

Fifth Edition

PRINCIPLES OF SEDIMENTOLOGY AND STRATIGRAPHY

Sam Boggs, Jr.
University of Oregon

Prentice Hall

Boston Columbus Indianapolis New York San Francisco Upper Saddle River
Amsterdam Cape Town Dubai London Madrid Milan Munich Paris Montréal Toronto
Delhi Mexico City São Paulo Sydney Hong Kong Seoul Singapore Taipei Tokyo

Library of Congress Cataloging-in-Publication Data

Boggs, Sam.

Principles of sedimentology and stratigraphy / Sam Boggs. — 5th ed.

p. cm.

Includes bibliographical references and index.

ISBN-13: 978-0-321-64318-6

ISBN-10: 0-321-64318-6

1. Sedimentation and deposition. 2. Geology, Stratigraphic. I. Title.

QE571.B66 2012

552'.5—dc22

2010041477

Acquisitions Editor: Andrew Dunaway

Marketing Manager: Maureen McLaughlin

Project Editor: Crissy Dudonis

Editorial Assistant: Michelle White

Marketing Assistant: Nicola Houston

Managing Editor, Geosciences and Chemistry: Gina M. Cheselka

Project Manager, Production: Shari Toron

Senior Manufacturing and Operations Manager: Nick Sklitsis

Operations Specialist: Maura Zaldivar

Art Director: Jayne Conte

Cover Designer: Suzanne Behnke

Senior Technical Art Specialist: Connie Long

Art Studio: Precision

Full-Service Project Management and Composition: Chitra Ganesan/PreMediaGlobal

Cover Photo Credit: Grand Teton National Park. Mt. Meek, Battleship Mountain (Butte). View to southwest from near summit of Grand Teton. Teton County, Wyoming. August 1966. U.S. Geological Survey Photographic Library/D.L. Gaskill

© 2012, 2006, 1995 Pearson Education, Inc.

Pearson Prentice Hall

Pearson Education, Inc.

Upper Saddle River, New Jersey 07458

All rights reserved. No part of this book may be reproduced, in any form or by any means, without permission in writing from the publisher.

Printed in the United States of America

18 17

Pearson Prentice Hall™ is a trademark of Pearson Education, Inc.

ISBN-10: 0-321-64318-6

ISBN-13: 978-0-321-64318-6

Prentice Hall
is an imprint of

PEARSON

www.pearsonhighered.com

Dedication

*Dedicated to my mother, Lucinda Caudill Boggs,
to whom I owe everything*

Sedimentary rocks form through a complex set of processes that begins with **weathering**, the physical disintegration and chemical decomposition of older rock to produce solid particulate residues (resistant minerals and rock fragments) and dissolved chemical substances. Some solid products of weathering may accumulate *in situ* to form soils that can be preserved in the geologic record (paleosols). Ultimately, most weathering residues are removed from weathering sites by erosion and subsequently transported, possibly along with fragmental products of explosive volcanism, to more distant depositional sites.

Transport of siliciclastic detritus to depositional basins can involve a variety of processes. Mass-transport processes such as slumps, debris flows, and mud flows are important agents in the initial stages of sediment transport from weathering sites to valley floors. Fluid-flow processes, which include moving water, glacial ice, and wind, move sediment from valley floors to depositional basins at lower elevations. When transport processes are no longer capable of moving sediment, **deposition** of sand, gravel, and mud takes place, either subaerially (e.g., in desert dune fields) or subaqueously in river systems, lakes, or the marginal ocean. Sediment deposited at the ocean margin may be reentrained and retransported tens to hundreds of kilometers into deeper water by turbidity currents or other transport processes. Sediments deposited in basins are eventually buried and undergo physical and chemical changes (diagenesis) resulting from increased temperature, pressure, and the presence of chemically active fluids. Burial diagenetic processes convert siliciclastic sediments to lithified sedimentary rock: conglomerate, sandstone, shale.

Weathering processes also release from source rocks soluble constituents such as calcium, magnesium, and silica that make their way in surface water and groundwater to lakes or the ocean. When concentrations of these chemical elements become sufficiently high, they are removed from water by chemical and biochemical processes to form "chemical" sediments. Subsequent burial and diagenetic alteration of these sediments generates lithified sedimentary rock: limestone, chert, evaporites, and other chemical/biochemical sedimentary rocks.

In summary, the origin of sedimentary rocks involves weathering of older rock to generate the materials that make up sedimentary rock, erosion and transport of weathered debris and soluble constituents to depositional basins, deposition of this material in continental (terrigenous) or marine environments, and diagenetic alteration during burial to ultimately produce lithified sedimentary rock. Because weathering plays such a critical role in generating the solid particles and chemical constituents that make up sedimentary rocks, Chapter 1 focuses on the physical and chemical processes of weathering, the nature of the resulting weathering products, and a brief discussion of soils. Chapter 2 continues with a detailed discussion of the various processes by which sediment grains are transported from weathering sites to depositional basins. Other aspects of the origin of sedimentary rocks are introduced and discussed in succeeding chapters, as appropriate.

Weathering and Soils

1.1 INTRODUCTION

Weathering involves chemical, physical, and biological processes, although chemical processes are by far the most important. A brief summary of weathering processes is presented here to illustrate how weathering acts to decompose and disintegrate exposed rocks, producing particulate residues and dissolved constituents. These weathering products are the source materials of soils and sedimentary rocks; thus, weathering constitutes the first step in the chain of processes that produce sedimentary rocks.

It is important to understand how weathering attacks exposed source rocks and what remains after weathering to form soils and be transported as sediment and dissolved constituents to depositional basins. The ultimate composition of soil and terrigenous sedimentary rock bears a relationship to the composition of their source rock; however, study of residual soil profiles shows that both the mineral composition and the bulk chemical composition of soils may differ greatly from those of the bedrock on which they form. Some minerals in the source rock are destroyed completely during weathering, whereas more chemically resistant or stable minerals are loosened from the fabric of the decomposing and disintegrating rock and accumulate as residues. During this process, new minerals such as iron oxides and clay minerals may form *in situ* in the soils from chemical elements released during breakdown of the source rocks. Thus, soils are composed of survival assemblages of minerals and rock fragments derived from the parent rocks plus any new minerals formed at the weathering site. Soil composition is governed not only by the parent-rock composition but also by the nature, intensity, and duration of weathering and soil-forming processes. It follows from this premise that the composition of terrigenous sedimentary rocks such as sandstones, which are derived from soils and other weathered materials, is also controlled by parent-rock composition and weathering processes.

Most ancient soils were probably eroded and their constituents transported to furnish the materials of sedimentary rocks; however, some survived to become part of the geologic record. We call these ancient soils **paleosols**. Weathering and soil-forming processes are significantly influenced by climatic conditions. Geologists are greatly interested in the study of past climates, called paleoclimatology, because of this relationship and because paleoclimates also influenced past sea levels and sedimentation processes as well as the life forms on Earth at various times.

In this chapter, we examine the principal processes of subaerial weathering and discuss the nature of the particulate residues and dissolved constituents that result from weathering. We also consider the less important but highly interesting processes of submarine weathering. Submarine weathering includes both the interaction of seawater with hot oceanic rocks along midocean ridges—a process that leaches important amounts of chemical constituents from hot crustal rocks—and low-temperature alteration of volcanic rocks and sediments on the ocean floor. Finally, we take a brief look at soils and paleosols and discuss important soil-forming processes and the factors, such as climate, that influence soil development.

1.2 SUBAERIAL WEATHERING PROCESSES

Physical Weathering

Physical (mechanical) weathering is the process by which rocks are broken into smaller fragments through a variety of causes, but without significant change in chemical or mineralogical composition. Except in extremely cold or very dry climates, physical and chemical weathering act together, and it is difficult to separate their effects.

FREEZE-THAW (FROST) WEATHERING Disruption of rock fabrics owing to stresses generated by freezing and thawing of water in rock fractures is an important physical weathering process in climates where recurring, short-term changes from freezing to thawing temperatures take place. Water increases in volume by about 9 percent when it changes to ice, creating enough pressure in tortuous rock fractures to crack most types of rock. To be effective, water must be trapped (sealed by freezing) within the rock body, and repeated freezing and thawing are necessary to allow progressive disintegration of the rock, which occurs very slowly. Other processes, such as the movement of water into a freezing zone rather than conversion of water in place to ice, may also, or alternatively, cause freeze-thaw expansion of cracks (Bland and Rolls, 1998, 89).

Freeze-thaw weathering commonly produces large, angular blocks of rock (Fig. 1.1) but may also cause granular disintegration of coarse-grained rocks such as granites. The presence of microfractures and other microstructures exerts an important control on the sizes and shapes of shattered blocks. Mechanically weak rocks such as shales and sandstones tend to disintegrate more readily than do hard, more strongly cemented rocks such as quartzites and igneous rocks (e.g., Nicholson and Nicholson, 2000).



FIGURE 1.1 Large, angular blocks of rock generated by freeze-thaw weathering of thin-bedded sandstones and mudstones of the Canning Formation (Paleocene) exposed along the Canning River, Arctic National Wildlife Refuge, Alaska. [Photograph by C. J. Schenk, U.S. Geological Survey *Open File Report 98-34*, The oil and gas resource potential of the Arctic National Wildlife Refuge 1002 Area, Alaska, 1999.]

INSOLATION WEATHERING Expansion of rock surfaces heated by the Sun (insolation) followed by contraction as temperature falls can allegedly weaken bonds along grain boundaries and cause subsequent flaking off of rock fragments or dislodging of mineral grains. A thermal gradient is set up between the surface and interior of a rock that has been heated; the rock surface expands more than the interior, creating stresses. These stresses presumably lead to formation of small cracks and possibly granular disintegration (Ollier and Pain, 1996, 26). Once a small crack in a rock's surface expands with heating, silt or sand particles may sift into the crack and prevent it from closing when the rock cools. Repeated heating and cooling causes the crack to grow wider and wider, resulting in small-scale disruption of the rock surface. These kinds of physical changes are caused mainly by heating from sunshine but may also result from fires (e.g., Allison and Goudie, 1994). Halsey et al., 1998, suggest that short-term, climatically induced heating and cooling cycles may be an important aspect of insolation weathering.

SALT WEATHERING High temperatures in desert environments also tend to promote weathering caused by the crystallization of salts in pore spaces and fractures (Sperling and Cooke, 1980; Watson, 1992; Bland and Rolls, 1998; Wright, 2007). Evaporation of water concentrates dissolved salts in saline solutions that have access to rock fractures and pores. Growth of salt crystals generates internal pressures (crystallization pressures) that can force cracks apart or cause

granular disintegration of weakly cemented rocks. Expansion pressures may also be generated when salts in fractures become hydrated (absorb water) and expand. Salt weathering is most common in semiarid regions but can occur also along seacoasts where salt spray is blown onto sea cliffs.

WETTING AND DRYING Alternate wetting and drying of soft or poorly cemented rocks such as shales causes fairly rapid breakdown of the rocks, and most disintegration may occur during the drying cycle. The exact causes of disintegration are not well understood, but drying may lead to negative pore pressures and consequent tensile stresses (contraction) that tend to pull the rock apart. On the other hand, absorption of water during wetting phases creates "swelling" pressures that push cracks apart. Disintegration by wetting and drying appears to be particularly effective on well-exposed, steep cliff faces where loosened fragments fall off and expose fresh surfaces.

STRESS-RELEASE WEATHERING A rock unit buried below a land surface experiences high compressional stresses because of the weight of the overlying rock. If some of the overlying rock is removed by erosion, compressional stresses on the rock unit are reduced and the rock unit "rebounds" upward. Expansion of the rock upward creates tensile stresses (pulls the rock apart), causing fractures to develop that are oriented nearly parallel to the topographic surface. These fractures divide the rock into a series of layers or sheets; hence, this process of crack formation is often called sheeting. These layers increase in thickness with depth and may exist for several tens of meters below Earth's surface. Sheeting is most conspicuous in homogeneous rocks such as granite (e.g., Fig. 1.2) but may occur also in layered rock, such as massive sandstone.

OTHER PHYSICAL PROCESSES Other factors that may contribute to mechanical weathering under certain conditions include volume increases caused by absorption of water (hydration) by clay minerals or other minerals; volume changes caused by alteration of minerals such as biotite and plagioclase to clay minerals; growth of plant roots in the cracks of rocks; plucking of mineral grains and rock fragments from rock surfaces by lichens as they expand and contract in response to wetting and drying; and burrowing and ingestion of soils and loosened rock materials by worms or other organisms.

Some physical weathering effects may be the result of two or more processes operating together. **Exfoliation**, the peeling off of large, curved sheets or slabs of rock from the weathered surfaces of an outcrop, is an apposite example. Stress release may create initial fractures, which then allow the entry of water that further widens fractures by freeze-thaw or other processes.

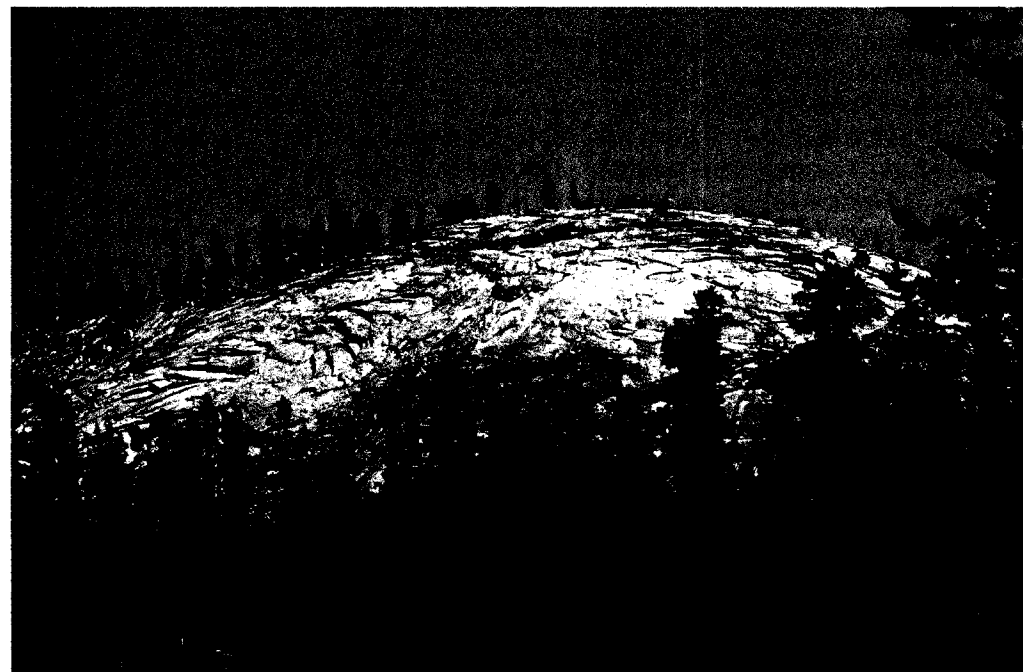


FIGURE 1.2 Conspicuous convex sheeting in granite, Yosemite National Park, California. Note that sheeting follows topographic surfaces. (Huber, N. K., 1987, *The Geologic Story of Yosemite National Park*, U. S. Geological Survey Bull. 1595, Figure 40-B.)

Spheroidal weathering is smaller-scale weathering of roughly cubic rock masses, cut by intersecting joints, causing layers or "skins" to spall off to produce spheroidal cores (Fig. 1.3). The fractures that separate the weathering rinds may form in response to stress release or possibly thermal changes (Taylor and Eggleton, 2001, 166); entry of water into fractures promotes additional physical stresses arising from freeze-thaw or chemical processes such as those mentioned in the preceding paragraph.

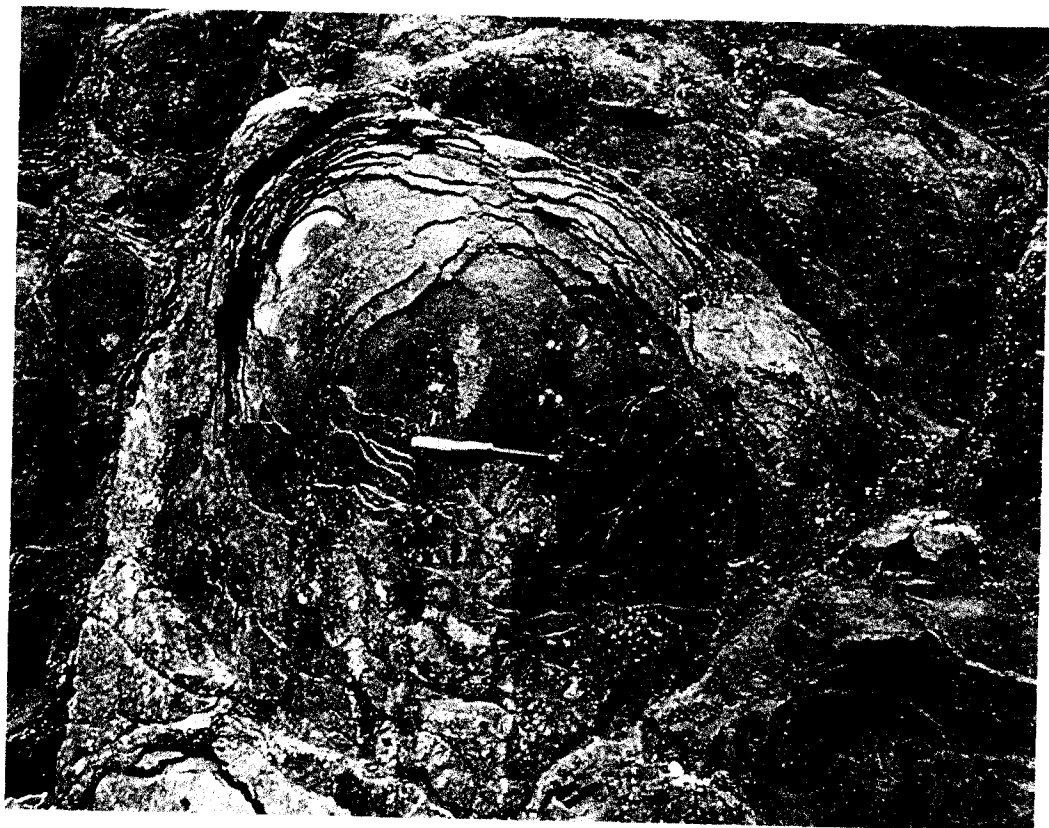


FIGURE 1.3 Spheroidal weathering in granite. Note how successive, thin layers of weathered rock are spalled off to produce a spheroidal core.

Chemical Weathering

Chemical weathering involves changes that can alter both the chemical and the mineralogical composition of rocks. Minerals in the rocks are attacked by water and dissolved atmospheric gases (oxygen, carbon dioxide), causing some components of the minerals to dissolve and be removed in solution. Other mineral constituents recombine *in situ* and crystallize to form new mineral phases. These chemical changes, along with changes caused by physical weathering, as discussed, disrupt the fabric of the weathered rock, eventually producing residual blocks and a loose residue of resistant grains and secondary minerals (e.g., Fig. 1.4). Water and dissolved gases play a dominant role in every aspect of chemical weathering. Because some water is present in almost every environment, chemical weathering processes are commonly far more important than physical weathering processes, even in arid climates. Nevertheless, owing to the low temperatures of the weathering environment (<30°C), chemical weathering occurs very slowly. The processes of chemical weathering are listed and briefly described in Table 1.1, along with selected examples of new minerals formed *in situ* during the weathering processes.

MAJOR CHEMICAL WEATHERING PROCESSES

Simple Solution Simple solution (*congruent dissolution*) occurs when a mineral goes into solution completely without precipitation of other substances (e.g., Birkland, 1999, 59). Simple solution of highly soluble minerals such as calcite, dolomite, gypsum, and halite, and even less soluble minerals such as quartz, occurs during exposure to meteoric water (rainwater). Chemical bonds between ions in the minerals are broken, destroying the minerals and releasing constituent ions into solution in surface and ground waters. If carbon dioxide is dissolved in the rainwater through interaction with atmospheric or soil CO₂, the usual case in the weathering environment,



FIGURE 1.4 Combined physical and chemical weathering of coarse-grained Pikes Peak Granite on the summit of Bison Mountain, Colorado. Note enhanced weathering of the granite along joints and fractures, producing residual blocks as well as fine-size weathered debris. (Lovering, T. S. and E. N. Goddard, 1950, *Geology and Ore Deposits of the Front Range, Colorado, U. S. Geological Survey Professional Paper 223, Fig. 9-D.*)

the solubilizing ability of water is enhanced. Dissolution of CO₂ in water forms carbonic acid (H₂CO₃—this is what makes soft drinks effervesce), which subsequently dissociates to produce hydrogen ions and carbonate ions (CO₂ + H₂O ↔ H₂CO₃ ↔ H⁺ + HCO₃⁻). Increase in H⁺ ions, relative to OH⁻ ions, makes meteoric waters more acidic and thus more aggressive dissolution agents, particularly for carbonate minerals. Simple solution of this type is an important weathering process, particularly in moderately wet climates where carbonate rocks or evaporites are present near the surface or at the water table.

Hydrolysis Hydrolysis is an extremely important chemical reaction between silicate minerals and acids that leads to breakdown of the silicate minerals and release of metal cations and silica, but the reaction does not lead to complete dissolution of the minerals. In other words, the amount of ions from the mineral that are taken into solution during weathering does not correspond to the formula of the weathering mineral. This kind of incomplete dissolution is called *incongruent dissolution*. If aluminum is present in the minerals undergoing incongruent dissolution during weathering, clay minerals such as kaolinite, illite, and smectite may form as a by-product of hydrolysis. For example, orthoclase feldspar can break down to yield kaolinite or illite, albite (plagioclase feldspar) can decompose to kaolinite or smectite, and so on, as illustrated by the reactions in Table 1.1. As mentioned, the H⁺ ions shown in Table 1.1 are commonly supplied by the dissociation of CO₂ in water. Thus, the more CO₂ that is dissolved in water, the more aggressive the hydrolysis reaction. Hydrolysis can also take place in water containing little or no dissolved CO₂, with H⁺ ions being supplied either by clay minerals that have a high proportion of H⁺ ions in cation exchange sites or by living plants, which create an acid environment. Most of the silica set free during hydrolysis goes into solution as silicic acid (H₄SiO₄); however, some of the silica may separate as colloidal or amorphous SiO₂ and be left behind during weathering to combine with aluminum to form clay minerals. Hydrolysis is the primary process by

BOX 1.1

pH

The acidity or alkalinity of a solution is expressed by its pH. The pH is defined as the negative logarithm to the base 10 of the approximate hydrogen-ion concentration in moles per liter. The pH scale extends from 0 to 14, corresponding to H⁺ concentrations ranging from 10⁰ to 10⁻¹⁴. For example, a solution containing a H⁺ concentration of 10⁻¹ moles per liter has a pH of 1, an H⁺ concentration of 10⁻⁷ yields a pH of 7, and so forth. Solutions with a pH of 7 are considered neutral. Acids have pH values lower than 7 and bases have values greater than 7.

TABLE 1.1 Principal processes of chemical weathering

Most important processes	Examples	Principal kinds of rock materials affected
Simple (congruent) Solution—Dissolution of soluble minerals in H ₂ O (direct solution) or in H ₂ O + CO ₂ (carbonation) to yield cations and anions in solution	SiO ₂ + 2H ₂ O → H ₄ SiO ₄ (direct solution) (quartz) (silicic acid) aq CaCO ₃ + H ₂ O + CO ₂ → Ca ²⁺ + 2HCO ₃ ⁻ (Carbonation) (calcite) aq	Highly soluble minerals (e.g., gypsum, halite), quartz Carbonate rocks
Hydrolysis (incongruent dissolution)—Reaction between H ⁺ and OH ⁻ ions of water and the ions of silicate minerals, yielding soluble cations, silicic acid, and clay minerals (if Al present)	2KAlSi ₃ O ₈ + 2H ⁺ + 9H ₂ O → H ₄ Al ₂ Si ₂ O ₉ + 4H ₄ SiO ₄ + 2K ⁺ (orthoclase) aq (kaolinite) (silicic acid) aq 2NaAlSi ₃ O ₈ + 2H ⁺ + 9H ₂ O → H ₄ Al ₂ Si ₂ O ₉ + 4H ₄ SiO ₄ + 2Na ⁺ (albite) aq (kaolinite) (silicic acid) aq	Silicate minerals
Oxidation—Loss of an electron from an element (commonly Fe or Mn) in a mineral, resulting in the formation of oxides or hydroxides (if water present)	2FeS ₂ + 15/2O ₂ + 4H ₂ O → Fe ₂ O ₃ + 4SO ₄ ²⁻ + 8H ⁺ (pyrite) (hematite) aq MnSiO ₃ + 1/2O ₂ + 2H ₂ O → MnO ₂ + H ₄ SiO ₄ (rhodonite) (pyrolusite) (silicic acid)	Iron- and manganese-bearing silicate minerals, iron sulfides
Other Processes		
Hydration and Dehydration—Gain (hydration) or loss (dehydration) of water molecules from a mineral, resulting in formation of a new mineral	Fe ₂ O ₃ + H ₂ O → 2FeOOH (hydration) (hematite) (goethite) CaSO ₄ · 2H ₂ O → CaSO ₄ + 2H ₂ O (dehydration) (gypsum) (anhydrite)	Ferric oxides Evaporites
Ion Exchange—Exchange of ions, principally cations, between solutions and minerals	K-clay + Mg ²⁺ + Na ⁺ → Mg-clay + K ⁺ Ca-zeolite + Na ⁺ → Na-zeolite + Ca ²⁺	Clay minerals and zeolites
Chelation—Bonding of metal ions to organic molecules having ring structures	Metal ions (cations) + chelating agent (e.g., secreted by lichens) → H ⁺ ions + chelate (metal ions/organic molecules in solution)	Silicate minerals

Note: aq = aqueous

which silicate minerals decompose during weathering. A more rigorous and detailed discussion of this process is given by Nahon (1991, 7).

Oxidation and Reduction Chemical alteration of iron and manganese in silicate minerals such as biotite and pyroxenes, caused by oxygen dissolved in water, is an important weathering process because of the abundance of iron in the common rock-forming silicate minerals. An electron is lost from iron during oxidation (Fe²⁺ to Fe³⁺ + e⁻, where e⁻ = electron transfer), which causes loss of other cations such as Si⁴⁺ from crystal lattices to maintain electrical neutrality. Cation loss leaves vacancies in the crystal lattice that either bring about the collapse of the lattice or make the mineral more susceptible to attack by other weathering processes. Oxidation of manganese minerals to form oxides and silicic acid or other soluble products is a less important but common weathering process. Another element that oxidizes during weathering is sulfur. For example, pyrite (FeS₂) is oxidized to form hematite (Fe₂O₃), with release of soluble sulfate ions. Under some conditions where material undergoing weathering is water saturated, oxygen supply may be low and oxygen demand by organisms high. These conditions can bring about **reduction** of iron (gain of an electron) from Fe³⁺ to Fe²⁺. Ferrous iron (Fe²⁺) is more soluble, and thus more mobile, than ferric iron (Fe³⁺) and may be lost from the weathering system in solution.

OTHER CHEMICAL WEATHERING PROCESSES Although simple solution, hydrolysis, and oxidation are the most important chemical weathering processes, under certain conditions several other processes can facilitate chemical weathering of minerals. **Hydration** is the process whereby water molecules are added to a mineral to form a new mineral. Common examples of hydration are the addition of water to hematite to form goethite, or to anhydrite to form gypsum. Hydration is accompanied by volume changes that may lead to physical disruption of rocks. Under some conditions, hydrated minerals may lose their water, a process called **dehydration**, and be converted to the anhydrous forms, with accompanying decrease in mineral volume. Dehydration is relatively uncommon in the weathering environment because some water is generally present.

Ion exchange is a process whereby ions in a mineral are exchanged with ions in solution, for example, the exchange of sodium for calcium. Most ion exchange takes place between cations (positively charged ions), but anion exchange also occurs. This reaction causes one mineral to be altered to another (new) mineral and, in the process, releases soluble ions into solution. Ion exchange is particularly important in alteration of one clay mineral to another (e.g., alteration of smectite to illite). Ion exchange also plays a role in alteration of one kind of zeolite to another (e.g., alteration of heulandite, a Ca-zeolite, to analcime, a Na-zeolite).

Chelation involves the bonding of metal ions to organic substances to form organic molecules having a ring structure (e.g., Boggs, Livermore, and Seitz, 1985). During weathering, chelation (i.e., organic complexing) performs the dual role of removing cations from mineral lattices and also keeping the cations in solution until they are removed from the weathering site. Chelated metal ions will remain in solution under pH conditions and at concentrations at which nonchelated ions would normally be precipitated. The bonding of aluminum or iron with a complexing agent and subsequent removal of these elements from a rock are of particular importance. A good example of natural chelation is provided by lichens that increase the rate of chemical weathering on rock surfaces on which they grow by secreting organic chelating agents. In addition to their role as chelating agents, plants also enhance chemical weathering processes by retaining soil moisture and by acidifying waters by release of CO₂ and various types of organic acids during decay.

WEATHERING RATES Determining the rate at which weathering takes place is a difficult and uncertain task. Various techniques are used to evaluate weathering rates: estimating the rate at which the landscape is lowered, estimating the rate at which bedrock is converted into soil, estimating the volume of solid detritus removed from weathering sites by streams, making chemical mass-balance calculations to evaluate the amount of soluble material removed in surface and ground water (e.g., Turkington and Paradise, 2005). Weathering processes proceed at different rates depending upon the climate and the mineral composition and grain size of the rocks undergoing weathering. Physical weathering processes may be quite effective in moderately cold climates (freeze-thaw) or arid climates (salt weathering), whereas chemical weathering processes are accelerated in humid, hot climates. Average rainfall is known to be a controlling factor in the

rate of chemical weathering (Nahon, 1991, 4); however, the influence of temperature on weathering rate is difficult to quantify, although we know that the rate of chemical reactions is accelerated by increasing temperature. Slope of the land surface is also important. Weathering tends to be more effective on low to moderate slopes as compared to steep slopes. Water is more likely to be retained on low slopes, and material undergoing weathering remains for a longer time before being removed by erosion.

The rate of weathering of silicate rocks, such as granite and gneiss, of a given grain size may be related to the relative chemical stabilities of the common rock-forming silicate minerals. Table 1.2 shows the order of relative stability to weathering of the most important mafic and felsic minerals, as determined by Goldich (1938) through empirical study of sand- and silt-sized particles in soil profiles. Readers will recognize this order as the same as that in which minerals crystallize in Bowen's reaction series. Minerals that crystallize at high temperatures (e.g., olivine) have the greatest degree of disequilibrium with surface weathering temperatures and thus tend to be less stable than minerals that crystallize at lower temperatures (e.g., quartz). Furthermore, the high-temperature minerals are bonded with weaker ionic or ionic-covalent bonds, whereas quartz is bonded with strong covalent bonds. Jackson (1968) suggests that the stability of very fine-size (clay-size) particles may differ somewhat from that of larger particles (Table 1.2). Rates of chemical weathering of silicate minerals are discussed in detail by White and Brantley, 1995.

TABLE 1.2 Relative stability of common sand-size minerals and various clay-size minerals under conditions of weathering

Sand- and silt-size minerals*		Clay-size minerals**
Mafic minerals	Felsic minerals	
Olivine		1. Gypsum, halite
	Ca plagioclase	2. Calcite, dolomite, apatite
Pyroxene		3. Olivine, amphiboles, pyroxenes
	Ca-Na plagioclase	4. Biotite
Amphibole		5. Na plagioclase, Ca plagioclase, K-feldspar, volcanic glass
	Na-Ca plagioclase	6. Quartz
	Na plagioclase	7. Muscovite
		8. Vermiculite (clay mineral)
Biotite		9. Smectite (clay mineral)
	K-feldspar, muscovite, quartz	10. Pedogenic (soil) chlorite
		11. Allophane (clay mineral)
		12. Kaolinite, halloysite (clay minerals)
		13. Gibbsite, boehmite (clay minerals)
		14. Hematite, goethite, magnetite
		15. Anatase, titanite, rutile, ilmenite (all, titanium-bearing minerals), zircon

(Increasing stability)

Source: *Goldich (1938); ** Jackson (1968).

Rates of weathering must take into account both physical and chemical processes, and they are very likely to be site-specific. Therefore, it is probably unwise to generalize too much about weathering rates. In particular, there is no rule of weathering susceptibility that can be applied generally to sedimentary rocks. Rates of weathering of these rocks are a function of the mineralogy, the amount and type of cement in the rocks, and the climate. For example, limestones weather rapidly by solution in wet climates and much more slowly in very arid or very cold climates. Quartz-rich sandstones cemented with silica cement weather very slowly under most climatic conditions. Finally, it is likely that rates of weathering have varied throughout geologic time depending upon climatic conditions and vegetative cover. Prior to the development of land plants in early Paleozoic time, absence of plant cover to hold soil moisture and contribute organic acids probably slowed rates of chemical weathering while contributing to increased rates of physical erosion.

PRODUCTS OF SUBAERIAL WEATHERING Subaerial weathering generates three types of weathering products that are important to the formation of sedimentary rocks (Table 1.3): (1) source-rock residues consisting of chemically resistant minerals and rock fragments derived

TABLE 1.3 Principal kinds of products formed by subaerial weathering processes and the types of sedimentary rocks ultimately formed from these products

Weathering process	Type of weathering product	Example	Ultimate depositional product
Physical weathering	Particulate residues	Silicate minerals such as quartz and feldspar; all types of rock fragments	Sandstones, conglomerates, mudrocks
Chemical weathering	Hydrolysis	Silicic acid (H_4SiO_4); K^+ , Na^+ , Mg^{2+} , Ca^{2+} , etc.	Cherts, limestones, etc.
	Secondary minerals	Clay minerals	Mudrocks (shales)
Simple solution	Soluble constituents	Silicic acid; K^+ , Na^+ , Mg^{2+} , Ca^{2+} , HCO_3^- , SO_4^{2-} , etc.	Limestones, evaporites, chert, etc.
	Oxidation	Secondary minerals	Ferric oxides (Fe_2O_3); manganese oxides (MnO_2)
	Soluble constituents	Silicic acid; SO_4^{2-}	Chert, evaporites, etc.

particularly from siliceous rocks such as granite, rhyolite, gneiss, and schist, (2) secondary minerals formed *in situ* by chemical recombination and crystallization, largely as a result of hydrolysis and oxidation, and (3) soluble constituents released from parent rocks mainly by hydrolysis and solution. Until they are removed by erosion, residues and secondary minerals accumulate at the weathering site to form a soil mantle composed of particles of various compositions and of grain sizes ranging from clay to gravel. Grain size and composition depend upon the grain size and composition of the parent rock and upon the nature and intensity of the weathering process. These characteristics of the weathering environment are in turn functions of climate, topography, and duration of the weathering process.

Source Rock Residues The residual particles in young or immature soils developed on igneous or metamorphic rocks may include, in addition to rock fragments, assemblages of minerals with low chemical stability: e.g., biotite, pyroxenes, hornblende, and calcic plagioclase. Mature soils, developed after more prolonged or intensive weathering of these rocks, commonly contain only the most stable minerals: quartz, muscovite, and perhaps potassium feldspars. Because the silicate minerals that make up siliciclastic sedimentary rocks such as sandstones have already passed through a weathering cycle before the siliciclastic rocks were formed, the weathering products of these rocks tend to be depleted in easily weathered minerals. Thus, even young soils developed on siliciclastic sedimentary rocks may have assemblages of mature minerals. Weathering of limestones by solution produces thin soils composed of the fine-size insoluble silicate and iron oxide residues of these rocks.

Secondary Minerals Secondary minerals developed at the weathering site are dominantly clay minerals, iron oxides or hydroxides, and aluminum hydroxides. The common secondary iron minerals include goethite, limonite, and hematite. The weathering products reflect both the nature and the intensity of the weathering process and the composition of the parent rock. Clay minerals formed in immature soils under only moderately intense chemical weathering conditions may be illites or smectites. More prolonged and intense leaching conditions lead to formation of kaolinite. Under extremely intense chemical weathering conditions, aluminum hydroxides such as gibbsite and diaspore are formed. These latter clay minerals are aluminum ores.

Comparing the chemical composition of unweathered silicate rocks with that of the weathering products of these rocks shows a net loss attributed to weathering of all major cations except aluminum and iron (e.g., Krauskopf, 1979). In the oxidized state, aluminum and ferric iron (Fe^{3+}) are both relatively insoluble. Although considerable silica is lost as soluble silicic acid during weathering, loss of Mg, Ca, Na, and K is comparatively much greater. Therefore, the relative abundance of silica, aluminum, and ferric iron in the particulate weathering residues of silicate rocks is greater than that in the parent source rocks.

Soluble Materials Soluble materials extracted from parent rocks by chemical weathering processes are removed from the weathering site in surface water or soil groundwater more or less continuously throughout the weathering process. Ultimately these soluble products make their way into rivers and are carried to the ocean. The most abundant inorganic constituents of rivers, representing the principal soluble products of weathering, are, in order of decreasing abundance, HCO_3^- (bicarbonate), Ca^{2+} , H_4SiO_4 (silicic acid), SO_4^{2-} (sulfate), Cl^- , Na^+ , Mg^{2+} , and K^+ (Garrels and McKenzie, 1971). These constituents are the raw materials from which chemically and biochemically deposited rocks such as limestones and cherts are formed in the oceans.

1.3 SUBMARINE WEATHERING PROCESSES AND PRODUCTS

Although we commonly think of weathering as being a subaerial process, an important kind of weathering also takes place on the ocean floor. Geologists have long recognized that sediments and rocks on the seafloor are altered by reaction with seawater, a process called **halmyrolysis** or submarine weathering. Halmyrolysis includes alteration of clay minerals of one type to another, formation of glauconite from feldspars and micas, and formation of phillipsite (a zeolite mineral) and palagonite (altered volcanic glass) from volcanic ash. Dissolution of the siliceous and calcareous tests of organisms may also be considered a type of submarine weathering. Prior to the

1970s, submarine weathering processes had not received a great deal of research, and it was not recognized that they might have a significant effect on the overall chemical composition of the oceans. Our concept of the importance of submarine weathering has changed dramatically since the mid-1970s because studies of volcanic rocks and weathering processes on the seafloor show that submarine weathering of basalts, particularly on midocean ridges, is an extremely important chemical phenomenon. This process results in both widespread hydration and leaching of basalts as well as changes in composition of seawater owing to ion exchange during the reaction of seawater with basalt.

Alteration of oceanic rocks occurs both at low temperatures (less than 20°C) and at higher temperatures ranging to $\sim 350^\circ\text{C}$. Low temperature alteration takes place as seawater percolates through fractures and voids in the upper part of the ocean crust, perhaps extending to depths of 2–5 km. Olivine and interstitial glass in the basalts are replaced by smectite clay minerals, and further alteration may lead to formation of zeolite minerals and chlorite. As a result of these changes, chemical elements are exchanged between rock and water, and large volumes of seawater become fixed in the oceanic crust in hydrous clay minerals and zeolites.

The discovery in 1977 of submarine thermal springs along the Galapagos Rift (Corliss et al., 1979) led to the awareness that large-scale hydrothermal activity takes place in the ocean. Since that initial discovery, scientists using submersible vehicles and water-sampling techniques have located many additional hot springs along midocean ridges in both the Pacific and Atlantic oceans, as well as along convergent plate margins, in backarc basins, and even on midplate volcanoes in the Hawaiian chain (e.g., Karl et al., 1988; Parson, Walker, and Dixon, 1995). These hot springs originate where seawater enters the ocean crust along fractures or other voids and comes in contact with hot volcanic rock. The heated water then flows out into the ocean through vents on the ocean floor and mixes with the overlying water. The heated water rises as hydrothermal plumes 100–300 m above the vent field. Exceptional plumes rising to heights of 1000 m have also been reported (e.g., Cann and Strens, 1989).

At the sites of many oceanic hot springs, investigators have found spectacular vents composed of sulfide, sulfate, and oxide deposits up to 10 m or more tall that discharge plumes of hot solutions (Fig. 1.5). These vents or chimneys are called **black smokers** if they discharge water

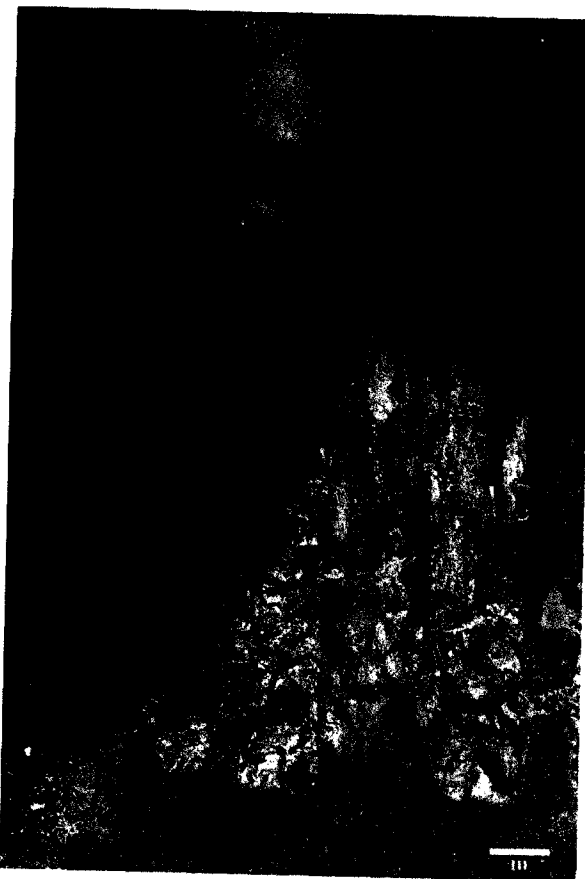


FIGURE 1.5 A multiple-orifice black smoker, Faulty Towers complex, Mothra hydrothermal vent field, Endeavour Segment, Juan de Fuca ridge. The constructional chimneys in the foreground were built by precipitation of sulfides and other minerals from heated water issuing from the vents at temperatures exceeding 250°C . [Photograph courtesy of John R. Delaney and Deborah S. Kelley, University of Washington School of Oceanography.]

containing suspended, fine-grained, dark-colored minerals or **white smokers** if the water contains no suspended dark minerals (McDonald, Spiess, and Ballard, 1980). The temperature of the water when it emerges from the vents may exceed 350°C . When these hot solutions mix with seawater of ambient temperature, they precipitate various minerals, particularly pyrite (FeS_2) and chalcopyrite (CuFeS_2), to build sulfide deposits around the vents. The deposits of fossil hydrothermal systems have now been observed in ancient oceanic ophiolite complexes exposed on land (e.g., Cann and Strens, 1989).

Reactions between hot basalt and seawater play a role in regulating the chemical composition of seawater. Magnesium, sulfate, and sodium ions are removed from seawater during this exchange, whereas many other elements such as calcium, iron, manganese, silicon, potassium, lithium, and strontium are enriched in the seawater (Edmond et al., 1982; Palmer and Edmond, 1989; Von Damm, 1990). The entire ocean apparently circulates through ocean-floor hydrothermal systems on a time scale of 10^6 – 10^7 years, which has a significant impact on the budget of several elements, including silica (Kadko et al., 1995).

The magnitude of hydrothermal alteration of basalts along midocean ridges and its effect on ocean chemistry is still being investigated and uncertainties remain; however, it now appears that circulation of ocean water through hydrothermal systems throughout geologic time has added significant quantities of certain ions to the ocean and removed others. Thus, both seafloor hydrothermal reactions and continental weathering processes supply ions to the ocean that may eventually be extracted to form chemically deposited rocks such as limestones, iron-rich sedimentary rocks, and cherts. Stanley and Hardie (1999) argue that changes in spreading rates along midocean ridges, where hydrothermal activity takes place, have exerted a major control on the calcium and magnesium content of seawater throughout geologic time. High spreading rates result in significant adsorption and loss of magnesium with concomitant increase in calcium, thus causing a decrease in the ratio of magnesium to calcium (Mg/Ca). Low spreading rates have the opposite effect of increasing the Mg/Ca ratio. As discussed in Chapters 6 and 11, these changes have important implications regarding the kinds of calcium-carbonate minerals deposited in the ocean.

1.4 SOILS

Considered from the standpoint of sedimentary-rock origin, we are perhaps more interested in the products of weathering than in the processes that bring about weathering, although it is useful for students to understand just how weathering processes operate to generate these products. The materials that make up sedimentary rocks are either siliciclastic grains derived from the land as a result of weathering (or explosive volcanism in some cases) or they are so-called "chemical" minerals that were precipitated from ocean or lake water. The elements that make up these chemical minerals were released from parent rocks by chemical weathering processes operating on land and in the ocean. Thus, it is quite reasonable to consider that the generation of both siliciclastic and chemical/biochemical sedimentary rocks begins with weathering.

Subaerial weathering products initially form soils of varied thickness over weathered bedrock. Throughout geologic time, most of these soils have ultimately been stripped away and transported as sediment to sedimentary basins; however, some soils are preserved to become part of the sedimentary record. Thus, because soils represent an incipient stage in the generation of siliciclastic sedimentary rocks and some are preserved in their own right, a discussion of soil-forming processes and the various kinds of soils that result from these processes is pertinent.

Soil-Forming Processes

Subaerial weathering processes generate a mantle of soil above bedrock. The characteristics and thickness of this soil mantle are a function of the bedrock lithology, the climate (rainfall, temperature), and the slope of the bedrock surface. These factors govern the intensity of weathering and determine which minerals survive to become part of the soil profile, what new minerals are created in the soil, and the length of time soil materials remain before being eroded and transported to depositional basins. On very steep slopes, for example, the weathered mantle may be removed so rapidly by erosion that little soil accumulates.

In addition to the chemical and physical weathering processes that cause breakdown of bedrock to form soils, several other biologic and chemical processes operate within soils over time to modify their characteristics (e.g., Birkland, 1999, 105; Shaetzl and Anderson, 2005):

1. *Additions to the ground surface*—precipitation of dissolved ions in rainwater; influx of solid particles such as windblown dust; addition of organic matter from surface vegetation
2. *Transformations*
 - a. Decomposition of organic matter within soils to produce organic compounds
 - b. Weathering of primary minerals; formation of secondary minerals, including iron oxides
3. *Transfers*
 - a. Movement of solid or suspended material downward from one soil horizon to a lower horizon by groundwater percolation (eluviation)
 - b. Accumulation of soluble or suspended material in a lower horizon (illuviation)
 - c. Transfer of ions upward by capillary movement of water and precipitation of ions in the soil profile
4. *Removals*—Removal of substances still in solution to become part of the dissolved constituents in groundwater or surface water
5. *Bioturbation of soil*—Soil disruption by animals (e.g., ants, termites) and plants

This list of soil-forming processes is highly simplified. Buol et al. (1997, 112) recognize and define more than two dozen soil-forming processes. These processes generate distinct soil horizons, which are collectively referred to as the soil profile. Further details of soil-forming processes may be found in additional readings listed at the end of this chapter.

Soil Profiles and Soil Classification

Soils are classified on the basis of the characteristic horizontal layers or horizons that are visible in road cuts, pits, and so on. The thickness and nature of these **soil horizons** are determined by the various soil-forming processes mentioned and may vary widely. Soil profiles can be divided crudely into five *major* horizons: O, A, E, B, and C. The **O-horizon** is the surface accumulation of mainly organic matter. The **A horizon**, which occurs at the surface or below the O-horizon, consists of a dark-colored accumulation of organic matter (e.g., leaf litter) that is decaying and mixing with mineral soil. The **E horizon**, which underlies an O or A horizon, is a light colored eluvial horizon (a horizon from which material was removed by downward movement) characterized by less organic matter, fewer iron and aluminum compounds, and/or less clay than the underlying horizon. The **B horizon** underlies an O, A, or E horizon and may contain illuvial (added material derived from an upper horizon) concentrations of fine organic matter, clay, and so forth; most of the original rock structures have been obliterated by soil-forming processes. The **C horizon**, which lies above bedrock, is partly altered bedrock that can be deeply weathered but is relatively unaffected by soil-forming processes. Studies of soil profiles show, however, that soil layers are commonly much more complex than indicated by this simple scheme. As many as 24 different kinds of soil horizons have been described (e.g., Birkland, 1999, 5).

Numerous systems for more detailed classification of soils are in existence: e.g., the Australian handbook classification, the U.S. Soil Taxonomy classification, and the FAO (UNESCO) world map classification (Eswaran et al., 2003). One of the more widely used soil classifications in the United States appears in *Soil Taxonomy: A Basic System for Making Independent Soil Surveys, 2nd ed.* (Soil Survey Staff, 1999), which recognizes 12 major classes or orders of soils with names such as aridosol (soils of arid regions) and ultisol (leached soils of warm, humid regions). These soil types are differentiated on the basis of a variety of complex criteria, such as the amount of contained organic material, the presence of clay layers, and the presence of oxic (iron-rich) horizons.

The factors that influence soil formation, and thus the kinds of soils that form, include the parent rock material, length of soil-forming process, climate (e.g., wet or dry), topography (steep or gentle slopes), and organisms (vegetation cover and soil fauna such as earthworms). Climate plays a particularly important role in soil formation.

PALEOSOLS In the context of this book, we are concerned primarily with ancient soils, called paleosols, rather than modern soils. Paleosols, sometimes referred to as *fossil soils*, are buried soils or horizons of the geologic past. Most soil horizons that developed in the past on elevated landscapes were eventually destroyed as erosion lowered the landscape. Nonetheless, some soils, presumably

those formed mainly in low-lying areas, escaped erosion to become part of the stratigraphic record. Quaternary soils that formed particularly on glacial or fluvial deposits are most common (e.g., Catt, 1986). Such soils that have not been buried are called relict soils. Many buried soils of Quaternary and much older age are also known. Old paleosols occur in the stratigraphic record at major unconformities, including unconformities in Precambrian rocks, where their presence may reflect the combined processes of soil formation, erosional landscape lowering, reorganization of preexisting soil horizons, and changing flow of groundwater (Retallack, 1990, 14). Paleosols are also present as interbeds in sedimentary successions, particularly in alluvial successions, that are at least as old as the Ordovician (e.g., Reinhardt and Sigleo, 1988). Geologists are becoming increasingly interested in paleosols as indicators of paleoenvironments and ancient climatic conditions.

RECOGNITION OF PALEOSOLS Because interbedded paleosols in sedimentary successions superficially resemble sediments or sedimentary rocks, many paleosols have unquestionably gone unrecognized in the past. Many of us have simply identified them as gray, red, or green mudstones. As awareness of paleosols has increased, however, more and more paleosols are being recognized. How can the ordinary geologist, not specifically trained in soil science, recognize paleosols in the field? Retallack (1988, 1997) suggests three principal kinds of diagnostic characteristics of paleosols that help distinguish them from sedimentary rocks: traces of life, soil horizons, and soil structure (Fig. 1.6).

Root traces are the most important **traces of life** preserved in paleosols. Root traces provide diagnostic evidence that rock was exposed to the atmosphere and colonized by plants, thus forming a soil. The top of a paleosol is the surface from which root traces emanate. Root traces mostly taper and branch downward (Fig. 1.7), which helps to distinguish them from burrows. On the other hand, some root traces spread laterally over hardpans in soils, and some kinds branch upward and out of the soil. Root traces are most easily recognized when their original organic matter is preserved, which occurs mostly in paleosols formed in waterlogged, anoxic lowland environments. Root traces in red, oxidized paleosols consist mainly of tubular features filled with material different from the surrounding paleosol matrix.

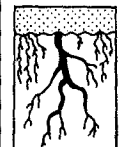
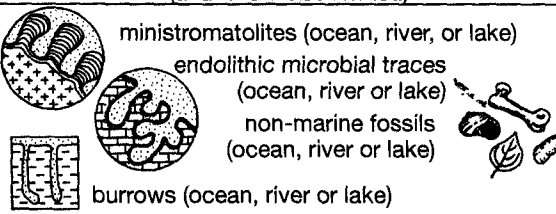
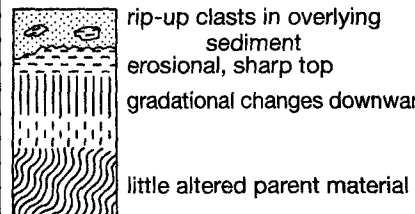
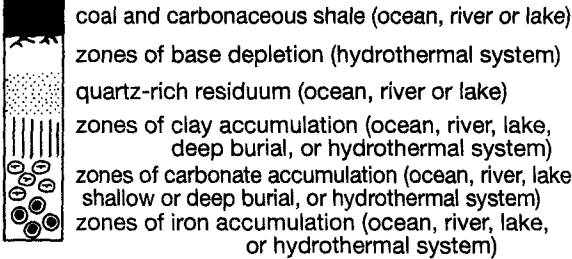
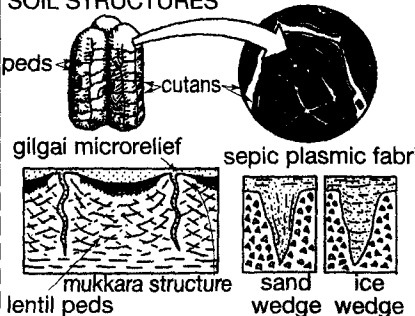
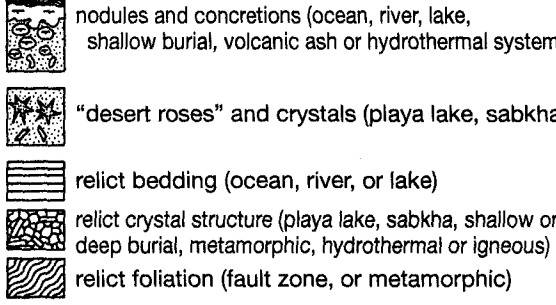
DIAGNOSTIC FEATURES OF PALEOSOLS	PROMINENT OR COMMON FEATURES OF PALEOSOLS (and where else formed)
ROOT TRACES  <ul style="list-style-type: none"> truncated tops downward taper downward branching 	 <ul style="list-style-type: none"> minitromatolites (ocean, river, or lake) endolithic microbial traces (ocean, river or lake) non-marine fossils (ocean, river or lake) burrows (ocean, river or lake)
SOIL HORIZONS  <ul style="list-style-type: none"> rip-up clasts in overlying sediment erosional, sharp top gradational changes downward little altered parent material 	 <ul style="list-style-type: none"> coal and carbonaceous shale (ocean, river or lake) zones of base depletion (hydrothermal system) quartz-rich residuum (ocean, river or lake) zones of clay accumulation (ocean, river, lake, deep burial, or hydrothermal system) zones of carbonate accumulation (ocean, river, lake, shallow or deep burial, or hydrothermal system) zones of iron accumulation (ocean, river, lake, or hydrothermal system)
SOIL STRUCTURES  <ul style="list-style-type: none"> peds gilgai microrelief mukgara structure sand wedge ice wedge cutans sepic plasmic fabric 	 <ul style="list-style-type: none"> nodules and concretions (ocean, river, lake, shallow burial, volcanic ash or hydrothermal system) "desert roses" and crystals (playa lake, sabkha) relict bedding (ocean, river, or lake) relict crystal structure (playa lake, sabkha, shallow or deep burial, metamorphic, hydrothermal or igneous) relict foliation (fault zone, or metamorphic)

FIGURE 1.6 Characteristic and common features useful in recognition of paleosols. [From Retallack, G. J., 1992, How to find a Precambrian paleosol, in Schidlowski, M., et al. (eds.), *Early organic evolution: Implications for mineral and energy resources*, Springer-Verlag, Berlin, Fig. 10, p. 27, reproduced by permission.]



FIGURE 1.7 An example of root traces in a paleosol. The original organic matter has been partially replaced by iron oxides. Early Miocene, Molalla Formation, western Oregon. [Photograph courtesy of G. J. Retallack.]

The presence of **soil horizons** is a second general feature of paleosols. The top of the uppermost horizon of a paleosol is commonly sharply truncated by an erosional surface, but soil horizons typically show gradational changes in texture, color, or mineral content downward into the parent material. Differences in grain size, color, reaction with weak hydrochloric acid (to test for the presence of carbonates), and the nature of the boundaries must all be examined to detect soil horizons (Retallack, 1988). Comparison with modern soil horizons aids in recognition.

Bioturbation (disruption) by plants and animals, wetting and drying, and other soil-forming processes cause paleosols to develop characteristic **soil structures** at the expense of the original bedding and structures in the parent rock. One of the characteristic kinds of soil structure is a network of irregular planes (called **cutans**) surrounded by more stable aggregates of soil material called **peds**. This structure gives a hackly appearance to the soil. Peds occur in a variety of sizes and shapes (Fig. 1.8). Their recognition in the field depends upon recognition of the cutans that bound them, which commonly form clay skins around the peds. Other kinds of soil structure include concentrations of specific minerals that form hard, distinct, calcareous, ferruginous, or sideritic lumps called **gleabules** (a general term including nodules and concretions). More diffuse, irregular, or weakly mineralized concentrations are called mottles. Figure 1.9 shows the field appearance of some Miocene paleosols. These paleosols are red; however, paleosols can have a variety of colors and properties (Retallack, 1997).

Paleosols can be recognized to have characteristics similar to those of modern soils; thus, U.S. Soil Taxonomy names such as aridosol and ultisol can be applied to paleosols (e.g., Retallack, 1992). Because the characteristics of paleosols reflect the conditions under which they formed, including climatic conditions, study of paleosols is an important tool in paleoenvironmental analysis. For example, aridosols suggest formation under desert conditions whereas ultisols reflect weathering under warm, moist conditions. Clearly, the processes of weathering that lead to

generation of sedimentary particles and soil formation are intimately tied up with climatic conditions. Weathering did not begin on Earth until an atmosphere containing water vapor and carbon

TYPE	PLATY	PRISMATIC	COLUMNAR	ANGULAR BLOCKY	SUBANGULAR BLOCKY	GRANULAR	CRUMB
SKETCH							
DESCRIPTION	tabular and horizontal to land surface	elongate with flat top and vertical to land surface	elongate with domed top and vertical to surface	equant with sharp interlocking edges	equant with dull interlocking edges	spheroidal with slightly interlocking edges	rounded and spheroidal but not interlocking
USUAL HORIZON	E, Bs, K, C	Bt	Bn	Bt	Bt	A	A
MAIN LIKELY CAUSES	initial disruption of relict bedding; accretion of cementing material	swelling and shrinking on wetting and drying	as for prismatic, but with greater erosion by percolating water, and greater swelling of clay	cracking around roots and burrows, swelling and shrinking on wetting and drying	as for angular blocky, but with more erosion and deposition of material in cracks	active bioturbation and coating of soil with films of clay, sesquioxides, and organic matter	as for granular; including fecal pellets and relict soil clasts
SIZE CLASS	very thin < 1 mm	very fine < 1 cm	very fine < 1 cm	very fine < 0.5 cm	very fine < 0.5 cm	very fine < 1 mm	very fine < 1 mm
	thin 1 to 2 mm	fine 1 to 2 cm	fine 1 to 2 cm	fine 0.5 to 1 cm	fine 0.5 to 1 cm	fine 1 to 2 mm	fine 1 to 2 mm
	medium 2 to 5 mm	medium 2 to 5 cm	medium 2 to 5 cm	medium 1 to 2 cm	medium 1 to 2 cm	medium 2 to 5 mm	medium 2 to 5 mm
	thick 5 to 10 mm	coarse 5 to 10 cm	coarse 5 to 10 cm	coarse 2 to 5 cm	coarse 2 to 5 cm	coarse 5 to 10 mm	not found
	very thick > 10 mm	very coarse > 10 cm	very coarse > 10 cm	very coarse > 5 cm	very coarse > 5 cm	very coarse > 10 mm	not found

FIGURE 1.8 Characteristics of various kinds of soil peds. [From Retallack, G. J., 1988, in Reinhardt, J., and W. R. Sigleo (eds.), *Field recognition of paleosols*: Geol. Soc. America Spec. Paper 216, Fig. 9, p. 216. Reproduced by permission of Geol. Soc. America, Boulder, Colo.]

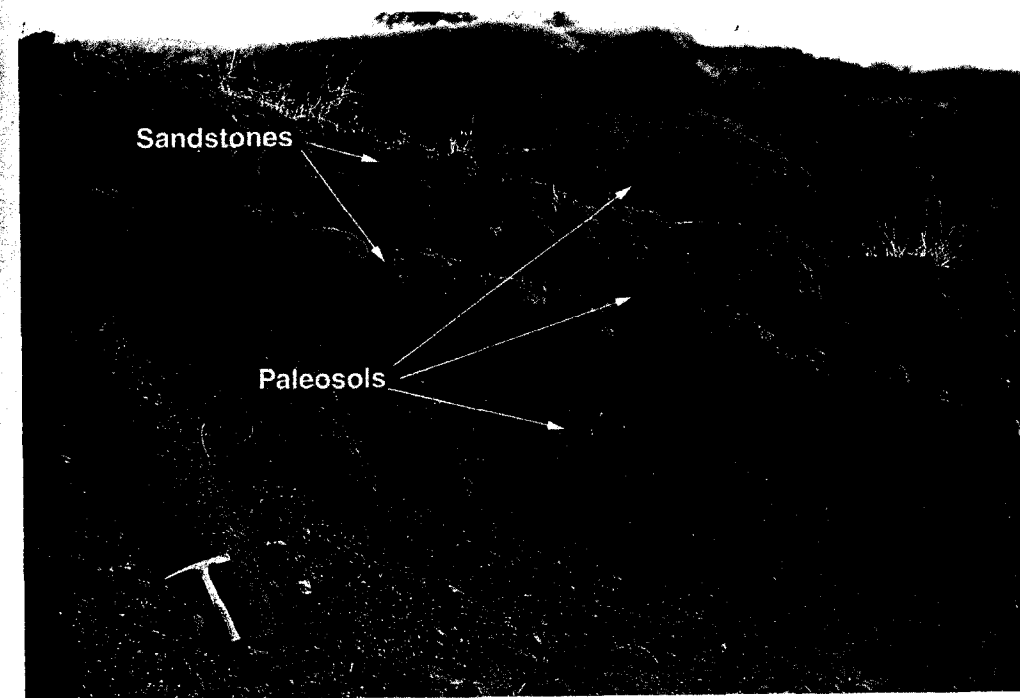


FIGURE 1.9 Red paleosols exposed below bedded sandstones in the Middle Miocene, Chinji Formation, Siwalik Group, in a creek bed 3 km south of Khaur, Potwar Plateau, Pakistan. The hammer is 25 cm long. From Retallack, Gregory, J.: *A Colour Guide to Paleosols*, 1997. Copyright John Wiley & Sons, Ltd. Reproduced with permission.

dioxide had accumulated sometime during the early Precambrian; subsequent addition of oxygen also had an important bearing on the weathering processes. Geologists are becoming increasingly aware of the need to study Earth's past climates (paleoclimatology).

This short, generalized description of paleosols is intended only to pique reader interest in fossil soils. Several of the books listed under Additional Readings at the end of the chapter provide further details.

1.5 CONCLUDING REMARKS

The processes that form sedimentary rocks can be considered to begin with weathering, a process strongly influenced by climatic conditions. Weathering brings about breakdown of older rocks exposed in upland areas to yield soluble ions, which are transported to the ocean in solution, and insoluble, chemically resistant minerals such as quartz that may accumulate at the weathering site for a time as soils. Soil formation, like weathering, is intimately related to climatic conditions. Some soils, called paleosols, are preserved to become part of the sedimentary record; however, most insoluble soil materials are removed by erosion and transported by gravity processes, water, glaciers, or wind to basins at lower elevations, where deposition takes place. Succeeding chapters of this book describe the processes of sediment transport, deposition, and burial that ultimately result in generation of lithified sedimentary rocks.

Further Readings

Weathering

Bland, W., and D. Rolls. 1998. *Weathering: An introduction to the scientific principles*. New York: Oxford University Press Inc.
 Humphris, S. E., et al. (eds.). 1995. *Seafloor hydrothermal systems*. Geophysical Monograph 91. Washington, D.C.: American Geophysical Union.
 Martini, I. P., and W. Chesworth (eds.). 1992. *Weathering, soils and paleosols*. Amsterdam: Elsevier.
 Nahon, D. B. 1991. *Introduction to the petrology of soils and chemical weathering*. New York: John Wiley & Sons.

Parson, L. M., C. L. Walker, and D. R. Dixon (eds.). 1995. *Hydrothermal vents and processes*. London: The Geological Society.
 Robinson, D. A., and R. B. G. Williams (eds.). 1994. *Rock weathering and landform evolution*. Chichester: John Wiley & Sons.
 Turkington, A. V. 2004. Sandstone weathering: a century of research and innovation. *Geomorphology*. v. 67. 229–253.
 White, A. F., and S. L. Brantley (eds.). 1995. Chemical weathering rates of silicate minerals. *Mineralogical Society of America Reviews in Mineralogy*. v. 31.

Soils and Paleosols

- Birkland, P. W. 1999. *Soils and geomorphology*. 3rd ed. New York: Oxford University Press.
- Bronger, A., and J. A. Catt (eds.). 1989. *Paleopedology: Nature and application of paleosols*. Destedt, Germany: Catena Verlag.
- Buol, S. W., et al. 1997. *Soil genesis and classification*. 4th ed. Ames, Iowa: Iowa State University Press.
- Meyer, R., 1997. *Paleoalterites and paleosols*. Rotterdam: A.A. Balkema.
- Ollier, C., and C. Pain. 1996. *Regolith, soils and landforms*. Chichester: John Wiley & Sons.
- Paquet, H., and N. Clauer (eds.). 1997. *Soils and sediments: Mineralogy and geochemistry*. Berlin: Springer-Verlag.
- Reinhardt, J., and W. R. Sigleo (eds.). 1988. *Paleosols and weathering through geologic time: Principles and applications*. Geological Society of America Special Paper 216.
- Schaetzl, R., and S. Anderson. 2005. *Soils: Genesis and geomorphology*. Cambridge: Cambridge University Press.
- Retallack, G. J. 1997. *A colour guide to paleosols*. Chichester: John Wiley & Sons.
- Retallack, G. J. 2001. *Soils of the past*. Oxford: Blackwell Science.

References Cited

- Allison, R. J., and A. S. Goudie. 1994. The effects of fire on rock weathering: an experimental study. in Robinson, D. A., and R. B. G. Williams (eds.). *Rock weathering and landform evolution*. Chichester: John Wiley & Sons. 1–56.
- Birkland, P.W. 1999. *Soils and geomorphology*. 3rd ed. New York: Oxford University Press.
- Bland, W., and D. Rolls. 1998. *Weathering: An introduction to the scientific principles*. New York: Oxford University Press.
- Boggs, S., Jr., D.G. Livermore, and M.G. Seitz. 1985. Humic macromolecules in natural waters. *Jour. Macromolecular Science* C25(4):599–657.
- Buol, S. W., et al. 1997. *Soil genesis and classification*. 4th ed. Ames, Iowa: Iowa State University Press.
- Cann, J. R., and M. R. Strens. 1989. Modeling periodic megaplume emission by black smoker systems: *Jour. Geophys. Research* 94:12,227–12,237.
- Catt, J. A. 1986. *Soils and Quaternary geology: A handbook for field scientists*. Oxford: Clarendon Press.
- Corliss, J. B., et al. 1979. Submarine thermal springs on the Galápagos Rift. *Science* 203:1073–1083.
- Edmond, J. M., et al. 1982. Chemistry of hot springs on the East Pacific Rise and their effluent dispersal. *Nature* 297:187–191.
- Eswaran, H., et al. (eds.). 2003. *Soil classification: a global desk reference*. Boca Raton: CRC Press.
- Garrels, R. M., and F. T. McKenzie. 1971. *Evolution of sedimentary rocks*. New York: W.W. Norton.
- Goldich, S. S. 1938. A study of rock weathering. *Jour. Geology* 46:17–58.
- Halsey, D. P., D. J. Mitchell, and S. J. Dews. 1998. Influence of climatically induced cycles in physical weathering. *Quarterly Journal of Engineering Geology* 31: 359–367.
- Jackson, M. L. 1968. Weathering of primary and secondary minerals in soils. *Transactions, 9th International Congress Soil Science*. 4:281–292.
- Kadko, D., J. Baross, and J., Alt. 1995. The magnitude and global implications of hydrothermal flux, in Humphris, S.E., et al. (eds.), *Seafloor hydrothermal systems*. Geophysical Monograph 91, American Washington, D.C.: Geophysical Union. 446–466.
- Karl, D. M., et al. 1988. Loihi Seamount, Hawaii: A mid-plate volcano with a distinctive hydrothermal system. *Nature* 335:532–535.
- Krauskopf, K. B. 1979. *Introduction to geochemistry*. 2nd ed. New York: McGraw-Hill.
- Macdonald, K. C., K. B. F. N. Spiess, and R. D. Ballard. 1980. Hydrothermal flux of the “black smoker” vents on the East Pacific Rise. *Earth and Planetary Sci. Letters* 48:1–7.
- Nahon, D. B. 1991. *Introduction to the petrology of soils and chemical weathering*. New York: John Wiley & Sons.
- Nicholson, D. T., and F. H. Nicholson. 2000. Physical deterioration of sedimentary rocks subjected to experimental freeze-thaw weathering. *Earth Surface Processes and Landforms* 25:1295–1307.
- Ollier, C., and C. Pain. 1996. *Regolith, soils and landforms*. New York: John Wiley & Sons.
- Palmer, M. R., and J. M. Edmond. 1989. The strontium isotope budget of the modern ocean. *Earth and Planetary Science Letters*. 92:11–26.
- Parson, L. M., C. L. Walker, and D. R. Dixon (eds.). 1995. *Hydrothermal vents and processes*. London: The Geol. Society.
- Reinhardt, J., and W. R. Sigleo (eds.). 1988. *Paleosols and weathering through geologic time: Principles and applications*. *Geol. Soc. America Spec. Paper 216*.
- Retallack, G. J. 1990. *Soils of the past*. Boston: Unwin Hyman.
- Retallack, G. J., 1992. Paleozoic paleosols, in Martini, I. P., and W. Chesworth (eds.). *Weathering, soils and paleosols*. Amsterdam: Elsevier. 543–564.
- Retallack, G. J. 1997. *A colour guide to paleosols*. Chichester: John Wiley & Sons.
- . 1988. Field recognition of paleosols, in Reinhardt, J., and W. R. Sigleo (eds.). *Paleosols and weathering through geologic time*. *Geol. Soc. Amer. Spec. Paper 216*. 1–20.
- Schaetzl, R., and S. Anderson. 2005. *Soils: Genesis and geomorphology*. Cambridge: Cambridge University Press.
- Soil Survey Staff. 1999. *Soil taxonomy: A basic system of soil classification for making independent soil surveys*. 2nd ed. Agricultural Handbook No. 436. U.S. Department of Agriculture, Natural Resources Conservation Service.
- Sperling, C. H. B., and R. U. Cooke. 1980. Salt weathering in an arid environment: Experimental investigations of the relative importance of hydration and recrystallization processes. *Papers in Geography* 9. London: Bedford College.
- Stanley, S. M., and L. A. Hardie. 1999. Hypercalcification: Paleontology links plate tectonics and geochemistry to sedimentology. *GSA Today* 9:1–7.
- Taylor, G., and R. A. Eggleton. 2001. *Regolith geology and geomorphology*. Chichester: John Wiley & Sons.
- Turkington, A. V., and T. R. Paradise. 2005. Sandstone weathering: A century of research and innovation. *Geomorphology* 67:229–253.
- Von Damm, K. L. 1990. Seafloor hydrothermal activity: Black smoker chemistry and chimneys. *Ann. Rev. Earth and Planetary Sciences*. 18:173–204.
- Watson, A. 1992. Desert soils. in Martini, I. P., and W. Chesworth (eds.). *Weathering, soils and paleosols*. Amsterdam: Elsevier. 225–260.
- White, A. F., and S. L. Brantley. 1995. Chemical weathering rates of silicate minerals: An overview. in White, A. F., and S. L. Brantley (eds.). *Chemical weathering rates of silicate minerals*. *Mineralogical Society of America, Reviews in Mineralogy* 31:1–22.
- Wright, J. S. 2007. An overview of the role of weathering in the production of quartz silt. *Sedimentary Geology*. 202:337–351.

Transport and Deposition of Siliciclastic Sediment

2.1 INTRODUCTION

Silicate minerals and rock fragments weathered from older rocks on land, together with pyroclastic particles generated by explosive volcanism, are the source materials of siliciclastic sedimentary rocks—conglomerates, sandstones, shales. These materials are eroded from highlands and transported to depositional basins at lower elevations, where they may undergo additional transport before final deposition. Mass-wasting processes such as slides and slumps commonly play an initial role in moving sediment short distances down steep slopes to sites where other transport processes take over. Subsequent transport may involve fluid flows (e.g., moving water) or sediment-gravity flows, such as mud flows, that may behave like fluids. Thus, study of sediment transport requires some understanding of the principles of fluid flow.

The fundamental laws of fluid dynamics are moderately complex when applied to fluid flow alone. These complexities are magnified when particles are entrained in the flow during sediment transport. Sediment transport can take place under a variety of conditions: subaerially by wind and certain kinds of sediment-gravity flows, and subaqueously in rivers, lakes, and the ocean by currents, waves, tides, and sediment-gravity flows.

In this chapter, we investigate sediment transport processes by first examining some of the properties of fluids and the basic concepts of fluid flow and sediment-gravity flow. We then consider the problems involved in entrainment and transport of particles by fluid- and sediment-gravity-flow processes. No attempt is made here to give a comprehensive review of fluid mechanics. Only those concepts of flow that are important to understanding sediment transport and deposition are discussed, and these concepts are presented in very simplified form. More rigorous treatment of fluid dynamics is available in numerous specialized books, e.g., Middleton and Southard (1984), Leeder (1999), Rowan et al. (2006). Details of sediment transport peculiar to various depositional environments (e.g., river systems, lakes, the ocean) are discussed in appropriate sections of subsequent chapters.

2.2 FUNDAMENTALS OF FLUID FLOW

Fluids are substances that change shape easily under their own weight. Air, water, and water containing various amounts of suspended sediment are the fluids of interest in sediment transport. The basic physical properties of these fluids are density and viscosity. Differences in these properties markedly affect the ability of fluids to erode and transport sediment.

Fluid **density**, commonly referred to as ρ (rho), is defined as mass per unit fluid volume. Density affects the magnitude of forces that act within a fluid and on the bed as well as the rate at which particles fall or settle through a fluid (slower in denser fluids). Density particularly influences the movement of fluids downslope under the influence of gravity. Density varies with different fluids and increases with decreasing temperature of a fluid. The density of water (0.998 g/mL at 20°C) is more than 700 times greater than that of air. This density difference influences the relative abilities of water and air to transport sediment; for example, water can transport particles of much larger size than those transported by wind.

Fluid **viscosity** is a measure of the ability of fluids to flow. Put simply, fluids with low viscosity flow readily and fluids with high viscosity flow sluggishly. For example, air has very low viscosity and ice has very high viscosity. Water has low viscosity; honey has high viscosity. The viscosity of water at 20°C is almost 55 times greater than that of air (Blatt, Middleton, and Murray, 1980, 91). Like density, viscosity increases with decreasing temperature of the fluid. Viscosity has a particularly important influence on water turbulence. Increasing viscosity tends to suppress turbulence (random movement of water molecules), thereby slowing the rate at which particles settle through water—a significant factor in transport of suspended sediment. Decreased turbulence also reduces the ability of running water to erode and entrain sediment.