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PRINCIPLES OF  
SEDIMENTOLOGY  
AND  
STRATIGRAPHY

FOURTH EDITION



SAM BOGGS, JR.

# ***Principles of Sedimentology and Stratigraphy***

# *Principles of Sedimentology and Stratigraphy*

Fourth Edition

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**Sam Boggs, Jr.**  
University of Oregon



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

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# Preface

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The roots of sedimentology and stratigraphy extend back to the 16<sup>th</sup> century; however, these disciplines are still growing and changing. Geologists continue to “fine tune” sedimentologic and stratigraphic concepts through a variety of research avenues and by using an array of increasingly sophisticated research tools. The result is a continuous outpouring of fresh data and new ideas. In fact, it is becoming increasingly difficult to keep abreast of the flood of new information appearing in the geological literature. A glance through recent issues of a well-known sedimentology journal reveals important new papers on sedimentation and tectonics, depositional systems, carbonates, biosedimentology, diagenesis, provenance, geochemistry, sediment transport and sedimentary structures, stratigraphic architecture, chronostratigraphy, numerical modeling, paleoclimatology, sequence stratigraphy, and basin analysis—to name but a few research areas.

I make no claim that I have, in this fourth edition of *Principles of Sedimentology and Stratigraphy*, fully evaluated all of these new data or captured all of the new ideas and concepts that may have been put forward since publication of the third edition. I have, however, tried to weave important new information into the basic structure of previous editions and revise concepts that may have become outdated. In addition, I have reorganized some of the chapters, added numerous references to pertinent new research articles and books, and added a significant number of new photographs, line drawings, and tables. I hope that these changes increase both the readability of the book for students and also keep them abreast of recent developments in the fields of sedimentology and stratigraphy.

As mentioned in the preface to the third edition, career opportunities for geology students are shifting away from the more traditional avenues of petroleum and mining geology toward environmental geology and other disciplinary areas that deal with problems of society. To be prepared for these careers, students need to gain a solid foundation in the basic principles of sedimentology, stratigraphy, and related sciences, as well as to develop insight into innovative applications of these principles to areas of study such as environmental analysis, paleoclimate evaluation, groundwater resources, and marine pollution. I hope that this book provides a useful part of the basic background that students need to advance into these exciting career fields.

I want to thank the following people for reviewing the Third Edition. I used their reviews to guide me in the preparation of the Fourth Edition:

Edwin J. Anderson, Temple University; Janok P. Bhattacharya, University of Texas – Dallas; Charles W. Byers, University of Wisconsin; Beth Christensen, Georgia State University; Joachim Dorsch, Saint Lewis University; James E. Evans, Bowling Green State University; Larry T. Middleton, Northern Arizona University; Michael R. Owen, Saint Lawrence University; Bruce Selleck, Colgate University; and Mark A. Wilson, College of Wooster.

SAM BOGGS, JR.

# Introduction

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## Kinds of Sedimentary Rocks

This book describes and discusses the physical, chemical, and biological characteristics of sedimentary rocks and the interpretations that we draw from these characteristics about the origin of sedimentary rocks. Geologists disagree somewhat about how the various kinds of sedimentary rocks should be classified; however, such rocks can conveniently be placed into three fundamental groups on the basis of composition and origin: siliciclastic, chemical/biochemical, and carbonaceous.

**Siliciclastic** sedimentary rocks are composed dominantly of silicate minerals, such as quartz and feldspar, and rock fragments (clasts). These materials originate mainly by the chemical and physical breakdown (weathering) of igneous, metamorphic, or (older) sedimentary rock. Conglomerates, sandstones, and shales belong to this group. Silicate detritus, including silicate minerals, rock fragments, and glass shards, can also be generated by explosive volcanism. Siliciclastic sedimentary rocks that formed mainly from the products of explosive volcanism are called **volcaniclastic** rocks. **Chemical/biochemical** sedimentary rocks are composed of minerals precipitated mainly from ocean or lake water by inorganic (chemical) and/or organic (biogenic) processes. They include limestone, chert, evaporites such as gypsum, phosphorites, and iron-rich sedimentary rocks. Evaporites are probably precipitated entirely by inorganic processes resulting from evaporation of lake or seawater. Biogenic processes, as well as inorganic processes, play an important role in the formation of many limestones and likely play some role in the origin of chert, phosphorites, and iron-rich sedimentary rocks. **Carbonaceous** sedimentary rocks contain a substantial amount (>~15%) of highly altered remains of the soft tissue of plants and animals, referred to as organic matter. The principal carbonaceous rocks are coal and oil shale. Carbonaceous sedimentary rocks make up only a small fraction of the total sedimentary record; however, these rocks (especially coals) have great economic importance as fossil fuels.

## Distribution of Sedimentary Rocks in Time and Space

Sedimentary rocks are confined to Earth's outer crust, where they make up only 5–10 percent of the outer 10 miles (16 km) or so of the crust. On the other hand, they are the most common rocks at Earth's surface. Sedimentary rocks and sediments cover nearly three-fourths of Earth's land surface and most of the ocean floor. They range in age from Precambrian to modern. The first sedimentary rocks were deposited nearly four billion years ago, at which time most of Earth's surface was covered with volcanic rocks. The relative proportion of sedimentary rocks at Earth's surface has increased progressively with time, as weathering processes brought about decomposition of other kinds of rock and deposition of the decomposition products to form sedimentary rocks.

## **Sedimentology Versus Stratigraphy**

The record of Earth history locked up in sedimentary rocks dates back almost four billion years. It is the study of this reservoir of Earth history that constitutes the sciences of sedimentology and stratigraphy. **Sedimentology** is the scientific study of the classification, origin, and interpretation of sediments and sedimentary rocks. It is often difficult to draw a sharp distinction between sedimentology and **stratigraphy**, which is defined simply and broadly as the science of rock strata. In general, however, sedimentology is concerned with the physical (textures, structures, mineralogy), chemical, and biologic (fossils) properties of sedimentary rocks and the processes by which these properties are generated. It is these properties that provide much of the basis for interpreting the physical features, climate, and environmental conditions of Earth in the geologic past. Stratigraphy, on the other hand, is concerned more with age relationships of strata, successions of beds, local and worldwide correlation of strata, and stratigraphic order and chronological arrangement of beds in the geologic column. Stratigraphy finds special applications in the study of plate reconstructions (plate tectonics) and in the unraveling of the intricate history of landward and seaward movements of ocean shorelines (transgressions and regressions) and rise and fall of sea level through time. Particularly exciting developments in stratigraphy have come about recently by applying the principles of seismology and paleomagnetism to stratigraphic problems.

## **Brief History of Sedimentology and Stratigraphy**

**Sedimentologic and stratigraphic study date** back to about A.D. 1500 with the observations of Leonardo da Vinci on fossils in sedimentary rocks of the Italian Apennines. Since that time, a steady drumbeat of progress in understanding sedimentary rocks has taken place, punctuated at intervals by significant new developments in tools and techniques for studying sedimentary rocks and emergence of new concepts and ideas about their origin. Especially noteworthy among these seminal events were (1) initiation of the use of the microscope to study fossils by Robert Hooke in the latter part of the 17<sup>th</sup> century, (2) elucidation of the concept of uniformitarianism (loosely, the present is the key to the past) by James Hutton in the late 18<sup>th</sup> century, (3) the birth of biostratigraphy (study and interpretation of sedimentary rocks on the basis of the fossils they contain) by William Smith in the early 19<sup>th</sup> century, (4) application of the petrographic microscope to study of sedimentary rocks by Henry Clifton Sorby around 1850, (5) development of one of the most far-reaching concepts in geologic philosophy—seafloor spreading and global plate tectonics—in the early 1960s, and (6) emergence of the concepts of seismic stratigraphy (study of seismic data for the purpose of extracting stratigraphic information), sequence stratigraphy (application of the concept of depositional sequences to stratigraphic interpretation), and magnetostratigraphy (study of rock magnetism as a stratigraphic tool) in the 1960s and 1970s. The pace of new developments in sedimentology and stratigraphy continues to the present time, spurred by the availability of technologically advanced laboratory tools such as the scanning electron microscope and mass spectrometer and development of advanced field procedures such as the ability to drill deep holes in the ocean floor and recover sediment cores in water several thousand meters deep.

## **Why Study Sedimentary Rocks?**

The sheer abundance of sedimentary rocks at Earth's surface provides a partial answer to a question frequently asked by students, "Why should we study sedimentary rocks; why bother?" In addition to their abundance, however, they are also important because of information they yield about Earth's history and because of the economic products they contain. All geologic study is aimed in one way or another at developing a better understanding of Earth's history. All rocks, whether sedimentary, igneous, or metamorphic, contain clues to some aspect of this history, but sedimentary rocks are unique with regard to the information they provide. From the composition, textures, structures, and fossils in sedimentary rocks, experienced geologists can decipher clues that provide insight into past climates, oceanic environments and ecosystems, the configurations of ancient land systems, and the locations and compositions of ancient mountain systems long since vanished. Thus, study of sedimentary rocks forms the primary basis for the sciences of paleoclimatology (study of climates throughout geologic time), paleogeography (study and description of the physical geography of Earth's past), paleoecology (study of the relationship between ancient organisms and their environment), and paleoceanography (study of the characteristics of ancient oceans). In addition, many sedimentary rocks have economic significance. Most of the world's oil and gas and all of its coal are contained in sedimentary rock successions. Iron-bearing minerals, uranium minerals, evaporite minerals, phosphate minerals, and many other economically valuable minerals also occur in these rocks.

Thus, the disciplines of sedimentology and stratigraphy, while having their roots in studies dating back to the early 16<sup>th</sup> century, are still vibrant, exciting, growing disciplines. I hope that this book will help students capture some of this sense of excitement: It provides an integrated view of sedimentology and stratigraphy. The first few chapters are devoted to description and discussion of the processes that form sedimentary rocks, the physical, chemical, and biological properties of rocks that result from these processes, and the principal kinds of sedimentary rocks. Succeeding chapters deal with sedimentary environment and their interpretation from the rock record; stratigraphic relationships revealed through study of lithology, seismic reflection characteristics, remanent magnetism, fossils, and radiometric ages; and basin analysis, which is the integrated sedimentological and stratigraphic study of sedimentary rocks.

## **Additional Sources of Information**

Numerous references are made throughout this book to research papers that provide detailed information about particular topics. In addition, a list of pertinent monographs is provided at the end of each chapter. Readers should find these research papers and books a useful starting point for additional literature research. Finally, Appendix E furnishes an extended list of Web sites where online information about sedimentology and stratigraphy is available.



# PART I

## *Origin and Transport of Sedimentary Materials*



Sediment transport in the braided Kongakut River, Arctic National Wildlife Refuge, Alaska

**S**edimentary 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.

# 1

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## *Weathering and Soils*

### 1.1 INTRODUCTION

**W**eathering 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 mid-ocean 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, p. 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.

#### *Insolation Weathering*

Expansion of rock surfaces heated by the Sun (insolation) followed by contraction as the 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, p. 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). Although observations



**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.]

in desert areas suggest that insolation weathering does occur, heating and cooling experiments in the laboratory have not yielded conclusive proof that insolation weathering is an important process. The concept remains controversial.

### *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). 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 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. **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. 1.2). The fractures that separate the weathering rinds may form in response to stress release or possibly thermal changes (Taylor and Eggleton, 2001, p. 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 1.2**

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, disrupt the fabric of the weathered rock, producing a loose residue of resistant grains and secondary minerals. 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 ( $<30^{\circ}\text{C}$ ), 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, p. 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 groundwaters. If carbon dioxide is dissolved in the rainwater through interaction with atmospheric or soil  $\text{CO}_2$ , the usual case in the weathering environment, the solubilizing ability of water is enhanced. Dissolution of  $\text{CO}_2$  in water forms carbonic acid ( $\text{H}_2\text{CO}_3$ —this is what you consume in your soft drinks), which subsequently dissociates to produce hydrogen ions and carbonate ions ( $\text{CO}_2 + \text{H}_2\text{O} \leftrightarrow \text{H}_2\text{CO}_3 \leftrightarrow \text{H}^+ + \text{HCO}_3^-$ ). Increase in  $\text{H}^+$  ions, relative to  $\text{OH}^-$  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 the 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

#### 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  $\text{H}^+$  concentrations ranging from  $10^0$  to  $10^{-14}$ . For example, a solution containing a  $\text{H}^+$  concentration of  $10^{-1}$  moles per liter has a pH of 1, an  $\text{H}^+$  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
<b>Simple (congruent) Solution</b> —Dissolution of soluble minerals in H <sub>2</sub> O (direct solution) or in H <sub>2</sub> O + CO <sub>2</sub> (carbonation) to yield cations and anions in solution	SiO <sub>2</sub> + 2H <sub>2</sub> O → H <sub>4</sub> SiO <sub>4</sub> (direct solution) (quartz) (silicic acid) aq CaCO <sub>3</sub> + H <sub>2</sub> O + CO <sub>2</sub> ↔ Ca <sup>2+</sup> + 2HCO <sub>3</sub> <sup>-</sup> (Carbonation) (calcite) aq aq	Highly soluble minerals (e.g., gypsum, halite), quartz Carbonate rocks
<b>Hydrolysis (incongruent dissolution)</b> —Reaction between H <sup>+</sup> and OH <sup>-</sup> ions of water and the ions of silicate minerals, yielding soluble cations, silicic acid, and clay minerals (if Al present)	2KAlSi <sub>3</sub> O <sub>8</sub> + 2H <sup>+</sup> + 9H <sub>2</sub> O → H <sub>4</sub> Al <sub>2</sub> Si <sub>2</sub> O <sub>9</sub> + 4H <sub>4</sub> SiO <sub>4</sub> + 2K <sup>+</sup> (orthoclase) aq (kaolinite) (silicic acid) aq 2NaAlSi <sub>3</sub> O <sub>8</sub> + 2H <sup>+</sup> + 9H <sub>2</sub> O → H <sub>4</sub> Al <sub>2</sub> Si <sub>2</sub> O <sub>9</sub> + 4H <sub>4</sub> SiO <sub>4</sub> + 2Na <sup>+</sup> (albite) aq (kaolinite) (silicic acid) aq	Silicate minerals
<b>Oxidation</b> —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 <sub>2</sub> + 15/2O <sub>2</sub> + 4H <sub>2</sub> O → Fe <sub>2</sub> O <sub>3</sub> + 4SO <sub>4</sub> <sup>2-</sup> + 8H <sup>+</sup> (pyrite) (hematite) aq aq MnSiO <sub>3</sub> + 1/2O <sub>2</sub> + 2H <sub>2</sub> O → MnO <sub>2</sub> + H <sub>4</sub> SiO <sub>4</sub> (rhodonite) (pyrolusite) (silicic acid)	Iron- and manganese-bearing silicate minerals, iron sulfides
<b>Other Processes</b>		
<b>Hydration and Dehydration</b> —Gain (hydration) or loss (dehydration) of water molecules from a mineral, resulting in formation of a new mineral	Fe <sub>2</sub> O <sub>3</sub> + H <sub>2</sub> O ↔ 2FeOOH (hydration) (hematite) (goethite) CaSO <sub>4</sub> · 2H <sub>2</sub> O ↔ CaSO <sub>4</sub> + 2H <sub>2</sub> O (dehydration) (gypsum) (anhydrite)	Ferric oxides Evaporites
<b>Ion Exchange</b> —Exchange of ions, principally cations, between solutions and minerals	K-clay + Mg <sup>2+</sup> ↔ Mg-clay + K <sup>+</sup> Ca-zeolite + Na <sup>+</sup> ↔ Na-zeolite + Ca <sup>2+</sup>	Clay minerals and zeolites
<b>Chelation</b> —Bonding of metal ions to organic molecules having ring structures	Metal ions (cations) + chelating agent (e.g., secreted by lichens) → H <sup>+</sup> ions + chelate (metal ions/organic molecules in solution)	Silicate minerals

Note: aq = aqueous



*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  $CO_2$  in water. Thus, the more  $CO_2$  that is dissolved in water, the more aggressive the hydrolysis reaction. Hydrolysis can also take place in water containing little or no dissolved  $CO_2$ , 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 ( $H_4SiO_4$ ); however, some of the silica may separate as colloidal or amorphous  $SiO_2$  and be left behind during weathering to combine with aluminum to form clay minerals. Hydrolysis is the primary process by which silicate minerals decompose during weathering. A more rigorous and detailed discussion of this process is given by Nahon (1991, p. 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 ( $Fe^{2+} \rightarrow Fe^{3+} + e^-$ , where  $e^-$  = electron transfer), which causes loss of other cations such as  $Si^{4+}$  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 ( $FeS_2$ ) is oxidized to form hematite ( $Fe_2O_3$ ), 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  $Fe^{3+}$  to  $Fe^{2+}$ . Ferrous iron ( $Fe^{2+}$ ) is more soluble, and thus more mobile, than ferric iron ( $Fe^{3+}$ ) 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, is 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 CO<sub>2</sub> 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, and making chemical mass-balance calculations to evaluate the amount of soluble material removed in surface water and groundwater. 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, p. 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-size 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 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

**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	1. Gypsum, halite
Olivine		2. Calcite, dolomite, apatite
	Ca plagioclase	3. Olivine, amphiboles, pyroxenes
Pyroxene		4. Biotite
	Ca-Na plagioclase	5. Na plagioclase, Ca plagioclase, K-feldspar, volcanic glass
Amphibole	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).

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 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,

**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	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 ( $Fe_2OOH$ ); manganese oxides ( $MnO_2$ )	Minor constituent in siliciclastic rocks
	Soluble constituents	Silicic acid; $SO_4^{2-}$	Chert, evaporites, etc.

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 are removed from the weathering site in surface water or soil groundwater more or less