CALIFORNIA'S AMAZING GEOLOGY

SECOND EDITION

DONALD R. PROTHERO

California's Amazing Geology

California has some of the most distinctive and unique geology in the United States. It is the only state with all three types of plate boundaries, an extraordinary history of earthquakes and volcanoes, and many rocks and minerals found nowhere else. The Golden State includes both the highest and lowest points in the continental US and practically every conceivable geological feature known. This book discusses not only the important geologic features of each region in California but also the complex geologic four-dimensional puzzle of how California was assembled, beginning over two billion years ago. The author provides an up-to-date and authoritative review of the geology and geomorphology of each geologic province, as well as recent revelations of the tectonic history of California's past. There are separate chapters on some of California's distinctive geologic resources, including gold, oil, water, coastlines, and fossils. An introductory section describes basic rock and mineral types and fundamental aspects of plate tectonics, so that students and other readers can make sense of the bizarre, wild, and crazy jigsaw puzzle that is California's geological history.

In this second edition, the book has an entirely new final section, "California's Environmental Hazards and Challenges," with new chapters on California's landslides, air and water pollution, renewable energy, and the future of climate change in California.

KEY FEATURES

- Thoroughly updates the market-leading textbook on California's geology
- Is written by an author with 30 years of teaching geology and leading field trips in California
- Introduces California's unique geological history
- Covers fundamentals of geology
- Characterizes specific geographical regions of California
- Describes major geological resources of California
- Summarizes the paleontology of California
- Reviews the likely impact of climate change on California's environment

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Preface

This book was inspired when I began to teach California geology at the college level and found there were no books suitable for such a course. All the books on the market either were grossly out of date or didn't discuss a modern understanding of California tectonics. California has one of the most amazing and complex histories of any state in the United States, and so the simplistic approaches of the books currently on the market do not do it justice. In this book, I try to give a taste of this incredible jigsaw puzzle of California geology without going into too much depth or requiring too much specialized geologic training. In particular, I am familiar with the backgrounds of most college students taking a course like this at a community college or many of the four-year universities in the state, and so I do not assume that they have taken a previous course in introductory physical geology. For that reason, the first five chapters of the book are a quick review of the basic concepts needed to understand the fundamentals of geology assumed in the book. Other concepts, such as glacial geomorphology, coastal geology, and volcanology, are introduced in later chapters as appropriate.

An updated book on California's fascinating geology should also interest the general reader who has no previous background on the state. Again, the introductory chapters are written to give any reader a grasp of fundamental geologic concepts so that the rest of the story makes sense.

Fate has placed me in an unusually good position to write this book. I have taught college-level geology for almost 45 years, for 27 years at Occidental College in Los Angeles, but also at the California Institute of Technology in Pasadena, the California State Polytechnic University in Pomona, and community colleges such as Pierce College in Woodland Hills, Glendale College in Glendale, and Mt. San Antonio College in Walnut. Thus, I have a wealth of experience in teaching this material at the college level and leading hundreds of field trips to localities all over California. These include not only trips I led as a college instructor but also field trips I have led for the Pacific Section SEPM (Society for Sedimentary Geology), for which I have served as president and vice president, and for groups such as the Skeptic Society. In addition, I'm a second-generation Californian, hooked on dinosaurs at age four, and I never outgrew my love of paleontology. My first field trip experiences were as a Cub Scout collecting fossils in Old Topanga Canyon, and I've had many such experiences since then. I received my undergraduate education from the University of California, Riverside, where a very active department ran field trips out into the Mojave Desert and Transverse Ranges dozens of times each year. Finally, for many years, I have collected fossils and published research and done fieldwork in the Cenozoic strata of the West Coast (Prothero 2001), so I have seen more outcrops in California than most geologists alive today. My roots in the state go very deep: my great-great-grandfather William E. Prothero was in the California Gold Rush but returned to his family farm in Illinois after he did not strike it rich (as did most of the Forty-Niners) and later served during the Civil War in Sherman's Illinois regiments as they marched to the sea.

I hope this book helps the student and interested amateur better understand the incredible geological jigsaw puzzle that is the Golden State.

RESOURCE

Prothero, D.R. (ed.). 2001. *Magnetic Stratigraphy of the Pacific Coast Cenozoic*. Pacific Section SEPM Special Publication 91: 394. Pacific Section SEPM, Fullerton, CA.

Acknowledgments or Credits List

About the Author

Donald R. Prothero has taught college geology and paleontology for 45 years, at Caltech, Columbia, Cal Poly Pomona, and Occidental, Knox, Vassar, Glendale, Mt. San Antonio, and Pierce Colleges. He earned his BA in geology and biology (highest honors, Phi Beta Kappa, College Award) from University of California, Riverside, in 1976, and his MA (1978), MPhil (1979), and PhD (1982) in geological sciences from Columbia University. He is the author of over 50 books (including eight leading geology textbooks and several trade books) and over 400 scientific papers, mostly on the evolution of fossil mammals (especially rhinos, camels, and pronghorns) and on using the Earth's magnetic field changes to date fossil-bearing strata. He has been on the editorial boards of journals such as *Geology*, *Paleobiology*, *Journal of Paleontology*, and *Skeptic* magazine. He is a fellow of the Linnean Society of London, the Paleontological Society, and the Geological Society of America and also received fellowships from the Guggenheim Foundation and National Science Foundation. He served as the president of Pacific Section SEPM (Society for Sedimentary Geology) in 2012 and served for five years as the program chair of the Society of Vertebrate Paleontology. In 1991, he received the Charles Schuchert Award for outstanding paleontologist under the age of 40. In 2013, he received the James Shea Award of the National Association of Geology Teachers for outstanding writing and editing in the geosciences. In 2015, he received the Joseph T. Gregory award for service to vertebrate paleontology. In 2016, he was named "Friend of Darwin" by the National Center for Science Education. In 2023, he was named "Distinguished Speaker" by the Palaeontological Association. He has been featured on numerous TV documentaries, including *Paleoworld*, *Walking with Prehistoric Beasts*, *Prehistoric Monsters Revealed*, *Monsterquest*, *Prehistoric Predators: Entelodon and Hyaenodon*, *Conspiracy Road Trip: Creationism*, as well as *Jeopardy!* and *Win Ben Stein's Money*.

1 The Golden State

Growing up in northern California has had a big influence on my love and respect for the outdoors. When I lived in Oakland, we would think nothing of driving to Half Moon Bay and Santa Cruz one day and then driving to the foothills of the Sierras the next day.

—Tom Hanks

Of all the 50 states of the United States, California has some of the most amazing geology and natural history. It has rocks dating back 1.8 billion years ago (Ga) as well as volcanoes that erupted just a century ago. It has a wide spectrum of minerals and rocks, from ancient volcanoes to young ones, granitic rocks that formed the mighty Sierra Nevada range, and peculiar rocks in the Coast Ranges found nowhere else in the United States. It has the highest mountains in the lower 48 states (Mt. Whitney, 4,421 m, or 14,495 ft, above sea level) and the lowest point in the Americas (Badwater in Death Valley, 86 m, or 283 ft, below sea level). Both are within a 135 mi running distance for ultramarathoners in the Badwater Marathon. Death Valley is one of the hottest and driest places on Earth, reaching 60°C (134°F) and receiving less than 2.5 cm (1 in) of average annual rainfall. The High Sierras are one of the snowiest places in the United States, with recorded snowfalls on Tamarack Mountain reaching 23 m (76 ft) in one year, 10 m (33 ft) in a single month, and 11 m (37 ft) on the ground at one time. California has the highest waterfall in North America (Yosemite Falls, 869 m, or 2,850 ft, of sheer drop). After Mt. Rainier, Mt. Shasta is the second tallest active volcano in the United States, reaching 4,317 m (14,162 ft). Until the eruption of Mt. Saint Helens in 1980, Mt. Lassen was the most recent active volcano in the lower 48 states, last erupting in 1914–1921.

Many of California's other geological features are extraordinary as well. It experienced the largest earthquakes in the United States outside of Alaska, including the 1906 San Francisco quake (magnitude 8.2), the 1872 Lone Pine quake (magnitude 8.25), and the 1857 Ft. Tejon quake (magnitude 8.2), as well as many smaller quakes, such as the 1994 Northridge quake (magnitude 6.7). One of the biggest explosive volcanic eruptions in the history of the planet occurred about 750,000 years ago, when Long Valley Caldera (near Mammoth Mountain ski resort and Bishop in the Owens Valley) blew its top and scattered 521 km³ (125 mi³) of ash across the entire western United States, reaching eastern Nebraska and Kansas. California is also one of the biggest producers of oil and gas in the United States, as well as yielding major deposits of gold, mercury, chromite, rare earth elements, and many other important natural resources.

Its natural history is also remarkable. It has the largest living things on Earth (the giant sequoias), the tallest on Earth (the coast redwoods), and some of the oldest living things on Earth (the bristlecone pines in the White Mountains are up to 5,000 years old). It has ecological regions ranging from coastal rain forests to dry deserts to snow-capped mountains to some of the most picturesque coastlines on Earth. It is home to some of the largest mammals, including large elk, cougar, and deer. There are immense elephant seals on its beaches, as well as tiny rodents and endemic species of foxes on its offshore islands. The grizzly bear once roamed widely over the whole state, which is why the state has the "Bear Flag" and the bear is the mascot of some of the University of California campuses. Due to the wide range of habitats and huge area, California is host to one of the widest diversities of birds, reptiles, amphibians, and fish in North America, as well as extremely diverse plant habitats, from misty redwood forests to cactus-filled desert.

California can claim many other distinctions as well, which make it one of the most important states. By population, California is the largest state in the United States (38 million people in the 2010 census), and it is the third largest state in area. In fact, if California were an independent nation, it would be the 35th largest nation in the world by population. This status as the largest state by population means California has major political clout: the largest delegation in Congress (53 representatives) and the most electoral votes in presidential elections (55). By itself, California provides about 20% of the electoral votes needed to win the presidency.

Even individual counties in California are larger than many states. San Bernardino County is the largest in the United States by area (52,070 km², or 20,105 mi²). It is bigger than nine states (Maryland, Hawaii, Massachusetts, Vermont, New Hampshire, New Jersey, Connecticut, Delaware, and Rhode Island), so if it were a state, it would be the 43rd largest in area. Los Angeles County has the largest population of any US county (almost 10 million people), more people than 42 of the 50 states. If Los Angeles County were to become a state, it would be the ninth largest in the United States by population.

California is home to 3 of the 10 largest cities in the United States, more than any other state except Texas, which also has 3. These include Los Angeles (number 2) and, surprisingly, San Diego (number 8) and San Jose (number 10). Although people think of San Francisco as huge, it is only number 13 on the list. Fresno is the 34th largest US city, Sacramento the 35th, Long Beach the 36th, and Oakland the 47th. That makes 8 of the top 50 largest cities, more than any other state. The Los Angeles Metropolitan Area is the second largest in the United States after New

York City, with 18 million people, larger than all but a few states. The San Francisco Bay Metropolitan Area ranks the 5th largest with 8 million, and the San Diego area is the 18th largest with 3 million.

This huge population and size are matched by incredible economic power, from the tech wizards of Silicon Valley to the center of the entertainment industry in Hollywood and environs (movies, TV, and music especially) to the incredible wealth in oil and agricultural products. California is the largest agricultural state in the union, producing 99% of all the almonds, walnuts, and pistachios; 95% of the broccoli and strawberries; 90% of the grapes and tomatoes; and 74% of all the lettuce consumed in the United States. California's gross domestic product (more than $14 trillion) is larger than that of all but eight countries in the world, so if California were a nation, it would be the world's ninth largest economy. When the governor of California speaks at international meetings of world leaders, he has more clout than any American official except the president or vice president. That influence has been translated into California leading the United States in clean energy policies, laws against air pollution, and many other progressive stances, such as political reform. California was the first state in the nation (and still one of the few) to institute measures such as initiatives (where voters can put measures on the ballot without their politicians), referendums, and recalls.

1.1 GEOGRAPHIC AND GEOLOGIC PROVINCES

California is particularly unusual and complicated in its geography and geology, especially compared to some of the flat states in the Midwest. Many of these states (such as Michigan or Illinois or Kansas) have almost no mountains or even tall hills, and the geology underlying them is fairly simple, with broad shallow dish-shaped basins and nearly flat-lying beds across most of the state. By contrast, California is the only state that has all three types of plate tectonic boundaries (subduction zone in the north, transform boundary along the San Andreas fault zone, and spreading ridge in the Salton Trough and Gulf of California) within it. It is also the only US state that straddles two tectonic plates.

Consequently, California has remarkably distinct geographic regions (Figure 1.1a), each of which has an almost entirely different geology from the other regions (Figure 1.1b). Each region has its own complicated geologic history, which we will look at one region at a time to better understand their local stories before we try to assemble the complete jigsaw puzzle of California in Chapter 15. We start with the regions with the simplest geology, such as the Cascade Range and Modoc Plateau, and then move to more complex regions, such as the Basin and Range and Sierras, before finishing with the wild and crazy geology of the Coast Ranges and Transverse Ranges.

The major geographic provinces of California are shown in Figure 1.1. We discuss each of them in the following chapters.

(a)
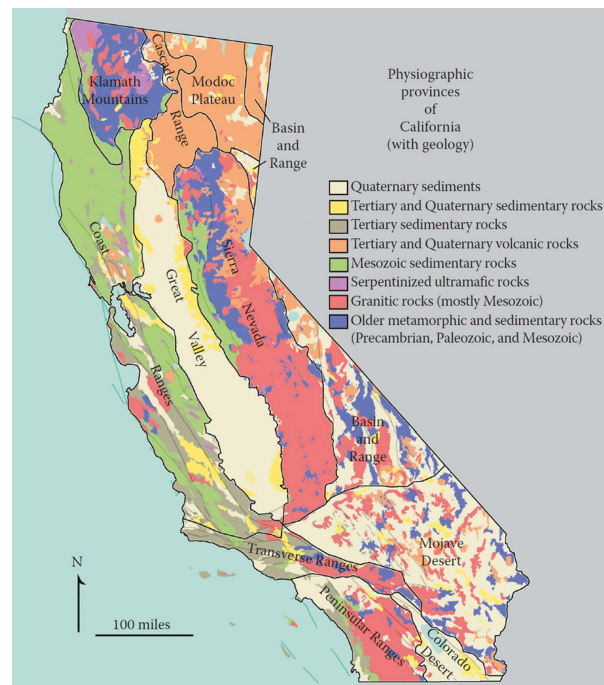
(b)

FIGURE 1.1 (a) Digital elevation map of California; (b) Geologic map and geographic provinces of California showing the geologic complexity of the state.

Source: Courtesy of California Division of Mines and Geology.

2 Building Blocks
Minerals and Rocks

The rock I'd seen in my life looked dull because in all ignorance I'd never thought to knock it open. People have cracked ordinary pegmatite—big, coarse granite—and laid bare clusters of red garnets, or topaz crystals, chrysoberyl, spodumene, emerald. They held in their hands crystals that had hung in a hole in the dark for a billion years unseen. I was all for it. I would lay about me right and left with a hammer, and bash the landscape to bits. I would crack the earth's crust like a piñata and spread to the light the vivid prizes in chunks within. Rock collecting was opening the mountains. It was like diving through my own interior blank blackness to remember the startling pieces of a dream: there was a blue lake, a witch, a lighthouse, a yellow path. It was like poking about in a grimy alley and finding an old, old coin. Nothing was as it seemed. The earth was like a shut eye. Mother's not dead, dear—she's only sleeping. Pry open the thin lid and find a crystalline intelligence inside, a rayed and sidereal beauty. Crystals grew inside rock like arithmetical flowers. They lengthened and spread, adding plane to plane in awed and perfect obedience to an absolute geometry that even the stones—maybe only the stones—understood.

—Annie Dillard
An American Childhood

Before we can discuss the details of California geology, we must briefly cover the fundamentals. For those who have already taken an introductory course on geology, this chapter may mostly be review or can be skipped altogether. However, this book is often used for college courses that assume no previous exposure to geology. In addition, many nongeologist readers of this book will be baffled if they encounter the names of important rocks and minerals without the proper explanation, so these are introduced here.

However, the material in this chapter is important to understand even if you have some geology background. California has some really unusual minerals and rocks, such as its state rock, serpentine, or some of its other peculiar rocks, such as ribbon cherts, blueschist, ophiolites, weird evaporite minerals, diatomites, and many others that are rare or unknown in most other states.

In the next few chapters, we also cover the basics of tectonics and structural geology, seismology, and other principles of geology that are usually covered in an introductory college course. When we discuss the geologic provinces, everyone should have at least some background to follow the discussion.

2.1 ATOMS AND ELEMENTS

First, let's quickly review the most basic principles of physics and chemistry. All matter is made of **atoms**, which are the smaller particle of matter that has the properties of a given **element**. For example, atoms of the element gold have its characteristic properties (such as its high density), but if you break a gold atom into its subatomic particles, they no longer have those same properties. The three main subatomic particles are the **proton** (which has a +1 charge and a mass of 1) and **neutron** (which has no charge and a mass of 1), both of which are found in the nucleus of the atom. The nucleus is surrounded by charged clouds of energy known as **electrons**, which have no mass but a −1 charge.

The proton is especially important because the number of protons in the nucleus is the **atomic number**, which determines which element you have. Any nucleus with one proton will be a hydrogen atom, a nucleus with two protons is part of a helium atom, and so on. If you change the number of protons in the nucleus, it becomes a different element. Because each proton carries a +1 charge, there should be an equal number of −1 electrons to balance the charge and make the atom electrically neutral. If the number of electrons does not match the number of protons, then the atom is a charged **ion**. If it has lost one or more electrons, then there is a net positive charge and it is a **cation**. If it has gained one or more electrons, then it has a net negative charge and it becomes an **anion**.

The number of neutrons in the nucleus has no effect on the charge of the atom (since neutrons are uncharged), but each neutron adds a mass of +1 to the nucleus, so they affect the **atomic weight**. For a given element (and thus for a fixed number of protons), there can be different atomic weights, depending on how many additional neutrons are in the nucleus. These different atoms with the same number of protons but different atomic weights are known as **isotopes**. For example, all isotopes of hydrogen have only one proton (they are no longer hydrogen if they don't), but hydrogen has several isotopes. Hydrogen with a mass of 1 (shown as ^1H, with the atomic weight in the left superscript) has only a single proton and no neutrons. This form of hydrogen (sometimes called hydrogen 1 or protium) is by far the most common in the universe and makes up 99.98% of the known hydrogen. However, about 0.01% of the hydrogen in the universe has not only a proton but also a neutron, giving an atomic weight of 2. It is known as hydrogen-2, or **deuterium**, and shown by the symbol ^2H. A third, very rare form of hydrogen is hydrogen-3, or **tritium**, with one proton and two neutrons. It is shown by the symbol ^3H. It is radioactive and produced only by nuclear reactions.

Nearly all the elements in nature have several different isotopes, and they are often very useful in geochemistry and in many other fields of science. For example, carbon has 15 known isotopes, only 2 of which are not radioactive but stable in geological settings. Normal carbon is carbon-12, or ^{12}C (six protons and six neutrons), and it makes up 99% of the carbon in your body and in almost anything carrying carbon. But carbon-13 (^{13}C, with six protons and seven neutrons) is also stable. Even though it makes up less than 1% of the carbon on the planet, its occurrence is a powerful tool in geology, geochemistry, and oceanography. Carbon-14 (^{14}C, with six protons and eight neutrons) is produced by bombardment of ^{14}N in the atmosphere that changes it into carbon-14. It is radioactive and unstable, decaying back to nitrogen at a known rate. This useful property allows it to be the basis of **carbon-14 dating** (or radiocarbon dating), the best tool geologists and archeologists have for measuring the age of human artifacts and anything less than about 60,000 years old (see Chapter 3).

There are more than a hundred elements in the periodic table, but most of them are extremely rare in geologic settings. Many of them exist only for milliseconds in a high-powered physics lab. In geology, the chemistry is even simpler. It turns out that there are only eight common elements (Table 2.1) in the Earth's crust that are worth remembering, because they make up more than 99% of all rocks on Earth.

The first surprise is that so many of most rocks in the Earth's crust are made up of oxygen (Table 2.1). An average rock is 46% oxygen by weight percent but 94% oxygen by volume. This is because not only is oxygen a light element, but also its ion has a large radius. You pick up a heavy rock, but in reality, it's made mostly of oxygen. Why is oxygen by far the most abundant? If you glance at Table 2.1,

you'll note that it is the only common anion on the table, and *something* has to balance all the positive charges of the remaining seven cations. Oxygen is very abundant in air and water, so it combines with almost any cation. Next in abundance is silicon, which (like carbon) is an element that readily bonds into long chains and forms complex three-dimensional structures. This is important in making the common rock-forming minerals, which are mostly combinations of silicon and oxygen, or **silicates**. In distant third place is aluminum, which is also an element that readily bonds into complex three-dimensional arrays with silicon, so many silicate minerals are rich in aluminum. These three elements just happened to be the common ones when the solar system and Earth formed, and they are found in most of the common rock-forming minerals as a result.

The next five elements in Table 2.1 are much less abundant than the "big three," with most of them making up only 5% or less in percentage of crustal rocks. All five are metallic cations, which bond with oxygen or complex silicate structures to make the huge variety of minerals. Note that they have different charges: sodium and potassium are both cations with a +1 charge. They can sometimes replace each other in a mineral, since they have the same charge. Likewise, calcium, magnesium, and ferrous iron (Fe^{+2}) are all common +2 cations and also can switch with each other in many minerals.

Even more surprising is what elements are *not* in the "big eight." Hydrogen and helium are the most common elements in much of the solar system, especially in the sun and outer planets (Jupiter, Saturn, Uranus, and Neptune). But they are rare on Earth, except where hydrogen is bonded with oxygen to make water. Why? When the Earth formed, it was not massive enough to have enough gravity to hold on to these elements and they floated off into space. Large planets like Jupiter and Saturn have much more gravity, and they held in their hydrogen and helium. Phosphorus and sulfur are also relatively rare and only concentrated in special settings. Perhaps most surprising of all is how rare carbon is in crustal rocks. After all, carbon is the building block of all life. There are a few minerals and rocks that have carbon in them, but they are rare compared to silicates.

TABLE 2.1

Average Chemical Composition of Earth's Crust, Hydrosphere, and Troposphere

Element (Symbol)	Crust		Hydrosphere	Troposphere
Percent by Mass	Percent by Volume	Percent by Volume	Percent by Volume	
Oxygen (O_2)	46.40	94.04	33.0	21.0
Silicon (Si)	28.15	0.88	—	—
Aluminum (Al)	8.23	0.48	—	—
Iron (Fe)	5.63	0.49	—	—
Calcium (Ca)	4.15	1.18	—	—
Sodium (Na)	2.36	1.11	—	—
Magnesium (Mg)	2.33	0.33	—	—
Potassium (K)	2.09	1.42	—	—
Nitrogen (N_2)	—	—	—	78.0
Hydrogen (H_2)	—	—	66.0	—
Other	0.66	0.07	1.0	1.0

2.2 MINERALS AND ROCKS

Now that we have the foundation of the elements available in the Earth's crust, let's see how they are combined into more complex molecules (groupings of atoms bonded together) known as **minerals**. The word *mineral* has all sorts of casual and inconsistent meanings in popular culture, but to geologists and chemists, the word *mineral* has a very strict and clear definition. A mineral is a

- Naturally occurring
- Inorganic
- Crystalline solid
- With a definite chemical composition and
- Characteristic physical properties

Let's discuss each of these components.

Naturally occurring. There are lots of complex compounds in the world, but if they are not produced naturally, they are not minerals. Thus, a synthetic diamond produced in a lab has all the properties of a diamond mined from the Earth, but it's not a mineral. Ice formed as ice crystals or snowflakes is a mineral, but not the ice in your ice cubes. Most of the stuff sold in a health food store that is called "mineral" was produced synthetically and therefore is not a mineral as a scientist uses the word.

Inorganic. Organic chemicals are built of the element carbon, so minerals are not made mostly of carbon. However, there are a handful of important minerals with carbon; we'll use *organic* in this context to mean compounds of carbon, oxygen, and hydrogen. Thus, sugar forms beautiful crystals, but it is organic and therefore not a mineral. Many of the "minerals" sold in a health food store are organic as well and thus not minerals.

Crystalline solid. Like the word *mineral*, the word *crystal* has a different meaning to a scientist than it does in popular culture. Typically, people use the word *crystal* to describe anything that sparkles. In a scientific definition, a *crystal* must have a *regular three-dimensional arrangement of atoms* in its internal structure, which repeats over and over. This three-dimensional array is called a **lattice**. It is analogous to the regular repeated pattern in wallpaper. For example, the atoms of the salt (sodium chloride, or NaCl) crystal (Figure 2.1a) are arranged in a cubic pattern, with each atom of sodium or chlorine forming 90° angles with the others. The same lattice is found in the mineral galena (Figure 2.1b), which is made of lead and sulfur in equal amounts (lead sulfide, or PbS). All minerals have a regular three-dimensional lattice of some kind, often very complex and with many other angles between atoms besides the 90° one seen in the simple cubic lattice.

Some things in nature may have a three-dimensional arrangement of atoms but they are not regular and repeating. Take, for example, volcanic glass or obsidian (Figure 2.2). At the molecular level, the atoms are not in any kind of repeated pattern but in a random tangle of long chains, like a bowl of spaghetti. Technically speaking, a glass isn't really a solid at all but a supercooled liquid. Over very long spans of time, the glass will slowly flow and change shape. This is clear if you ever see a piece of window glass in a very old house. If it has been in its window frame for about a century, the glass will be thicker at the bottom because it has slowly flowed downhill over the course of decades. Thus, a *glass is by definition not a crystal*. A popular item at many gift shops is "cut glass crystal" drink-
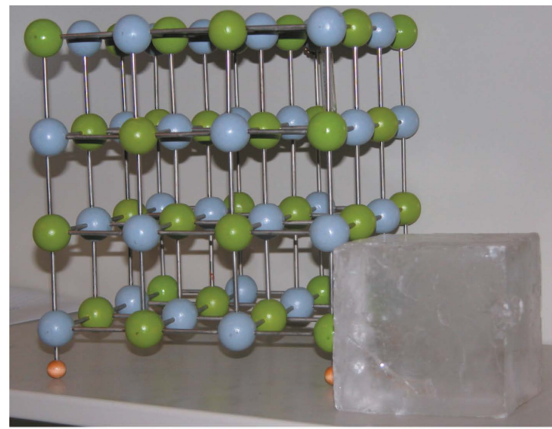
(a)

(b)

FIGURE 2.1   (a) Giant cubic crystal of salt (the mineral halite) next to a ball-and-stick model of the atomic structure of the lattice of a cubic mineral like halite. The blue balls could represent sodium atoms, while the green balls are chlorine atoms. (b) Crystals of the cubic mineral galena, or lead sulfide (PbS), showing its 90° cleavages and cubic habit.

Source: (a) Photo by the author. (b) Courtesy of Wikimedia Commons.

ing goblets and chandeliers, but this is not the definition of *crystal* that scientists use.

Definite chemical composition. Most minerals have a simple chemical formula, like most other compounds. There is a bit of substitution allowed if you replace one ion with another of a similar charge. For example, the mineral calcite (calcium carbonate, or $CaCO_3$) can have a certain percentage of magnesium replacing calcium sites in its lattice and still be calcite. However, if it gets to be 50:50 Ca/Mg, then it's no longer calcite but a different mineral, dolomite.

Characteristic physical properties. Most of the features of the minerals we have discussed occur at the atomic level. But to identify the mineral, you need to know what physical properties are typical of a hand sample of the mineral. These include its
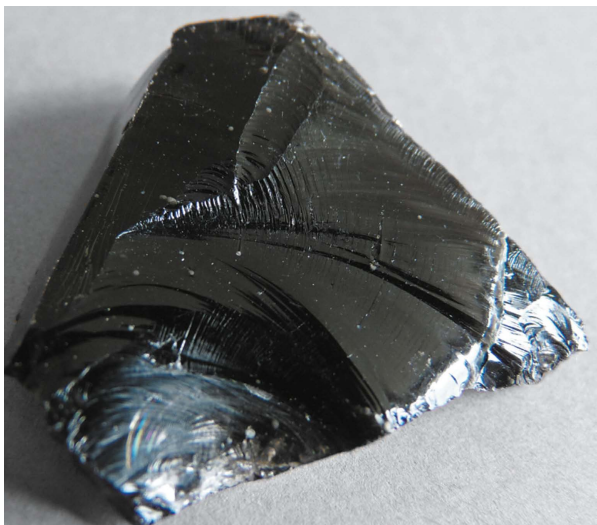
FIGURE 2.2 Chunk of obsidian or volcanic glass, with the characteristic scallop-shaped fracture pattern ("conchoidal fracture") found in glass, and also minerals without cleavage, such as quartz. This natural fracture pattern makes sharp edges, which is why obsidian and quartz are often used for stone arrowheads and spearheads.

Source: Courtesy of Wikimedia Commons.

TABLE 2.2

Mineral Classes Classified Based on Their Dominant Anion

Mineral Classes

Minerals are classified by their dominant anion.

➤ *Silicates*	SiO_2^{4-}	*Most rock-forming minerals*
Oxides	O^{2-}	Magnetite, hematite
Sulfides	S^-	Pyrite, galena
Sulfates	SiO_2^{4-}	Gypsum
Halides	Cl^- or F^-	Fluorite, halite
Carbonates	CO_3^{2-}	Calcite, dolomite
Native elements	Cu, Au, C	Copper, gold, graphite

pg12_1.jpg

Note: Word endings are very important in chemistry! *Silicon* is the element on the periodic table. A *silicate* is a mineral made of silicon plus any number of atoms of oxygen. *Silica* is the compound silicon dioxide, or silicon plus two atoms of oxygen (SiO_2). But *silicone* is a synthetic compound made of silicon in a lab, used for lubricants, breast implants, and other purposes. Don't mix them up!

color, its hardness (from soft minerals like talc and gypsum to the hardest mineral, diamond), whether it fractures with an irregular surface or cleaves into many parallel planes, and less commonly used properties, like density (lead sulfide or galena, for example, is unusually dense because it contains lead), reaction to acid (the mineral calcite fizzes in dilute hydrochloric acid), and magnetism (the mineral magnetite is naturally magnetic).

Many of these properties of minerals can be understood by knowing the crystal lattice. For example, the cubic lattice of minerals like salt (NaCl) or galena (PbS) is demonstrated at the hand sample level, since any time you hit and break a piece of these minerals, they will naturally cleave to form cubic faces with 90° angles (Figure 2.1).

The atomic-level properties and crystal lattice can make a huge difference in the behavior of a mineral at the macroscopic level. Let's take as an example the two common minerals formed of pure carbon: diamond and graphite. One is the hardest substance in nature, and the other is one of the softest, yet they are chemically identical. Why are they so different? Diamond has a crystal lattice with all the atoms of carbon tightly bonded together and very short, strong chemical bonds. This structure will survive huge amounts of pressure, and an expert cutting a diamond has to know exactly how to cleave a large stone into several smaller ones. If he misses, the diamond is ruined. Graphite (the "lead" in a pencil), on the other hand, has all its carbon atoms arranged in sheets, with very long, very weak molecular bonds between the sheets. Pushing the graphite tip of a pencil across paper is enough to break those weak bonds, leaving tiny flakes of graphite behind on the paper as pencil markings.

Another example is calcium carbonate, or $CaCO_3$, in two different mineral lattices: calcite (the common mineral in limestones and marbles) and aragonite (also known as mother-of-pearl). They are the same chemistry, but their lattices are very different and lead to very different properties. The most obvious of these is that aragonite is much more soluble than calcite, so in weakly acidic conditions, aragonite will dissolve, but calcite won't. This is why if you own pearls, it is important to wash your acidic sweat off them once you put them away after wearing them. For some minerals, knowing the chemical composition is not enough; the crystal lattice makes a huge difference in the properties of the mineral as well.

There are literally thousands of different kinds of minerals, but most can be organized into just a few classes based on chemical composition (Table 2.2). There is no room to discuss them all in a brief introduction like this, but their major characteristics are detailed in Table 2.2. As we saw in Table 2.1, however, most of the Earth's crust is made of silicon and oxygen with minor aluminum, so it's no surprise that the most important rock-forming minerals are the **silicates**, made of silicon plus oxygen.

The major classes of silicate minerals are given in Table 2.2. Each has the same basic building block: the silicon–oxygen tetrahedron (Figure 2.3), or SiO_4 unit. Each class of silicate minerals uses these fundamental building blocks over and over and links them together in more and more complex structures.

The simplest silicate mineral structures are built of **isolated tetrahedra**, where the silicon–oxygen (SiO_4) building
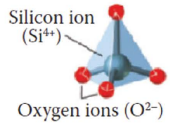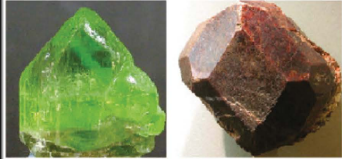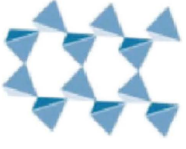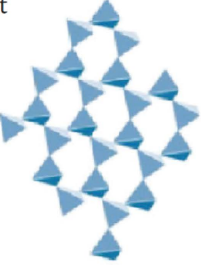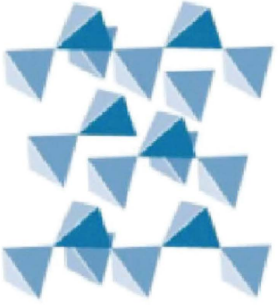
SILICATE STRUCTURE	MINERAL/FORMULA	EXAMPLE
Single tetrahedral Silicon ion (Si^{4+}) Oxygen ions (O^{2-})	Olivine: Mg_2SiO_4 Garnet: $Fe_3Al_2(SiO_4)_3$	
Single chain	Pyroxenes e.g., augite: $(Mg,Fe)SiO_3$	
Double chain	Amphiboles e.g., hornblende: $Ca_2(Mg,Fe,Al)_3Si_8O_{22}(OH)_2$	
Sheet	Micas and clays e.g., kaolin: $Al_2Si_2O_5(OH)_4$	
Framework	Quartz: SiO_2	

	Feldspars
	  Potassium feldspar:
	  $KAlSi_3O_8$	

	Plagioclase feldspar: $(Ca,Na)AlSi_2O_8$	

FIGURE 2.3   Classes of silicate minerals and their structures.

Source: Redrawn from several sources by E. T. Prothero; courtesy of Wikimedia Commons.

blocks do not bond directly to each other but are held in place by the electrostatic charges of the cations (especially Mg and Fe) between them in the lattice. Common single tetrahedral minerals are the green mineral **olivine**, which is (Mg, Fe)SiO_4, and **garnet**, which is built of SiO_4 tetrahedra with a variety of cations to make the six major types of garnets (from red-brown almandine to green grossular, and many others).

The next most complex arrangement is found in the class of minerals known as **pyroxenes** (PEER-ox-eens), which have linked the SiO_4 tetrahedra into long **single chains**. Among the common pyroxenes are the dull greenish-black mineral **augite** ($MgSiO_3$) as well as **jadeite**, one of the two minerals that can make the beautiful gemstone known as jade. Because of their crystal lattice made of single chains

stacked together, pyroxenes have a 90° cleavage in hand samples, the most reliable physical property to recognize them.

The next step in silicate complexity is to link two single chains side by side to make a **double-chain** silicate structure, similar to the way the rails of a train track are linked together by wooden ties. Double-chain silicates are known as **amphiboles**, and their lattice structure gives the hand samples cleavages that are roughly either 60° or 120° (technically, 57° and 124°), their most diagnostic property. The most common amphibole is the shiny jet-black prismatic mineral known as **hornblende**, which is common in many igneous and metamorphic rocks discussed later. There are others, such as the greenish amphiboles tremolite and actinolite and the blue amphibole glaucophane, which we will discuss later.

Once you have a double-chain structure, the next, more complex arrangement is to bond the double chains together side by side to form **sheet silicates**. Sheet silicates are made of two layers of silicon–oxygen tetrahedral (*t*) sandwiching a layer of aluminum-oxygen octahedra (*o*). This *t-o-t* structure is like an Oreo cookie, with the *t* layers represented by the chocolate cookie layers, and the *o* layer representing the aluminum–oxygen creamy filling. Most sheet silicates are built of stacks of *t-o-t* structures with other materials (different kinds of cations and water molecules) trapped between the *t-o-t* layers. The most familiar sheet silicates are a class of minerals known as **micas**, which are distinctive in that they cleave into large flat thin sheets. The silver-white mica is known as **muscovite**, and before it was possible to make glass windows, large sheets of muscovite were used as windows and curtains (such as the "isinglass curtains" in the song "The Surrey with the Fringe on Top" from the Rogers and Hammerstein musical *Oklahoma*). There is also a common black mica known as **biotite**, a green mica called **chlorite**, and a lithium-rich lavender mica known as **lepidolite** that we will see in many different types of rocks. In addition to the micas, all the clay minerals that make up the muds of the world are sheet silicates; they are the most common minerals on Earth for this reason.

We have progressed from isolated tetrahedral to single chains to double chains to sheets, each structure being more and more complex and linked together (in chemical terms, **polymerized**). The final step is to link the silicon–oxygen tetrahedral into a complex three-dimensional **framework** structure that is almost impossible to render in a two-dimensional illustration. However, there are ball-and-stick models that capture their shape well, and now there are animations online that give some idea of their geometrically complex structure. Some of the most common and important minerals of all are framework silicates. The most important of these is **quartz**, made of pure silica (SiO_2), one of the most abundant sedimentary minerals on Earth. The other important classes of framework silicates are the aluminum-rich **feldspars**, the most common minerals in igneous rocks. Two types of feldspar are particularly important:

the pink-colored **potassium feldspars** ($KAlSi_3O_8$), which exist in three mineral forms (orthoclase, microcline, and sanidine), and the **plagioclase feldspars**, which transform continuously from pure calcium-rich plagioclase known as anorthite ($CaAl_2Si_2O_8$, typically bluish-gray in color) to intermediate plagioclases with a mixture of calcium and sodium, to pure sodium-rich plagioclase known as albite ($NaAlSi_3O_8$, white in color). Plagioclase crystals typically show lots of fine parallel lines known as **striations** on their cleavage surfaces. Plagioclases are typically the most common minerals in most igneous rocks and are found in all but a few of them.

These are a lot of mineral names to master, especially for beginning students, yet this is the minimum number of minerals needed to understand the rocks that occur on the Earth's surface and its interior. If you have the opportunity to study hand samples of each of these minerals and test their properties and compare them to each other, the names make more sense and become easy to remember with lots of experience and practice. Nearly every geology student masters these minerals after their first few classes in geology, so it's not that hard to do. It just takes practice and study. Once you know your minerals, you can understand the rocks that are made from them.

2.3 IGNEOUS ROCKS

Igneous rocks are formed by **magma** (molten rock) from deep in the Earth that rises up from deep **plutons** (magma chambers) and crystallizes as it cools. The size of the crystals in the rock depends on how fast it cools and solidifies. If the molten rock is spewed out of a volcano, the magma cools quickly and the crystals have little time to form, so they are **microcrystalline** ("aphanitic"), too small to see with the naked eye. They are only visible in thin polished sections of rocks when viewed under a special microscope. If the crystals cool slowly over years to hundreds of years deep in an underground pluton, then they have time to grow larger. Sometimes they are still just barely visible to the naked eye, but still they are **macrocrystalline** ("phaneritic") nonetheless. A few magmas cool extremely slowly over the course of decades or centuries or longer, producing **pegmatites** full of giant crystals.

Some igneous rocks have a composite texture, with macrocrystalline crystals (**phenocrysts**) floating in a rock that is mostly microcrystalline (called the **groundmass**). This hybrid texture is called **porphyritic** (the noun form of the word is **porphyry**) and results from a magma that had a two-stage cooling history: the phenocrysts cooled slowly in a large pluton, and then the semi-crystallized magma was blown out of a volcano, where the rest of the melt cooled quickly to form the groundmass surrounding the phenocrysts. The important thing to remember is that the *crystal size is a result of the time and mode of cooling*, so microcrystalline rocks are volcanic, macrocrystalline rocks cool slowly in plutons (plutonic rocks), and porphyritic rocks have both stages in their history.

The crystal size is one element of a classification of igneous rocks (Figure 2.4). The other axis is based on their chemical composition. Rocks coming from the mantle are relatively rich in magnesium, iron, and calcium and are known as **mafic** rocks for short ("ma" for magnesium, plus the chemical abbreviation "Fe" for iron). This magnesium–iron–calcium chemistry produces minerals rich in these elements, such as olivine and pyroxenes (magnesium- and iron-rich silicates) and calcium plagioclase. If a mafic magma cools quickly, it is the familiar black lava called **basalt** that erupts out of Kilauea on the Big Island of Hawaii and other mantle-sourced volcanoes. If it cools slowly, the same magma that makes black basalt instead produces a macrocrystal-line rock with visible pyroxene and calcium plagioclase known as a **gabbro**.

(a)

(b)

FIGURE 2.4 Classification chart of (a) igneous rocks and (b) Bowen's reaction series.

Source: Redrawn from several sources by E. T. Prothero; photos by the author.

Some magmas are so rich in magnesium and iron that they are known as **ultramafics**. These contain almost nothing but olivine derived directly from the mantle or lower crust. A macrocrystalline rock made of pure olivine is called a **peridotite** and represents a direct sample of the upper mantle. (The gem name for olivine crystals is *peridot*, so a rock made of olivine is a peridotite.) Currently, there are no ultramafic volcanic lavas erupting anywhere on Earth, but back about three billion years ago (Ga), they were common and made an olivine-rich lava called **komatiite**.

The other extreme of magma chemistry are melts that are rich in silicon, aluminum, potassium, and sodium. These rocks are often called **silicic** or **felsic** (an abbreviation based on combining *feldspar* and *silica*). With this kind of magma chemistry, you produce minerals such as quartz (pure silica), sodium feldspars (sodium, aluminum, and silica), potassium feldspars, and micas such as biotite and muscovite (rich in potassium, aluminum, and silica). If the rock contains quartz and two-thirds of its total feldspar is potassium feldspar, then it is a true **granite**. True granites have so much pink or red potassium feldspar in them that they tend to be red as well. There are volcanic eruptions of magmas with the composition of granites. These are known as **rhyolite**, and they are extremely fine-grained and generally pink or red in color due to the rusting of the iron in them.

The rocks in the Sierra Nevada Mountains and other California mountains that laypeople call "granites" do not have enough potassium feldspar in them to be true granites, as a geologist defines the term. Many of them have less than 33% of their total feldspar as potassium feldspar, as well as about 20% quartz. Most geologists call these rocks **granodiorite** rather than granite. Most of the plutonic rocks in California are granodiorites or diorites. However, in this book, we will often use the term *granitics* as a category name to refer to the entire classes of felsic plutonic rocks, including granodiorites, diorites, and many other felsic rocks we will not discuss here. The volcanic equivalent of a granodiorite is called a **dacite**. There are more complicated schemes of subdividing felsic igneous rocks, but we will not discuss them here because that requires much more background in geochemistry and mineralogy than is appropriate for this book.

Rocks that are intermediate in composition between basalt–gabbro and granodiorite–dacite are usually made of hornblende plus an intermediate mixed sodium–calcium plagioclase, but no quartz or potassium feldspar. These rocks are known as **diorite** if they are macrocrystalline and usually have a speckled black-and-white, "salt-and-pepper" appearance. If they cooled quickly, a diorite composition becomes a microcrystalline volcanic rock known as **andesite**. They were named after the Andes Mountains of South America, although it turns out there are few actual andesites in the Andes by the modern definition of that term. Most andesites are light to dark gray in color and porphyritic, with tiny black phenocrysts of hornblende floating in
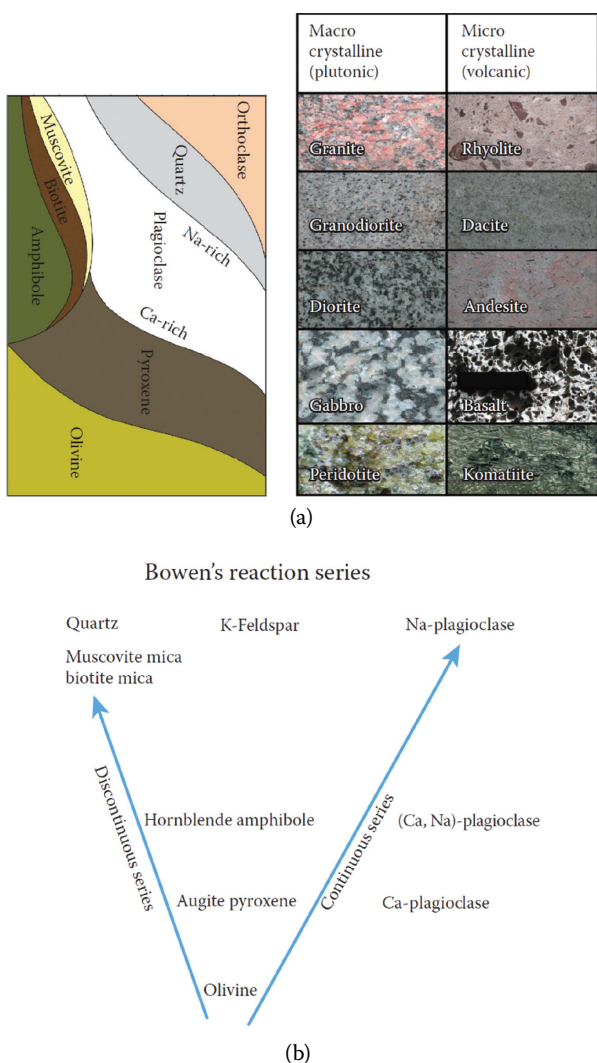
a microcrystalline groundmass of intermediate plagioclase plus hornblende.

This is the simplest possible list of names for ten basic types of igneous rocks (Figure 2.4a). Igneous petrologists recognize hundreds of different types, but that level of detail isn't required here. Now that we have some names and understand some of the basics of magma chemistry, the next question is, *Why are there so many different types of igneous rocks?* If their magmas were all coming straight from the mantle, they would all be peridotite or komatiite in composition. Something happens to that ultramafic magma as it rises from the mantle and through the crust so that it changes chemistry as it rises, losing magnesium, iron, and calcium and gaining silicon, aluminum, potassium, and sodium, in order to make all the different kinds of magmas that produce the hundreds of different kinds of igneous rocks (**magma differentiation**).

This is the central question of all of igneous petrology. It perplexed and puzzled geologists for most of the nineteenth century, when many types of igneous rocks were named and described but no one could explain how they all formed. The breakthrough came in the early twentieth century, when Norman L. Bowen tried to simulate the process of magma differentiation in the laboratory. He would take a chip of peridotite, melt it in a high-temperature furnace, and then drop the molten blob into liquid mercury to instantly quench and chill it. Once the blob had cooled, he would separate out all the material that had crystallized (mostly olivine) and then melt the rest and go through the process again and again. (Although he didn't know it, working with mercury is very dangerous, and eventually Bowen lost his mind due to mercury poisoning, just as the "mad hatters" who used mercury to make felt for hats did in the days of *Alice in Wonderland*.) As Bowen did his experiments, the remaining material became more and more depleted in magnesium, iron, and calcium as he removed the crystals of olivine, pyroxene, and calcium plagioclase that had formed first. As the residue became depleted in those elements, the magma that remained became richer in silicon, aluminum, potassium, and sodium.

Bowen's experiments simulate what happens in many magma chambers. In a process called **fractional crystallization**, the first mineral to cool and crystallize will settle out to the bottom or sides of the magma chamber and remove their magnesium, iron, and calcium from the remaining magma. Over time, this changes the composition of the magma left behind, so eventually it can no longer crystallize olivine but must make pyroxene instead, and its plagioclase goes from calcium-rich to more sodium-rich. After enough olivine and pyroxene have been removed from the melt, it would no longer have enough magnesium or iron left, and then hornblende would be the next mineral formed. In other words, as each fraction crystallizes out and changes the melt left behind, the magma changes chemistry (hence "fractional crystallization").

As an analogy, imagine a room full of students standing up from their chairs. Let's say that the room is 60% men

and 40% women. For this example, the men will represent iron and magnesium atoms, and the women silica atoms. Initially, all the students are standing, as if they were atoms floating around in a melt. Now we'll have two men sit down for every woman who sits down, and do this over and over. Sitting down represents a crystal that sinks out of the melt, removing itself from the chemistry of the magma. After enough time, all the men will be sitting, but quite a few women will still be standing. So the ratio of men to women at the start was about two men for every woman, but at the end, there are zero men standing and only women left (as if they were still floating in the melt). In the same manner, fractional crystallization pulls out the elements from the minerals that crystallize out first (such as iron, magnesium, and calcium) and deposits them on the walls and floor of the magma chamber. This leaves the remaining melt depleted in those elements and enriched in elements that were rare initially.

Bowen plotted this crystallization sequence of minerals, and today it is known as **Bowen's reaction series** (Figure 2.4b). This simple arrangement of minerals exactly parallels the mineral sequence of the igneous rock classification scheme (Figure 2.4a). It is considered one of the greatest discoveries in the science of geology and showed the power of experimental approaches and lab simulation when purely descriptive methods failed.

Bowen's reaction series has two pathways. The **discontinuous series** is the sequence of different classes of dark-colored minerals, from olivine to pyroxene to amphibole to biotite. At the same time, the magma is continuously crystallizing out plagioclase as well, forming a **continuous series**. The first plagioclase to form is calcium-rich, and each later plagioclase is poorer in calcium and richer in sodium. Eventually, the two series converge in felsic magmas, which crystallize out quartz, potassium feldspar, and sodium plagioclase, plus biotite or muscovite, or both—the minerals found in granites and granodiorites.

Bowen's reaction series was based on a lot of brilliant lab experiments derived from understanding the well-documented geochemical patterns of igneous rocks. It has since been confirmed by many field observations. There are many ancient layered magma chambers (now exposed to the surface and eroded) that cooled slowly and built up layers of crystals on their floors as they cooled. The bottom layers are always rich in ultramafic minerals and rocks (mostly peridotites and related rocks) and then gradually become enriched in gabbros as you go up through the layers into magmas cooled later in the process.

There are other proofs as well. When studying a thin section of a gabbro or basalt under the microscope, it is typical to see an early formed olivine crystal that is surrounded by a rim of pyroxene which crystallized around the olivine core. As the magma chemistry changed, it could no longer make olivine but made pyroxene instead. Also, microscopic examination of plagioclase crystals often shows that they have layers around the outside that are sodium-rich, while

the inner core is calcium-rich, also demonstrating that the surrounding magma chemistry changed as the crystal grew.

Fractional crystallization is an important way to change an ultramafic magma direct from the mantle into a mafic or even intermediate magma. Another mechanism is called **partial melting**. Imagine that deep in the Earth's crust, there was a mass of already-cooled gabbro or diorite that was reheated. The first materials to separate away from the original rock would be minerals that melt at the lowest temperatures, such as quartz, potassium feldspar, and sodium plagioclase. If this low-temperature melt were then cooled, it would become a granodiorite or a granite.

Another mechanism for getting different magmas is to melt a different country rock. Thus, if you start with a rock of felsic or intermediate composition that was already in the crust, melting it will only produce a rock that is just as felsic, if not even more so.

Another important consideration is what causes rocks to melt. It turns out that for rocks that are in subduction zones (Chapter 4) plunging deep in the mantle, there is a lot of water entrapped in the ocean floor basalts that once formed on a mid-ocean ridge. Water and other volatiles (gases) dramatically lower the melting temperature of a rock. Many of the andesite–dacite–rhyolite magmas we will see in later chapters are formed because of different amounts of volatiles in the source material that melted.

Yet another likely mechanism for changing a gabbro to a diorite to a granodiorite magma is to have it melt its way up through existing crustal rocks, which are typically much richer in silicon, potassium, aluminum, and sodium. As these wall rocks of the magma chamber were melted and digested into the magma, they could change its chemistry from mafic to felsic. If pieces of wall rock are ripped away and melted into the new magma, it is known as **assimilation**. If the wall rock around the hot magma chamber is partially melted into felsic magma and then mixed with the more mafic magma, it is known as **contamination**. Either way, magmas formed deep in the Earth's crust that must melt their way through more than 100 km (60 mi) of crustal rock above them will end up being very different in chemistry. At one time, assimilation and contamination were thought to be crucial in making magmas more silicic, but now partial melting and fractional crystallization are considered more important.

Finally, there are a few rare instances where a mafic magma chamber from one source melted its way into a felsic magma chamber from a different source (or vice versa), forming a **mixed magma chamber** that is intermediate in composition (Figure 2.5).

This is the most basic (and highly oversimplified) summary of how the chemistry of magmas changes to form the hundreds of different types of igneous rocks known on the planet. Most of these processes happen deep in the crustal magma chambers to form plutonic rocks. Plutons intrude into older rocks ("country rock" or "wall rock") and melt their way to higher levels. They can take many shapes (Figure 2.6). Most plutons intrude to form a body of magma

(a)

(b)

FIGURE 2.5 Evidence of magma mixing from Lassen Volcanic National Monument: (a) pink rhyolite with inclusions of dark andesite; (b) light andesite inclusions in dacite.

Source: Photos by the author.

cutting through older rock called a **dike** (spelled *dyke* in Britain). If a dike intrudes parallel to the bedding of a sedimentary rock, it is a special kind, known as a **sill**. A large chain of plutons that underlie a volcanic mountain range are known as a **batholith**. We will see all these features, as well as the features typical of volcanoes, in later chapters.

2.4   SEDIMENTARY ROCKS

The second major class of rocks is much more familiar to us, since they form at the Earth's surface, not in deep magma chambers or dangerous volcanoes. They are known as **sedimentary rocks**. They are of enormous importance, since nearly all the economic products we obtain from the Earth come from loose sediments or sedimentary rocks. These range from energy sources like oil, gas, coal, and uranium, to the groundwater we rely on, to materials for construction (building stone like sandstone or limestone, concrete made of crushed limestone plus sand and gravel,
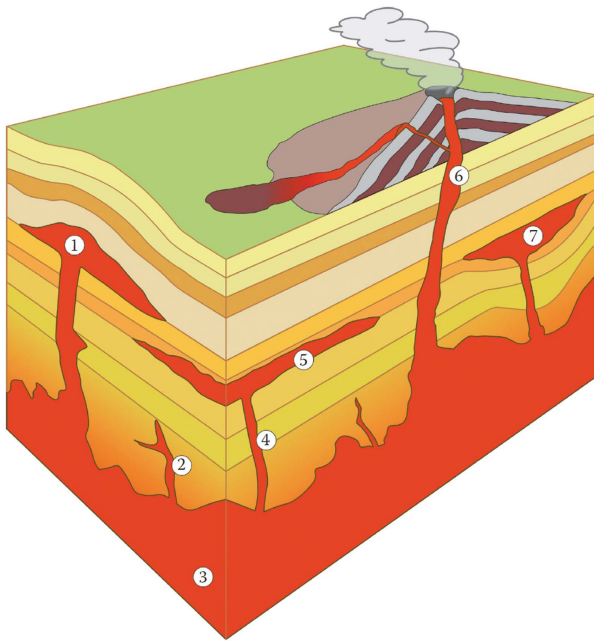
FIGURE 2.6 Terminology of dikes, sills, and other types of intrusions: (3), huge magma chamber known as a batholith; (2), (4), and (6), dikes that either feed an (4) intrusion or a (6) volcano or cool after they reach a (2) dead end; (5), dike intruded parallel to bedding known as a sill; (1), blister- or dome-shaped intrusion known as a laccolith; (7), flat-topped intrusion called a lopolith.

Source: Courtesy of Wikimedia Commons.

gypsum for drywall, quartz sand for glass, etc.), to many of our metallic mineral resources (especially uranium, iron, and steel).

Even without these economic incentives, understanding sedimentary rocks is still supremely important. Sedimentary rocks are the source of nearly all our information about Earth history and ancient environments and the only source of fossils that demonstrate the history of life and help us tell geologic time. Finally, carbon-rich rocks such as coal and limestone are the thermostats that make our planet livable. These crustal reservoirs lock up or release carbon to the atmosphere so Earth is neither a hellish super-greenhouse like Venus nor a frozen ice ball like Mars.

All sedimentary rocks undergo some version of a basic pathway. They always start as *weathered material* of a pre-existing rock that can be igneous, metamorphic, or sedimentary. Once that weathered material is picked up by wind or water and *transported*, it becomes loose sediment. Eventually, the sediment in motion comes to a stop and is *deposited*. But it's still loose sediment until, sometime later in its history (usually after some deep burial), the loose sand grains are cemented together, or the mud particles are compressed, and the loose sediment is *lithified* into a sedimentary rock. Every step of this history can be detected in clues in the rock, and expert sedimentary geologists are like detectives, gleaning clues about the past from a sandstone or limestone that no one else even notices.

There are two versions of this pathway. Most sedimentary rocks are made of broken particles of pre-existing rocks or minerals, known as **clasts** (Greek *klastos*, meaning "broken fragments"). Therefore, **clastic** (or "detrital") sedimentary rocks are made of different-sized pieces of other rocks, from huge boulders down to fine clay. Eventually, these loose grains of gravel, sand, or mud must be lithified into sedimentary rock. The second pathway is much simpler. Instead of fragments of rocks or minerals, the ions from the pre-existing rock weather out and are dissolved in water, where they stay until something causes them to precipitate from the water and crystallize into minerals like halite (forming rock salt), gypsum (hydrous calcium sulfate), and calcite or aragonite (calcium carbonate). Thus, when they crystallize, they are already lithified into solid rock. Since this is a purely chemical process, we call these **chemical** sedimentary rocks.

2.4.1 Clastic Sedimentary Rocks

The most important property of rock fragments is their size, so classifications of clastic sediments are based on size. It's also important because the size of grains is a good indicator of the agents of deposition (wind, water, and glacial ice), and the grain size decreases downstream from the source, so it helps tell us about transport. Rather than the exotic and obscure names of igneous rocks (try to remember what minerals are in a *lherzolite* or a *jacupirangite*!), most sedimentary rock classifications are based on the practical Anglo-Saxon words that English speakers have used all our lives—*sand, gravel*, and *mud*—only those words have a strict definition in classification schemes (Figure 2.7). Any particle larger than 2 mm in diameter is gravel, which can be further broken down into granules (2–4 mm in diameter), pebbles (4–64 mm), cobbles (64–256 mm), and boulders (larger than 256 mm). A rock made of lithified gravel and finer sand is called a **conglomerate** if it has rounded gravel and a **breccia** if its clasts are angular. Conglomerates are the product of flood energies or debris flows and are diagnostic of sedimentary settings near the mountains where such floods occur. Breccias are even more specific, since it's hard for large pebbles not to become rounded, even with a few kilometers of washing downhill. Thus, breccias occur in unusual settings (collapsed cave ceilings, impact debris, and landslides) and are powerful indicators of these odd conditions.

We all know what sand is, and what it feels like (think sandpaper or beach sand), but to a sedimentologist, *sand* is strictly defined as particles between 2 and 1/16 mm in diameter. Many sands become cemented to form the rock known as **sandstone**. Most sandstones are rich in quartz, the most stable mineral on the Earth's surface, so quartz sandstone is the most common type. However, rare sandstones can be rich in feldspars, which are common in igneous rocks but rapidly break down into clays when weathered on the Earth's surface. Thus, a sandstone rich in feldspars (called an **arkose**) is very unusual and typically indicates

Millimeters (mm)		Wentworth size class		Rock type			
	4096	Boulder	Gravel	Conglomerate/ Breccia			
	256	Cobble					
	64	Pebble					
	4	Granule					
	2.00	Very coarse sand	Sand	Sandstone/ Arkose			
	1.00	Coarse sand					
1/2	0.50	Medium sand					
1/4	0.25	Fine sand					
1/8	0.125	Very fine sand					
1/16	0.0625	Coarse silt	Silt	Mudstone	Siltstone	Shale	
1/32	0.031	Medium silt					
1/64	0.0156	Fine silt					
1/128	0.0078	Very fine silt					
1/256	0.0039	Clay	Mud	Claystone			
	0.00006						

FIGURE 2.7 Classification of clastic sedimentary rocks.

Source: Redrawn based on several sources by E. T. Prothero; photos by the author.

that the sand grains have not traveled very far or have not been deeply weathered in wet, humid conditions.

Sediment finer than 1/16 mm can be called by the familiar name *mud*, and if it is compressed into a rock, it is called a **mudstone**. Mud can be further subdivided into the coarser material known as silt (1/16–1/256 mm) and clay (finer than 1/256 mm). A rock made entirely of silt is a **siltstone**, and clay makes a **claystone**. In the field, you can tell silt from clay because silt is still close enough to the sand size range that it is gritty to chew, whereas clay is "creamy" to chew. Almost all mud rocks, however, undergo some burial and pressure. This squeezes down the clay minerals (layered sheet silicates), forces out the water between the mineral grains (often almost 70% of the volume of mud is water), and turns it into a slightly different rock known as **shale**, which breaks into flat sheets along bedding (so it is "fissile"). Shales are by far the most common sedimentary rock on the Earth's surface.

2.4.2 CHEMICAL SEDIMENTARY ROCKS

As mentioned earlier, chemical sedimentary rocks form when ions from pre-existing rocks dissolve by weathering and go into solution in the waters of the Earth (groundwater, rivers, lakes, or the ocean) and then precipitate back out to form new sedimentary minerals. The most common way to do this is for organisms (plants or animals) to pull ions out of seawater (such as calcium and carbonate) and precipitate their shells with calcium carbonate minerals, such as calcite or aragonite. When these carbonate shells and coral skeletons of sea creatures accumulate, they build up carbonate sediment that eventually can crystallize into a rock known as **limestone**. Unlike sand and gravel and mud, *limestones are born, not made*. They are always built

from fossils (Figure 2.8a), even if recrystallization might make the fossils invisible. Most limestones forming today are restricted to tropical or subtropical settings, with warm, shallow, clear water and no clastic sand or mud. Places such as Florida, the Bahamas, parts of the Caribbean, the Persian Gulf, and the South Pacific are the main locations forming carbonate sediments today. During the geologic past, huge shallow tropical seas drowned the continents for millions of years, accumulating huge thicknesses of limestone in much of the world.

Another chemical found in water is silica. It can precipitate to form a rock known as **chert**, which is made of submicroscopic crystals of quartz. Chert comes in many colors based on impurities, so if it is black, we call it flint; if it is red, it is jasper; white chert is novaculite; and so on. Chert in the form of flint or jasper was once important for arrowheads and spearheads, for flint was used to start fires and fire flintlock muskets, among many other purposes.

Chert forms in two main ways. In places where plankton that use silica in their skeletons are extremely abundant, they accumulate to form a silica-rich shale known as **bedded chert** or "ribbon chert" (Figure 2.8d). This kind of chert is precipitated by organisms, just as limestones are. The other kinds of chert form when silica-rich groundwater percolates through other rocks (usually limestone) and replaces calcite with silica. These are known as **nodular cherts** (Figure 2.8e).

Most of the chemical sedimentary rocks just discussed were formed by organisms precipitating ions out of water, so they are often called **biochemical** (or organic) **sedimentary rocks**. But there are also rocks that form by straight chemical precipitation, without the help of organisms to drive precipitation. These are **inorganic chemical sedimentary rocks**. The main mechanism required to make
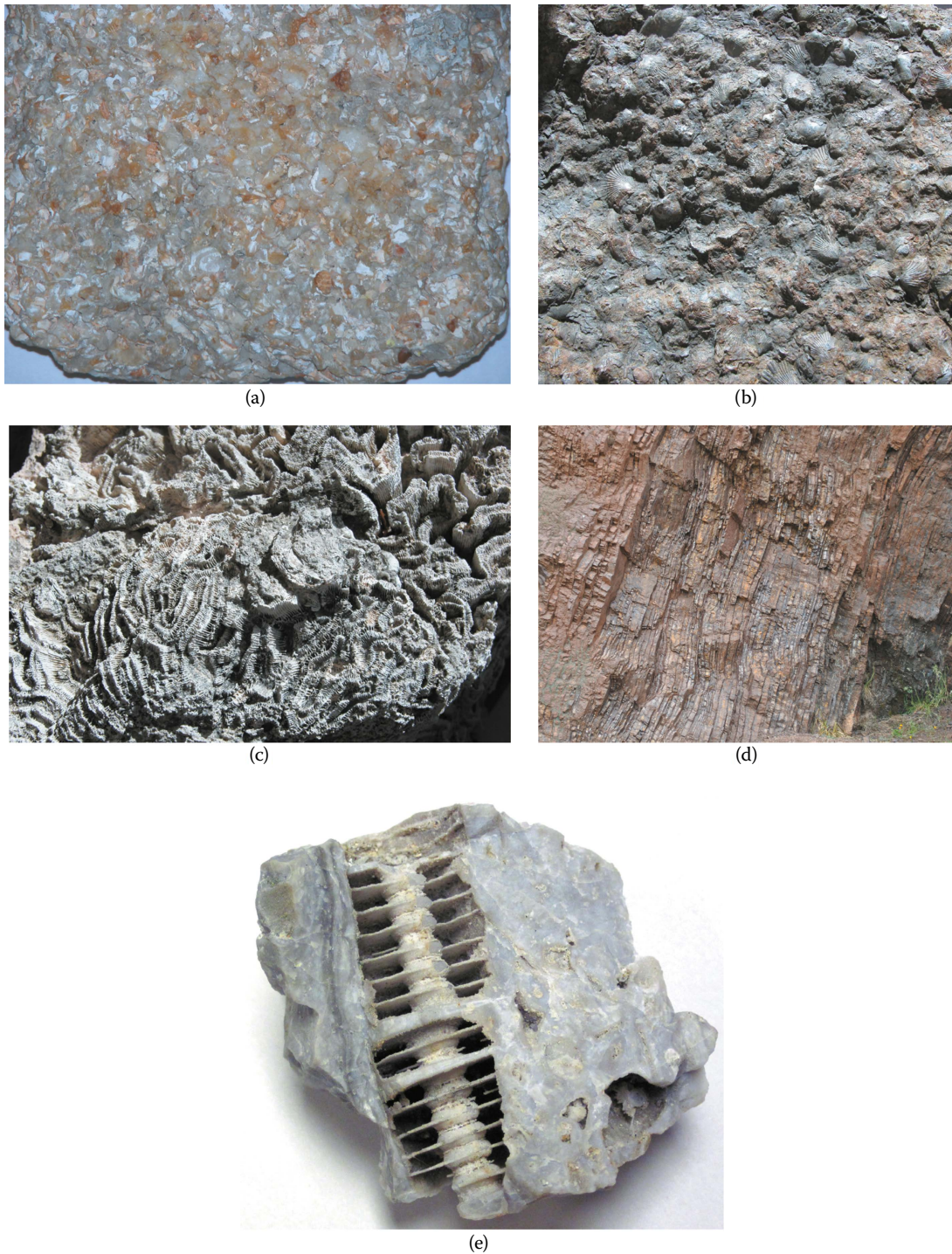
FIGURE 2.8 Selection of biochemical sedimentary rocks (limestones and cherts), including (a) a shell-rich coquina; (b) fossiliferous limestone full of brachiopods; (c) fossil coral, which was already a solid limestone while the corals were still alive; (d) bedded or "ribbon" cherts, which are made of layers rich in silica from plankton interbedded with shales formed from the deep-sea muds; and (e) nodular chert, in which the silica replaces fossiliferous limestone as it seeps through in the groundwater. Here, the chert nodule has replaced the limestone surrounding a calcite crinoid stem, which has dissolved away, leaving a crinoid stem-shaped void.

Source: Photos by the author.

minerals precipitate in this way is to evaporate all the water away and leave the minerals behind to crystallize out of a salty **brine**. Thus, these minerals are known as **evaporites**. Such evaporation is common in dry lake beds, but also in tropical lagoons and hot desert seas like the Persian Gulf, where the rate of evaporation is greater than the amount of seawater flowing in to replace the water evaporated. The most common evaporite minerals are salt (sodium chloride), gypsum (hydrous calcium sulfate), and sometimes calcite or aragonite, but there are hundreds of additional

evaporite minerals, many of which are unique to some of California's dry lakes (see Chapter 7).

2.5 METAMORPHIC ROCKS

The third major class of rocks is metamorphic rocks (*meta* meaning "after" or "changed," and *morphos* meaning "form" in Greek). We are familiar with the word *metamorphosis* to describe many kinds of changes in form, such as the caterpillar changing to a pupa and then a butterfly. Metamorphic rocks are transformed or changed from some original "parent rock" or **protolith** (usually an igneous rock or sedimentary rock) into an entirely new and different kind of rock with new minerals and new textures or fabrics. Some metamorphic rocks are so completely transformed that they are unrecognizable, and we may never know what kind of protolith they started from.

2.5.1 Pressure and Temperature

Metamorphism occurs when the protolith is buried deep in the crust and experiences extremely *high temperatures* and *directed pressure*. Thanks to the enormous heat flow coming up from the Earth's interior, the crustal rock below your feet gets hotter and hotter the deeper you go. In fact, the **geothermal gradient** is about a 30°C increase per kilometer of depth, so at 30 km, the crustal rocks would be about 900°C (above the melting temperature of many minerals). Most people can't even imagine this, but if you go down an old abandoned mine shaft with no air-conditioning, you can feel how it gets hot as you descend. South African diamond and gold miners work at depths close to 3.9 km (12,800 ft) and require a continuous supply of refrigerated fresh air to survive shifts of only a few hours, because the temperature of the rocks and air down there is 60°C (140°F).

The increase in pressure is also intense. The pressure gradient is 3 kbar per 10 km depth you descend. One kilobar (which is 1,000 bars in the metric system) is about 14,500 lb/in², so at 30 km down (very shallow crustal levels), pressures are about 9 kbar, or almost 45,000 lb/in². These pressures are far too great for any device to overcome by drilling (no drilling hole has gone much farther than 12 km, or 40,000 ft). You can laugh at any of those science fiction movies that imagine drilling to the Earth's interior or a "journey to the center of the Earth"! They are complete fantasy, because the pressures and heat are so intense even down 4 km that no human or machine or drill could survive.

The pressure is also **directed pressure**, or pressure applied in a specific direction, so that one axis of a rock is being crushed in the up–down direction, while there is less pressure in the perpendicular direction, so the rock can be squeezed sideways as it is flattened. This is very different from the uniform pressure that you experience on all sides of your body from the air around you, or from water around you when you dive. Directed pressure tends to squash or flatten most rocks, stretching and squeezing them outward. Old minerals that were present in the rock as it began metamorphism will be flattened out and rotate until they are perpendicular to the direction of maximum pressure. Any new minerals that grow during this metamorphic process (especially platy minerals like the micas muscovite, biotite, and chlorite) will tend to grow perpendicular to the direction of greatest stress as well. As a result, the rock will acquire a strongly planar fabric in the direction of least stress, which is called a **foliation** (after the Latin word *folium*, for "leaf"). Many metamorphic rocks have a planar fabric or layering and will split along this plane of foliation as a result.

2.5.2 Types of Metamorphic Rocks

Let's take a variety of different protoliths and see what happens during metamorphism (Figure 2.9). If we start with shale, for example, we have lots of chemistry (silicon, aluminum, potassium, sodium, and other elements) to work with. The first metamorphic product of a shale would be a rock known as a **slate**, which is platy and highly foliated, so it readily breaks into things like roofing tiles and (in the old days) slate for blackboards. As the temperature and directed pressure increase, the clay minerals in the shale are transformed into tiny flakes of micas like muscovite or biotite (not yet visible to the naked eye), and the rock acquires a distinctive sheen; this is called a **phyllite**. Further pressure and temperature allow the new metamorphic minerals (muscovite and biotite plus garnet, hornblende, and others) to grow large enough to be visible to the eye; this kind of rock is called a **schist**. Finally, at extremely high pressures and temperatures, some of the minerals begin to melt and segregate into bands of light-colored minerals (typically quartz and plagioclase) and dark-colored minerals (typically biotite and hornblende). This *compositional banding* is the diagnostic feature of a rock known as **gneiss**. Any further increase in pressure and temperature and the gneiss melts completely and can become a magma that could cool to form an igneous rock.

Let's try a different protolith (Figure 2.9). If we start with an olivine peridotite or pyroxene-rich gabbro, we have only magnesium, iron, and silica to work with. Under high pressures and temperatures, you transform olivine or pyroxene into a new mineral known as **serpentine**, which gets its name from its snake-like green color and waxy, smooth, scaly "snakeskin" feel to the touch. A rock made of the mineral serpentine is called a **serpentinite**. Serpentinite is the official state rock of California because it is common in the Coast Ranges, the Klamaths, and the Sierra Nevada Mountains, where ultramafic olivine-rich slices of oceanic crust have been metamorphosed to form serpentinite.

What if you start with a different protolith, such as a quartz sandstone? Quartz sandstone contains only one chemical, silicon dioxide, and you can't make anything but quartz from that chemistry, no matter how high the pressure and temperature. So under metamorphism, the quartz sandstone becomes a different quartz-rich rock, a **quartzite** (Figure 2.9). The mineralogy won't change (always forming quartz), but the original fabric of spherical quartz sand grains packed together will vanish, and the grain boundaries will fuse together and become interlocked, like pieces of a jigsaw puzzle. If you hit
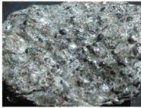
FIGURE 2.9   Different protoliths produce different kinds of metamorphic products as pressure and temperature increase.

Source: Redrawn from several sources by E. T. Prothero; photos by the author.

a sandstone, the rock will break between the grains, but if you hit a quartzite, it will fracture right through the original sand grains, since they are fused into one mass of quartz.

Let's consider one more protolith: limestone. Limestone has only one mineral available, calcite (calcium carbonate), so it can only form calcite no matter what the pressure and temperature. Thus, a fossil-rich limestone protolith transforms into a **marble** under metamorphism (Figure 2.9). A true marble is still made entirely of calcite, but the minerals and fossil fragments have completely recrystallized so that no fossils are visible anymore and only large shiny crystals of calcite are left.

So far, most of the metamorphic rocks we have discussed (shale, phyllite, schist, gneiss, and serpentinite) are highly foliated. But quartzite and marble never undergo foliation, no matter how high the pressures and temperatures. This is because they are not made of platy minerals like micas or elongated prismatic minerals like hornblende. Minerals like quartz and calcite have no long axis or preferred orientation, so no matter how much you cook or squeeze them, they will never foliate.

2.5.3   METAMORPHIC GRADE AND FACIES

We can plot all the different possible regimes of pressure and temperature on the graph shown in Figure 2.10. Temperature increases to the right on the *x*-axis, and pressure increases downward (as it does in the real world) on the *y*-axis. In the upper left part of the diagram is the region of lowest pressure and temperature (Earth surface conditions or very shallow burial), and no metamorphism takes place at this level. If we descend the geothermal gradient line until temperatures are around 330–500°C and pressures are about 2–8 kbar, we are in a region of relatively low pressures and temperatures, producing **low-grade metamorphism**. Geologists use a shorthand term for this region, calling it the **greenschist facies** (*facies* is the Latin word for "appearance"). Rocks that have undergone low pressures and temperatures *appear* as green schists because they usually grow the green mica chlorite, plus sometimes other green minerals, such as amphiboles known as actinolite–tremolite, and maybe even the pistachio-green mineral epidote. Moving farther down the geothermal gradient, we reach a region of intermediate pressures and temperatures (4–12 kbar and 500—700°C). The geologists' term for **intermediate-grade metamorphism** is **amphibolite facies**, because these rocks tend to be rich in the black amphibole hornblende. Finally, we go to the very deepest part of the continental crust, where there are extreme pressures (6–14 kbar) and temperatures (at least 700°C), making the highest-grade metamorphics. These rocks have a gneissic texture and are sometimes called **granulite facies**. All three of these facies are produced by **regional metamorphism**, such as when a collision between continents creates a huge uplifted mountain belt (such as the Himalayas today). Rocks that undergo shallow burial will only reach greenschist facies, while those with even deeper burial will become amphibolites or, finally, granulite gneisses, if they are many kilometers down into the crust at the root of the mountains.
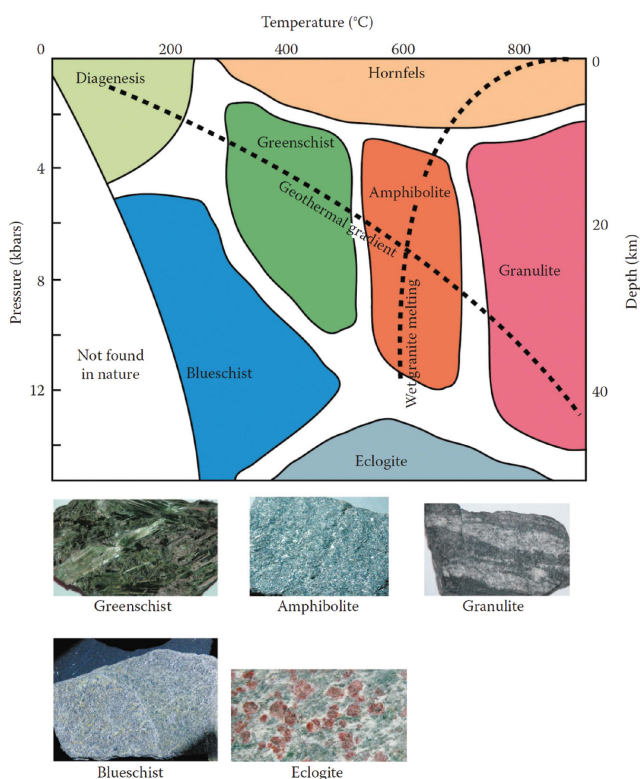
FIGURE 2.10 Metamorphic facies diagram showing how rocks change their characteristic appearance (*facies* in Latin) as they change minerals and texture. Rocks subjected to very low temperatures and pressure are only slightly changed ("diagenesis") but not truly metamorphic. Low temperatures and pressures produce greenschist-grade metamorphics. If those rocks are buried deeper in the roots of mountains ("regional metamorphism"), they experience higher pressures and temperatures as they proceed down the geothermal gradient (dashed line going from upper left to lower right), turning into amphibolite facies or granulite gneiss facies. Rocks that are heated by an igneous intrusion without much pressure of burial are known as the hornfels facies. Rocks formed in a subduction zone, which is relatively cool despite extreme pressure of burial, form blueschist metamorphics. The boundary for melting metamorphic rocks into magma is also shown.

Source: Redrawn based on several sources by E. T. Prothero; photos by the author.

Look at Figure 2.10 again. We can see how the three main facies are found in increasing depths and pressures as they descend deeper into the crust (and further down the geothermal gradient). But what about the peculiar region across the top of the diagram labeled **hornfels**? A glance at the axes of the plot suggests that these would be formed under high temperatures but low pressures. This could not happen by simple burial at greater depths, because both temperature and pressure increase together. Where do such peculiar conditions exist? The only possibility is to heat a rock to high temperatures without burying it deeply. This can only happen when a magma body or dike intrudes through the country rock, cooking it without much depth of burial. Such metamorphism caused by the contact of an intruded magma is called **contact metamorphism** and

produces hornfels. Such rocks often don't show much in the way of different minerals, but their fabric is welded and baked compared to the unheated rock at some distance from the magma intrusion.

Finally, there is one more peculiar region in Figure 2.10 to explain. On the lower left is a region marked **blueschist**. These rocks get this name because they are often a bluish-gray or even deep-blue color (Figure 2.10), due to the blue amphibole glaucophane, plus another blue or white mineral called lawsonite. The axes in Figure 2.10 indicate that it is a region of very high pressure but relatively low (less than 400°C) temperatures. How could this happen? Most rocks descending into the deep crust get hot when they are under such high pressures. The answer to the mystery of blueschists was discovered in California, because they are particularly common in the Coast Ranges (see Chapter 12) and rare elsewhere in the United States. The only place blueschists are found is in the remnants of ancient subduction zones (Chapter 4), where the cold down-going plate plunges into the hot mantle and reaches depths of 20 km or greater. At this point, the rocks are surrounded by the high pressures of such deep burial, but they are still relatively cold because the old oceanic plate retains a lot of water and heats up slowly. Under these unusual conditions, the pressures can get very high, but the temperatures low enough to form glaucophane, lawsonite, and some other distinctive minerals. Then when some of this plate gets scraped off in an accretionary wedge (see Chapter 4), the blueschist can rise to the surface after having been more than 20 km underground in a subduction zone.

2.6 ROCK CYCLE

As we have suggested already, minerals and rocks can transition from one category to another quite easily. Take the example we used earlier of the sedimentary rock known as shale. It can transition from sedimentary rock to metamorphic rock as it experiences high pressures and temperatures. It then goes from shale to slate, phyllite, schist, and gneiss. Eventually, it gets hotter and hotter until it melts. Then it has become a magma and can cool into an igneous rock (Figure 2.9). Thus, we have a sequence of sedimentary to metamorphic to igneous rock (Figure 2.11). Then if that igneous rock reaches the Earth's surface, it will weather and break down into loose sand and mud and return to the sedimentary beginning of this loop all over again.

This is the **rock cycle**, and it is a demonstration of the fact that rocks can transform from one category to the next and eventually return to their starting point over millions of years. One of the great lessons you learn from geology is not only that time is immense (millions and billions of years), but given enough time, ocean bottoms can turn into mountains and then weather down into sediment and return to the ocean—or be forced down into the lower crust and transform into metamorphic rocks or even melt into a magma. With enough time, any of these extremely slow processes are inevitable.
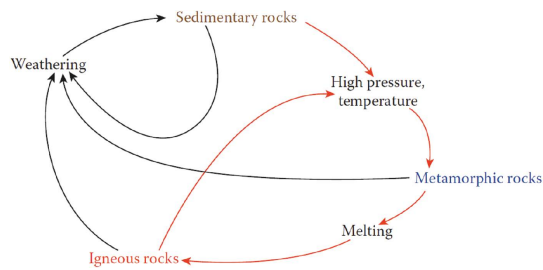
FIGURE 2.11 The rock cycle. Over millions of years, no rock is permanent, but it is part of a continuous slow cycle, changing from one class of rock (e.g., an igneous rock) to another (e.g., weathered sediments from the igneous rocks) to another (e.g., metamorphism of sedimentary rocks into metasedimentary rocks). If the metamorphic rock is heated enough, it melts and returns to the igneous part of the loop.

Source: Redrawn based on several sources by E. T. Prothero.

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VIDEOS

Three kinds of rocks: www.youtube.com/watch?v=sN7AficX9e0
Series on minerals and rocks:
www.youtube.com/watch?v=8a7p1NFn64s
www.youtube.com/watch?v=32NG9aeZ7_c
www.youtube.com/watch?v=ZkHp_nnU9DY
www.youtube.com/watch?v=aCnAF1Opt8M
www.youtube.com/watch?v=Etu9BWbuDlY
www.youtube.com/watch?v=1oQ1J0w3x0o

Animations

Clever site that allows you to build silicate structures with interactive animation: https://ees.as.uky.edu/sites/default/files/elearning/module09swf.swf

Excellent 3D animation of complex silicate structures: http://web.visionlearning.com/silica_molecules.shtml

Learning module on silicate structures: www.visionlearning.com/en/library/Earth-Science/6/The-Silicate-Minerals/140

3 Dating California
Stratigraphy and Geochronology

[The concept of geologic time] makes you schizophrenic. The two time scales—the one human and emotional, the other geologic—are so disparate. But a sense of geologic time is the important thing to get across to the non-geologist: the slow rate of geologic processes—centimeters per year—with huge effects if continued for enough years. A million years is a small number on the geologic time scale, while human experience is truly fleeting—all human experience, from its beginning, not just one lifetime. Only occasionally do the two time scales coincide.

—Eldridge Moores
in Assembling California**,** *by John McPhee*

3.1 STRATIGRAPHY

One of the oldest branches of geology is known as **stratigraphy**. It literally means "the study and description of layered rocks," but over the past 250 years, it has come to mean any study of rocks that helps us determine their age, how they were deposited, and what they tell us about geologic history. Today, we can think of stratigraphy as the toolbox we use to understand the geologic past.

3.2 DATING ROCKS

There are two basic ways of determining the age of a geologic event or rock body. They are **relative age** and **numerical age**.

Relative age is the age of one object or event in relation to another, that is, this rock is younger than that one. There are several common ways to determine this. Most of these concepts (Figure 3.1) were first formulated by the Danish physician Niels Steensen (Nicholas Steno in Latin) in 1669:

Principle of superposition. In any layered sequence of rocks, it makes sense that the lowest rocks in the stack have to be older than the upper ones; you can't put one thing on top of another if it isn't already there. Thus, in the layered sequences we will see repeatedly in this book, the rocks get younger as you move up the section, and this is the primary method by which we determine relative age. A good analogy for this is a stack of papers on a messy desk that have piled up for months undisturbed. The oldest papers would be at the base of the stack, and the most recently used papers would be at the top.

Crosscutting relationships. Whenever one rock unit or fault cuts across another, the rock that does the cutting must be younger than whatever it cuts through. This applies primarily to dikes of molten rock (magma) from the Earth's interior, which can cut through (intrude) the surrounding rock as they melt their way through the Earth's crust. Likewise, a fault can only cut through rocks that are already there.

Principle of original continuity. Whenever you see two rock units that are now separated by erosion, it is safe to assume that they were originally connected and continuous, and that the erosion has since cut through them. Thus, the erosion cuts through the layers, and it must be younger. For example, just below the rim of the Grand Canyon, you can see a distinctive white layer known as the Coconino Sandstone on both the North Rim and South Rim side. Even though they are separated by a huge canyon, we can visualize how they were once connected and conclude that the cutting of the canyon must be younger than the (now discontinuous) layers it cut through.

Principle of original horizontality. When we see sedimentary beds of sand or mud laid down in rivers or oceans today, they are nearly always in horizontal layers due to Earth's gravity. Thus, if we find these beds tilted or folded in any way, then the deformation must have happened after the sand or mud was laid down and turned into sandstone or shale.

In addition to determining the sequence of rock units and faults through these principles, we must also be aware of erosional gaps in the rock sequence, known as **unconformities**. These occur when one rock unit is uplifted and eroded away and then down-dropped again and the erosional surface covered by a new rock layer. Most of these unconformities take millions of years to develop and form, and each erosional surface represents a time gap that may be millions of years long, often longer than the time represented by actual rocks in the sequence. There are three common kinds of unconformities (Figure 3.2):

Angular unconformity. The rocks below the erosional surface are tilted and eroded off and then covered by the upper sequence (Figure 3.2b). This is the easiest unconformity to spot in the field.
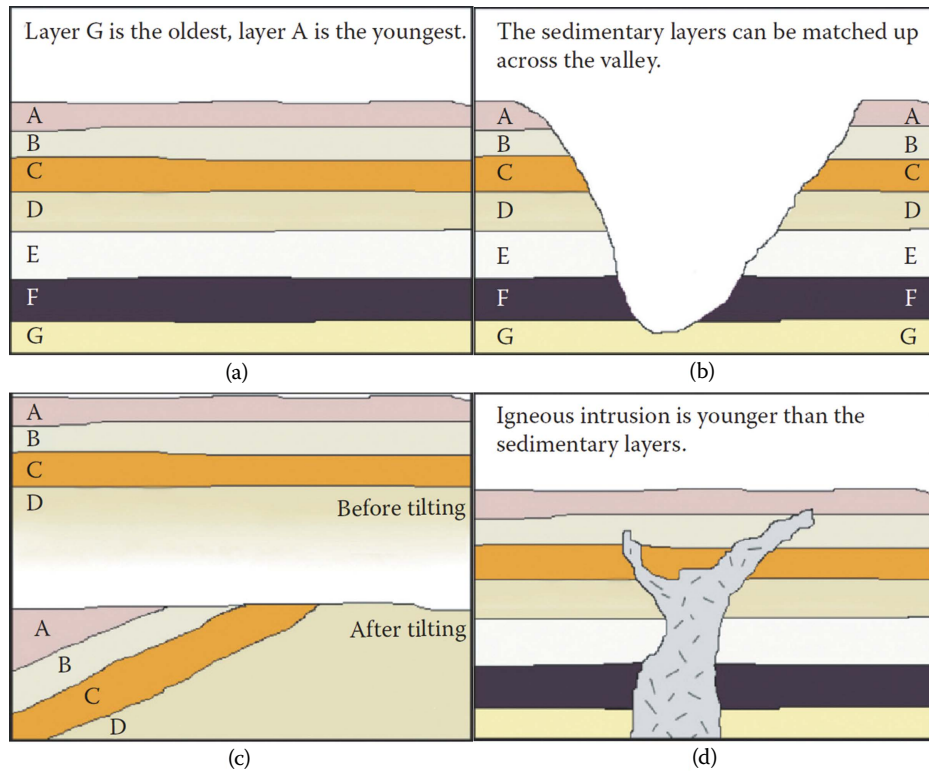
Layer G is the oldest, layer A is the youngest.

A
B
C
D
E
F
G
(a)

The sedimentary layers can be matched up across the valley.

A
B
C
D
E
F
G
(b)

A
B
C
D
Before tilting

A
B
C
D
After tilting
(c)

Igneous intrusion is younger than the sedimentary layers.
(d)

FIGURE 3.1 Steno's principles. (a) *Superposition*: the layers at the bottom of the stack are older than the layers placed above them. (b) *Original continuity*: if you find a canyon or valley cut through layers that match on both sides, then the canyon is younger than the layers it cuts. (c) *Original horizontality*: rock layers are normally deposited as horizontal sheets of sediments or horizontal lava flows and volcanic ash layers, so if you find them tilted or folded, their deformation occurred after they were deposited or erupted. (d) *Crosscutting relationships*: any event (igneous intrusion, fault, etc.) that cuts through pre-existing rocks must be younger than what it cuts.

Source: Drawing by M. P. Williams.

Angular
unconformity

Nonconformity

Disconformity

(a)
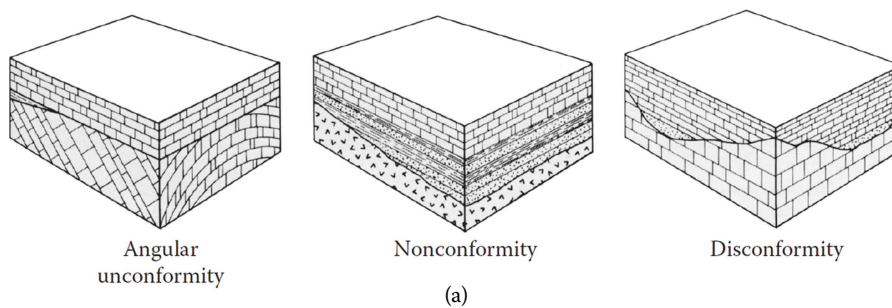
FIGURE 3.2 The major types of unconformities: (a) Diagram representing the three main types. (b) Angular unconformity between the 12-million-year-old Mint Canyon Formation (tilted gray sandstones) and 1-million-year-old Saugus Formation (horizontal gravels incised into the lower unit), so there is an 11-million-year time gap between the lower and upper units. This outcrop is on California Highway 14 northeast of the Soledad Canyon Road exit. (c) Nonconformity between Precambrian granitic rocks (deeply weathered and crumbling in lower left) and hard, resistant, layered Lower Cambrian Wood Canyon Formation, Marble Mountains. (d) Disconformity, showing one sequence of tan sandstones full of gravels eroded down into channels cut into a lower sequence of gray-pink sandstones and siltstones, lower Miocene Vasquez Formation, Vasquez Rocks.

Source: (a) redrawn from several sources by P. Linse. Photos (b) and (d) by the author; photo (c) by J. Foster.

(b)

(c)

(d)

FIGURE 3.2 (Continued)

An angular unconformity represents at least five separate events: the slow deposition of the lower sequence of sediments, the hardening of those loose sediments into sedimentary rock, tilting the lower sequence, eroding it off, and then depositing the upper sequence. Each angular unconformity by itself represents millions of years of time.

Nonconformity. Here the rocks below the erosional surface (Figure 3.2c) are a nonsedimentary rock (in other words, igneous or metamorphic) formed by crystallization or high temperatures and pressures deep in the Earth's crust. These must be uplifted, exposed, and eroded and then covered by sediments to form the erosional surface.

Disconformity. This is the subtlest type of unconformity, since the beds above and below the erosional surface are parallel. However, between them lies a slight erosional surface that usually must be detected by scouring or channeling into the lower unit (Figure 3.2d) that is then filled by sediments of the upper unit. In many cases, there is no obvious field evidence of erosion, and the gap can only be determined by the age of the fossils above and below.

In the field, geologists must reconstruct a long sequence of events using superposition of the bedded rocks (sedimentary rocks and lava flows) plus the crosscutting relationships of the igneous intrusions, faults, and unconformities. From this method, we can determine the relative age in any complex series of events.

3.3 GEOCHRONOLOGY

Numerical dating (formerly but incorrectly called by the obsolete term *absolute dating*) or **geochronology** is the method of determining the age of geologic event given in some sort of numerical age, for example, "this rock is X number of years old." Although the principles of relative dating and the geologic time scale were developed more than 200 years ago, there was no reliable way of determining the numerical age of events until about 1913, when radioactive dating was developed. All radioactive dating works on similar principles. Certain elements in nature (such as uranium-238 or potassium-40) are naturally radioactive and decay to a daughter element (e.g., lead-207 or argon-40) while giving off heat and radioactive particles (alpha, beta, and gamma radiation). This decay occurs in a predictable fashion, with an exponential or logarithmic decay curve, so that if you know the amount of parent material and daughter material and the decay rate for this system, you can determine the age.

There are many different radioactive elements, but only a few are useful in geologic settings. The most commonly used method is potassium–argon dating, where potassium-40 decays to argon-40. Because potassium is one of the eight most common elements in the Earth's crust (Table 2.1), it can be found in many different minerals (especially micas and potassium feldspars) and can date rocks as old as 4.6 billion years old and as young as a million years old. The main weakness of this system is that argon is a gas, which tends to escape from the system and give erroneous ages if the material has been weathered.

There are also systems of uranium–lead and rubidium–strontium dating, but these are used mostly for rocks that are more than 600 million years old, since their decay rates are very slow. In addition, uranium and rubidium are relatively rare elements, so a lot of time is needed to accumulate a measurable amount of daughter atoms.
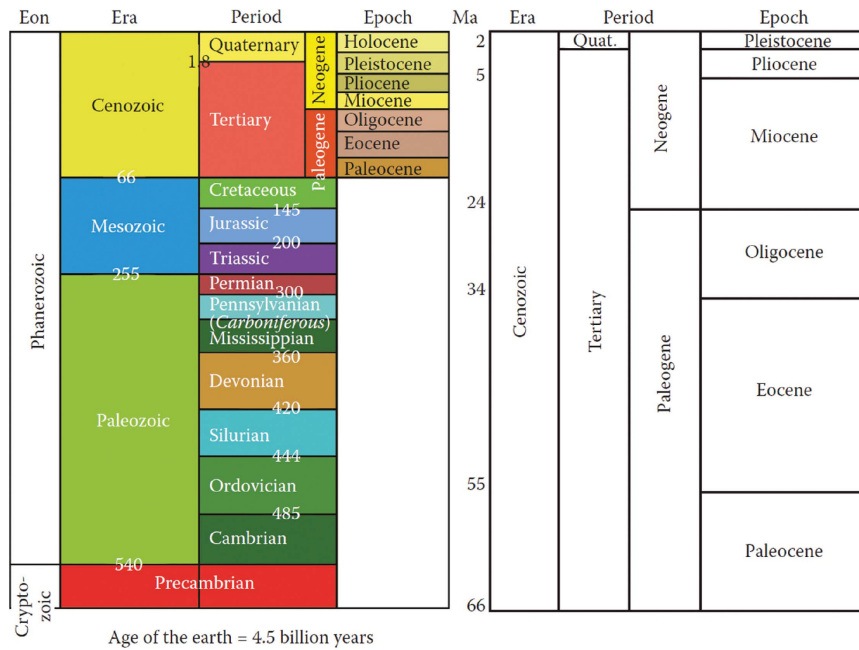
FIGURE 3.3 Standard geologic time scale. On the left is the entire Phanerozoic eon. On the right is a detail of the Cenozoic epochs.

Source: Drawn from several sources by E. T. Prothero.

Finally, most people have heard of carbon-14 dating, where you can measure the decay of radioactive carbon-14 back to nitrogen-14 in any carbon-bearing substance, including bones, wood, charcoal, pottery, baskets, and shells. The main limitation of this method is that the decay rate is very fast, so that nothing older than about 60,000–80,000 years old can be dated by carbon-14. Thus, it is a common method in archeology and in studying fossils and rocks from the last ice age, but useless to any other kind of geologic setting.

In all these systems (except carbon-14), you are measuring the ratio of parent and daughter elements that accumulate in a closed, sealed crystal. This means that you need the crystal to cool in a melt and lock in the radioactive materials in its crystal structure so that none can escape. For this reason, the only kinds of rocks that can be numerically dated by most methods (e.g., potassium–argon) are igneous rocks formed from a molted magma, such as volcanic rocks erupted out of the Earth, or magma that has cooled and crystallized in a deep underground magma chamber (plutonic rocks). Sedimentary rocks are not formed this way. Their grains of sand or mud are weathered and eroded out of pre-existing rock, so there is no way to radioactively date sedimentary rocks directly. When a geologist gives a numerical age estimate to a sedimentary rock, it is based on knowing the relative age (which is based on its distinctive fossils) and then finding some place on Earth where a volcanic lava flow or ash layer is interbedded with rocks bearing

the same fossils. That is how the international geologic time scale was produced (Figure 3.3).

RESOURCES

McPhee, J. 1993. *Assembling California*. Farrar, Straus, and Giroux, New York.

Prothero, D.R., and Schwab, F. 2013. *Sedimentary Geology: Principles of Sedimentology and Stratigraphy*. W.H. Freeman, New York.

VIDEOS

Numerical dating

Bill Nye on radiometric dating: www.youtube.com/watch?v=Q6O LBVDbJoI

NOVA on radiocarbon dating: www.pbs.org/wgbh/nova/tech/radio carbon-dating.html

Radiocarbon dating

www.youtube.com/watch?v=54wR-zwuDGo

www.youtube.com/watch?v=udkQwW6aLik

www.youtube.com/watch?v=phZeE7Att_s

Relative dating (note that these use the obsolete term *absolute dating*):

www.youtube.com/watch?v=fYSeM63Fv0s

Animations

http://science.jburroughs.org/mbahe/BioA/starranimations/ chapter16/videos_animations/carbon_14_dating_v2.swf