Geology, Soils, and Sediments

Within a given climatic regime, soil properties are the major factor governing ecosystem processes. This chapter provides background on the factors regulating those soil and sediment properties that most strongly influence ecosystems as well as the transport of materials from land to rivers, lakes, and the ocean.

Introduction

Soils form a thin film over Earth's surface in which geological and biological processes intersect. The soil consists of solids, liquids, and gases, with solids typically occupying about half the soil volume, and liquids and gases each occupying 15–35% of the volume (Ugolini and Spaltenstein 1992). The physical soil matrix provides a source of water and nutrients to plants and microbes and is the physical support system in which terrestrial vegetation is rooted. It is the medium in which most decomposer organisms and many animals live. For these reasons, the physical and chemical properties of soils strongly influence all aspects of ecosystem functioning, which, in turn, feed back to influence the physical, structural, and chemical properties of soils (see Fig. 1.5; Amundson et al. 2007). Soils play such an integral role in ecosystem processes that it is difficult to separate the study of soils from that of ecosystem processes. In open-water (**pelagic**) ecosystems, phytoplankton cannot directly tap resources from sediments, so sediment processes provide nutrient resources to primary producers only indirectly through mixing of the water column.

Soils are also a critical component of the total Earth System. They mediate many of the key reactions in the giant global reduction–oxidation cycles of carbon, nitrogen, and sulfur and provide essential resources to biological processes that drive these cycles. Soils represent the intersection of the "bio," "geo," and chemistry in biogeochemistry. Many of the later chapters in this book address the short-term dynamics of soil processes, particularly those processes that occur on timescales of hours to centuries. This chapter emphasizes soil processes that occur over longer timescales or that are strongly influenced by physical and chemical interactions with the environment. This is essential background for understanding the dynamics of ecosystems.

A Focal Issue

Human activities have massively increased nutrient and sediment inputs from terrestrial to aquatic ecosystems. Soils that developed over thousands of years can be eroded away in years to decades, causing loss of productive capacity in upland ecosystems and accumulation in reservoirs, lowland floodplains, estuaries, and coastal waters. On human timescales, this is an essentially permanent restructuring of regional landscapes. The extensive cultivation of drought-sensitive crops on marginal lands in the U.S. in the 1920s, for example, created a landscape vulnerable to

Fig. 3.1 Extensive cultivation replaced drought-resistant native vegetation with drought-sensitive crops in the midwestern U.S. in the 1920s. In the "Dustbowl" era, a drought in the 1930s killed these crops and generated massive dust storms, such as this one approaching Stratford Texas in 1935. Photograph courtesy of NOAA, http://www.photolib.noaa.gov/htmls/theb1365.htm

drought. Hot, dry weather combined with strong winds in the 1930s caused extensive wind erosion that reduced the productive potential of soils, modified regional climate, and triggered land abandonment and human migration (Fig. 3.1; see Chap. 12; Peters et al. 2004; Schubert et al. 2004). Erosion of the loess plateau in China and drylands in sub-Saharan Africa are current issues that threaten livelihoods of millions of people over extensive regions. What properties of vegetation and soils cause some soils to be more susceptible to erosion than others? Why are topsoils, which are the first layers to be eroded, so much more fertile than deeper soils? What are the consequences of wind and water erosion for those ecosystems where soil particles are deposited? What management practices sustain the productivity soils and reduce erosion rates? This chapter addresses these questions and other issues that are important for sustainability of ecosystems and managed landscapes.

Controls Over Soil Formation

The soil properties of an ecosystem result from the dynamic balance of two opposing forces: soil formation and soil loss. State factors differ in their effects on these opposing processes and therefore on soil and ecosystem properties (Jenny 1941; Amundson and Jenny 1997).

Parent Material

The physical and chemical properties of rocks and the rates at which they are uplifted and weathered strongly influence soil properties. The dynamics of the rock cycle, operating over billions of years, govern the variation and distribution of geological materials on Earth's surface. The rock cycle describes the cyclic process by which rocks are formed and **weathered**, i.e., chemically and physically altered near Earth's
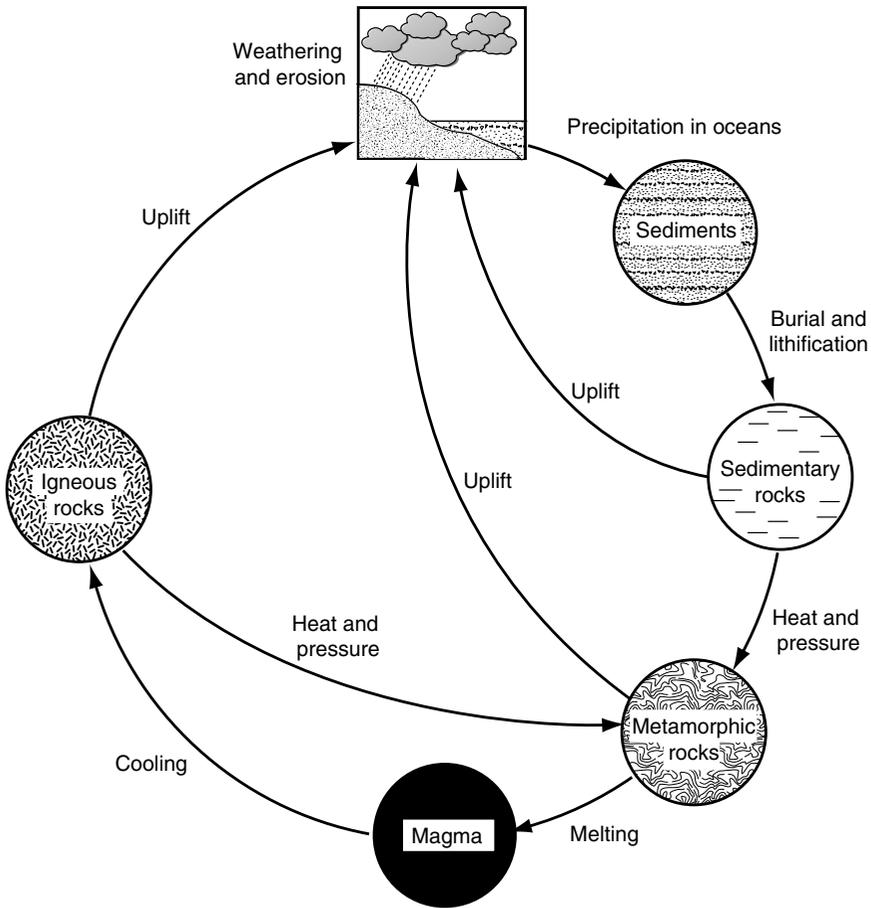
Fig. 3.2 The rock cycles as proposed by Hutton in 1785. Rocks are weathered to form sediment, which is then buried. After deep burial, the rocks undergo metamorphosis or melting, or both. Later, they are deformed and uplifted into mountain chains, only to be weathered again and recycled. Redrawn from Press and Siever (1986)

surface (Fig. 3.2). The rock cycle produces minerals that buffer the biological acidity that accounts for much of rock weathering but also provides many of the nutrients that allow biology to produce this acidity. The compounds produced by weathering move via rivers to lakes, reservoirs, and the ocean where they are deposited to form sediments, which are then buried to form **sedimentary rocks**. **Igneous rocks** form when **magma** from deep within Earth moves upward toward the surface in cracks or volcanoes. Either sedimentary or igneous rocks can be modified under heat or pressure to form **metamorphic rocks**. With additional heat and pressure, metamorphic rocks melt and become magma. Any of

these rock types can be raised to the surface via uplift, after which the material is again subjected to weathering and erosion (Fig. 3.2). Earth's crust cycles through the rock cycle every 100–200 million years, i.e., two to four times since plants first colonized the land (see Fig. 2.15). The timing and locations of uplift and the type of rock uplifted ultimately determine the distribution of different types of bedrock across Earth's surface.

Plate tectonics are the driving force behind the rock cycle. The **lithosphere** or crust, the strong outermost shell of Earth that rides on partially molten material beneath, is broken into large rigid plates, each of which moves independently. Where the plates converge and collide,
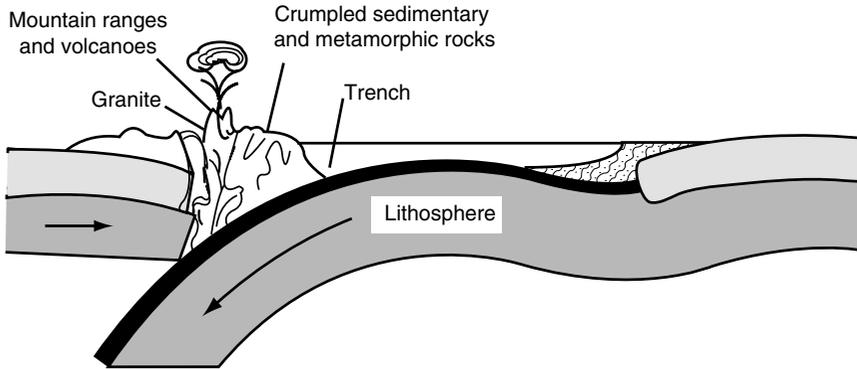
Fig. 3.3 Cross section of a zone of plate collision, in which the oceanic plate is subducted beneath a continental plate, forming an ocean trench in the zone of subduction and mountains and volcanoes in the zone of uplift. Redrawn from Press and Siever (1986)

portions of the lithosphere buckle downward and are **subducted**, leading to the formation of ocean trenches, while the overriding plate is **uplifted**, causing the formation of mountain ranges (Fig. 3.3). Regions of plate collision and active mountain building coincide with Earth's major earthquake belts. The Himalayan Mountains, for example, are still rising due to the collision of the Indian subcontinent with Asia 40 million years ago. If plates converge in one place, they must diverge or separate elsewhere. Throughout Earth history, massive super-continents have formed and broken apart, with continents rafting to new locations and forming new super-continents. This occurred most recently when the super-continent of Pangaea broke up 50–200 million years ago to form Eurasia, Africa, Antarctica, and the Americas. Australia, for example, is moving from its point of origin in Antarctica toward Southeast Asia at 5–6 cm year^{-1}. The mid-Atlantic and mid-Pacific ridges are zones of active divergence of today's ocean plates. Continental drift has rafted the world's biota and soils through multiple climate zones during their evolutionary history.

Climate

Temperature, moisture, carbon dioxide, and oxygen influence rates of chemical reactions that govern the rate and products of weathering, as well as biological activity, and therefore the development of soils from rocks. Temperature, moisture, and oxygen also influence biological processes such as the production of organic matter by plants and its decomposition by microbes and therefore the amount and quality of organic matter in the soil (see Chaps. 5–7). Soil carbon, for example, increases with decreasing temperature and with increasing precipitation along global and regional climate gradients (Post et al. 1982; Burke et al. 1989; Jobbágy and Jackson 2000). Precipitation is one pathway by which materials enter ecosystems. **Oligotrophic** (nutrient-poor) bogs are isolated from mineral soils and depend entirely on precipitation to supply new minerals. The movement of water is also crucial in determining whether the products of weathering accumulate or are lost from a soil and transported to other places. In summary, climate affects virtually all soil properties at scales ranging from local to global.

Topography

Topography influences soils through its effect on climate, moisture availability, and differential transport of fine soil particles. Topographic gradients form a hillslope complex or **catena** from ridge top to valley bottom. These gradients and the **aspect** (compass direction) of the slope strongly influence soil properties (Amundson and

Fig. 3.4 Relationship between hillslope position, likelihood of erosion or deposition, and soil organic carbon concentration. Redrawn from Birkeland (1999)
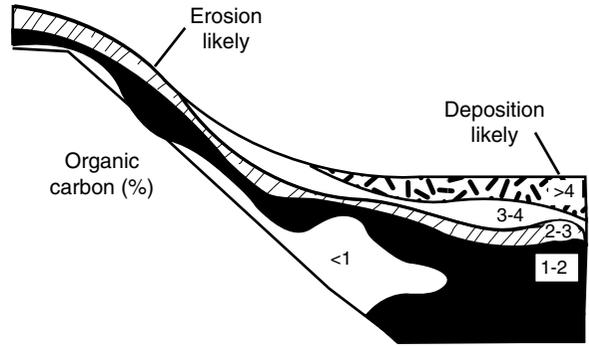
Jenny 1997). Erosion, for example, preferentially moves fine-grained materials downslope and deposits them at lower slope positions, where they tend to form deep fine-textured soils with a high soil organic content (Fig. 3.4) and high water-holding capacity. These valley-bottom soils supply more resources to plants and microbes and provide greater physical stability, typically leading to higher rates of most ecosystem processes than on ridges or shoulders of slopes. Soils in lower slope positions in sagebrush ecosystems, for example, have greater soil moisture, higher soil organic matter content, and higher rates of nitrogen mineralization and gaseous losses than do upslope soils (Burke et al. 1990; Matson et al. 1991).

The aspect of a slope influences solar input (see Chap. 2) and therefore soil temperature, rates of evapotranspiration, and soil moisture. At high latitudes and in wet climates, the cool wet environment of poleward-facing slopes reduces rates of decomposition and mineralization (Van Cleve et al. 1991). At low latitudes and in dry climates, the greater retention of soil moisture on these slopes allows a longer growing season and supports forests, whereas slopes facing the equator are more likely to support desert or shrub vegetation (Whittaker and Niering 1965).

Finally, slope position determines patterns of snow redistribution in cold climates, with deepest accumulations beneath ridges and in the protected lower slopes. These differential accumulations alter effective precipitation and length of growing season enough to influence plant and microbial processes well into the summer.

Time

Many soil-forming processes occur slowly, so the time over which soils develop influences their properties. Rocks and minerals are weathered over time, and important nutrient elements are transferred among soil layers or transported out of the ecosystem. Hillslopes erode, and valley bottoms accumulate materials, and biological processes add organic matter and critical nutrient elements like carbon and nitrogen. Phosphorus availability is high early in soil development and declines in availability over time due to losses from the system and phosphorus fixation in mineral forms that are unavailable to plants (Fig. 3.5; Walker and Syers 1976). This process plays out over millions of years of soil development in Hawai'i, despite a warm moist climate, changing the system from nitrogen limitation on young soils to phosphorus limitation on older soils (Hobbie and Vitousek 2000; Vitousek 2004).

Some changes in soil properties happen relatively quickly. Retreating glaciers and river floodplains often deposit phosphorus-rich till. If seed sources are available, these soils are colonized by plants with symbiotic nitrogen-fixing microbes, allowing these ecosystems to accumulate their maximum pool sizes of carbon and nitrogen within 50–100 years (Crocker and Major 1955; Van Cleve et al. 1991). Other soil-forming processes occur slowly. Young marine terraces in coastal California have relatively high phosphorus availability but low carbon and nitrogen content. Over at least tens of thousands of years, these terraces accumulate organic matter and
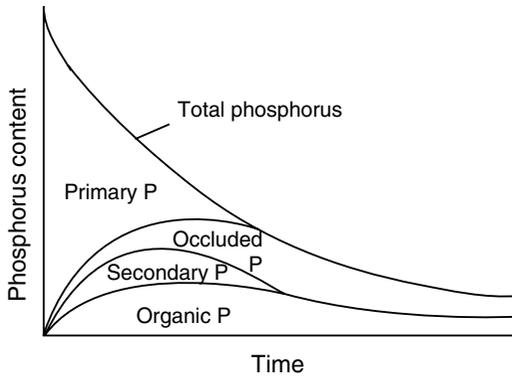
Fig. 3.5 Effects of long-term weathering and soil development on the distribution and availability of phosphorus (P). Newly exposed geologic substrate is relatively rich in weatherable minerals, which release phosphorus. This release leads to accumulation of both organic and readily soluble forms (secondary phosphorus such as calcium phosphate). As primary minerals disappear and secondary minerals capable of sorbing phosphorus accumulate, an increasing proportion of the phosphorus remaining in the system is held in unavailable (occluded) forms. Availability of phosphorus to plants peaks relatively early in this sequences and declines thereafter. Redrawn from Walker and Syers (1976)

nitrogen, causing a change from coastal grassland to productive redwood forest (Jenny et al. 1969). Over several 100,000 years, silicates are leached out, leaving behind a hardpan of iron and aluminum oxides with very low fertility and seasonally anaerobic soils. The pygmy cypress forests that develop on these old terraces have very low productivity. The phenolic compounds produced by these trees as defenses against herbivores also retard decomposition, further reducing soil fertility (see Chap. 7; Northup et al. 1995).

Potential Biota

The past and present organisms at a site strongly influence soil chemical and physical properties. Most soil development occurs in the presence of living organisms. Plants are the sources of organic carbon that enter soils, and functionally different types of plants (e.g., grasses, deciduous trees, evergreen shrubs) strongly influence the amount and especially the depth distribution of soil carbon (Jobbágy and Jackson 2000). Carbon-containing soil organic matter, in turn, influences most functional properties of soils, as described later.

Plants also strongly influence mineral properties of soils. They are geochemical pumps that remove bio-essential elements from soils, store them in tissues, and return them to the soil through litterfall and decomposition (Amundson et al. 2007). In the process, soluble forms of rock-derived minerals such as phosphorus, calcium, potassium, and silicon can be moved upward in the soil profile and are most available in the upper portion of the soil. This is partially offset by downward leaching. Upward movement generally predominates unless minerals precipitate out in less available forms at depth (e.g., calcium in desert soils or iron and aluminum in wet soils), as described later. CO_2 from plant and microbial respiration and the organic acids produced by many plants generate soil acidity that contributes to rock weathering. Vegetation differences in either absorption of minerals or release of organics strongly influences soil properties (see Chap. 7). It is often difficult, however, to separate the chicken from the egg. Did the vegetation determine soil properties or vice versa (Berner et al. 2004; Dietrich and Perron 2006; Amundson et al. 2007)?

One approach to determining vegetation effects on soils has been to plant monocultures or species mixes into initially homogeneous sites. Rapidly growing grasses in a nitrogen-poor perennial grassland enhanced the nitrogen mineralization (or reduced microbial immobilization) of nitrogen by soils within 3 years (see Fig. 11.5; Wedin and Tilman 1990), as did deep-rooted forbs in an annual grassland (Hooper and Vitousek 1998). Another approach is to examine the consequences of species invasions or extinctions on soil processes. The invasion of a non-native nitrogen fixer into Hawaiian rainforests, for example, increased nitrogen inputs to the system more than fivefold, altering the characteristics of soils and the colonization and competitive balance among native plant species (see Fig. 11.3; Vitousek et al. 1987). Yet another approach is to examine weathering and erosion rates in places without biota (Mars or early Precambrian soils) or with minimal biotic

effects (e.g., Antarctic dry valleys; Amundson et al. 2007).

Animals also influence soil properties. Earthworms, termites, and invertebrate shredders, for example, stimulate decomposition (see Chap. 7), thereby modifying soil properties that are influenced by soil organic content. Grazers such as North American bison concentrate sodium in their wallows, which disperses clays and creates water-holding pans. Other grazers like African rhinos generate large dung middens that concentrate nutrients, whereas termites form large termitaria that concentrate soil resources and vertically redistribute nutrients. Microorganisms also influence the structure and properties of soils through the types of organic compounds they release into the soil environment.

Human Activities

Over the past 40 years, the doubling of human population and associated agricultural and industrial activities have strongly influenced soil development worldwide. Human activities directly influence soils through changes in nutrient inputs, irrigation, alteration of soil microenvironment, and increased erosional loss of soils. Human activities also indirectly affect soils through changes in other drivers, including changes in atmospheric composition and the additions and deletions of species.

Controls Over Soil Loss

Soil formation depends on the balance between deposition, erosion, and soil development (i.e., the changes that soils undergo in place). Soil thickness varies with hillslope position, with erosion dominating on steep slopes, deposition in valley bottoms, and soil development on gentle slopes and terraces where the lateral transport of materials is minimal (Fig. 3.4). Much of Earth's surface is in hilly or mountainous terrain where erosion and deposition are important processes. Erosion removes the products of weathering and biological activity. In young soils, erosional

losses reduce soil fertility by removing clays and organic matter that store water and nutrients. On highly weathered landscapes, however, erosion renews soil fertility by removing the highly weathered remnants (sands and iron oxides) that contribute little to soil fertility and exposing less weathered materials that provide a new source of essential nutrients (Porder et al. 2005).

The dominant erosional processes depend on topography, the properties of surface materials, and the pathways by which water leaves the landscape. Mass wasting is a major erosional process in most regions. This is the downslope movement of soil or rock material under the influence of gravity without the direct aid of other media such as water, air, or ice. Mass wasting includes both fine-scale processes such as the movement of individual soil particles (**soil creep**) and massive events such as landslides or debris flows that can rapidly transport cubic meters to cubic kilometers of material. Mass wasting occurs most rapidly on steep slopes, regardless of the underlying mechanism. Any process that moves soil particles (e.g., freeze–thaw events or animal burrowing) contributes to their net downhill movement. Erosion caused by soil creep is the aggregate result of millions of tiny events. Gophers, for example, as a result of their preference for deep soils, burrow more actively and increase erosion from deep soils, reducing the variability in soil thickness across landscapes (Yoo et al. 2005). Landslides, on the other hand, are rare but massive events. The probability of a landslide depends on the **shear stress** that the soil experiences, i.e., the force parallel to the slope that drives mass wasting events such as landslides. It is the balance between the gravitational driving force for downslope movement (F_t) and the friction that resists this movement (F_n; Fig. 3.6).

Many factors influence the **shear strength** of a soil mass (i.e., the shear stress that a soil can sustain without slope failure; Selby 1993). Sometimes the sliding friction between the material and some well-defined plane (such as a frozen soil layer) determines whether a landslide occurs. More commonly, however, it is the internal friction among individual components within the soil
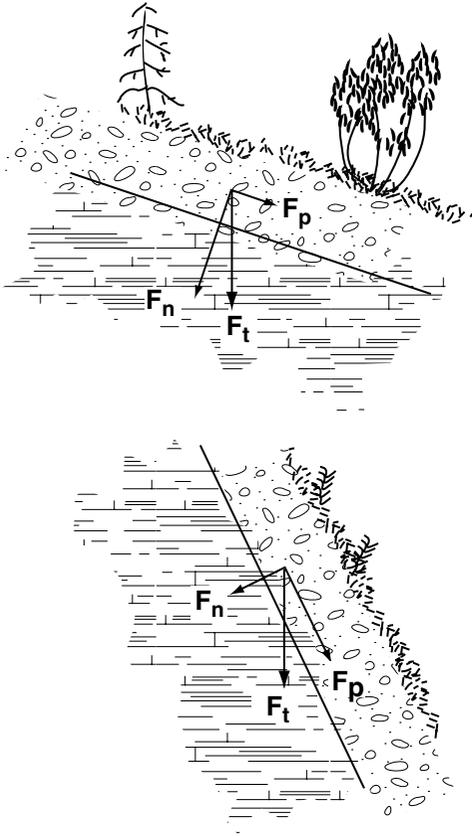
Fig. 3.6 The effect of slope angle on the partitioning of the total gravitational force (F_t) into a component (F_n) that is normal (perpendicular) to the slope (and therefore contributes to friction that resists erosion) and a component (F_p, shear stress) that is parallel to the slope (and therefore promotes erosion). Steep slopes have a larger value of F_p and lower values of F_n and therefore a greater tendency for mass wasting

matrix that largely determines its resistance to mass wasting. Cohesion among soil particles and water molecules enhances the internal friction that resists mass wasting. A small amount of water enhances cohesion among particles, explaining why sand castles are easier to make with moist than with dry sand. High water content, however, increases the weight of the soil, makes soil grains more buoyant, and reduces the frictional strength. Wet soils become unstable, leading to liquefaction of the soil mass, which can flow down slope. Fine-particle soils have lower slope thresholds of instability and are more likely to lead to slope failure than are coarse-textured

soils. Roots also increase the resistance of soils to downslope movement, so deforestation and other land-use changes that reduce root biomass increase the probability of landslides.

The pathways by which water leaves the landscape strongly influence erosion. Water can leave a landscape via several pathways: evaporation and transpiration to the atmosphere, groundwater flow, shallow subsurface flow, and overland flow (when precipitation exceeds infiltration rate; see Fig. 4.4). The relative importance of these pathways depends on topography, vegetation, and material properties such as the hydraulic conductivity of soils. Groundwater and shallow subsurface flow dissolve and remove ions and small particles. At the opposite extreme, overland flow causes erosion primarily by surface sheet wash, rills, and rain splash. This often occurs in sparsely vegetated arid and semi-arid soil-mantled landscapes and on disturbed ground. Overland flow rates of 0.15–3 cm s^{-1} are enough to suspend clay and silt particles and move them downhill (Selby 1993). As water collects into gullies, its velocity, and therefore erosion potential, increases. A doubling of velocity causes a 60-fold increase in the size of particles that can be eroded. Vegetation and a litter layers greatly increase infiltration into the soil by reducing the velocity with which raindrops hit the soil, thereby preventing surface compaction by raindrops. Vegetated soils are also less compact because roots and soil animals create channels in the soil. In these ways, vegetation and a litter layer substantially increase infiltration and therefore groundwater and subsurface flow.

High wind speeds at the soil surface are another important agent of erosion. This often occurs after vegetation removal. Some agricultural areas in China have lost meters of soil to wind erosion and have become a major source of iron to phytoplankton in the Pacific Ocean (see Chap. 9).

Streams and rivers play an important role in soil redistribution across landscapes. At the scale of large river basins, three broad geomorphic zones can be identified (Naiman et al. 2005): an **erosional zone**, where erosion dominates over deposition, a **transfer zone**, where erosion and
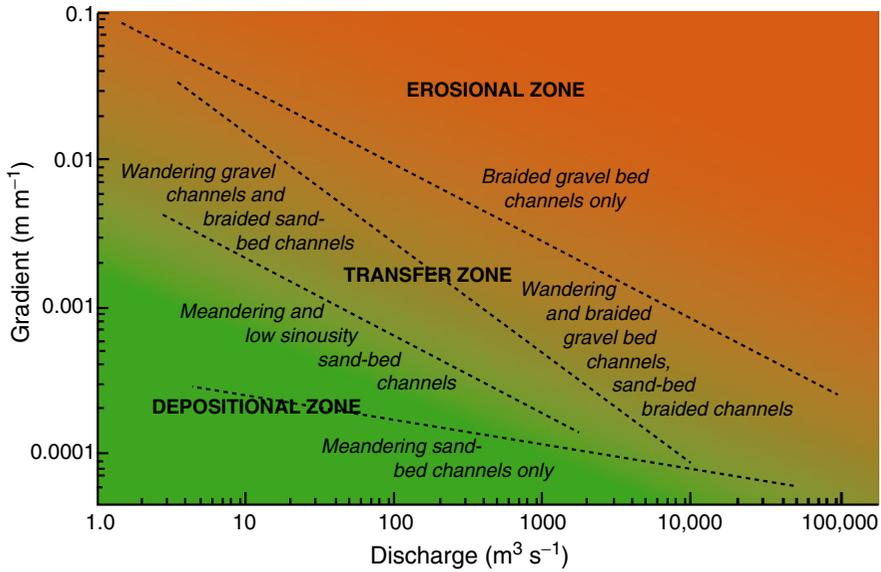
Fig. 3.7 Effects of river discharge and elevational gradient on patterns of river channel morphology. There is a gradual transition from a predominance of erosion in streams with steep gradients or high discharge (Erosional Zone shown in *red*) to a predominance of deposition in streams with shallow gradients or low discharge (Depositional Zone, shown in *green*). Modified from Church (2002)

deposition are in dynamic balance over long timescales, and a **depositional zone**, where deposition rate exceeds erosion rate and the capacity of the river to transport suspended materials (Fig. 3.7). Most sediments delivered to the ocean originate in the erosional zone (Milliman and Syvitski 1992). Here, slopes become steeper as headwater streams downcut into their beds, increasing the shear stress on adjacent soils and the rate of mass wasting. As materials are delivered to the stream by mass wasting and erosion of the streambed, they are transported downstream at a rate that depends on flow velocity and grain size of the sediments, with fine particles moving downstream faster than gravel and boulders. Glaciers, mining, or vegetation removal substantially augments sediment delivery in the erosional zone.

In the transfer zone, there is less delivery of primary sediments to the stream or river and the dominant processes are the sorting of sediments according to grain size and the downstream transport of materials as a result of a balance of erosion and deposition. When stream energy

increases, for example during a flood, progressively larger particles are mobilized, and, as river energy declines, the larger particles are deposited first. This produces a heterogeneous patchwork of gravel bars, sand bars, and silt-filled side channels (Naiman et al. 2005). Stream energy, and therefore the size of particles transported, is greater during flood events, in steep gradients (e.g., riffles), and in deep narrow channels. The transfer zone that links zones of erosion and deposition may shift through time as a result of (1) mountain uplift or sea-level change, which together determine the vertical gradient in the river basin; (2) discharge, which depends on prevailing climate and water inputs or removals from streams; and (3) sediment inputs, which may be influenced by human activities and other factors. Floodplains form during periods when deposition predominates, and channel incision occurs when erosion predominates.

In the depositional zone, rivers tend to meander and develop broad alluvial floodplains and deltas. Rivers in the depositional zone tend to show larger peak discharges (floods) than upstream

Table 3.1 Climatic and topographic effects on long-term erosion rates

Climate zone	Relief	Erosion rate[a] (mm century^{-1})
Glacial	Gentle (ice sheets)	5–20
	Steep (valleys)	100–500
Polar montane	Steep	1–100
Temperate maritime	Mostly gentle	0.5–10
Temperate continental	Gentle	1–10
	Steep	10–20+
Mediterranean	–	1–?
Semi-arid	Gentle	10–100
Arid	–	1–?
Wet subtropics	–	1–100?
Wet tropics	Gentle	1–10
	Steep	10–100

Data from Selby (1993)

[a]Erosion rates are estimated from average sediment yields of rivers in different climatic and topographic regimes. Extreme uncertainty in maximum values is indicated (?)

because of the accumulation of water from a large drainage basin into a single channel. During these floods, the river overflows its banks and fills low-lying areas. Flooding accounts for most of the deposition in this river zone. In the Amazon, for example, more sediment is transported laterally to the floodplain than to the ocean (Dunne et al. 1998). Other finer-scale dynamics that occur within the floodplain involve the erosion of sediments on the outer bends of meanders, where river velocity is greatest, and deposition as new sand or silt bars on the inner sides of river bends. These dynamics cause the river to redistribute materials within the floodplain, creating habitat mosaics of different-aged stands.

Sediments that enter the ocean are deposited near the river mouth, forming a delta or tidal mudflats or are redistributed by coastal currents. Soft (non-rocky) coastlines, including sandy beaches and barrier islands are maintained by the dynamic balance between the delivery of sediments to the coastal zone, their horizontal redistribution by coastal currents and storms, and export (particularly of fine particles offshore). Dredging of harbors to maintain shipping channels and "armoring" of coastlines to prevent erosion in one location reduces sediment inputs elsewhere, often with disastrous unintended consequences. Redistributing sediment delivery from the Mississippi River by routing river flow

offshore, for example, contributed to subsidence of wetlands and loss of barrier islands that would otherwise have helped to protect New Orleans during the 2005 Hurricane Katrina.

Erosion of landscapes results from the combined action of wind, water, ice, and mass wasting. On average, erosion of terrestrial material to the ocean is about 1–10 mm century^{-1} (Selby 1993). However, erosion rates vary regionally by two to three orders of magnitude, depending on topography, climate, human activities, and the sensitivity of rocks and soils to erosion (Table 3.1). Erosion rates tend to approach rates of tectonic uplift, so regions with active tectonic uplift and steep slopes generally have higher erosion rates than flat, weathered terrain. Climate influences erosion primarily through its effects on vegetation cover. In arid, semi-arid, and polar regions with minimal vegetation, for example, surface wash from raindrop impacts and overland flow during intense rains cause most erosion. In contrast, ecosystems with greater vegetative cover lose material primarily through the dissolution of rocks (weathering) to produce soluble compounds that leach out of the system. Low vegetation cover also makes lands more prone to soil loss from wind erosion. The contribution of large rare events like landslides to long-term erosion rates is poorly known. They may be more important in redistributing materials within a drainage

basin than in causing loss from the land to the ocean. For example, 90% of the materials eroded from the upland Piedmont region in the southeastern U.S. since 1700 is still stored on hillslopes, valley bottoms, and reservoirs (Selby 1993). At a global scale, human activities have increased erosion and sediment flux in rivers by 2.3 billion metric tons per year, but have reduced sediment flux to the ocean by 1.4 billion metric tons per year because of sediment trapping in reservoirs (Syvitski et al. 2005). These patterns are regionally variable, however. Indonesia, for example, has considerable land-use change and sediment transport but very few reservoirs to prevent these sediments from reaching the ocean. Much of the erosion on natural landscapes probably occurs during high-rainfall events or after disturbances have reduced vegetation cover rather than during average conditions.

Development of Soil Profiles

Soils develop through the additions of materials to the system, transformations of those materials within the system, transfers down and up in the soil profile, and losses of materials from the system (Fig. 3.8; Richter and Markewitz 2001).

Additions to Soils

Direct inputs to the soil system come from both outside and inside the ecosystem. Inputs from *outside* the ecosystem come from precipitation and wind, which deposit ions and dust particles, and floods and tidal exchange, which deposit sediments and solutes (see Chap. 9). The source of these materials determines their size distribution and chemistry, leading to the development of soils with specific textural and chemical characteristics. Sometimes these inputs are huge, for example, hundreds to thousands of g m^{-2} of dust inputs to loess-accumulating regions of North America and Asia during the Pleistocene (Sun et al. 2000; Bettis et al. 2003). Organisms *within* the ecosystem add organic matter and nitrogen to
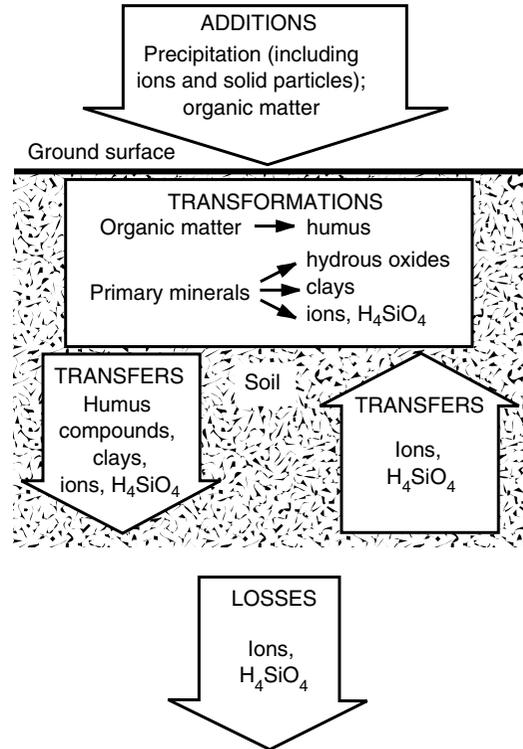

Fig. 3.8 Processes leading to additions, transformations, transfers, and losses of materials from soils. Redrawn from Birkeland (1999)

the soil as dead organic matter, including the above- and belowground portions of plants, animals, and soil microbes.

Soil Transformations

Within the soil, materials are transformed through an interaction of physical, chemical, and biological processes. Freshly deposited dead organic matter is transformed in the soil by **decomposition** to soil organic matter, releasing carbon dioxide and nutrients such as nitrogen and phosphorus (see Chap. 7). Recalcitrant plant and microbial organic compounds undergo physicochemical interactions with soil minerals that contribute to the long-term storage of soil organic matter. The quantity of soil carbon in deep soils, for example, correlates more closely with clay content than with climate (Jobbágy and Jackson 2000).

Weathering is the change of parent rocks and minerals to produce more stable forms. This occurs when rocks and minerals become exposed to physical and chemical conditions different from those under which they formed (Ugolini and Spaltenstein 1992). Weathering involves both physical and chemical processes and is influenced by characteristics of the parent material, environmental conditions (temperature and moisture), and the activities of organisms. **Physical weathering** is the fragmentation of parent material without chemical change. This can occur when rocks are fractured by expansion and contraction during cycles of freeze–thaw, heating–cooling, or wetting–drying or when roots grow into rock fissures. Fire, for example, is a potent force for physical weathering because it rapidly heats exposed rock surfaces while leaving the deeper layers cool. In addition, soil particles and rock fragments are abraded by wind, or ground against one another by glaciers, landslides, or floods. Physical weathering is especially important in extreme and highly seasonal climates. Wherever it occurs, it opens channels in rocks for penetration by water and air and increases the surface area for chemical weathering reactions.

Chemical weathering occurs when parent rock materials react with acidic or oxidizing substances, usually in the presence of water. During chemical weathering, **primary minerals** (unmodified minerals present in the rock or unconsolidated parent material) dissolve, releasing ions and forming **secondary minerals** (insoluble reaction products of weathering). Chemical weathering most commonly involves the reaction of water and acid on a mineral. Carbonic acid is the most important of these acids. It forms through the reaction of CO_2 with water and then ionizes to produce a hydrogen ion and a bicarbonate ion. The CO_2 concentration in soil, which drives the formation of carbonic acid, is 10- to 500-fold higher than in air, due to the **respiration** (CO_2 production) by plants, soil animals, and microbes and the low diffusivity of gases in soil. Weathering rates are particularly high adjacent to roots because high rates of biological activity produce abundant CO_2 and organic acids in the **rhizosphere**, the zone of soil that is directly influenced by roots. Carbonic acid, for example, attacks potassium feldspar, which is converted to a secondary mineral, kaolinite by the removal of soluble silica and potassium (3.1).

$$2KAlSi_3O_8 + 2(H^+ + HCO_3^-) + H_2O \rightarrow Al_2Si_2O_5(OH)_4 + 4SiO_2 + 2K^+ + 2HCO_3^- \tag{3.1}$$
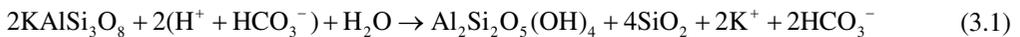
Other sources of acidity that promote chemical weathering include organic acids, nitric acid, sulfuric acid, and the hydrogen ions excreted by plant roots when cations are absorbed (Richter and Markewitz 2001). Plant roots and microbes secrete many organic acids into the soil, which influence chemical weathering through their contribution to soil acidity and their capacity to **chelate** ions. In the chelation process, organic acids combine with metallic ions, such as Fe^{3+} and Al^{3+}, making them soluble and mobile. Chelation lowers the concentration of unchelated inorganic ions at the mineral surface, so dissolved and primary mineral forms are no longer in equilibrium with one another. This accelerates the rate of weathering.

Warm climates promote chemical weathering because temperature speeds chemical reactions and enhances the activities of plants and microbes. Wet conditions promote weathering through their

direct effects on weathering reactions and their effects on biological processes. Not surprisingly, the hot, wet conditions of the humid tropics yield the highest rates of chemical weathering.

The physical and chemical properties of rock minerals determine their susceptibility to weathering and the chemical products that result. Sedimentary rocks like shale that form by chemical precipitation, for example, have more basic cations like calcium (Ca^{2+}), sodium (Na^+), and potassium (K^+) than do igneous rocks. Sedimentary rocks tend to produce soils with a relatively high pH and a high capacity to supply mineral cations to plants. Igneous rocks form more acidic soils.
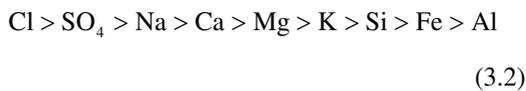
Minerals weather in the same order in which they crystallized during formation (Schlesinger 1997; Birkeland 1999). Olivine, for example, is one of the first minerals to crystallize as magma

Table 3.2 Stability of common minerals under weathering conditions at Earth's surface

Most stable	Fe^{3+} oxides	Secondary mineral
	Al^{3+} oxides	Secondary mineral
	Quartz	Primary mineral
	Clay minerals	Secondary mineral
	K$^+$ feldspar	Primary mineral
	Na$^+$ feldspar	Primary mineral
	Ca^{2+} feldspar	Primary mineral
Least stable	Olivine	Primary mineral

Data from Press and Siever (1986)

$\bigcirc$ and $\odot$ = Oxygen

$\circ$ and $\bullet$ = Silicon

$\bigcirc$ and $\odot$ = Hydroxyl

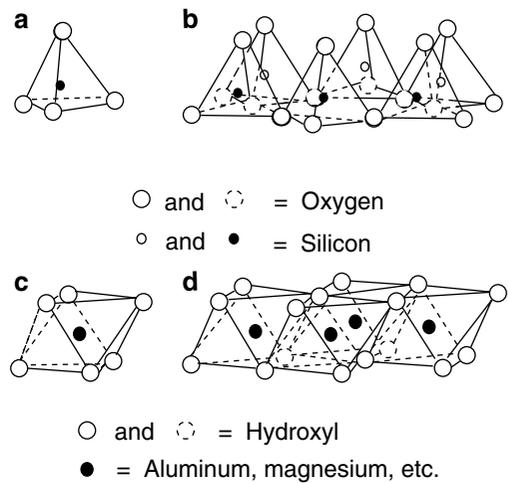$\bullet$ = Aluminum, magnesium, etc.

Fig. 3.9 Diagram showing the molecular structure of a simple clay layer: (**a**) a tetrahedral unit, (**b**) a tetrahedral sheet, (**c**) an octahedral unit, and (**d**) an octahedral sheet. Redrawn from Grim (1968)

cools, forms relatively few bonds, and weathers easily. Feldspar forms and weathers more slowly than olivine, and quartz is one of the last minerals to crystallize, has strong bonds that create a crystalline structure, and is highly resistant to weathering (Table 3.2). Secondary minerals such as the silicate clay minerals and iron and aluminum oxides are among the most resistant minerals to weathering. Differences among elements in their susceptibility to weathering and solubility in water leads to the following sequence in which elements are weathered from rocks and leached into rivers:

$$Cl > SO_4 > Na > Ca > Mg > K > Si > Fe > Al$$

(3.2)

Moderately weathered soils therefore have relatively high concentrations of Ca$^+$, Mg$^+$, and K$^+$ (elements essential for plant growth) and low concentrations of soluble Al^{3+} (a slowly weathered element often toxic to plants). In contrast, in the ancient soils of the wet tropics, the relatively mobile ions of Si and Mg^{2+} as well as Ca^{2+}, K$^+$, and Na$^+$ have leached away, leaving behind the less mobile ions of Al^{3+} and Fe^{3+}.

The secondary minerals formed in weathering reactions play critical roles in soils and ecosystem processes. Insoluble products of chemical weathering are fine clay particles consisting of hydrated silicates of aluminum, iron, and magnesium arranged in layers (sheets). Two types of sheets make up these minerals: A tetrahedral sheet consists of units with one silicon atom surrounded by four O$^-$ groups (Fig. 3.9a). An octahedral sheet consists of units with six O$^-$ or OH$^-$ groups surrounding an Al^{3+}, Mg^{2+}, or Fe^{3+} ion (Fig. 3.9c). Various combinations of these

sheets give rise to a wide variety of clay minerals with different exchange properties. Montmorillonite or illite, for example, which have 2:1 ratios of silica- to aluminum-dominated layers, have a higher **cation exchange capacity** (CEC) than does kaolinite, which has a 1:1 ratio of silica- to aluminum-dominated layers (Fig. 3.10). Some exchange sites on soil minerals, particularly silicate clays with surface oxygen layers, have a permanently fixed charge. Other exchange sites, particularly iron and aluminum clays with surface hydroxyl layers, vary between positive and negative depending on pH.

In tropical climates, silica is preferentially leached from secondary clay minerals, producing red iron and aluminum oxide clays like gibbsite, which has only aluminum-dominated octahedral sheets. Highly weathered minerals dominated by octahedral sheets strongly bind anions like phosphate. In cold, wet climates, however, iron and aluminum are preferentially leached, leaving behind silica-dominated quartz sand. CEC tends to decline with weathering, whereas anion exchange capacity increases (Fig. 3.10), as discussed later. Most soils contain mixtures of several secondary minerals. The structure and concentration of clay minerals strongly influence the CEC, water-holding capacity, and other characteristics of soils.
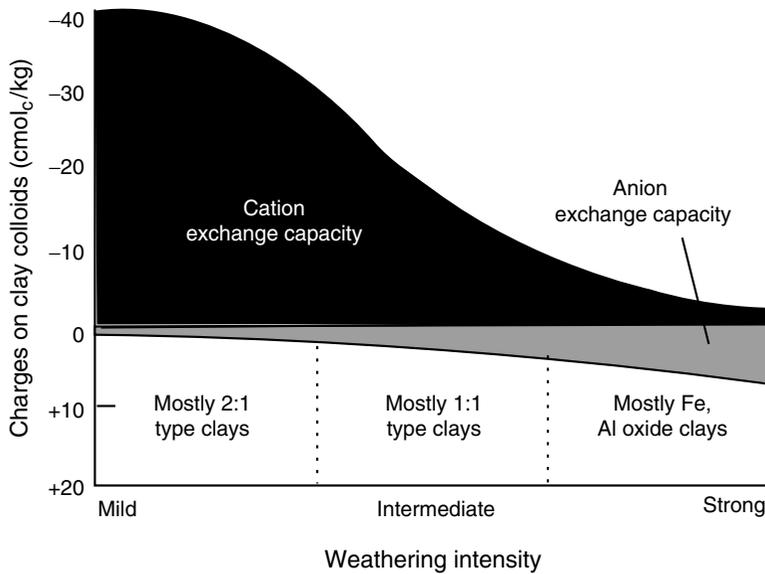
Fig. 3.10 The effect of weathering intensity on the charges on clay minerals and in turn on their cation and anion exchange capacities. Redrawn from Brady and Weil (2008)

Secondary minerals that form in soils can be either **crystalline**, with highly regular arrangements of atoms, as in the silicate clay minerals described earlier, or **amorphous**, with no regular arrangement of atoms. Allophane ($Al_2O_3 \cdot 2SiO_2 \cdot nH_2O$), for example, is an amorphous secondary mineral characteristic of volcanic ash deposits. With time, allophane transforms through loss of silica to crystalline aluminum oxide minerals like gibbsite ($Al(OH)_3$). Allophane has a high anion exchange capacity due to a surplus of positive charges. It also strongly binds phosphorus and can lead to phosphorus limitation in relatively young volcanic soils.

Soil Transfers

Vertical transfers of materials through soils generate distinctive soil profiles, i.e., the vertical layering of soils. These transfers typically occur by **leaching** (the downward movement of dissolved materials) and particulate transport in water. Soluble ions that are added in precipitation or released by weathering in upper layers of the soil profile can move downward in solution until a change in chemical environment causes them to become reactants in chemical processes, leading to insoluble products, or until dehydration causes them to precipitate out of solution. The quantity of base cations in secondary minerals therefore often increases with depth within the upper meter of soil. These cations are leached from upper layers (termed **horizons**) and form new minerals under the new conditions of pH and ionic content encountered at depth. Chelated complexes of organic compounds and iron or aluminum ions are also water soluble and can move in water to deeper layers of the soil profile. Slight changes in ionic content or the microbial breakdown of the organic matter are among the processes that can cause the metal ions to precipitate as oxides. Clay particles like silicates and iron and aluminum oxides can also be transported downward in solution, sometimes forming deep horizons with high clay content in wet climates. Soil texture affects the rate and depth of leaching and thus the translocation and accumulation of materials in soil profiles. Constituents released during weathering of coarse-textured glacial till, for example, may be leached from the soil before they have a chance to chemically react to form secondary minerals.

Soils of arid and semi-arid environments also accumulate materials in specific horizons. These systems often have a hard calcium carbonate-rich

calcic horizon or caliche. Downward-moving soil water carries dissolved Ca^{2+} and bicarbonate (HCO_3^-). Precipitation as calcium carbonate occurs under conditions of increasing pH, which drives reaction (3.3) to the left. Precipitation can also occur under saturating concentrations of carbonate, or with evaporation of soil water.

$$CaCO_3 + H_2CO_3 \leftrightarrow Ca^{2+} + 2HCO_3^- \quad (3.3)$$

Although most of the transfers in soils occur through the downward movement of water, materials can also move upward in water. The capillary rise of water from a shallow water table, for example, transfers water and ions from lower to upper soil layers (see Chap. 4). Because capillary water movement depends on adhesive properties of soil particles, the potential distance for capillary rise is greater in clay soils with small pore sizes than in sandy soils (Birkeland 1999), as explained later. Soluble ions or compounds may accumulate in layers at the top of the capillary fringe. **Salt pans**, for example, form at the soil surface in low-lying areas of deserts, forming extensive **salt flats**, where the water evaporates rather than running off. Minerals that are added to soils in irrigation water in dry regions can also accumulate at the soil surface, as the water evaporates. This **salinization** has led to widespread abandonment of farmland in dry regions of the world, as in many parts of Australia.

Some minerals accumulate at the interface between waterlogged and aerobic soils. Poor drainage often leads to low oxygen availability because oxygen diffuses 10,000 times more slowly in water than in air and is easily depleted in waterlogged soils by root and microbial respiration. Low oxygen concentration creates reducing conditions that convert ions with multiple oxidation states to their reduced forms. Iron and manganese, for example, are more soluble in their reduced states (Fe^{2+} and Mn^{2+}, respectively) than in their oxidized states (Fe^{3+} and Mn^{4+}, respectively). Fe^{2+} and Mn^{2+} diffuse through waterlogged soils to the surface of the water table, where there is enough oxygen to convert them to their oxidized forms. Here they precipitate out of solution to form a distinct iron- and manganese-rich layer. This layering of iron and manganese is particularly pronounced in lake sediments where there is a strong gradient in oxygen concentration from the sediment surface. The conversion from ferric (Fe^{3+}) to ferrous (Fe^{2+}) iron gives rise to the characteristic gray and bluish colors of waterlogged **gley soils**.

Soils that are subjected to repeated wetting and drying and saturation during some seasons can also develop characteristic accumulations of minerals. **Plinthite**, for example, is an iron- and aluminum-rich material in tropical soils that can harden irreversibly with repeated cycles of wetting and drying. Depending on their location within the profile, these layers can impede water drainage and root growth.

The actions of plant roots and soil animals transfer materials up and down the soil profile (Paton et al. 1995). Organic matter inputs to soil occur primarily at the surface and in upper soil horizons. When leaves or roots are shed or plants die, the minerals acquired by deep roots are also deposited on or near the soil surface. This contributes, for example, to the base-rich soils and unique ground flora beneath deep-rooted oak trees in southern Sweden (Andersson 1991) or dogwood trees in the eastern U.S. (Thomas 1969). Tree windthrow, which occurs when large trees are toppled by strong winds, also redistributes roots and associated soil upward. Finally, animals such as gophers transfer materials up and down in the soil profile as they tunnel and feed on plant roots. Earthworms in temperate soils and termites in tropical soils are particularly important in transferring surface organic matter deep into the soil profile and, at the same time, bringing mineral soil from depth to the surface. These processes play critical roles in the redistribution of nutrients and in the control of net primary productivity.

Losses from Soils

Materials are lost from soil profiles primarily as solutions and gases. The quantity of minerals leached from an ecosystem depends on both the amount of water flowing through the soil profile and its solute concentration. Many factors influence these concentrations, including plant

demand, microbial mineralization rate, cation or anion exchange capacity, and previous losses via leaching or gas fluxes. As water moves through the soil, exchange reactions with mineral and organic surfaces replace loosely bound ions on the exchange complex with ions that bind more tightly, as explained later. In this way, **monovalent** (ions with a single charge) cations such as Na^+, NH_4^+, and K^+ and anions such as Cl^- and NO_3^- are easily released from the exchange complex into the soil solution and are particularly prone to leaching loss. The maintenance of charge balance of soil solutions requires that the leaching of negatively charged ions (**anions**) be accompanied by an equal charge of positive ions (**cations**). Inputs of H_2SO_4 in acid rain therefore increase leaching losses of readily exchangeable base cations like Na^+, NH_4^+, and K^+, which leach downward with SO_4^{2-}.

Materials can also be lost from soils as gases. Gas emissions depend on the rate of gas production by microbes, the diffusional paths through soils, and the exchange at the soil–air interface (Livingston and Hutchinson 1995). The controls over these losses are discussed in Chap. 9.

Soil Horizons and Soil Classification

Ecosystem differences in additions, transformations, transfers, accumulations, and loss give rise to distinct soils and soil profiles. Soils include organic, mineral, gaseous, and aqueous constituents arranged in a relatively predictable vertical structure. The number and depth of horizons (layers) and the characteristics of each layer in a soil profile vary widely among soils. Nonetheless, a series of horizons can be described that is typical of many soils (Fig. 3.11). The organic or **O horizon** of soil consists of organic material that accumulates above the mineral soil. This organic layer is derived from the **litter** of dead plants and animals and can be subdivided based on the degree of decomposition that most material has undergone, with the lower portion of the organic horizon being more decomposed. The **A horizon** is the uppermost mineral soil horizon. Being adjacent to the O horizon, it typi-

cally contains substantial organic matter and is therefore dark in color. The O and A horizons are the zones of most active plant and microbial processes and therefore have highest nutrient supply rates (see Chap. 9). Many soils in wet climates have an **E horizon** beneath the A horizon that is strongly leached. Most clay minerals and iron and aluminum oxides have been leached from the horizon, leaving behind resistant minerals like quartz, among other sand and silt-size particles. The **B horizon** beneath the A and E is the zone of maximum accumulation of iron and aluminum oxides and clays. Salts and precipitates sometimes also accumulate here, especially in arid and semi-arid environments. The **C horizon** lies beneath the A and B horizons. Although it may accumulate some of the leached material from above, it is relatively unaffected by soil-forming processes and typically includes a significant portion of unweathered parent material. Finally, at some depth, there is an unweathered layer of bedrock (R). Leaching and cation loss predominate in wet environments, producing acid soils. Salt inputs and accumulation predominate in dry environments, producing basic soils.

Despite the large variation among the world's soils, they can be classified into major groups that have formed in response to similar soil-forming factors and processes and therefore share many of the same properties. Soil classification systems rely on the diagnostic characteristics of specific horizons and on organic matter content, base saturation, and properties that indicate wetness or dryness. The soil taxonomy used in the U.S. recognizes 12 major soil groups, called **soil orders** (Table 3.3). Most agronomic and ecosystem studies classify soils to the level of a **soil series**, a group of soil profiles with similar profile characteristics such as type, thickness, and properties of the soil horizons. Soil series can be further subdivided into **types** based on the texture of the A horizon, and into **phases** based on information such as landscape position, stoniness, and salinity. A comparison of soil profiles from the major soil orders illustrates the impact of different climatic regimes on soil development (Figs. 3.12 and 3.13). More detailed descriptions of soil orders are presented by Brady and Weil (2008).

Fig. 3.11 A generic soil profile, showing the major horizons that are formed during soil development. Density of dots reflects concentration of soil organic matter
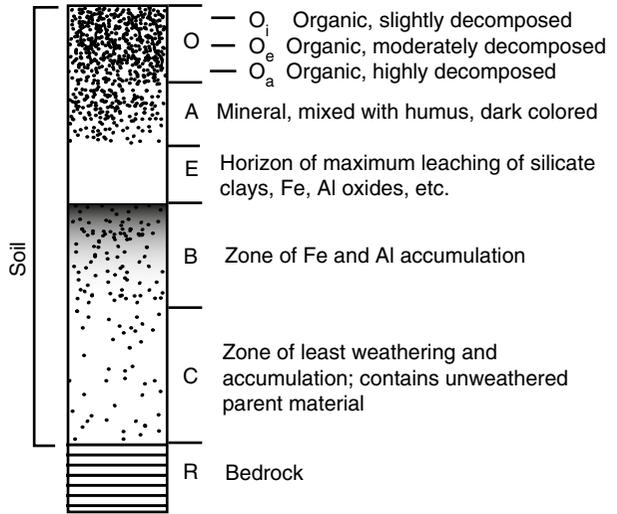

O — O_i Organic, slightly decomposed
  — O_e Organic, moderately decomposed
  — O_a Organic, highly decomposed

A Mineral, mixed with humus, dark colored

E Horizon of maximum leaching of silicate clays, Fe, Al oxides, etc.

B Zone of Fe and Al accumulation

C Zone of least weathering and accumulation; contains unweathered parent material

R Bedrock

Table 3.3 Names of the soil orders in the U.S. soil taxonomy and their characteristics and typical locations

Soil order	Area (% of ice-free land)	Major characteristics	Typical occurrence
Rock and sand	14.1		
Entisols	16.3	No well-developed horizons	Sand deposits, plowed fields
Inceptisols	9.9	Weakly developed soils	Young or eroded soils
Histosols	1.2	Highly organic; low oxygen	Peatland, bog
Gelisol	8.6	Presence of permafrost	Tundra, boreal forest
Andisols	0.7	From volcanic ejecta; moderately developed horizons	Recent volcanic areas
Aridisols	12.1	Dry soils with little leaching	Arid areas
Mollisols	6.9	Deep, dark-colored A horizon with >50% base saturation	Grasslands, some deciduous forests
Vertisols	2.4	High content (>30%) of swelling clays; crack deeply when dry	Grassland with distinct wet and dry seasons
Alfisols	9.7	Enough precipitation to leach clays into a B horizon; >50% base saturation	Moist forests; shrublands
Spodosols	2.6	Sandy leached (E) horizon; acidic B horizon; surface organic accumulation	Cold, wet climates, usually beneath conifer forests
Ultisols	8.5	Clay-rich B horizon, low base saturation	Wet tropical/subtropical climate; forest or savanna
Oxisols	7.6	Highly leached horizon on old landforms	Hot, humid tropics beneath forests

Data from Miller and Donahue (1990) and Brady and Weil (2008)

Entisols are soils with minimal soil development. They occur either because the soils are recent, or processes that disrupt soil structure dominate over soil-forming processes. This is the most widespread soil type in the world, occupying 16% of the ice-free surface. **Inceptisols**, in which the soil profile has only begun to develop, occupy an additional 10% of

the ice-free surface. Thus, including rock and shifting sand, about 40% of the ice-free surface of Earth shows minimal soil development (Table 3.3; Fig. 3.12).

Histosols are highly organic soils that develop in any climate zone under waterlogged conditions that restrict oxygen diffusion into the soil, leading to slow rates of decomposition and accumulation
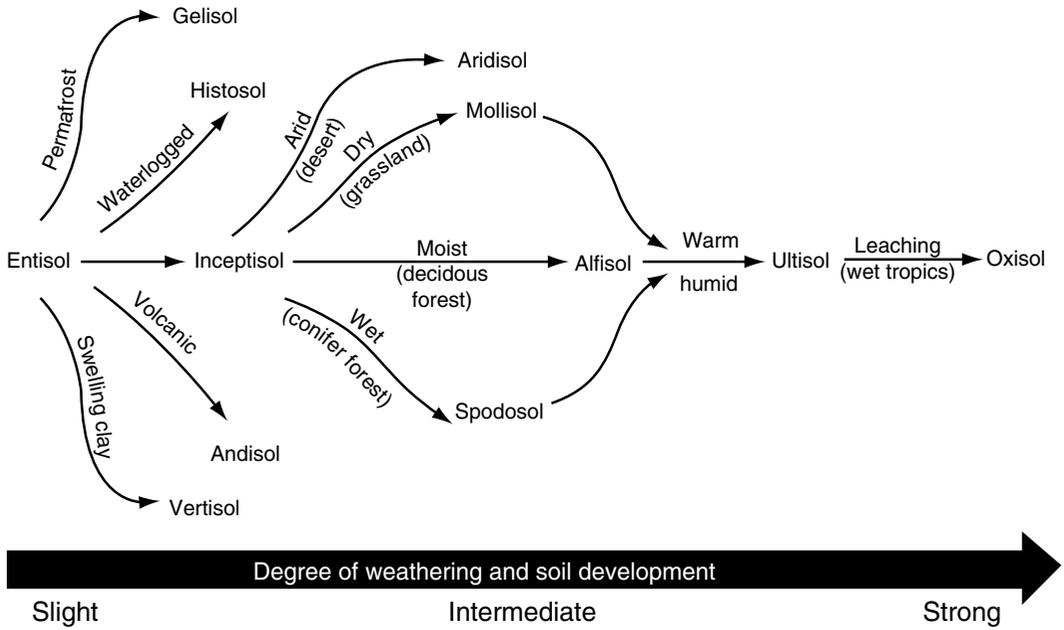
Fig. 3.12 Relationships among the major soil orders, showing the conditions under which they form, relative time required for formation, and the types of ecosystems with which they are most commonly associated. Based on Birkeland (1999) and Brady and Weil (2008)
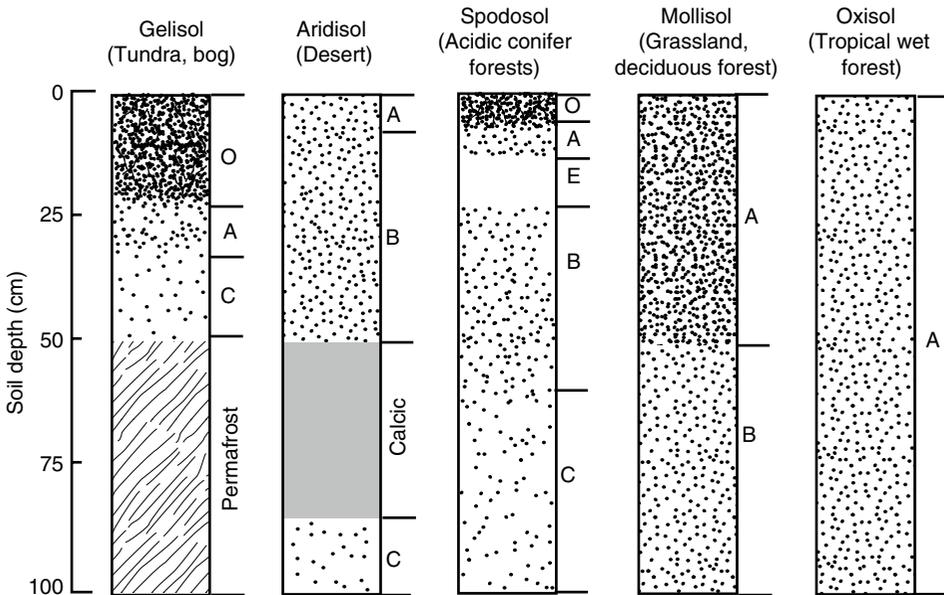

Fig. 3.13 Typical profiles of five contrasting soil orders, showing differences in the types and depths of horizons. Symbols as in Fig. 3.11

Fig. 3.14 Diagram showing the general soil moisture and temperature regimes that characterize the most extensive soils of seven soil orders. Soils of other soil orders (Andisols, Entisols, Inceptisols, and Histosols) can occur across this entire spectrum of environmental conditions. Vertisols (not shown) occur only where clay materials are abundant, under intermediate temperature, and moisture conditions. Data from Brady and Weil (2008)
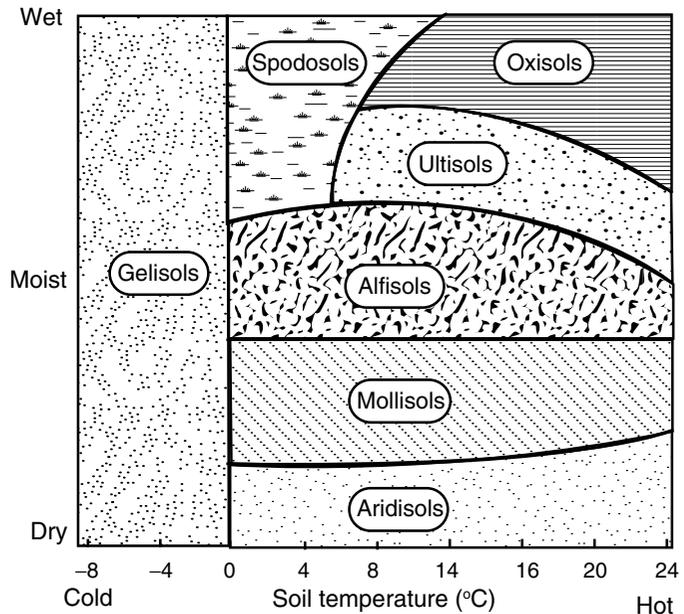
of organic matter. Histosols have a well-developed O horizon of undecomposed organic material where most plants are rooted. The high water table prevents the vertical leaching required for soil development, so these soils have weak development of mineral soil horizons. **Gelisols** are soils that develop in climates with an average annual temperature below 0°C that are underlain by a layer of permanently frozen soil (permafrost). They typically have a surface organic horizon or are frost churned (Figs. 3.13 and 3.14).

Andisols are young soils on volcanic substrates that tend to produce amorphous clays.

Aridisols, as the name implies, develop in arid climates. The low rainfall minimizes weathering and deep leaching, causing accumulation of soluble salts. There is no surface O horizon. The shallow A horizon has little organic matter due to low productivity and rapid decomposition. Low precipitation results in a poorly developed B horizon. Many of these soils form a calcic layer of calcium and magnesium carbonates that precipitate at depth because there is insufficient water to leach them out of the system. Desert calcic layers can greatly reduce root penetration, restricting the roots of many desert plants to surface soils. Aridisols are a widespread soil type, accounting

for 12% of the terrestrial surface (Miller and Donahue 1990).

Vertisols are characterized by swelling and shrinking clays. These clay-rich soils tend to occur in warm regions with a moist to dry climate, often on limestone or other base-rich parent materials. Vertisols often have no B horizon because the swelling and shrinking leads to a vertically well-mixed soil. Trees are often excluded from vertisols due to the frequent soil disturbance.

Mollisols are fertile soils that develop beneath grasslands and some deciduous forests. They have a deep, organic-rich A horizon with a high nutrient content that grades into a B horizon. Due to their high fertility, mollisols have been extensively cultivated and support the major grain-growing regions of the world. They account for 22% of U.S. soils and 7% of soils worldwide (Miller and Donahue 1990).

Spodosols (or podzols by European terminology) are highly leached soils that develop most commonly in cool, wet climates, usually beneath conifer stands. Beneath the A horizon is usually a highly leached, almost white, E horizon and a dark brown or black B horizon, where leaching products accumulate. These soils are often coarse textured and acidic. **Alfisols** usually develop

beneath temperate and subtropical forests, especially deciduous forests that receive less precipitation. They are less strongly leached than spodosols and have a base-rich zone of clay accumulation in the B horizon.

Ultisols develop in warm, wet climates, where there is substantial leaching. The B horizon of these soils often has a high clay content and a low base saturation. **Oxisols** are the most highly weathered and leached group of soils. They occur on old landforms in the wet tropics. The A horizon is so highly weathered that it contains iron and aluminum oxides, largely as clay particles with very little silica and extremely low fertility. This horizon often extends several meters in depth.

Four generalizations emerge from this broad comparison of soil orders:

1. Nearly half (40%) of Earth's soils show minimal soil development and therefore largely reflect the properties of their parent material and current climate.
2. Wet environments tend to produce acidic leached soils, whereas dry environments produce basic ones in which cations accumulate.
3. Weathering and soil formation occur most rapidly in warm, wet climates, where plant productivity is greatest. Weathering is accentuated with time.
4. The quantity, quality, and turnover rate of soil organic matter are sensitive to climate and strongly influence soil fertility and other soil properties

Soil Properties and Ecosystem Functioning

Soil Physical Properties

Spatial and temporal variations in soil development generate large variations in soil properties. In the following paragraphs, we discuss how the properties of soil particles and the configuration of intervening spaces govern the availability of water and nutrient resources for plant growth and therefore their cycling through ecosystems.

Particle size distribution (**soil texture**) is important because it determines the surface area in a given soil volume. Soil texture is defined by the relative proportion of three sizes of particles: **clay** (<0.002 mm), **silt** (0.002–0.05 mm), and **sand** (0.05–2.0 mm; Fig. 3.15). **Loam** soils, which constitute the majority of soils, are mixtures of these three size classes and exhibit some properties of each size class. Rocks and gravel are larger (>2 mm) particles that also occupy a substantial proportion of the volume of many soils. Most gravel and sand particles are unweathered primary minerals, whereas clay particles are mostly secondary minerals. Silt particles are intermediate in composition (Fig. 3.16).

Soil texture depends on the balance between soil development that occurs in place, deposition by wind or water, and erosional loss of materials. As soils weather in place, the conversion of primary to secondary minerals (mostly small particles) increases the proportion of small soil particles. For this reason, high-latitude soils, with their slow rates of chemical weathering, often have low clay content, often about 10%, compared to temperate or tropical soils. Weathering rate and texture also depend on parent material, as discussed earlier. Small particles are particularly susceptible to erosion by wind or water. Water erosion transports clay from hilltops to valley bottoms, producing fine-textured soils in river valleys and leaving coarser-textured soils on the slopes. If river valleys are poorly vegetated, as in braided rivers that drain glaciated landscapes, wind can then move fine particles back to hillslopes to form **loess** soils with a high silt content. Over millions of years, minerals dissolve and are lost from the soil.

Clay particles have about 10,000 times greater surface area than the same weight of medium-sized sand particles (Brady and Weil 2008). Organic matter also has a high surface-volume ratio. Surface area, in turn, determines the amount of water that adsorbs to particle surfaces and therefore the capacity of soils to retain water. Surface charge and CEC also depend on particle surface area, as described later. Soil texture influences these and so many other important soil characteristics that it is a good general
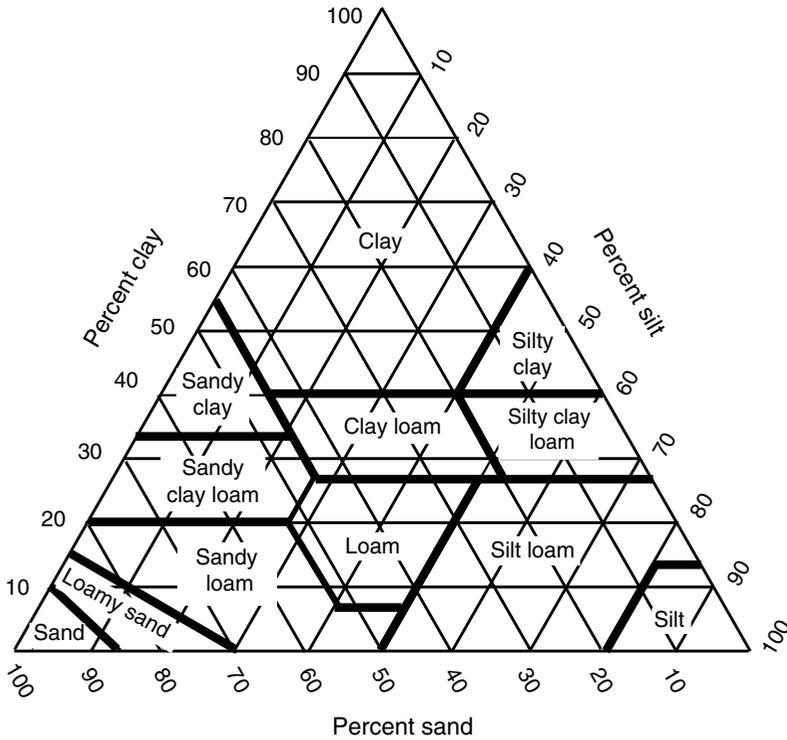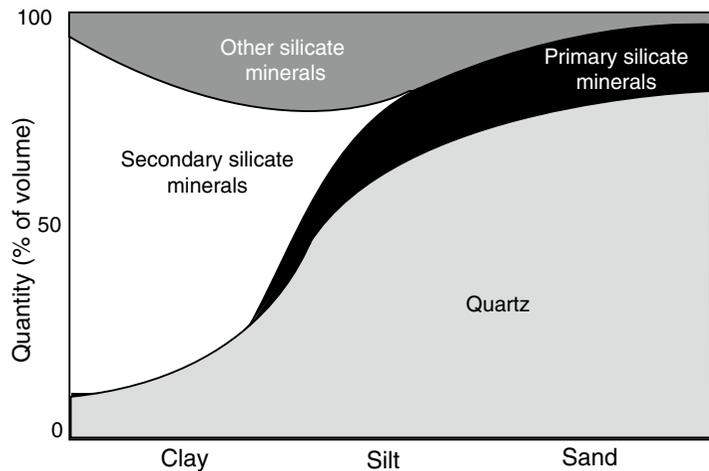
Fig. 3.15 Percentages of sand, silt, and clay in the major soil textural classes. Redrawn from Birkeland (1999)

Fig. 3.16 General relationship between particle size and kinds of minerals present. Redrawn from Brady and Weil (2008)

predictor of many ecosystem properties (Parton et al. 1987).

The physical properties of soils depend on the properties of both particles and the spaces between them. **Bulk density**, the mass of dry soil per unit volume, is an easily measured index of the relative proportion of particles and **voids** (spaces) in the soil. Bulk densities of mineral soil horizons (1.0–2.0 g cm⁻³) are typically higher than those of organic horizons (0.05–0.4 g cm⁻³).

Fig. 3.17 Volume distribution of organic matter, sand, silt, clay, micropores, and macropores in a representative silt loam. Redrawn from Brady and Weil (2008)
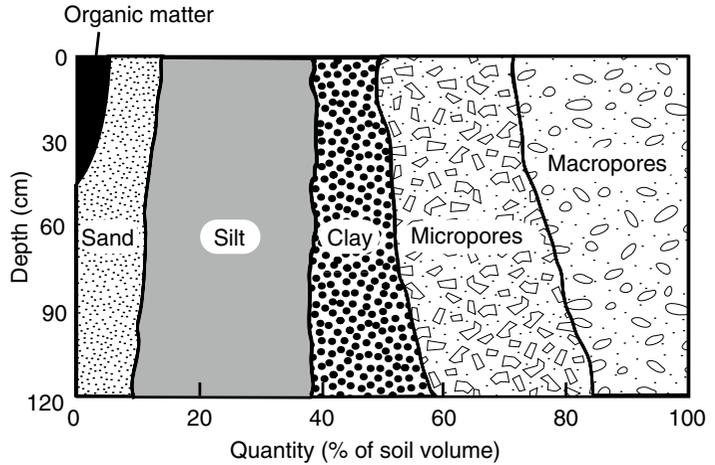
Fine-textured soils usually have aggregated groupings of particles, as described later, with intervening spaces and therefore generally have greater pore volume and lower bulk density than coarse-textured soils. If compacted, however, clay soils can have higher bulk density than coarse-textured soils.

Soil structure reflects the aggregation of soil particles into larger units. **Aggregates** form when soil particles become cemented together and then crack into larger units as soils dry or freeze. Soil aggregates are hierarchically structured with large aggregates (>3 mm diameter) consisting of progressively smaller aggregates and sub-aggregates down to clusters of a few clay and humus particles less than 0.001 mm in diameter. Aggregates form more easily in loam and clay soils than in sandy soils. Materials that glue soil particles together to form aggregates include organic matter, iron oxides, polyvalent cations, and silica. Iron and aluminum oxides are particularly important to aggregate formation in highly weathered tropical soils. For example, iron oxides can cement clay particles to produce very stable **pseudosand** aggregates that improve soil drainage in some clay-rich tropical oxisols and ultisols. In contrast, organic matter, such as polysaccharide secreted by roots and bacteria, is more important in aggregate formation in temperate soils. Fungal hyphae also contribute strongly to aggregation in many soils. For these reasons,

disturbances that reduce soil organic content and its associated microbes can lead to a loss of soil structure, which contributes to further soil degradation. Earthworms and other soil invertebrates contribute to aggregate formation by ingesting soil and producing feces that retain a coherent structure. Plant species and their microbial associates differ in the capacity of their exudates to form aggregates. Some mycorrhizal fungi, for example, produce a glycoprotein **glomalin** that is particularly effective in cementing microaggregates to form macroaggregates (Wilson et al. 2009). In summary, soil texture, mineral chemistry, organic matter content, and species composition all influence soil structure.

The pore structure of soils is critical to their functioning. Pores account for about half the soil volume (Fig. 3.17) and range in size from micropores that are too small for bacteria or root hairs to penetrate to macropores (>0.08 mm diameter). Micropores include both the original spaces between soil particles and the spaces that form as clays swell and shrink due to absorption and loss of water between clay platelets. This swelling and shrinking creates a wide range of pore sizes, from the smallest micropores to large macropores between soil aggregate. Macropores also form when roots and soil animals, especially earthworms, move through the soil, often along previously formed cracks or fractures. The resulting cracks and channels are important pathways for

water infiltration, gas diffusion, and root growth, thus affecting water availability, soil aeration, oxidation–reduction processes, and plant growth. The fine-scale heterogeneity in pore structure is critical to the functioning of soils. Slow gas diffusion through the partially cemented pores within aggregates creates anaerobic conditions immediately adjacent to aerobic surfaces of soil pores. This allows anaerobic processes (e.g., denitrification) that require the products of aerobic processes (nitrification, in this case) to occur even in well-aerated soils (see Chap. 9). The surfaces of macropores are hot spots of biological activity, including the secretion of exudates by roots, rapid growth and turnover of bacterial biofilms, and predation by soil animals (see Chap. 7).

Human activities substantially alter the soil structure of many ecosystems. Compaction by animals and machinery compresses many of the larger cracks and pores between aggregates, reducing oxygen diffusion into the soil. Compaction also reduces infiltration of rainwater, increasing the likelihood of overland flow and erosion. Conversely, plowing mechanically disrupts aggregates, creates new macropores, and disrupts the macropores that were previously present. Depending on the initial condition of the soil, plowing can either improve or degrade soil structure. In native prairie, for example, plowing reduces total pore volume by disrupting aggregates and initial macropores, whereas plowing of compacted soils may increase macropore volume (Brady and Weil 2008). Regardless of the effect of plowing on macropore volume, the breaking up of previously cemented aggregates increases oxygen diffusion and decomposition of soil organic matter that was previously "protected" from decomposition by anaerobic conditions, leading to loss of soil carbon (Fisher and Binkley 2000; Baker et al. 2007).

Water is a critical resource for most ecosystem processes, and its availability depends critically on soil structure. In soils, water is held in the pore spaces as films of water adsorbed to soil particles. The soil is **water saturated** when all pore spaces are filled with water. When the larger pores fill with water, water begins to drain under

the influence of gravity (**saturated flow**), even when some of the smaller pores within aggregates have not yet filled. Water drainage continues until, often after several days, the adhesive forces that hold water in films on soil particles equals the gravitational pressure. At this point, called **field capacity**, water no longer freely drains.

At water contents below field capacity, water moves through the soil by **unsaturated flow** in response to gradients of **water potential**, i.e., the potential energy of water relative to pure water (see Chap. 4). When plant roots absorb water from the soil to replace water that is lost in transpiration, this reduces the thickness of water films adjacent to roots, causing the remaining water to adhere more tightly to soil particles. The net effect is to reduce the soil water potential at the root surface. Water moves along water films through the soil pores toward the root in response to this gradient in water potential. As plants continue to transpire, water continues moving toward the root until some minimal water potential is reached, when roots can no longer extract water from the particle surfaces or the continuity of the water film is broken (see Chap. 4). This point is called the **permanent wilting point**. **Water-holding capacity** is the difference in water content between field capacity and permanent wilting point (see Fig. 4.8). Water-holding capacity is substantially enhanced by presence of clay and soil organic matter because of their large surface area. The water-holding capacity of an organic soil might, for example, be 300% (3 kg H_2O per kg dry soil), while that of a clay soil may be 30% and that of a sandy soil could be less than 20%. On a volumetric basis, water-holding capacity is normally highest in loam soils. One consequence of this difference is that, for a given amount of rainfall, coarse-textured soils will be wetted more deeply than soils without large pores (e.g., many clay soils) but will retain less water in surface soil horizons that are accessible to most plants. The water-holding characteristics of soils help determine the amount of water available for plant absorption and growth and for microbial processes, including decomposition and nutrient cycling and loss.

Soil Chemical Properties

In addition to its effects as a resource that supports plant growth, water strongly influences chemical and biological processes in soils through its effects on oxygen availability. **Oxidation–reduction** reactions involve the transfer of electrons from one reactant to another, yielding chemical energy that can be used by organisms (Schlesinger 1997). In these reactions, the energy source (often organic matter) gives up one or more electrons (**oxidation**). These electrons are transferred to electron acceptors (**reduction**). A handy mnemonic is: "LEO (loss of electrons = oxidation) the lion says GER (gain of electrons = reduction)." **Redox potential** is the tendency of an environment to receive or supply electrons (Schlesinger 1997; Fisher and Binkley 2000). More precisely, it is the net oxidation state of a pair of chemicals such as sulfate and sulfide or water and oxygen. Any soil or sediment has a mixture of chemicals that support complex patterns of electron transfer among chemicals. In addition, soils and sediments differ widely in redox potential (net oxidation state) due to their chemical composition and oxygen availability. Under the most aerobic conditions, which occur inside the mitochondria of live, eukaryotic cells, redox reactions transfer electrons from carbohydrates through a series of reactions to oxygen. This series of reactions releases the energy that supports cellular growth and maintenance. Other redox reactions occur in the cells of soil or benthic organisms, when electrons are transferred from electron donors to acceptors other than oxygen (Table 3.4). Organisms harvest the most energy by transferring electrons to oxygen, so this reaction predominates when oxygen is present. However, under anaerobic conditions, which commonly occur in flooded soils with high organic matter contents, in the interior of soil aggregates, or in lake or coastal ocean sediments, electrons must be transferred to other electron acceptors, with progressively less energy being released with transfer to each of the following electron acceptors (Table 3.4):
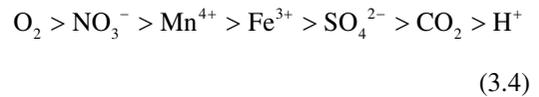
$$O_2 > NO_3^- > Mn^{4+} > Fe^{3+} > SO_4^{2-} > CO_2 > H^+$$

$$(3.4)$$

Table 3.4 Sequence of H^+-consuming redox reactions that occur with progressive declines in redox potential

Reaction[a]	Redox potential[b] (mV)	Energy release[b] (Kcal mol^{-1} per e$^-$)
Reduction of O_2 $O_2 + 4H^+ + 4e^- \rightarrow 2H_2O$	812	29.9
Reduction of NO_3^- $NO_3^- + 2H^+ + 2e^- \rightarrow NO_2^- + H_2O$	747	28.4
Reduction of Mn^{4+} to Mn^{2+} $MnO_2 + 4H^+ + 2e^- \rightarrow Mn^{2+} + 2H_2O$	526	23.3
Reduction of Fe^{3+} to Fe^{2+} $Fe(OH)_3 + 3H^+ + e^- \rightarrow Fe^{2+} + 3H_2O$	−47	10.1
Reduction of SO_4^{2-} to H_2S $SO_4^{2-} + 10H^+ + 8e^- \rightarrow H_2S + 4H_2O$	−221	5.9
Reduction of CO_2 to CH_4 $CO_2 + 8H^+ + 8e^- \rightarrow CH_4 + 2H_2O$	−244	5.6

Data from Schlesinger (1997)

[a]The reactions at the top of the table occur in soils with high redox potential and release more energy (and are therefore favored) when the electron acceptors are available. The reactions at the bottom of the table release less energy and therefore occur only if other electron acceptors are absent or have already been consumed by redox reactions. Abbreviations include electrons (e$^-$), nitrite ion (NO_2^-), manganese dioxide (MnO_2), ferric hydroxide ($Fe(OH)_3$), organic matter (CH_2O), universal gas constant (R), temperature (T), and equilibrium constant (K)

[b]Assumes that all reactants and products are at molar concentrations, which is seldom true, and complete coupling to the oxidation reaction:

$CH_2O + H_2O \rightarrow CO_2 + 4H^+ + 4e^-$ and that the energy released $= RT \ln(K)$
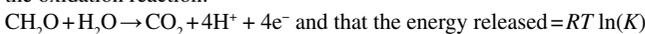
Organic matter is abundant enough in most soils and sediments to serve as the major electron donor, although reduced iron, sulfide, etc. are also important electron donors. In the absence of oxygen, reduced compounds such as Fe^{2+} become increasingly important electron donors. As soil oxygen becomes depleted and redox potential declines, the preferred electron acceptors are gradually consumed (Table 3.4). A similar gradient in redox potential and preferred electron acceptors occurs with depth in flooded soils or in lake sediments. As oxygen becomes depleted with depth or time, for example, the redox reaction that generates the most energy is initially denitrification (transfer of electrons to nitrate) followed by reduction of Mn^{4+} to Mn^{2+}. These reactions are typically carried out by facultatively anaerobic bacteria, i.e., bacteria that can metabolize and grow under either aerobic or anaerobic conditions. Facultative anaerobes use oxygen when present because of the greater energy return (Table 3.4), but switch to nitrate or Mn^{4+} as electron acceptors when oxygen is depleted by decomposer respiration. Nitrate is produced in soils or water through nitrification by obligate aerobic nitrifying bacteria (see Chap. 9). Denitrification is therefore most important in redox reactions in situations with substantial temporal or spatial variation in oxygen availability or external nitrate inputs. Denitrification is particularly important, for example, in the interior of soil aggregates, in seasonally flooded soils, and in the hypolimnion (bottom water) and sediments of seasonally stratified lakes.

Below the zone of Mn^{4+} reduction (or after Mn^{4+} has been depleted), most redox reactions are performed by obligately anaerobic bacteria or occur abiotically (Howarth 1984; Schlesinger 1997). Under these conditions, Fe^{3+} is reduced to Fe^{2+}, with organic matter as the most common electron donor in bacterial reduction. Alternatively, in salt marshes sulfate is reduced to sulfide by sulfate-reducing bacteria, and sulfide serves as the electron donor in abiotic reduction of Fe^{3+} (Howarth 1984). In either case, there is a visible transition from red (Fe^{3+}) to black or gray (Fe^{2+}) color of the soil or sediment. These reactions provide less than half as much energy per unit of organic matter decomposed as does denitrification or manganese reduction (Table 3.4), so iron- and sulfate-reducing bacteria can compete effectively only when nitrate and Mn^{4+} have been depleted from soil. Finally, as other electron acceptors are depleted, methanogenic bacteria reduce CO_2 to methane (CH_4), often in combination with continued sulfate reduction. In general, **hypoxic** (weakly oxygenated) environments often support high rates of denitrification because of the juxtaposition in space or time of aerobic and anaerobic microenvironments, whereas environments that are permanently anaerobic are more important in sulfate reduction (marine environments or salt marshes) or methanogenesis (low-sulfur environments). In coastal marine sediments, methanogenesis and sulfate reduction are linked, with methanogenesis producing methane that is consumed by sulfate-reducing bacteria; this results in net reduction of sulfate and very little net emission of methane (Howarth 1984). Many soils, deep lake waters, and some sediments experience substantial seasonal fluctuations in oxygen availability and therefore in the relative importance of each redox reaction. The biological bases of redox reactions are described in Chap. 7, and their role in element cycles is described in Chap. 9.

Soil organic matter content is a critical component of soils and sediments. It provides the energy and carbon base for heterotrophic soil organisms (see Chap. 7) and is an important reservoir of essential nutrients required for plant growth (see Chap. 8). In addition, it strongly affects rates of weathering and soil development, soil water-holding capacity, soil structure, and nutrient retention. Soil organic matter originates from dead plant, animal, and microbial tissues, but includes materials ranging from new, undecomposed plant tissues to charcoal to resynthesized humic substances that are thousands of years old, whose origins are chemically and physically unrecognizable (see Chap. 7). Because soil organic matter is critical to so many soil properties, loss of soil organic matter through inappropriate land management is a major cause of land degradation and loss of biological productivity.

The capacity of ecosystems to provide cations to support biological activity depends, over the

long term, on parent material and rates of weathering and loss, as discussed earlier. Over days to decades, however, the cations that are loosely bound to the soil exchange complex (primarily clay particles and soil organic matter) are the primary source of supply. **Cation exchange capacity (CEC)** reflects the capacity of a soil to form loose electrostatic bonds between positively charged cations and the negatively charged sites on the surfaces of soil minerals and organic matter. Cation exchange occurs when a cation in solution displaces a cation on the exchange complex. Values for CEC vary more than 100-fold among clay minerals and tend to decline with weathering (Fig. 3.10). The negative charge on clay minerals originates from an excess of negative charges on their surfaces and exposed edges. Soil organic matter also has a very high CEC due to the presence of –OH and –COOH groups and contributes substantially to the total CEC of some soils. Organic matter, for example, accounts for most CEC in those tropical soils that consist primarily of iron and aluminum oxides and 1:1 silicate clay minerals, which have a relatively low CEC. High-latitude soils also derive a large proportion of their CEC from organic matter due to their high organic content and low clay content. The pool of exchangeable cations in the soil is many times larger than the pool of soluble cations and represents the major short-term reservoir of cations for plant and microbial absorption.

Base saturation is the percentage of the total exchangeable cation pool that is accounted for by **base cations** (the non-hydrogen, non-aluminum cations). The identity of the cations on the exchange sites depends on the concentrations of cations in the soil solution and on the strength with which different cations are held to the exchange complex. In general, cations occupy exchange sites and displace other ions in the sequence

$$H(Al^{3+}) > H^+ > Ca^{2+} > Mg^{2+} > K^+ \approx NH_4^+ > Na^+$$

$$(3.5)$$
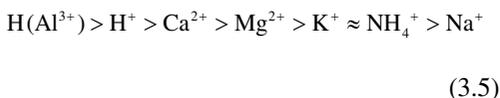
so leached soils tend to lose Na^+ and NH_4^+ but retain Al^{3+} and H^+. This displacement series is a consequence of differences among ions in charge and hydrated radius. Ions with more positive charges bind more tightly to the exchange complex than do ions with a single charge. Ions with a smaller hydrated radius have their charge concentrated in a smaller volume and also tend to bind tightly to the exchange complex.

Minerals and organic matter have both positively and negatively charged groups and therefore electrostatically bind both anions and cations, although CEC is generally much greater than **anion exchange capacity** (Fig. 3.10). In some soils, especially those in the tropics, iron and aluminum oxide minerals have a positive surface charge at their typical pHs. In these soils, there is enough anion exchange capacity to attract anions more strongly than cations (Uehara and Gillman 1981). As with cations, anion adsorption depends on the concentration of anions and their relative capacities to be held or to displace other anions. Anions generally occupy exchange sites and displace other ions in the sequence

$$PO_4^{3-} > SO_4^{3-} > Cl^- > NO_3^- \qquad (3.6)$$
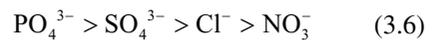
so leached soils tend to lose NO_3^- and Cl^- but retain PO_4^{3-} and SO_4^{3-}.

In addition to weak electrostatic bonds associated with cation and anion exchange, minerals can strongly bind both cations (e.g., K^+) and anions (e.g., PO_4^{3-}). The ecologically most important of these strong chemical bonds causes **phosphorus fixation**, which is particularly pronounced in highly weathered tropical soils and in some volcanic soils, explaining why ecosystems with these soils often show strong phosphorus limitation of plant growth and decomposition (Uehara and Gillman 1981). Phosphorus fixation is also sensitive to pH, causing phosphorus availability in soils to decline substantially at both high pH (e.g., limestone soils) and low pH (e.g., highly weathered soils).

Ecosystems often maintain a relatively stable pH despite continuous inputs of H^+ from precipitation, decomposition, and more recently from anthropogenic acid rain. This **buffering capacity** results from a multitude of soil chemical reactions that produce or consume H^+. These include reactions of H^+ with aluminum compounds like gibbsite at low pH and with carbonates at high pH. Many of these reactions are a normal component

of chemical weathering. Reactions of H^+ with organic matter occur over a wide pH range. Exchange of H^+ with cations on organic and mineral exchange complexes also contributes to buffering, particularly at intermediate pH. Soils differ in the relative importance and the capacity of these reactions to buffer pH, but soils with a high CEC and base saturation often have the greatest buffering capacity. Buffering capacity is important because it maintains soil pH within a relatively narrow range for long periods even with chronic exposure to acid rain. When the buffering capacity is exceeded, the soil pH begins to drop, which can solubilize $Al(OH)_x$, Al^{3+}, and other cations, with potentially toxic effects in both terrestrial and downstream aquatic ecosystems (Schulze 1989; Driscoll et al. 2001). Acidic temperate and tropical soils, for example, have a relatively low CEC and buffering capacity, and some of the reactions that consume H^+ release aluminum in solution, making these soils toxic to those plants and microbes that are not adapted to acidic conditions.

Summary

Five state factors control the formation and characteristics of soils. (1) Parent material is generated by the rock cycle, in which rocks are formed, uplifted, and weathered to produce the materials from which soil is derived. (2) Climate is the factor that most strongly determines the rates of soil-forming processes and therefore rates of soil development. (3) Topography modifies these rates at a local scale through its effects on microclimate and the balance between soil development and erosion. (4) Organisms also strongly influence soil development through their effects on the physical and chemical environment. (5) Time integrates the impact of all state factors in determining the long-term trajectory of soil development. In recent decades, human activities have modified the relative importance of these state factors and substantially altered Earth's soils.

The development of soil profiles represents the balance between profile development, soil mixing, erosion, and deposition. Profile development occurs through the input, transformation, vertical transfer, and loss of materials from soils.

Inputs to soils come from both outside the ecosystem (e.g., dust or precipitation inputs) and inside the ecosystem (e.g., litter inputs). The organic matter inputs are decomposed to produce CO_2 and nutrients or are transformed into recalcitrant organic compounds. The carbonic acid derived from CO_2 and the organic acids produced during decomposition convert primary minerals into secondary clay minerals with greater surface area and CEC. Water moves these secondary minerals and the soluble weathering products down through the soil profile until new chemical conditions cause them to become reactants or precipitate out of solution. Leaching of materials into groundwater or erosion and gaseous losses to the atmosphere are the major avenues of loss of materials from soils. The net effect of these processes is to form soil horizons that vary with climate, parent material, biota, and soil age and have distinctive physical, chemical, and biological properties.

Review Questions

1. What processes are responsible for the cycling of rock material in Earth's crust?
2. At large geographic scales, which state factors control soil formation? How might interactive controls modify the effects of these state factors?
3. What processes determine erosion rate? Which of these processes are most strongly influenced by human activities?
4. What processes cause soil profiles to develop? Explain how differences in climate, drainage, and biota might affect profile development.
5. What are the processes involved in physical and chemical weathering? Give examples of each. How do plants and plant products contribute to each?
6. How does soil texture affect other soil properties? Why does it influence ecosystem processes so strongly?
7. What is cation exchange capacity (CEC), and what determines its magnitude in temperate soils? How would you expect the determinants of CEC to differ between histosols, alfisols, and oxisols?

8. In a warm climate, how do soil processes and properties differ between sites with extremely high and extremely low precipitation? In a moist climate, how do soil processes and properties differ between sites with extremely high and extremely low soil temperature?

9. If global warming caused only an increase in temperature, how would you expect this to affect soil properties after 100 years? After a million years?

Additional Reading

Amundson, R., and H. Jenny. 1997. On a state factor model of ecosystems. *BioScience* 47:536–543.

Birkeland, P.W. 1999. *Soils and Geomorphology*. 3rd Edition. Oxford University Press, New York.

Brady, N.C. and R.R. Weil. 2008. *The Nature and Properties of Soils*. 14th edition. Pearson Education, Inc., Upper Saddle River, NJ.

Chadwick, O.A., L.A. Derry, P.M. Vitousek, B.J. Huebert, and L.O. Hedin. 1999. Changing sources of nutrients during 4 million years of soil and ecosystem development. *Nature* 397:491–497.

Jenny, H. 1941. *Factors of Soil Formation*. McGraw-Hill, New York.

Selby, M.J. 1993. *Hillslope Materials and Processes*. 2nd edition. Oxford University Press, Oxford.

Ugolini, F.C. and H. Spaltenstein. 1992. Pedosphere. Pages 123–153 *in* S.S. Butcher, R.J. Charlson, G.H. Orians, and G.V. Wolfe, editors. *Global Biogeochemical Cycles*. Academic Press, London.

Vitousek, P.M. 2004. *Nutrient Cycling and Limitation: Hawai'i as a Model System*. Princeton University Press, Princeton.