils can develop on different parts of a single slope because of local variability in soil-forming



Strong environmental controls on weathering (i.e., precipitation, temperature, and evaporation) and organic material inputs (i.e., leaf litter fall) lead to strong latitudinal gradients in soil depth and the mineralogy of weathering products. Soils tend to be thickest and most intensively weathered in tropical regions.

FIGURE 3.10 Soil Properties and Depths with Latitude. Soil properties, including average soil depth and characteristic weathering minerals, vary by latitude because climate and vegetation (important for soil formation) are latitude-dependent. [Adapted from Strakov (1967).]



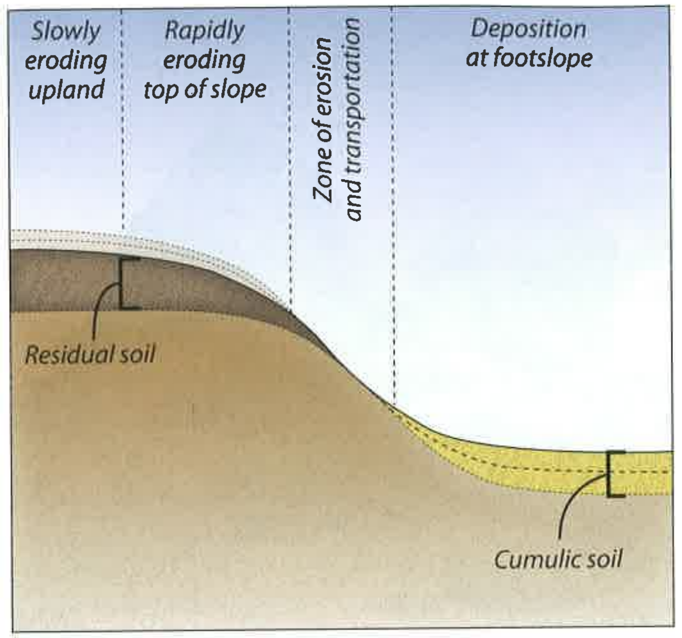
Soil properties change over time and thus landforms in one area with different ages (a **chronosequence**) may have soils with a progression of physical properties. Consider, for example, a set of river **terraces** in which the oldest terrace is preserved high above younger terraces and the modern river. Soil properties such as soil thickness or the percent of CaCO$_3$, clay, or Fe$_2$O$_3$ will typically increase with increasing terrace age.

FIGURE 3.11 Soil Chronosequences. Soil characteristics, such as horizon development, color, clay content, and chemical composition, change predictably with age in a soil chronosequence when other soil-forming variables such as parent material are constant.

processes, erosion, and deposition. A **soil catena** (Latin for "chain or series") is a suite of soils from different positions in a landscape—each of which has different characteristic soil moisture conditions, sediment transport rates, slope steepness, and chemical weathering rates [**Figure 3.12**]. Catenas reflect local differences in soil-forming processes that result from topographic position and the hydrological and geomorphological processes that govern infiltration, runoff, and soil erosion. For example, soils at the top of slopes where water readily washes over or through the soil profile may be well drained, oxidized, and reddish in color, whereas those soils at the base of slopes tend to have a higher water table and reducing conditions that produce more gray to blue colors. Soils developed on the steep slopes around a plateau may be quite different from those developed on the flat plateau surface. In many environments, sloping surfaces are the rule and distinct soils develop on hilltops at mid-slope, and on the base of slopes as erosion thins the soils on the upper slopes and accumulation stacks the soils a the base of slopes.

Paleosols

Unlike soils actively developing at Earth's surface, **paleosols** are ancient soils in which soil development processes have ceased. Some paleosols are buried and thus preserved in the geologic record. **Relict soils** are those that have remained at the surface since their formation but may have formed under different climatic or hydrologic conditions. **Buried soils** are readily recognized, typically by an abrupt change in color, texture, or mineralogy, and can be found in geologic materials of almost any age [**Photograph 3.11**].



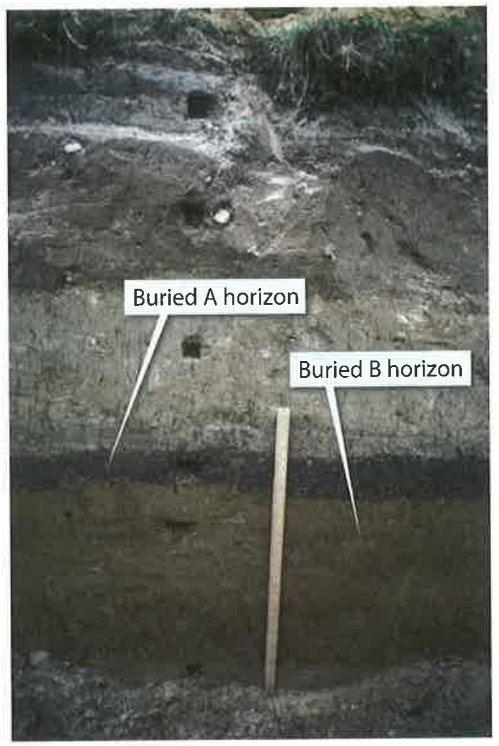
The suite of soils associated with landforms in a particular region form a soil **catena** in which soil properties vary systematically with landscape position. The slowly eroding upland consists of soils developed on in situ parent material. Steeper hillslopes may display truncated soil profiles and represent a zone of translation and more rapid erosion (transport slopes). Lower gradient slopes may have soils developed on deposits of material transported from upland environments.

FIGURE 3.12 Soil Catena. Soils are generally eroded from upland areas, transported down hillslopes, and deposited at low-gradient areas below slopes. A sequence of soils with properties that vary as a function of hillslope position is known as a catena.

When found in the geologic record, paleosols indicate a land surface on which a soil was able to form. They thus represent a period during which soil production and development outpaced both erosion and burial by deposition. Because the characteristics of paleosols reflect the soil-forming factors of the landscape in which they developed, they can be useful indicators of past climatic or environmental conditions. Some soil features, such as organic matter, do not survive extended burial and are poorly preserved, whereas other characteristics, like soil structure and texture of B horizons, are geologically more robust.

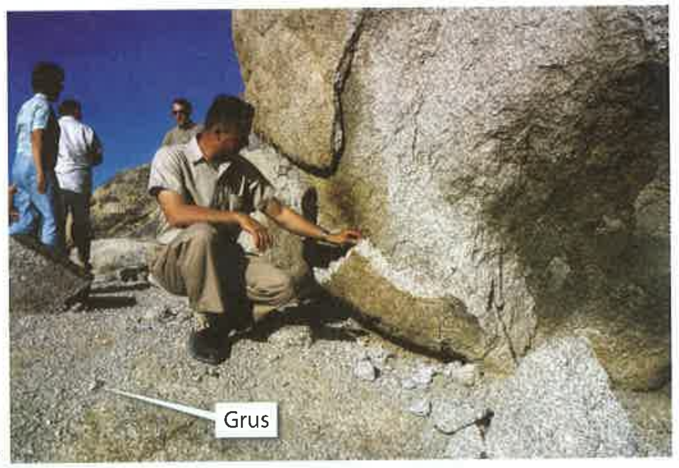
Weathering-Dominated Landforms

Weathering processes produce a variety of distinctive coarse- and fine-scale landforms. At fine scales, physical and chemical processes wear away the edges and corners of rock outcrops, joint-bounded blocks, and boulders fastest; they thus promote **spheroidal weathering**, which converts angular blocks of rock that result from tectonic and unloading fractures into rounded cobbles, boulders, and monoliths. Outward expansion of clay minerals on chemically altered rock surfaces enhances spheroidal weathering and produces weathering rinds that peel off like onion skins [**Photograph 3.12**].



P. Bierman

PHOTOGRAPH 3.11 Paleosol. Wall of trench cut into historic alluvial fan that covers a paleosol. Well-preserved buried A horizon (black) and B horizon (brown) buried by tan and gray post-settlement sediment resulting from clearance of hillslopes above. Meter stick for scale.

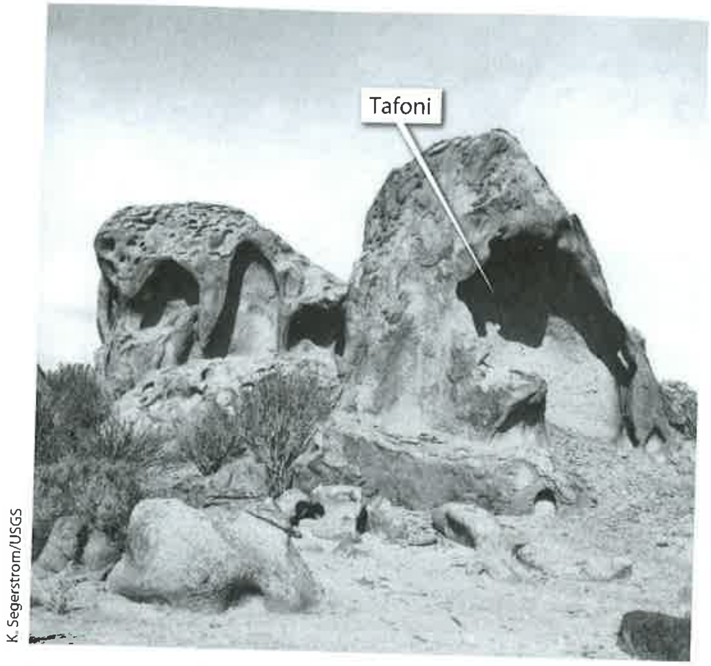


D. Trent

PHOTOGRAPH 3.12 Spheroidal Weathering. Spheroidal weathering of granitic rock at Goldstone in the Mojave Desert, was largely brought on by hydrolysis of feldspar and by wetting-drying and oxidation of biotite and expandable clay. Rock is surrounded by grus, weathered, disarticulated mineral grains that once made up the granite rock.

P. Bierman

PHOTOGRAPH 3.13 Weathering Pit. A weathering pit on the top of granitic Pildappa Rock, on the Eyre Peninsula in south central Australia, is more than a meter wide.



K. Segerstrom/USGS

PHOTOGRAPH 3.14 Tafoni. Tafoni developed in granite near Caldera, Chile.

Wetting and drying or freezing of water causes localized spalling and granular disintegration that creates **weathering pits** on bare rock surfaces of certain rock types [Photograph 3.13]. Extreme cases of weathering pit development result in cavernous or honeycomb textures known as **tafoni**, from the Greek for tomb or the French for window [Photograph 3.14]. Dissolution pits and cavities often characterize highly soluble carbonate outcrops. At coarser scales, weathering-dominated landforms include karst topography (discussed in later chapters), inselbergs, tors, and duricrusts. Differential weathering can also operate on a much larger scale. For example, in the valley and ridge province of the eastern United States, differential weathering and subsequent erosion results in quartzite ridge tops and limestone valley bottoms.

Inselbergs and Tors

Inselbergs (German for "island mountains") and **tors** are high-standing bodies of exposed rock that rise above surrounding terrain and result from spatial variability in the rates of rock weathering and erosion [Photographs 3.15 and 3.16]. Inselbergs are large residual rock masses still attached to bedrock after episodes of deep weathering and subsequent erosion removed the surrounding material. Such stripping can reveal **etchplains,** extensive bedrock surfaces formed under a mantle of weathered material. Tors are smaller features consisting of bedrock on which may sit multiple, smaller exhumed core stones not necessarily still attached to the rock. In some places, field evidence suggests that inselbergs and tors weathered more slowly in the subsurface than did surrounding areas because of differences in either mineral composition or fracture density that rendered them less susceptible to chemical weathering than the surrounding rock [Figure 3.13]. In other places, a chance exposure of one part of the rock mass at the surface renders it less susceptible to further weathering, providing a positive feedback to amplify the topography (until the outcrop is destroyed by erosion from the sides).

Variability in the progress of deep weathering produces an uneven **weathering front,** the boundary between fresh rock below and that altered by weathering closer to Earth's surface. Inselbergs and tors represent exhumation of less-weathered material protruding above the weathering front. These exhumed landforms record a changing balance between weathering and erosion. Specifically, they represent the change from an erosion-limited landscape to one where the pace of weathering limits erosion rates.

Tors are common landforms in weathered landscapes around the world; the rock towers of southwest England and Wales are some of the best-known examples of tors. Australia's famous Uluru (also known as Ayers Rock), a massive sandstone outcrop that rises hundreds of meters above the desert plain of the central outback, is perhaps the largest and most dramatic example of an inselberg in an arid setting. In a more humid climate, Rio de Janiero's famous Sugarloaf, a granitic inselberg, towers above deeply weathered surroundings [Photograph 3.17]. Along Brazil's coast, mechanically weak fault zones along which weathering is enhanced and where shattered rock is easily eroded have been mapped between the towering inselbergs. It appears that the steep sides of these inselbergs likely result from enhanced weathering along these near-vertical fault zones.

P. Bierman

PHOTOGRAPH 3.15 Inselbergs. The Olgas of central Australia are sandstone inselbergs, also known as domes or bornhardts.

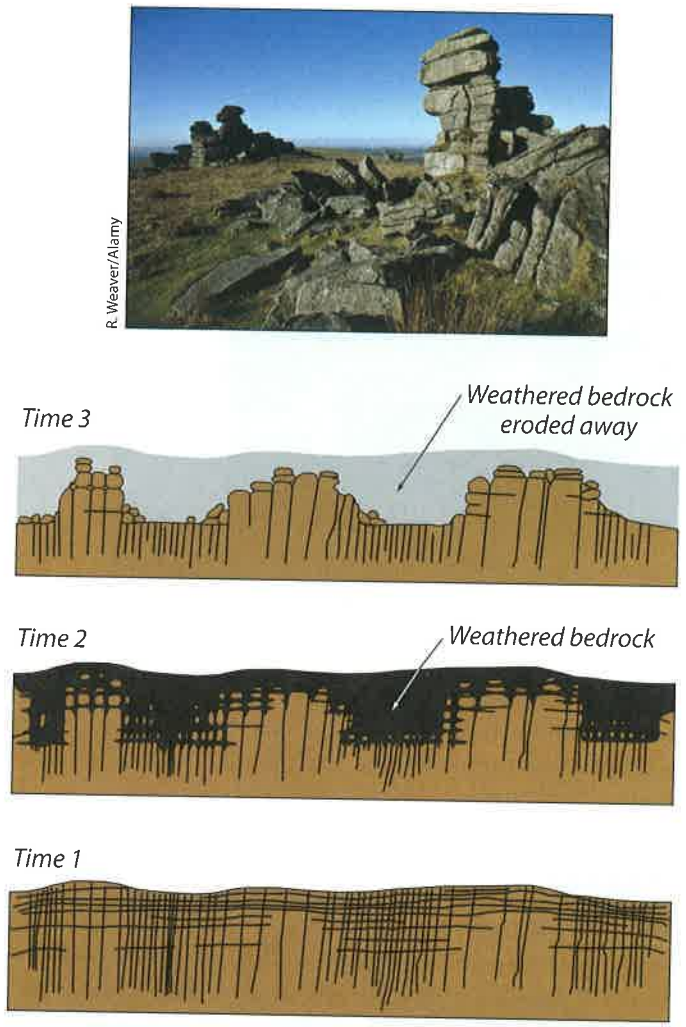
P. Bierman

PHOTOGRAPH 3.16 Tor. Devil's Marbles, a tor with boulders, is the residuum of spheroidal weathering in central Australia.

Duricrusts

Duricrusts are erosion-resistant soils hardened and turned back into rock by cementation within the pedogenic zone. In arid regions, enough calcium carbonate ($CaCO_3$) may accumulate within a soil profile to form a cementlike layer of caliche (a Spanish word for "porous material cemented by calcium carbonate"), or **calcrete,** that is erosion resistant when it is exposed at the land surface. **Silcrete,** a hard silica-rich (SiO_2) layer, behaves similarly [**Photograph 3.18**] and also forms in arid and semi-arid regions as silica in groundwater reprecipitates. Duricrusts also form by evaporation of solute-rich groundwater along stream valleys. These protective shells can even be tough enough to cause **topographic inversions,** which occur when formerly low-lying terrain becomes more resistant to erosion than the neighboring uplands and eventually becomes elevated as the surrounding terrain erodes away.

Extreme oxidation, intense weathering, drying, and leaching produce iron-rich residual soils known as **laterite** and **ferricrete.** Ferricrete is an iron-rich duricrust made of highly concentrated iron oxides as a result of intensive weathering. Laterite is clay-rich soil containing large proportions of iron and aluminum oxides. These erosion-resistant crusts form where intense weathering removes all but the least mobile elements, leaving behind just aluminum and iron oxides. Laterite and ferricrete are often found on slowly eroding topographic highs in the tropics and form the caprock for mesas in central Australia and the Amazon.



The relief developed on **etchplains,** regional bedrock surfaces characterized by **tors** and **inselbergs,** reflects differential weathering and subsequent erosion of an irregular **regolith.** As weathering proceeds faster in more highly fractured portions of a landscape, the interface between regolith and bedrock may develop substantial relief. Subsequent erosion of weaker, weathered material, perhaps triggered by a change in climate or base level, can leave less-highly fractured and weathered bedrock standing above surrounding terrain as **core stones,** tors, and inselbergs.

FIGURE 3.13 Etchplain Formation. Erosion of deeply weathered regolith exposes an irregular bedrock surface of tors and inselbergs that was shaped in the subsurface by physical and chemical alteration of bedrock along the weathering front. [Adapted from Linton (1955).]

PHOTOGRAPH 3.17 The Sugarloaf. Rio de Janiero's famous Sugarloaf is a granitic inselberg towering above deeply weathered surroundings. Its steep margins are defined by near-vertical fault zones along which chemical weathering proceeded more quickly, isolating the more resistant rock.

Applications

Weathering is of fundamental geomorphological and societal importance because it transforms hard rock into material that can be moved by surface processes and used by living organisms. The physical breakdown of rock into smaller pieces and the chemical transformation of rock into secondary minerals and dissolved elements influence the types and rates of geomorphological processes that shape topography.

Soils are the frontier between geology and biology. The thin skin of weathered rock and decomposing organic matter provides the nutrients that nourish the plants on which all terrestrial life depends. Cation-exchange capacity and base saturation of soils are the basis for soil fertility because plants can readily extract nutrients from soils with high base saturation. Sustained cultivation without replenishing soil nutrients leads to declining crop yields. This is one reason why floodplains that receive annual deposits of fresh minerals and volcanic soils that are periodically replenished by ash fall are prized and highly productive agricultural lands around the world. Modern industrial agriculture uses tremendous amounts of chemical fertilizers (principally nitrogen, phosphorus, and potassium) to supplement native soil fertility in place of traditional crop rotations, applications of manure, and organic farming techniques that are based on soil ecology. Understanding soil-forming processes and soil fertility are important for evaluating options for maintaining soil fertility and sustaining agricultural productivity.

PHOTOGRAPH 3.18 Silcrete. Silica-cemented rocks at Coober Pedy in central Australia. Such silica-cemented layers (silcrete) hold up the mesa in the distance, acting as caprocks.

Maintenance of soil fertility depends not only on sustaining soil nutrient levels, but also on conserving the soil itself. Extensive soil loss to erosion can follow deforestation, tillage, and destruction of vegetative cover by fire or overgrazing. On some now-barren Caribbean islands, sugarcane cultivation on steep slopes sent most of the topsoil into the ocean in less than a century. Recent Earth history is rife with truncated and thinned soil profiles that record examples of ancient societies (such as classical Greece, Rome, and Easter Island) that failed to prevent rates of soil erosion from exceeding rates of soil production. Globally, the average rate of net soil loss from agricultural fields has increased 10- to 20-fold as a result of tillage and exposure of bare soil to the effects of wind, rainfall, and runoff. Basic training in geomorphology can help practitioners tailor sustainable agricultural practices to particular soils and landforms.

Farming practices have reduced soil organic matter across vast areas of the continents, particularly in mollisols. About one-third of the carbon dioxide added to the atmosphere by human activity since the Industrial Revolution came from the decay of soil organic matter, the result of plowing fertile grassland soils. But people can also improve soil fertility. The recently discovered organic-rich, incredibly fertile "terra preta" soils in the Amazon jungle formed over millennia as indigenous people burned trash and broken pottery in their fields. Today these soils form islands of fertility in otherwise infertile tropical soils.

Weathering also provides minerals critical for modern life. Deep weathering on ancient land surfaces produced iron and aluminum ores through pervasive leaching and removal of other common elements, thereby concentrating relatively immobile elements in residual soils. Aluminum is a common element in terms of its distribution in Earth's crust, but it is dispersed in most rocks such that it is not economically extractable. It takes millions of years to dissolve away everything else and make laterite soils that are enriched enough in aluminum that they constitute aluminum ores. Few people realize that we wrap our soda and beer in metal mined from ancient soils.

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DIGGING DEEPER How Fast Do Soils Form?

How quickly does soil, the discontinuous mantle of weathered rock and organic material that covers large parts of Earth's surface, form? This is an important and practical question, because soil and its fertility are so important to agriculture and thus to human survival. It is also a difficult question to answer because the answer varies not only by where one looks but also by how one defines soil formation.

There are many different ways of approaching the concept of soil formation, the processes and rates by which soil is created from rock and sediment. In many landscapes, soils develop on stable deposits of unconsolidated material such as blankets of loess, bouldery colluvium, or the sand and gravel of river terraces. In these cases, the rate of **soil development** describes the speed at which pedogenic processes alter the parent material. In such settings, it is the pedogenic alteration of primary to secondary minerals and the accumulation of organic matter that makes soils fertile and useful for agriculture.

This rate of soil development (pedogenesis) has been quantified in many different environments by geomorphologists studying **chronosequences**, soils formed in similar parent materials but having different ages [**Figure DD3.1**].

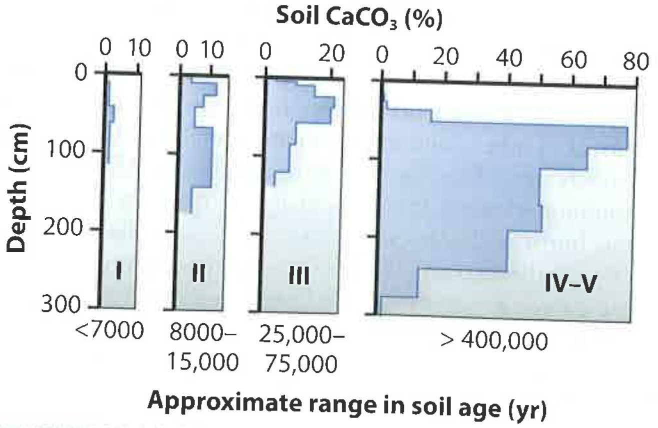

FIGURE DD3.1 Soil chronosequences are constructed from observations of soils that developed in similar climates and on similar parent materials but differ in age. Here, the soil chronosequence (soils I through V) is used to indicate that as arid soils develop over time, the amount of calcium carbonate ($CaCO_3$) in the soil increases predictably. This increase reflects both in situ weathering of calcium-bearing minerals and the addition by wind of carbonate-bearing dust. Over time, carbonate is carried more deeply into the soil profile by infiltrating water. [From Birkeland, (1984).]

In some areas, the rate of soil development has been quantified well enough that soil ages are estimated using a **soil development index** that takes into account soil characteristics that change over time, such as reddening, structure, and clay content (Harden, 1982).

In the last work of his long, stimulating, and controversial career, Darwin (1881) concluded that soils formed at a pace of less than a millimeter per year. He arrived at this conclusion by measuring the thickness of new soil that lay on top of the foundations and floors of Roman buildings in England. Darwin attributed the development of new soil to the action of worms gradually bringing up material to the ground surface; thus, Darwin's soil was cumulic—it was not formed by the weathering of bedrock from below but rather by the addition of material from above.

In contrast to soil development on unconsolidated parent material, **soil production**, or the production of regolith, refers to the in situ development of soil by the loosening of individual mineral grains from bedrock (Heimsath et al., 1997). Such loosening is the result of both physical and chemical weathering, the rate of which depends on many factors including soil moisture, rock type, and biological disturbance by burrowing animals and tree roots. The concept of soil production from rock is applicable to upland, often hilly, soil-mantled bedrock landscapes where the limiting factor for soil formation is the weathering of the underlying rock. The concept of soil production does not apply to agricultural lowlands where soils develop on unconsolidated deposits.

Two centuries ago, Playfair (1802) recognized that the long-term rates of soil production and soil erosion must balance for soil-mantled, upland landscapes to persist over time. If soil erosion occurred faster than soil production, the soil would thin until bare rock was exposed at the surface, but soil blankets slopes in most humid and temperate landscapes. More than a century and a half later, a thorough review of the literature found no reliable, quantitative estimates of how fast soil formed from bedrock (Smith and Stamey, 1965). Since then, a variety of geochemical measurements have been used to quantify rates of soil production from rock.

Measurements of the bulk composition of rocks, soils, and the **dissolved load** of runoff collected in studies of small watersheds, along with the assumption of steady state, allows estimation of rates of soil production from a **mass balance** of major rock-forming elements (Alexander, 1985; 1988). Inherent in this approach is the assumption that cations are released from minerals as rock chemically weathers into regolith.

Elemental mass fluxes and thus rates of chemical weathering and soil production in 18 North American, European, Australian, and African watersheds, each underlain by a single rock type, were positively correlated with runoff; that is, rock weathered faster in wetter climates. Assuming all cations came from weathering of bedrock, rates of rock weathering and thus soil production varied from 0.002 to 0.158 mm/yr. Similar application of a global geochemical mass balance, based on the mean compositions of major elements (aluminum, calcium, iron, potassium, magnesium, sodium, silicon) in Earth's crust, soils, and the dissolved load of rivers, found that the global average rate of soil production from rock weathering was between 0.03 and 0.10 mm/yr with a mean rate of 0.06 mm/yr (Wakatsuki and Rasyidin, 1992). Not all of this weathering was conversion of bedrock to regolith and eventually soil. Some of the cations may have been released as soils developed on unconsolidated parent materials.

Analyses of cosmogenic isotopes (see Chapter 2) in subsoil bedrock allows direct measurement of rates of soil production (Heimsath et al., 1997) on upland landscapes where hillslope processes dominate the movement of material downslope. This approach assumes steady-state soil production and steady soil thickness; that is, the rate of bedrock weathering and new soil production matches the rate of downslope transport of soil [**Figure DD3.2**].

Investigations at several different field sites show that ^{10}Be concentrations vary systematically with the thickness of soil cover. Applications of this approach to soil profiles in California and Australia support an exponential decline of soil production from rock with increasing soil depth

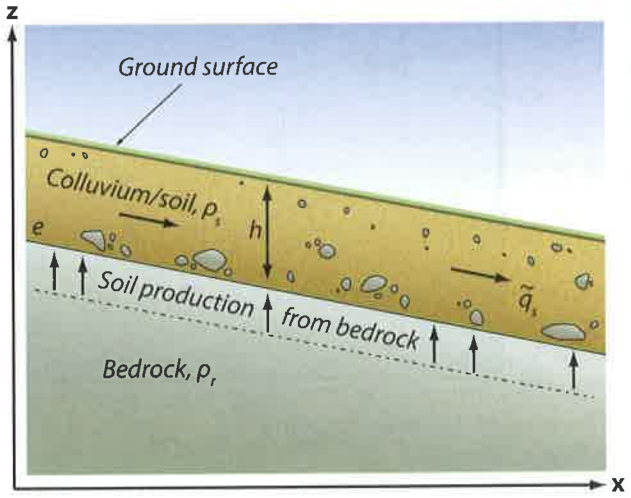

FIGURE DD3.2 This schematic diagram illustrates the relationship between soil production and soil erosion on an idealized cross section of a hillslope. The soil thickness *h* will be maintained if soil production from bedrock matches mass transport, q_s, of the active layer of soil. The area shown between the base of the soil at elevation *e* and the dash-dot line is the amount of bedrock that would be converted to soil over some specified time interval. Here ρ_s and ρ_r are the densities of soil and bedrock, respectively. [From Heimsath et al. (1997).]

DIGGING DEEPER How Fast Do Soils Form? *(continued)*

[Figure DD3.3]. Heimsath et al. (1997) attribute this decline in soil production rates with increasing depth to decreased effectiveness of mechanical (physical) weathering processes such as freeze–thaw, tree-throw, and biogenic disturbance in breaking up bedrock as soil cover thickens. The cosmogenic approach allows direct determination of soil production rates from bedrock and provides a means to map their variability across soil-mantled, upland landscapes around the world.

A global compilation of soil formation rate estimates, which includes both rates of soil production and soil development made by various methods, finds they range widely between 0.0001 to 0.5 mm/yr, with global median and mean values of 0.017 and 0.036 mm/yr, respectively (Montgomery, 2007). The same study found that short-term rates of soil formation, short-term rates of soil erosion under native vegetation, and long-term geological erosion rates were similar. Before people got involved, it appeared as though soils formed at about the same pace as they eroded **[Figure DD3.4]**.

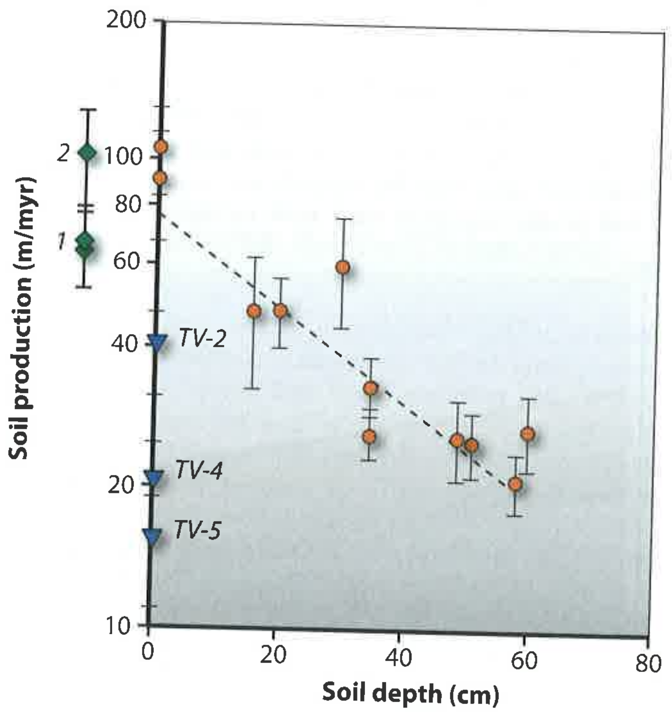

FIGURE DD3.3 This figure shows erosion rates on three different parts of the same landscape, Tennessee Valley, in northern California. Left of the vertical axis are diamonds that represent basin-scale erosion rates estimated by measuring in situ cosmogenic ^{10}Be in river sediment. These rates are similar to the rates of soil production measured in rock outcropping at the soil surface (filled circles plotted at zero soil depth). As the soil thickens, soil production rates decline. Outcrops standing well above the soil surface had the lowest erosion rates (plotted as upside-down triangles, labeled TV). [From Heimsath et al. (1997).]

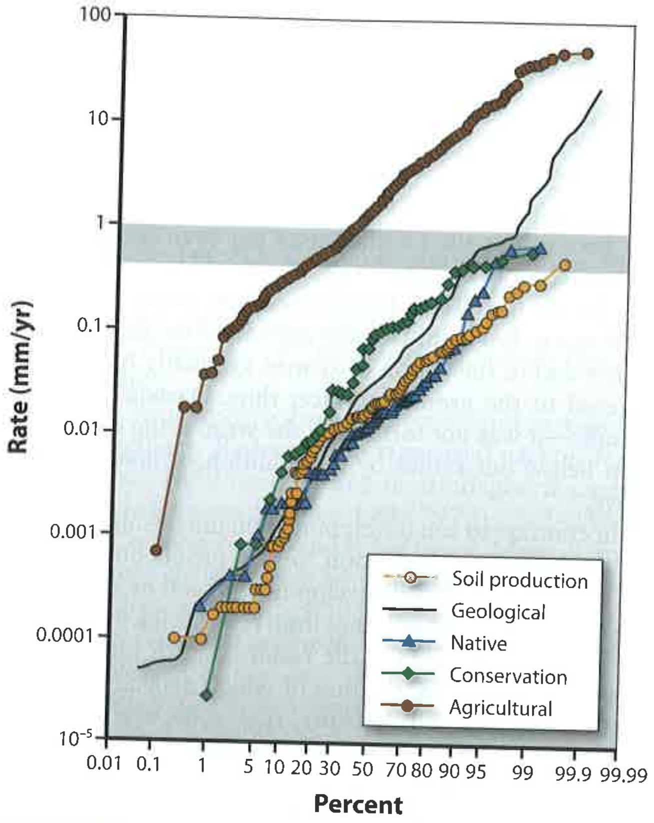

FIGURE DD3.4 Five cumulative probability plots from a global compilation show that rates of soil erosion from agricultural fields under conventional tillage regimes (red circles; sample size, n = 448) are far higher than other rates of erosion: where soil conservation techniques are used (green diamonds, n = 47), where plots are under native vegetation (blue triangles, n = 65), long-term, geological rates of erosion (solid line, n = 925), and rates of soil production (orange circle, n = 188). The shaded area represents the range of soil erosion rates that the U.S. Department of Agriculture defines as tolerable soil loss from agricultural fields. Even these "tolerable" rates of soil erosion are unsustainable in terms of geological rates of soil production and erosion. To read the graph, note that the 50 percent value is the median; the lowest rate plots on the far left and the highest rate plots on the far right. [From Montgomery (2007).]

The maximum reported pace of soil production (0.5 mm/yr) is less than the average rate of soil erosion reported for U.S. and global croplands (0.89 to >1 mm/yr) (Montgomery, 2007; Wilkinson and McElroy, 2007). This discrepancy between geological rates of soil production and contemporary agricultural erosion rates leads to the unsettling conclusion that conventional farming practices are not sustainable. Fortunately, however, soil erosion rates under conservation agriculture, such as "no till" farming, are much closer to soil production rates, showing that agriculture need not result in accelerated rates of soil loss.

Unless better soil management practices are adopted worldwide, the gradual, ongoing loss of fertile topsoil and degradation of agricultural lands present a growing problem for the world, especially as population continues to rise. Not only does the loss of soil impact agricultural productivity, but some of the eroded, nutrient-rich topsoil ends up in rivers and lakes, impacting water quality and encouraging the growth of nuisance organisms. Although soil erosion (coupled with deforestation and climate shifts) plagued many ancient societies, soil conservation measures rooted in geomorphological principles as well as agronomy can be employed to make agriculture sustainable. Because even unsustainable rates of soil erosion appear slow on human timescales, there is little urgency surrounding the issue of soil loss. Reducing the rate of soil erosion is important because healthy soil and fertile land are arguably society's most overlooked strategic resource.

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WORKED PROBLEM

Question: For an arid landscape and for a humid landscape, developed on the same type of granitic rock, contrast rock weathering on the respective hillslopes and soil development in the valleys. Consider the germane rock-weathering processes and describe the resulting soil profiles.

Answer: In the humid landscape, hillslopes will be soil-mantled and much of the weathering will be chemical, occurring below the cover of soil. There will be some physical weathering, much of it accomplished by tree roots. In the arid landscape, there will be less weathering overall, chemical weathering will be less intensive, the slopes will be rocky, and a greater proportion of the weathering will be physical, occurring on bare rock surfaces. In both cases, the granite will decompose to its constituent minerals; however, in the humid landscape more of the minerals will have been chemically altered, with the feldspars weathering to clay minerals and the iron-rich minerals oxidizing. The chemical alteration of the granite will assist physical decomposition to grus, probably through the alteration of biotite.

Soils in the two landscapes will be distinctive. In the arid landscape, soils will likely be capped with a vesicular A horizon, reflecting the accumulation of aeolian dust and the formation of small vesicles from wetting and drying. This will be underlain by a B horizon, characterized by reddening and perhaps the accumulation of calcium carbonate. If the soil is old enough, the carbonate will coat B-horizon clasts and create a durable and impermeable petrocalcic horizon (calcrete). The humid soil may have an organic-rich O horizon at the surface, which is likely to be underlain by an A horizon and possibly an E horizon if the soil is permeable and if there are trees growing on the surface that produce acidic leaf litter. Below the upper horizons, there will be a B horizon typified by reddening and an increase in iron and clay content.

KNOWLEDGE ASSESSMENT Chapter 3

○ 1. Define and differentiate saprolite, soil, and regolith.
○ 2. Describe how physical, chemical, and biological weathering differ.
○ 3. How does chemical weathering influence physical weathering and vice versa?
○ 4. List three mechanisms of physical weathering and describe how each works.
○ 5. What are borndhardts, tors, and inselbergs?
○ 6. Propose a series of physical processes and rock history that could lead to exfoliation.
○ 7. What process likely created felsenmeer?
○ 8. Predict what will happen to boulders or outcrops of rock exposed to a forest or range fire or to extended heating in the desert sun.
○ 9. Explain how salts can physically weather rock.
○ 10. Why does chemical weathering occur?
○ 11. Explain how silica and most rock-forming minerals differ from calcium carbonate in terms of chemical weathering?
○ 12. Why are iron and aluminum oxides characteristic of materials left as residue after extensive chemical weathering?
○ 13. List the most important factors influencing the amount of leaching by solution of material from soil or rock.
○ 14. Discuss how various types of biological activity affect the physical and chemical weathering rates of rocks.
○ 15. What controls the swelling potential of clay minerals?
○ 16. Define cation exchange, identify which ions are most likely to be exchanged, and explain why cation exchange is an important process.
○ 17. What is oxidation of minerals and where can you see it occur?
○ 18. Sketch and explain Goldich's Weathering Series.
○ 19. Starting with feldspar, explain how it weathers over time to eventually become gibbsite.
○ 20. What is chelation and why is it important in soil formation?
○ 21. Why is the density of saprolite less than that of the rock from which it was derived?
○ 22. Describe the patterns of and balance between chemical and physical weathering on a global scale.
○ 23. Explain how vegetation type and density affect soil formation rates and the type of soils that result.
○ 24. Explain how topography (slope steepness) affects soil thickness.
○ 25. How is soil development related to the age of a soil?
○ 26. List three ways by which material can be added to and removed from soils.
○ 27. Define translocation and explain how it can occur in soils.
○ 28. Explain how bioturbation occurs and why it is important.
○ 29. List the most common soil horizons and explain how each forms.
○ 30. Describe the two general models relating soil production rates to soil depth.
○ 31. Compare and contrast pedogenesis and soil production.
○ 32. Make a table listing the 12 soil orders, describing their salient characteristics, and suggesting in what environment each might be found.
○ 33. List the five factors thought to control rates of pedogenesis (soil development).
○ 34. Define a soil catena and explain differences observed across a landscape in terms of soil-forming processes.
○ 35. Suggest in what geomorphic environment you might find a chronosequence and explain why defining a chronosequence could be useful to you as a geomorphologist.
○ 36. Compare a paleosol and a cumulic soil.
○ 37. List and describe several examples of weathering-induced landforms at both small and large scales.
○ 38. List several types of duricrusts and explain how they help shape landscapes.
○ 39. Compare rates of soil production to contemporary rates of soil loss and explain why this comparison is important.