Crustal Deformation

CHAPTER

10

Deformed strata in the Panamint Range, Death Valley National Park, California. (Photo by Michael Collier)

Earth is a dynamic planet. In the preceding chapters, you learned that weathering, mass wasting, and erosion by water, wind, and ice continually sculpture the landscape. In addition, tectonic forces deform rocks in the crust. Evidence demonstrating the operation of enormous forces within Earth includes thousands of kilometers of rock layers that are bent, contorted, overturned, and sometimes riddled with fractures (Figure 10.1). In the Canadian Rockies, for example, some rock units have been thrust for hundreds of kilometers over other layers. On a smaller scale, crustal movements of a few meters occur along faults during major earthquakes. In addition, rifting (spreading) and extension of the crust produce elongated depressions and over long spans of geologic time may even create ocean basins.

Structural Geology: A Study of Earth's Architecture

The results of tectonic activity are strikingly apparent in Earth's major mountain belts (see chapter-opening photo). Here, rocks containing fossils of marine organisms may be found thousands of meters above sea level, and massive rock units show evidence of having been intensely fractured and folded, as though they were made of putty. Even in the stable interiors of the continents, rocks reveal a history of deformation that shows they have been uplifted from much deeper levels in the crust.

Structural geologists study the architecture of Earth's crust and "how it got this way" insofar as it resulted from deformation. By studying the orientations of faults and folds, as well as small-scale features of deformed rocks, structural geologists can often reconstruct the original geologic setting and the nature of the forces that generated these rock structures. In this way, the complex events of Earth's geologic history are unraveled.

An understanding of rock structures is not only important in deciphering Earth history, it is also basic to our economic well-being. For example, most occurrences of oil and natural gas are associated with geologic structures that act to trap these fluids in valuable "reservoirs" (see Chapter 23).

FIGURE 10.1 Uplifted and folded sedimentary strata at Stair Hole, near Lulworth, Dorset, England. These layers of Jurassic-age rock, originally deposited in horizontal beds, have been folded as a result of the collision between the African and European crustal plates. (Photo by Tom & Susan Bean, Inc.)

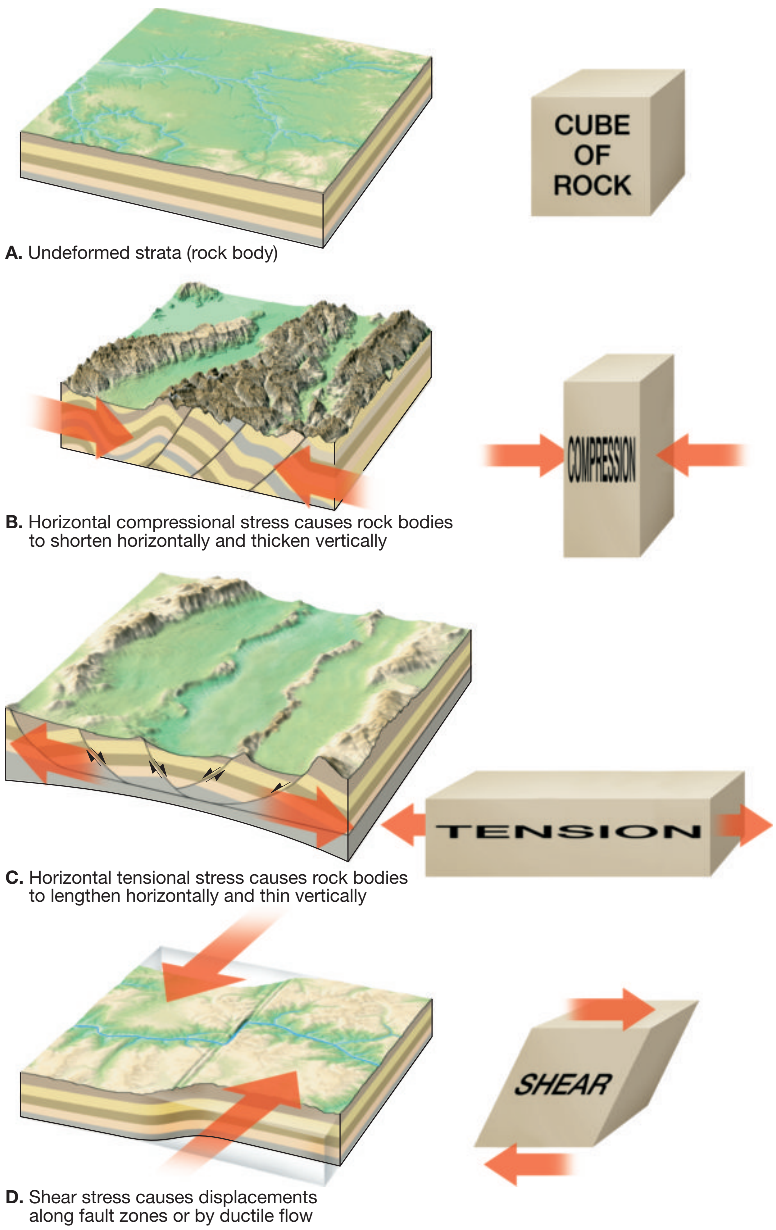


FIGURE 10.2 Deformation of Earth's crust caused by tectonic forces and associated stresses resulting from the movement of lithospheric plates. **A.** Strata before deformation. **B.** Compressional stresses associated with plate collisions tend to shorten and thicken Earth's crust by folding, flowing, and faulting. **C.** Tensional stresses at divergent plate boundaries tend to lengthen rock bodies by displacement along faults in the upper crust and ductile flow at depth. **D.** Shear stresses at transform plate boundaries tend to produce offsets along fault zones. The right side of each diagram illustrates the deformation (strain) of a cube of rock in response to the differential stresses illustrated in corresponding diagram to the left.

Furthermore, rock fractures are sites of hydrothermal mineralization, which means they can be sites of metallic ore deposits. Moreover, the orientation of fracture surfaces, which represent zones of weakness in rocks, must be considered when selecting sites for major construction projects such as bridges, hydroelectric dams, and nuclear power plants. In short, a working knowledge of rock structures is essential to our modern way of life.

In this chapter we will examine the forces that deform rock, as well as the structures that result. The basic geologic structures associated with deformation are folds, faults, joints, and foliation (including rock cleavage). Because rock cleavage and foliation were examined in Chapter 8, this chapter will be devoted to the remaining rock structures and the tectonic forces that produce them.

Deformation

Crustal Deformation
▸ Deformation

Every body of rock, no matter how strong, has a point at which it will fracture or flow. **Deformation** (*de* = out, *forma* = form) is a general term that refers to all changes in the size, shape, orientation, or position of a rock mass. Most crustal deformation occurs along plate margins. Plate motions and the interactions along plate boundaries generate the tectonic forces that cause rock units to deform.

Force and Stress

Force is that which tends to put stationary objects in motion or change the motions of moving bodies. From everyday experience you know that if a door is stuck (stationary), you apply force to open it (get it in motion).

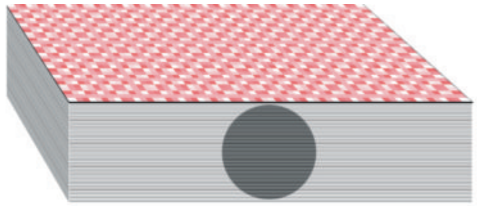
To describe the forces that deform rocks, structural geologists use the term **stress**—the amount of force applied to a given area. The magnitude of stress is not simply a function of the amount of force applied but also relates to the area on which the force acts. For example, if you are walking barefoot on a hard surface, the force (weight) of your body is distributed across your entire foot, so the stress acting on any one point of your foot is low. However, if you step on a small pointed rock (ouch!), the stress concentration at a point on your foot will be high. Thus, you can think of stress as a measure of how concentrated the force is. As you saw in Chapter 8, stress may be applied uniformly in all directions (*confining pressure*) or nonuniformly (*differential stress*).

Types of Stress

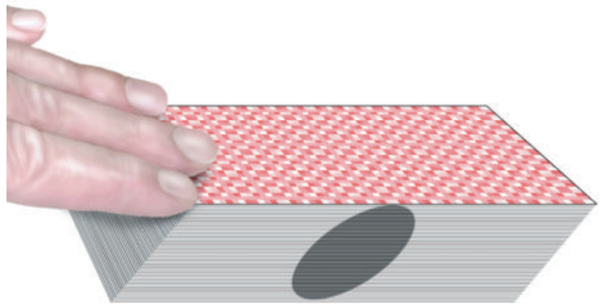
When stress is applied unequally in different directions, it is termed **differential stress.** Differential stress that shortens a rock body is known as **compressional** (*com* = together, *premere* = to press) **stress.** Compressional stresses associated with plate collisions tend to shorten and thicken Earth's crust by folding, flowing, and faulting (Figure 10.2B). Recall from our discussion of metamorphic rocks that compressional stress is more concentrated at points where mineral grains are in contact, causing mineral matter to migrate from areas of high stress to areas of low stress (see Figure 8.7). As a result, the mineral grains (and the rock unit) tend to shorten in the direction parallel to the plane of maximum stress and elongate perpendicular to the direction of greatest stress.

When stress tends to elongate or pull apart a rock unit, it is known as **tensional** (*tendere* = to stretch) **stress** (Figure 10.2C). Where plates are being rifted apart (divergent plate boundaries), tensional stresses tend to lengthen those rock bodies located in the upper crust by displacement along faults. At depth, on the other hand, displacement is accomplished by a type of puttylike flow.

Differential stress can also cause rock to **shear** (Figure 10.2D). One type of shearing is similar to the slippage that occurs between individual playing cards when the top of the deck is moved relative to the bottom (Figure 10.3). In near-surface environments, shearing often occurs on closely spaced parallel surfaces of weakness, such as bedding planes, foliation, and microfaults. Further, at transform fault boundaries, shearing stresses produce large-scale offsets along major fault zones. By contrast, at great depths where temperatures and confining pressures are high, shearing is accomplished by solid-state flow.



A. Deck of playing cards



B. Shearing occurs when hand pushes top of deck

FIGURE 10.3 Illustration of shearing and the resulting deformation (strain). **A.** An ordinary deck of playing cards with a circle embossed on its side. **B.** By sliding the top of the deck relative to the bottom, we can illustrate the type of shearing that commonly occurs along closely spaced planes of weakness in rocks. Notice that the circle becomes an ellipse, which can be used to measure the amount and type of strain. Additional displacement (shearing) of the cards would result in further strain and would be indicated by a change in the shape of the ellipse.

Strain

Perhaps the easiest type of deformation to visualize occurs along small fault surfaces where differential stress causes rocks to move relative to each other in such a way that their original size and shape are preserved. Stress can also cause an irreversible change in the shape and size of a rock body, referred to as **strain.** Like the circle shown in Figure 10.3B, *strained bodies do not retain their original configuration during deformation.* Figure 10.1 illustrates the strain (deformation) exhibited by rock units near Dorset, England. When studying strained rock units like those shown in Figure 10.1, geologists ask, "What do these deformed structures indicate about the original arrangement of these rocks, and how have they been deformed?"

How Rocks Deform

When rocks are subjected to stresses greater than their own strength, they begin to deform, usually by folding, flowing, or fracturing (Figure 10.4). It is easy to visualize how rocks break, because we normally think of them as being brittle. But how can rock units be *bent* into intricate folds without being broken during the process? To answer this question, structural geologists performed laboratory experiments in which rocks were subjected to differential stress under conditions that simulated those existing at various depths within the crust (Figure 10.5).

Although each rock type deforms somewhat differently, the general characteristics of rock deformation were determined from these experiments. Geologists discovered that when stress is gradually applied, rocks first respond by deforming elastically. Changes that result from *elastic deformation* are recoverable; that is, like a rubber band, the rock will return to nearly its original size and shape when the stress is removed. (As you will see in the next chapter, the energy for most earthquakes comes from stored elastic energy that is released as rock snaps back to its original shape.)

FIGURE 10.4 Deformed sedimentary strata exposed in a road cut near Palmdale, California. In addition to the obvious folding, light-colored beds are offset along a fault located on the right side of the photograph. (Photo by E. J. Tarbuck)

Undeformed | Low confining pressure | Intermediate confining pressure | High confining pressure

FIGURE 10.5 A marble cylinder deformed in the laboratory by applying thousands of pounds of load from above. Each sample was deformed in an environment that duplicated the confining pressure found at various depths. Notice that when the confining pressure was low, the sample deformed by brittle fracture, whereas when the confining pressure was high, the sample deformed plastically. (Photo courtesy of M. S. Patterson, Australian National University)

Once the elastic limit (strength) of a rock is surpassed, it either flows (*ductile deformation*) or fractures (*brittle deformation*). The factors that influence the strength of a rock and thus how it will deform include temperature, confining pressure, rock type, availability of fluids, and time.

Temperature and Confining Pressure Rocks near the surface, where temperatures and confining pressures are low, tend to behave like a brittle solid and fracture once their strength is exceeded. This type of deformation is called **brittle** (*bryttian* = to shatter) **failure** or **brittle deformation.** From our everyday experience, we know that glass objects, wooden pencils, china plates, and even our bones exhibit brittle failure once their strength is surpassed. By contrast, at depth, where temperatures and confining pressures are high, rocks exhibit *ductile* behavior. **Ductile deformation** is a type of solid-state flow that produces a change in the size and shape of an object without fracturing, as occurs with brittle failure. Ordinary objects that display ductile behavior include modeling clay, beeswax, caramel candy, and most metals. For example, a copper penny placed on a railroad track will be flattened and deformed (without breaking) by the force applied by a passing train. Ductile deformation of a rock—strongly aided by high temperature and high confining pressure—is somewhat similar to the deformation of a penny flattened by a train. One way this type of solid-state flow is accomplished within a rock is by gradual slippage and recrystallization along planes of weakness within the crystal lattice of mineral grains (see Figure 8.7B). This microscopic form of gradual solid-state flow involves slippage that disrupts the crystal lattice and immediate recrystallization that repairs the structure. Rocks that display evidence of ductile flow usually were deformed at great depth and may exhibit contorted folds that give the impression that the strength of the rock was akin to soft putty (Figure 10.6).

Rock Type In addition to the physical environment, the mineral composition and texture of a rock greatly influence how it will deform. For example, crystalline rocks composed of minerals that have strong internal molecular bonds tend to fail by brittle fracture. By contrast, sedimentary rocks that are weakly cemented, or metamorphic rocks that contain zones of weakness, such as foliation, are more susceptible to ductile flow. Rocks that are weak and thus most likely to behave in a ductile manner when subjected to differential stress, include rock salt, gypsum, and shale whereas limestone, schist, and marble are of intermediate

Students Sometimes Ask . . .

I'm confused. Aren't stress and strain the same thing?

No. Although commonly used in similar situations (such as "I'm under stress," and "I can't take the strain"), the terms *stress* and *strain* have specific—and different—meanings in geology. Stress is an applied force; strain is the deformation (bending or breaking) that occurs due to stress. For example, squeezing a tennis ball is subjecting it to a force (stress). The result of this squeezing changes the shape of the ball (strain). Put another way, stress is the action that strains rocks. Strain is a result that can be measured.

FIGURE 10.6 Rocks in Colorado exhibiting the results of ductile behavior. These rocks were deformed at great depth and were subsequently exposed at the surface. (Photo by Marli Miller)

strength. In fact, rock salt is so weak that it deforms under small amounts of differential stress and rises like stone pillars through beds of sediment that lie in and around the Gulf of Mexico. Perhaps the weakest naturally occurring solid to exhibit ductile flow on a large scale is glacial ice. By comparison, granite and basalt are strong and brittle. In a near-surface environment, strong, brittle rocks will fail by fracturing when subjected to stresses that exceed their strength. It is important to note, however, that the presence of even tiny quantities of water in rocks greatly reduces their resistance to ductile deformation.

Time One key factor that researchers are unable to duplicate in the laboratory is how rocks respond to small amounts of stress applied over long spans of *geologic time.* However, insights into the effects of time on deformation are provided in everyday settings. For example, marble benches have been known to sag under their own weight over a span of a hundred years or so, and wooden bookshelves may bend after being loaded with books for a relatively short period. In nature small stresses applied over long time spans surely play an important role in the deformation of rock. Forces that are unable to deform rock when initially applied may cause rock to flow if the stress is maintained over an extended period of time.

It is important to note that the processes by which rocks deform occur along a continuum that ranges from pure brittle fracture at one end to ductile (viscous) flow at the other. There are no sharp boundaries between different types of deformation. We also need to remind ourselves that the elegant folds and flow patterns we see in deformed rocks are generally achieved by the combined effect of distortion, sliding, and rotation of the individual grains that make up a rock. Further, this distortion and reorganization of mineral grains takes place in rock that is essentially solid.

Mapping Geologic Structures

GEODe EARTH
Crustal Deformation
▸ Mapping Geologic Structures

The processes of deformation generate features at many different scales. At one extreme are Earth's major mountain systems. At the other extreme, highly localized stresses create minor fractures in bedrock. All of these phenomena, from the largest folds in the Alps to the smallest fractures in a slab of rock, are referred to as **rock structures.** Before beginning our discussion of rock structures, let us examine the way geologists describe and map them.

When conducting a study of a region, a geologist identifies and describes the dominant structures. A structure often is so large that only a small portion is visible from any particular vantage point. In many situations, most of the bedrock is concealed by vegetation or by recent sedimentation. Consequently, the reconstruction must be done using data gathered from a limited number of *outcrops,* which are sites where bedrock is exposed at the surface (see Box 10.1). Despite such difficulties, a number of mapping techniques enable geologists to reconstruct the orientation and shape of the existing structures. In recent years this work has been aided by advances in aerial photography, satellite imagery, and the development of the global positioning system (GPS). In addition, seismic reflection profiling (see Chapter 12) and drill holes provide data on the composition and structure of rocks that lie at depth.

Geologic mapping is most easily accomplished where sedimentary strata are exposed. This is because sediments are usually deposited in horizontal layers. If the sedimentary rock layers are still horizontal, this tells geologists that the area is probably undisturbed structurally. But if strata are inclined, bent, or broken, this indicates that a period of deformation occurred following deposition.

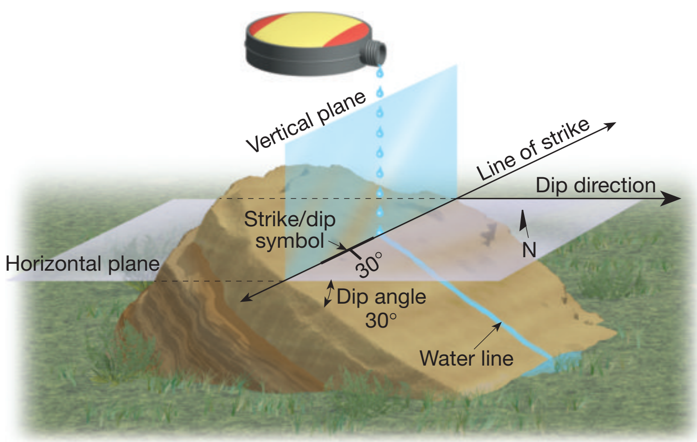
Strike and Dip

Geologists use measurements called *strike* (trend) and *dip* (inclination) to help determine the orientation or attitude of a rock layer or fault surface (Figure 10.7). By knowing the strike and dip of rocks at the surface, geologists can predict the nature and structure of rock units and faults that are hidden beneath the surface, beyond their view.

Strike is the compass direction of the line produced by the intersection of an inclined rock layer or fault, with a horizontal plane (Figure 10.7). The strike, or compass bearing, is generally expressed as an angle relative to north. For example, N10° E means the line of strike is 10° to the east of north. The strike of the rock units illustrated in Figure 10.7 is approximately north 75° east (N75° E).

Dip is the angle of inclination of the surface of a rock unit or fault measured from a horizontal plane. Dip includes both an angle of inclination and a direction toward which the rock is inclined. In Figure 10.7 the dip angle of the rock layer is 30°. A good way to visualize dip is to imagine that water will always run down the rock surface parallel to the dip. The direction of dip will always be at a 90° angle to the strike.

FIGURE 10.7 Strike and dip of a rock layer.



BOX 10.1 ▶ UNDERSTANDING EARTH

Naming Local Rock Units

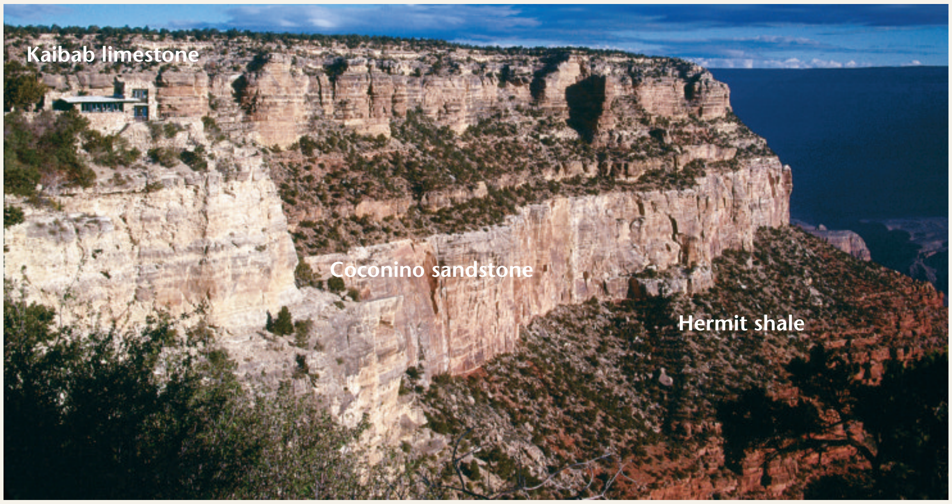


FIGURE 10.A Grand Canyon with a few of its rock units (formations) named. (Photo by E. J. Tarbuck)

One of the primary goals of geology is to reconstruct Earth's long and complex history through the systematic study of rocks. In most areas, exposures of rocks are not continuous over great distances. Consequently, the study of rock layers must be done locally and then correlated with data from adjoining areas to produce a larger and more complete picture. The first step in the effort to unravel past geologic events is that of describing and mapping rock units exposed in local outcrops.

Describing anything as complex as a thick sequence of rocks requires subdividing the layers into units of manageable size. The most basic rock division is called a *formation*, which is simply a distinctive series of strata that originated through the same geologic processes. More precisely, a formation is a mappable rock unit that has definite boundaries (or contacts with other units) and certain obvious characteristics (rock type) by which it may be traced from place to place and distinguished from other rock units.

Figure 10.A shows several named formations that are exposed in the walls of the Grand Canyon. Just as these rock strata in the Grand Canyon were subdivided, geologists subdivide rock sequences throughout the world into formations.

Those who have had the opportunity to travel to some of the national parks in the West may already be familiar with the names of certain formations. Well-known formations include the Navajo Sandstone in Zion National Park, the Redwall Limestone in the Grand Canyon, the Entrada Sandstone in Arches National Park, and the Wasatch Formation in Bryce Canyon National Park.

Although formations can consist of igneous or metamorphic rocks, the vast majority of established formations are sedimentary rocks. A formation may be relatively thin and composed of a single rock type, for example, a 1-meter-thick layer of limestone. At the other extreme, formations can be thousands of meters thick and consist of an interbedded sequence of rock types such as sandstones and shales. The most important condition to be met when establishing a formation is that *it constitutes a unit of rock produced by uniform or uniformly alternating conditions.*

In most regions of the world, the name of each formation consists of two parts—for example, the Oswego Sandstone and the Carmel Formation. The first part of the name is generally taken from a geologic structure or a locality where the formation is clearly and completely exposed. For instance, the expansive Morrison Formation is well exposed at Morrison, Colorado. As a result, this particular exposure is known as the *type locality.* Ideally, the second part of the name indicates the dominant rock type as exemplified by such names as the Dakota Sandstone, the Kaibab Limestone, and the Burgess Shale. When no single rock type dominates, the term *formation* is used, such as the well-known Chinle Formation, exposed in Arizona's Petrified Forest National Park.

In summary, describing and naming formations is an important first step in the process of organizing and simplifying the study and analysis of Earth's history.

In the field, geologists measure the strike (trend) and dip (inclination) of sedimentary rocks at as many outcrops as practical. These data are then plotted on a topographic map or an aerial photograph along with a color-coded description of the rock. From the orientation of the strata, an inferred orientation and shape of the structure can be established, as shown in Figure 10.8. Using this information, the geologist can reconstruct the pre-erosional structures and begin to interpret the region's geologic history.

Folds

Crustal Deformation
▶ Folds

During mountain building, flat-lying sedimentary and volcanic rocks are often bent into a series of wavelike undulations called **folds.** Folds in sedimentary strata are much like those that would form if you were to hold the ends of a

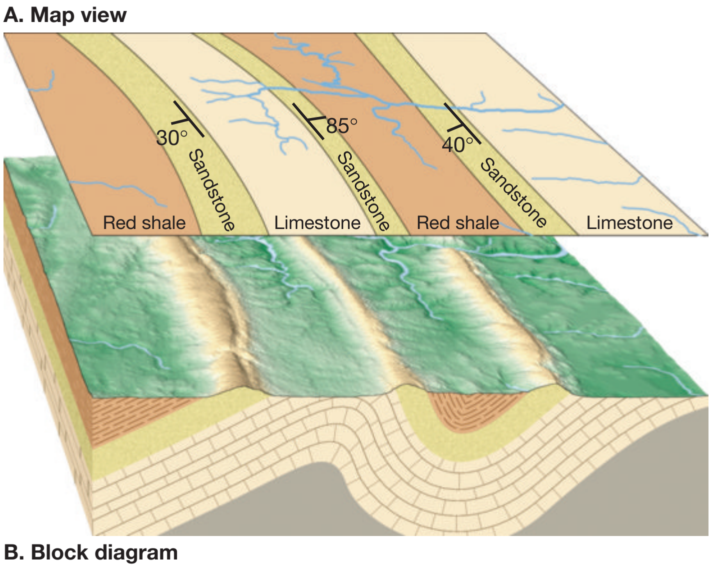


FIGURE 10.8 By establishing the strike and dip of outcropping sedimentary beds on a map (A), geologists can infer the orientation of the structure below ground (B).

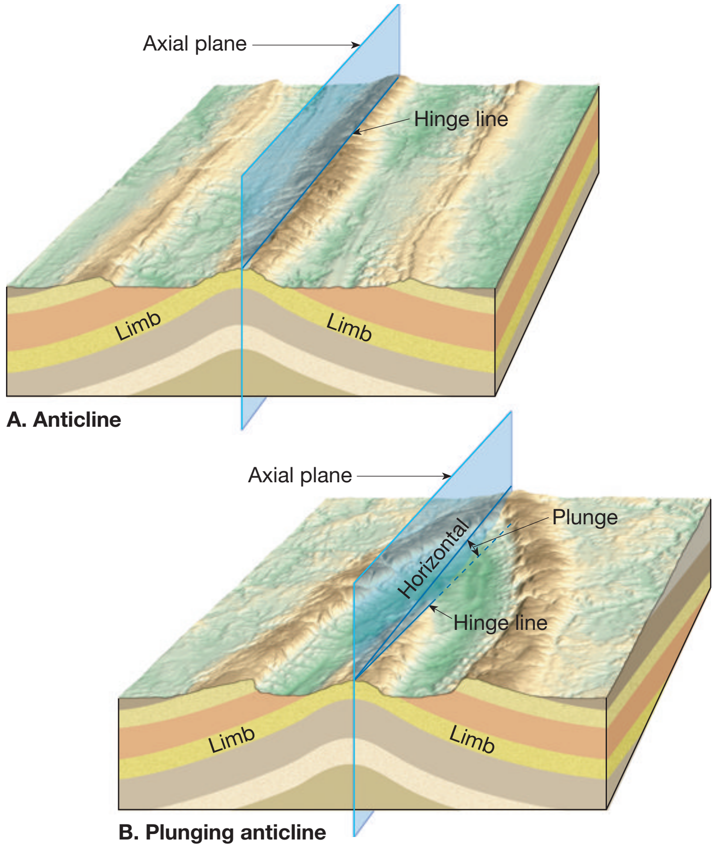


FIGURE 10.9 Idealized sketches illustrating the features associated with symmetrical folds. The hinge line of the fold in **A** is horizontal, whereas the hinge line of the fold in **B** is plunging.

sheet of paper and then push them together. In nature, folds come in a wide variety of sizes and configurations. Some folds are broad flexures in which rock units hundreds of meters thick have been slightly warped. Others are very tight microscopic structures found in metamorphic rocks. Size differences notwithstanding, most folds are the result of compressional stresses that result in the shortening and thickening of the crust. Occasionally, folds are found singly, but most often they occur as a series of undulations.

To aid our understanding of folds and folding, we need to become familiar with the terminology used to name the parts of a fold. As shown in Figure 10.9, the two sides of a fold are called *limbs.* A line drawn along the points of maximum curvature of each layer is termed the hinge line, or simply the *hinge.* In some folds, as Figure 10.9A illustrates, the hinge is horizontal, or parallel to the surface. However, in more complex folding, the hinge is often inclined at an angle known as the *plunge* (Figure 10.9B). Further, the *axial plane* is an imaginary surface that divides a fold as symmetrically as possible.

Types of Folds

The two most common types of folds are called anticlines and synclines (Figure 10.10). An **anticline** is most commonly formed by the upfolding, or arching, of rock layers.* Figure 10.9 is an example of an anticline. Anticlines are sometimes spectacularly displayed where highways have been cut through deformed strata. Often found in association with anticlines are downfolds, or troughs, called **synclines.** Notice in Figure 10.11 that the limb of an anticline is also a limb of the adjacent syncline.

Depending on their orientation, these basic folds are described as *symmetrical* when the limbs are mirror images of each other and *asymmetrical* when they are not. An asymmetrical fold is said to be *overturned* if one limb is tilted beyond the vertical (Figure 10.10). An overturned fold can also "lie on its side" so that a plane extending through the axis of the fold actually would be horizontal. These *recumbent* folds are common in mountainous regions such as the Alps (Figure 10.12).

Folds do not continue forever; rather, their ends die out much like the wrinkles in cloth. Some folds *plunge* because the axis of the fold penetrates into the ground (Figure 10.13A, B). As the figure shows, both anticlines and synclines can plunge. Figure 10.13C shows an example of a plunging anticline and the pattern produced when erosion removes the upper layers of the structure and exposes its interior. Note that the outcrop pattern of an anticline points in the direction it is plunging, whereas the opposite is true for a syncline. A good example of the kind of topography that results when erosional forces attack folded sedimentary strata is found in the Valley and Ridge Province of the Appalachians (see Figure 14.13).

*By strict definition, an anticline is a structure in which the oldest strata are found in the center. This most typically occurs when strata are upfolded. Further, a syncline is strictly defined as a structure in which the youngest strata are found in the center. This occurs most commonly when strata are downfolded.

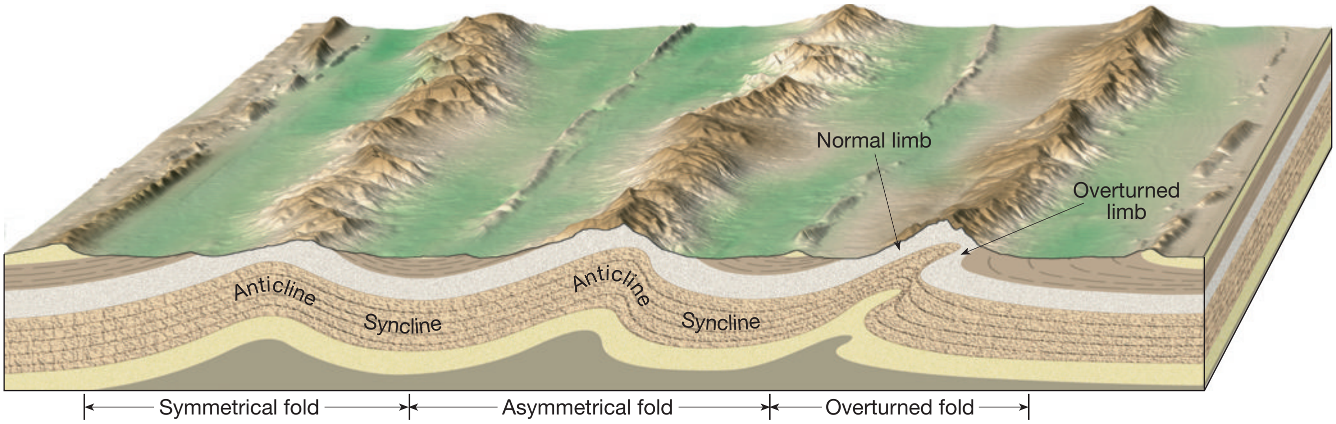
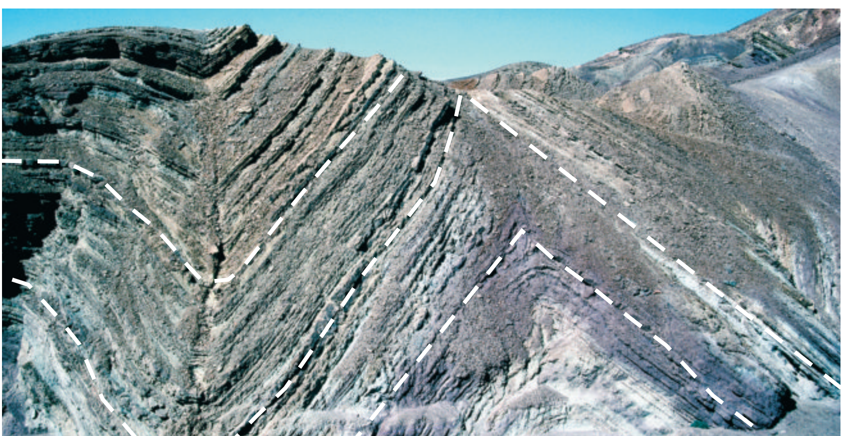


FIGURE 10.10 Block diagram of principal types of folded strata. The upfolded, or arched, structures are anticlines. The downfolds, or troughs, are synclines. Notice that the limb of an anticline is also the limb of the adjacent syncline.

It is important to understand that ridges are not necessarily associated with anticlines, nor are valleys related to synclines. Rather, ridges and valleys result because of differential weathering and erosion. For example, in the Valley and Ridge Province, resistant sandstone beds remain as imposing ridges separated by valleys cut into more easily eroded shale or limestone beds.

Although we have separated our discussion of folds and faults, in the real world folds are generally intimately coupled with faults. Examples of this close association are broad, regional features called *monoclines.* Particularly prominent features of the Colorado Plateau, **monoclines** (*mono* = one, *kleinen* = incline) are large, steplike folds in otherwise horizontal sedimentary strata (Figure 10.14). These folds appear to be the result of the reactivation of steeply dipping fault zones located in basement rocks beneath the plateau. As large blocks of basement rock were displaced upward along ancient faults, the comparatively ductile sedimentary strata above responded by folding. On the Colorado Plateau, monoclines display a narrow zone of steeply inclined beds that flatten out to form the uppermost layers of large elevated areas, including the Zuni Uplift, Echo Cliffs Uplift, and San Rafael Swell (Figure 10.14). Displacement along these reactivated faults often exceeds 1 kilometer (0.6 miles), and the very largest monoclines exhibit displacements that approach 3 kilometers (2 miles).

FIGURE 10.11 Syncline (left) and anticline (right) share a common limb. (Photo by E. J. Tarbuck).



Domes and Basins

Broad upwarps in basement rock may deform the overlying cover of sedimentary strata and generate large folds. When this upwarping produces a circular or elongated structure,

FIGURE 10.12 Recumbent folds in the Swiss Alps. (Photo by Mike Andrews)

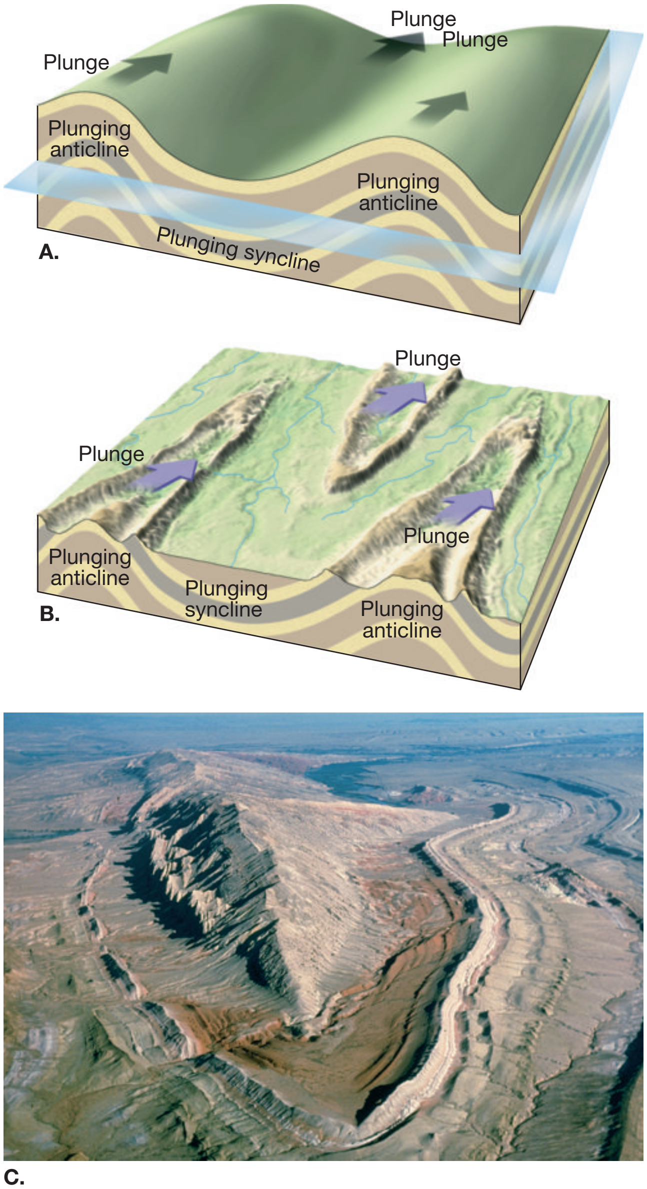


FIGURE 10.13 Plunging folds. **A.** Idealized view of plunging folds in which a horizontal surface has been added. **B.** View of plunging folds as they might appear after extensive erosion. Notice that in a plunging anticline the outcrop pattern "points" in the direction of the plunge, while the opposite is true of plunging synclines. **C.** Sheep Mountain, a doubly plunging anticline. (Photo by John S. Shelton)

the feature is called a **dome** (Figure 10.15A). Downwarped structures having a similar shape are termed **basins** (Figure 10.15B).

The Black Hills of western South Dakota is a large domed structure thought to be generated by upwarping. Here erosion has stripped away the highest portions of the unwarped sedimentary beds, exposing older igneous and metamorphic rocks in the center (Figure 10.16). Remnants of these once continuous sedimentary layers are visible, flanking the crystalline core of this mountain range. The more resistant strata are easy to identify because differential erosion has left them outcropping as prominent angular ridges called **hogbacks.** Because hogbacks can form whenever resistant strata are steeply inclined, they are also associated with other types of folds.

Domes can also be formed by the intrusion of magma (laccoliths) as shown in Figure 5.32. In addition, the upward migration of salt formations can produce salt domes that are common in the Gulf of Mexico.

Several large basins exist in the United States (Figure 10.17). The basins of Michigan and Illinois have very gently sloping beds similar to saucers. These basins are thought to be the result of large accumulations of sediment, whose weight caused the crust to subside (see section on isostasy in Chapter 14). A few structural basins may have been the result of giant asteroid impacts.

Because large basins contain sedimentary beds sloping at such low angles, they are usually identified by the age of the rocks composing them. The youngest rocks are found near the center, and the oldest rocks are at the flanks. This is just the opposite order of a domed structure, such as the Black Hills, where the oldest rocks form the core.

Faults

GEODe EARTH
Crustal Deformation
▸ **Faults and Fractures**

Faults are fractures in the crust along which appreciable displacement has taken place. Occasionally, small faults can be recognized in road cuts where sedimentary beds have been offset a few meters, as shown in Figure 10.18. Faults of this scale usually occur as single discrete breaks. By contrast, large faults, like the San Andreas Fault in California, have displacements of hundreds of kilometers and consist of many interconnecting fault surfaces. These *fault zones* can be several kilometers wide and are often easier to identify from high-altitude photographs than at ground level.

Students Sometimes Ask . . .

How do geologists determine which side of a fault has moved?

Surprisingly, for many faults it cannot be definitively established. For example, in the picture of a fault shown in Figure 10.18, did the left side move down, or did the right side move up? Since the surface (at the top of the photo) has been eroded flat, either side could have moved; or both could have moved, with one side moving more than the other (for instance, they both may have moved up, it's just that the right side moved up more than the left side). That's why geologists talk about *relative* motion across faults. In this case, the left side moved down *relative to* the right side, and the right side moved up *relative to* the left side (note arrows on photo).

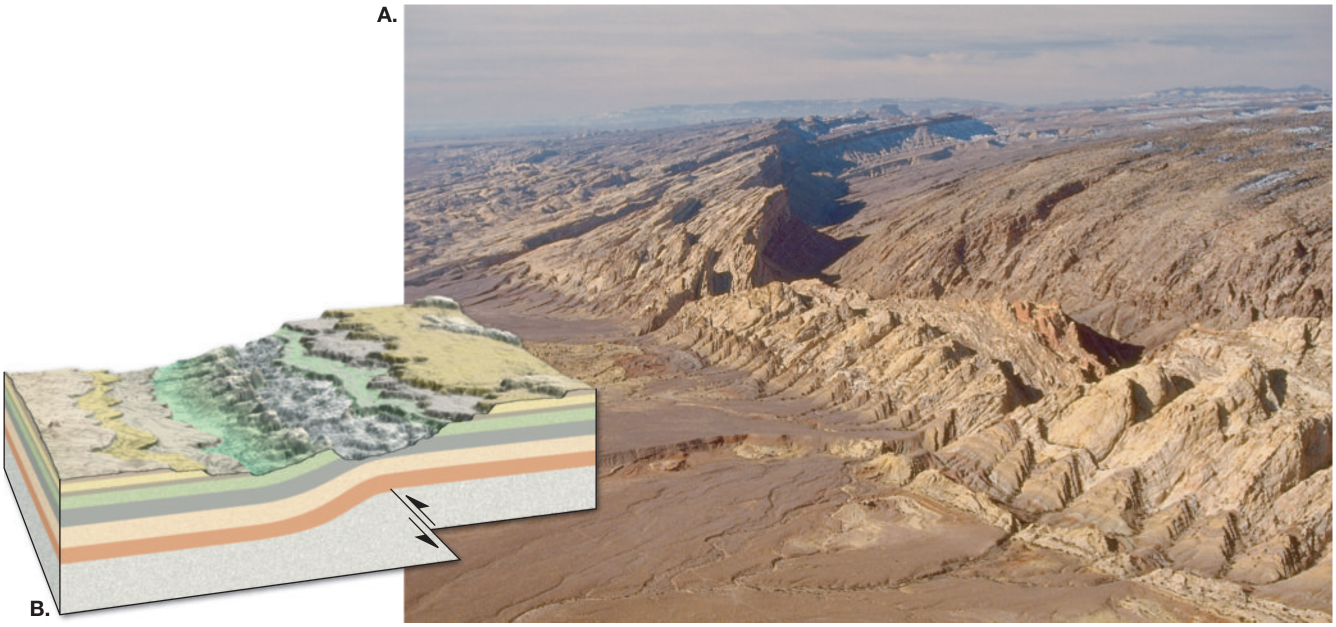


FIGURE 10.14 Monocline. **A.** The San Rafael monocline, Utah. (Photo by Stephen Trimble) **B.** Monocline consisting of bent sedimentary beds that were deformed by faulting in the bedrock below. The thrust fault in this diagram is called a *blind thrust* because it does not reach the surface.

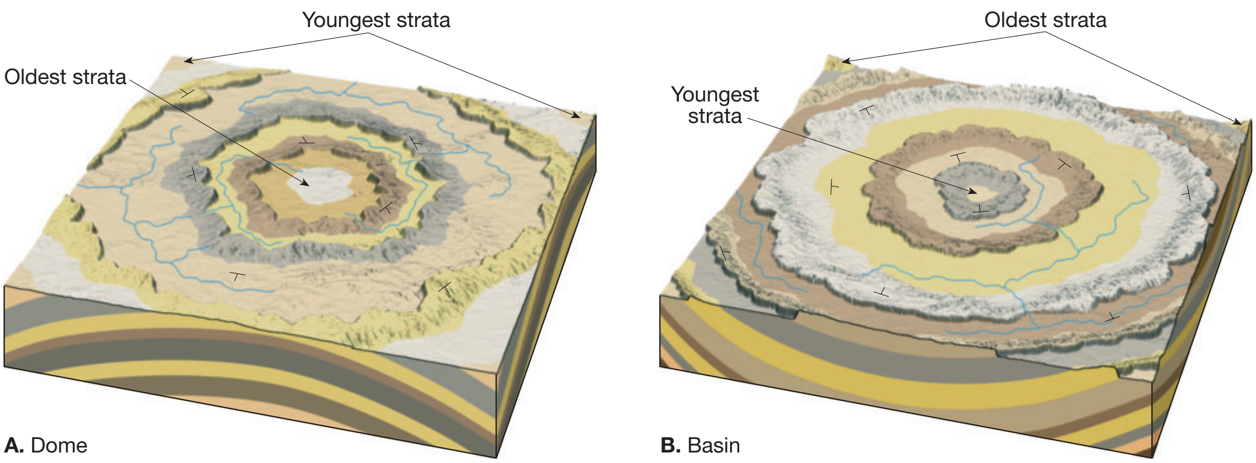
Sudden movements along faults are the cause of most earthquakes. However, the vast majority of faults are inactive and thus are remnants of past deformation. The rock along faults often becomes broken and pulverized as crustal blocks on opposite sides of a fault grind past one another. The loosely coherent, clayish material that results from this activity is called *fault gouge.* On some fault surfaces the rocks become highly polished and striated, or grooved, as the crustal blocks slide past one another. These polished and striated surfaces, called *slickensides* (*slicken* = smooth), provide geologists with evidence for the direction of the most recent displacement along the fault. Geologists classify faults by these relative movements, which can be predominantly horizontal, vertical, or oblique.

Dip-Slip Faults

Faults in which the movement is primarily parallel to the dip (or inclination) of the fault surface are called **dip-slip faults.** Vertical displacements along dip-slip faults may produce long, low cliffs called **fault scarps** (*scarpe* = a slope). Fault scarps, such as the one shown in Figure 10.19, are produced by displacements that generate earthquakes.

It has become common practice to call the rock surface that is immediately above the fault the *hanging wall* and to call the rock surface below, the *footwall* (Figure 10.20). This nomenclature arose from prospectors and miners who excavated shafts and tunnels along fault zones because these are frequently sites of ore deposits. In these tunnels, the miners would walk

FIGURE 10.15 Gentle upwarping and downwarping of crustal rocks produce domes (A) and basins (B). Erosion of these structures results in an outcrop pattern that is roughly circular or elongate.



A. Dome **B.** Basin

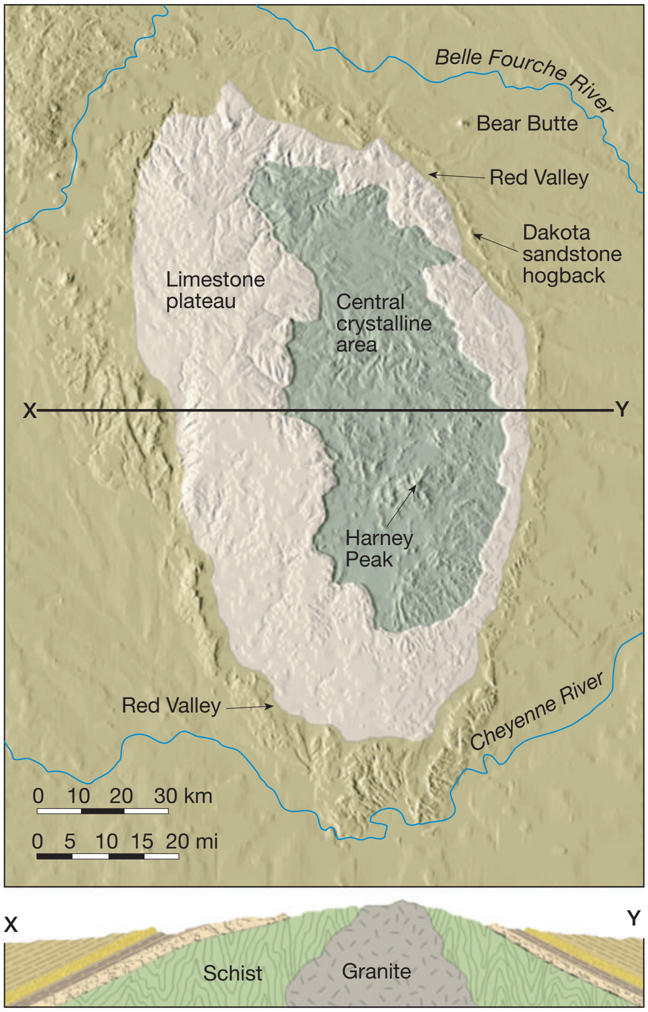


FIGURE 10.16 The Black Hills of South Dakota, a large domal structure with resistant igneous and metamorphic rocks exposed in the core.

on the rocks below the mineralized fault zone (the footwall) and hang their lanterns on the rocks above (the hanging wall).

Two major types of dip-slip faults are **normal faults** and **reverse faults.** In addition, when a reverse fault has an angle of dip (inclination) less than 45°, it is called a *thrust fault.* We will consider these types of dip-slip faults next.

Normal Faults Dip-slip faults are classified as normal faults when the hanging wall block moves down relative to the footwall block (Figure 10.21). Most normal faults have steep dips of about 60°, which tend to flatten out with depth. However, some dip-slip faults have much lower dips, with some approaching horizontal. Because of the downward motion of the hanging wall, normal faults accommodate lengthening, or extension, of the crust.

Most normal faults are small, having displacements of only a meter or so, like the one shown in the road cut in Figure 10.18. Others extend for tens of kilometers where they may sinuously trace the boundary of a mountain front. In the western United States, large-scale normal faults like these are associated with structures called **fault-block mountains.**

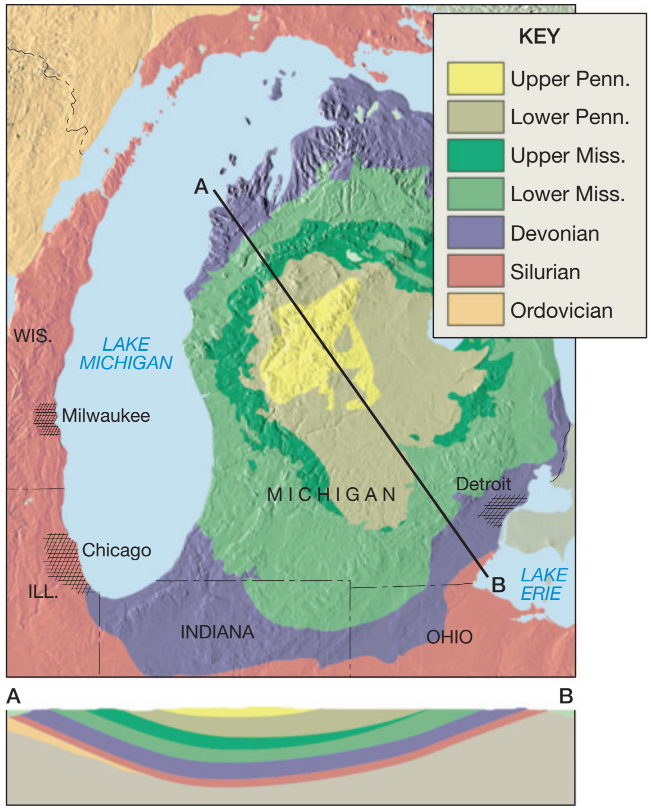


FIGURE 10.17 The bedrock geology of the Michigan Basin. Notice that the youngest rocks are centrally located, while the oldest beds flank this structure.

Examples of fault-block mountains include the Teton Range of Wyoming and the Sierra Nevada of California. Both are faulted along their eastern flanks, which were uplifted as the blocks tilted downward to the west. These precipitous mountain fronts were produced over a period of 5 million to 10 million years by many irregularly spaced

FIGURE 10.18 Faulting caused the vertical displacement of these beds located near Kanab, Utah. Arrows show relative motion of rock units. (Photo by Tom Bean/DRK Photo)

FIGURE 10.19 A fault scarp located near Joshua Tree National Monument, California. (Photo by A. P. Trujillo/APT Photos)

episodes of faulting. Each event was responsible for just a few meters of displacement.

Other excellent examples of fault-block mountains are found in the Basin and Range Province, a region that encompasses Nevada and portions of the surrounding states (Figure 10.22). Here the crust has been elongated and broken to create more than 200 relatively small mountain ranges. Averaging about 80 kilometers in length, the ranges rise 900 to 1500 meters above the adjacent down-faulted basins.

The topography of the Basin and Range Province has been generated by a system of roughly north to south trending normal faults. Movements along these faults have produced alternating uplifted fault blocks called **horsts** and down-dropped blocks called **grabens** (*graben* = ditch). Horsts generate elevated ranges, whereas grabens form many of the basins. As Figure 10.22 illustrates, structures called **half-grabens,** which are titled fault blocks, also contribute to the alternating topographic highs and lows in the Basin and Range Province. The horsts and higher ends of

FIGURE 10.20 The rock immediately above a fault surface is the *hanging wall*, and that below is called the *footwall.* These names came from miners who excavated ore along fault zones. The miners hung their lanterns on the rocks above the fault trace (hanging wall) and walked on the rocks below the fault trace (footwall).

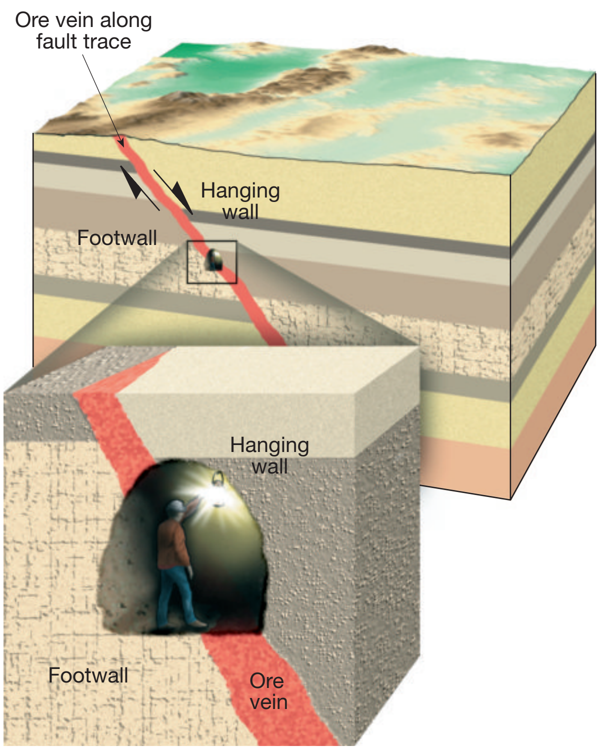
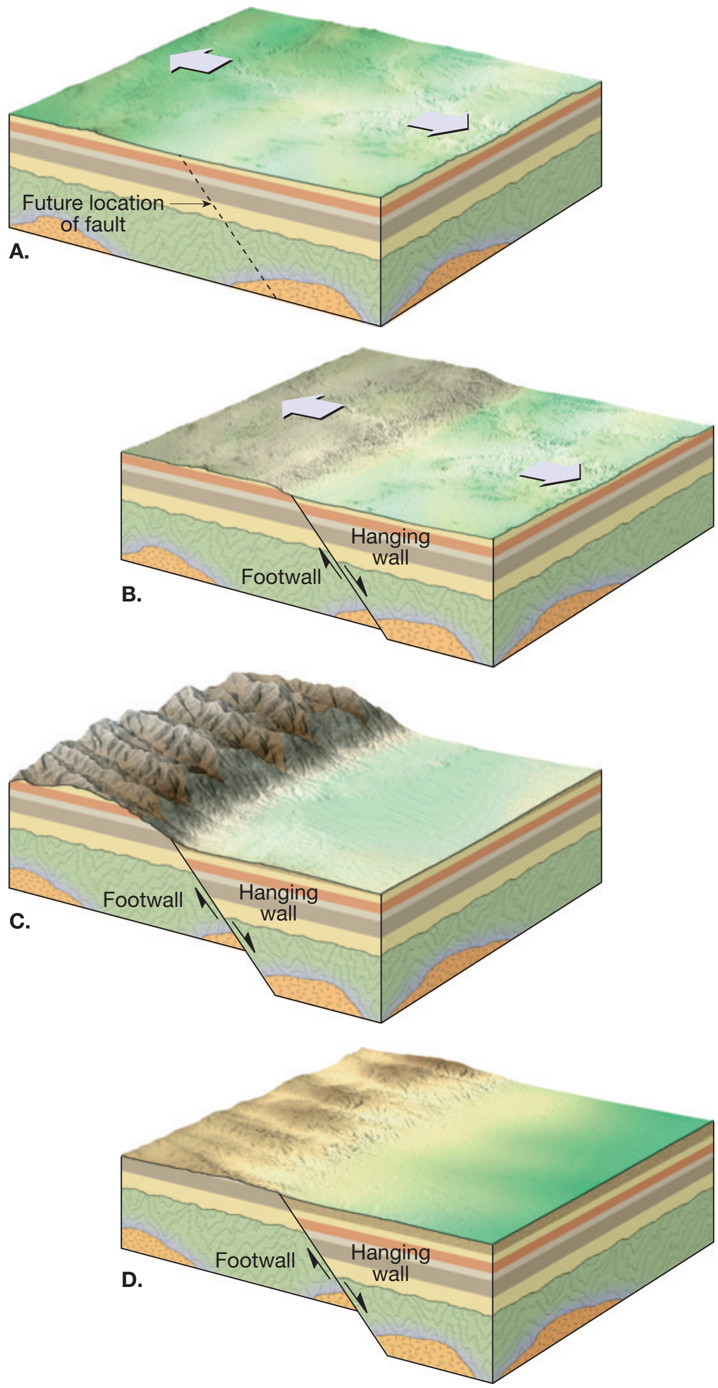


FIGURE 10.21 Block diagrams illustrating a normal fault. **A.** Rock strata prior to faulting. **B.** The relative movement of displaced blocks. Displacement may continue in a fault-block mountain range over millions of years and consist of many widely spaced episodes of faulting. **C.** How erosion might alter the upfaulted block. **D.** Eventually the period of deformation ends and erosion becomes the dominant geologic process.



the tilted fault blocks are the source for sediments that have accumulated in the basins that were created by the grabens and lower ends of the tilted blocks.

Notice in Figure 10.22 that the slopes of the normal faults in the Basin and Range Province decrease with depth and eventually join together to form a nearly horizontal fault called a **detachment fault.** These faults may extend for hundreds of kilometers below the surface. Here they form a major boundary between the rocks below, which exhibit ductile deformation, and the rocks above, which demonstrate brittle deformation via faulting.

Normal faulting is also prevalent at spreading centers where plate divergence occurs. Here, a central block (graben) is bounded by normal faults and subsides as the plates separate. These grabens produce an elongated valley bounded by uplifted fault blocks (horsts).

The Rift Valley of East Africa is made up of several large grabens, above which tilted horsts produce a linear mountainous topography. This valley, nearly 6000 kilometers (3600 miles) long, contains the excavation sites of some of the earliest human fossils. Examples of inactive rift valleys include the Rhine Valley in Germany and the Triassic-age grabens of the eastern United States. Even larger inactive normal fault systems are the rifted continental margins, such as along the east coasts of the Americas and the west coasts of Europe and Africa.

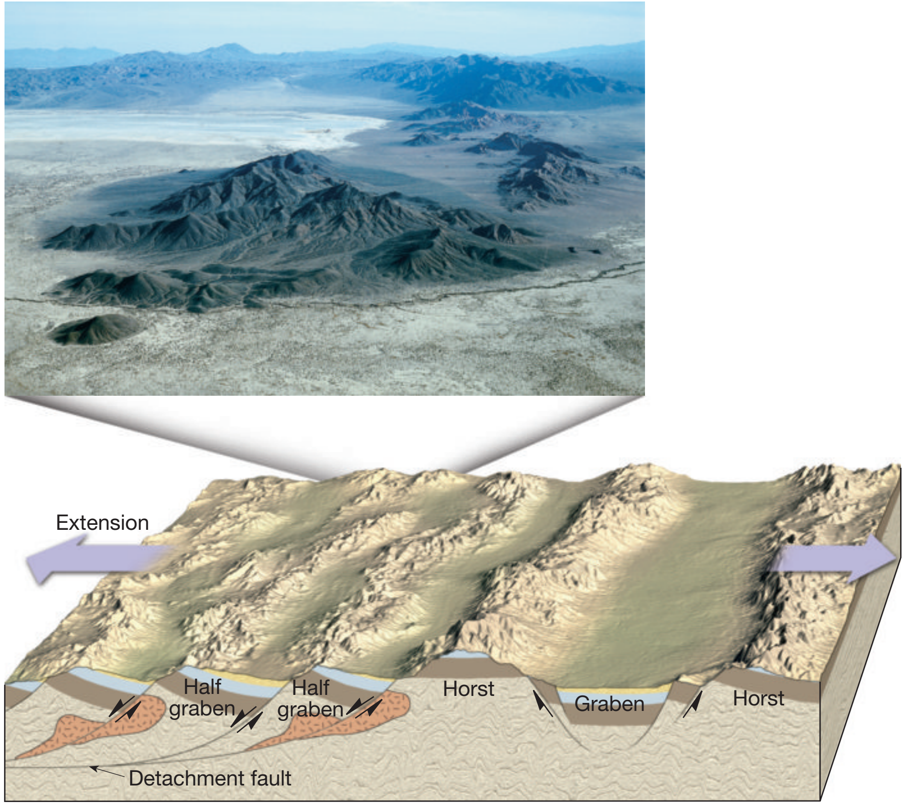


FIGURE 10.22 Normal faulting in the Basin and Range Province. Here, tensional stresses have elongated and fractured the crust into numerous blocks. Movement along these fractures has tilted the blocks, producing parallel mountain ranges called fault-block mountains. The downfaulted blocks (grabens) form basins, whereas the upfaulted blocks (horsts) are eroded to form rugged mountainous topography. In addition, numerous tilted blocks (half-grabens) form both basins and mountains. (Photo by Michael Collier)

Fault motion provides geologists with a method of determining the nature of the forces at work within Earth. Normal faults indicate the existence of extensional forces that pull the crust apart. This "pulling apart" can be accomplished either by uplifting that causes the surfaces to stretch and break or by opposing horizontal forces.

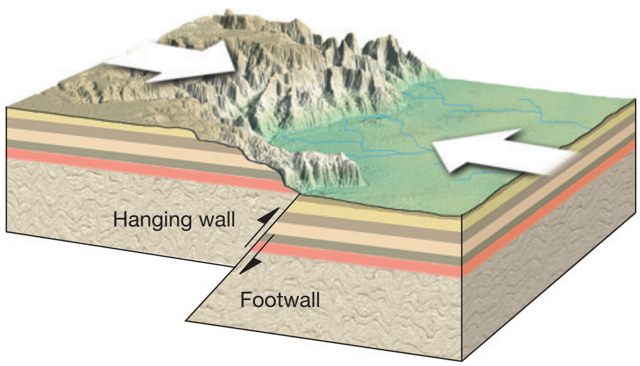
Reverse and Thrust Faults Reverse faults and **thrust faults** are dip-slip faults in which the hanging wall block moves up relative to the footwall block (Figure 10.23). Recall that reverse faults have dips greater than 45° and thrust faults have dips less than 45°. Because the hanging wall block moves up and over the footwall block, reverse and thrust faults accommodate horizontal shortening of the crust.

Most high-angle reverse faults are small and accommodate local displacements in regions dominated by other types of faulting. Thrust faults, on the other hand, exist at all scales. Small thrust faults exhibit displacement on the order of millimeters to a few meters. Some large thrust faults have displacements on the order of tens to hundreds of kilometers.

Whereas normal faults occur in tensional environments, thrust faults result from strong compressional stresses. In these settings, crustal blocks are displaced *toward* one another, with the hanging wall being displaced upward relative to the footwall. Thrust faulting is most pronounced in subduction zones and other convergent boundaries where plates are colliding. Compressional forces generally produce folds as well as faults and result in a thickening and shortening of the material involved.

In mountainous regions such as the Alps, Northern Rockies, Himalayas, and Appalachians, thrust faults have displaced

FIGURE 10.23 Block diagram showing the relative movement along a reverse fault.



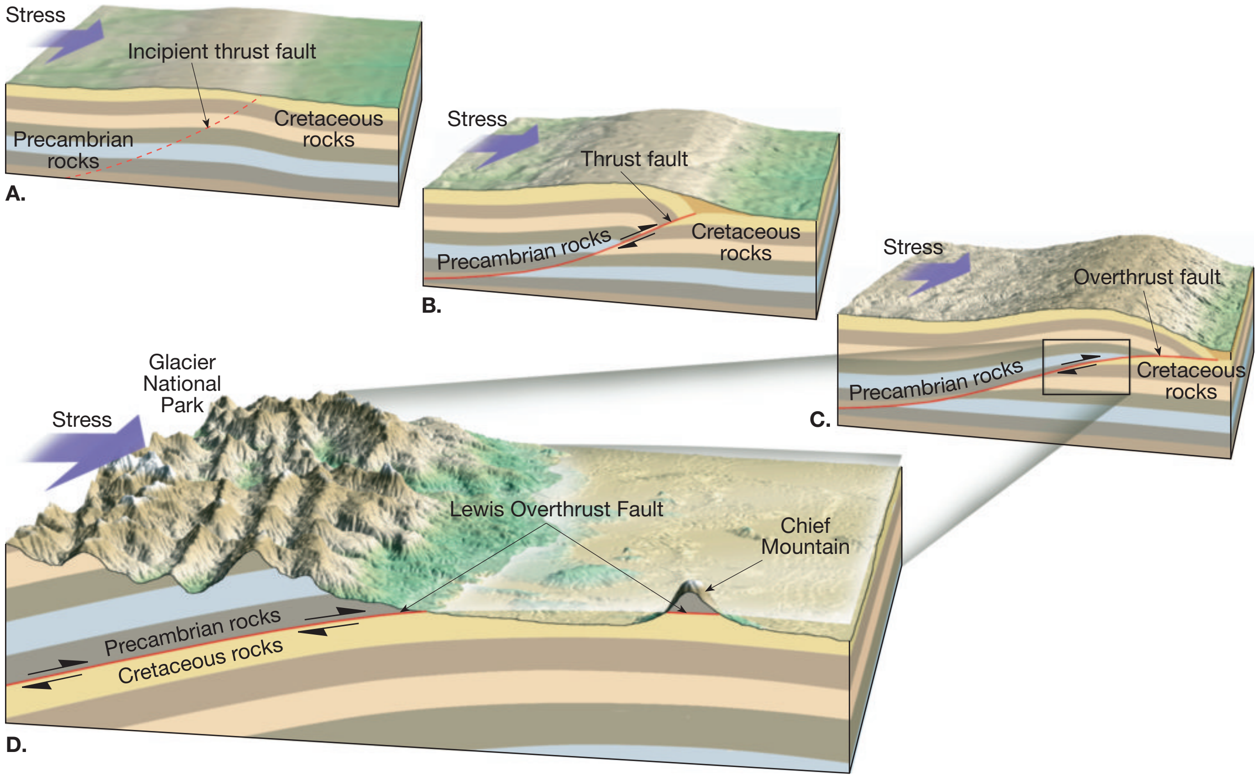


FIGURE 10.24 Idealized development of Lewis Overthrust fault. **A.** Geologic setting prior to deformation. **B., C.** Large-scale movement along a thrust fault displaced Precambrian rock over Cretaceous strata in the region of Glacier National Park. **D.** Erosion by glacial ice and running water sculptured the thrust sheet into a majestic landscape and isolated a remnant of the thrust sheet called Chief Mountain.

strata as far as 50 kilometers over adjacent rock units. The result of this large-scale movement is that older strata end up overlying younger rocks. A classic site of thrust faulting occurs in Glacier National Park (Figure 10.24). Here, mountain peaks that provide the park's majestic scenery have been carved from Precambrian rocks that were displaced over much younger Cretaceous strata. At the eastern edge of Glacier National Park, there is an outlying peak called Chief Mountain. This structure is an isolated remnant of the thrust sheet that was severed by the erosional forces of glacial ice and running water. An isolated block, such as Chief Mountain, is called a **klippe** (*klippe* = cliff) (Figure 10.25).

Strike-Slip Faults

Faults in which the dominant displacement is horizontal and parallel to the strike of the fault surface are called **strike-slip faults.** Because of their large size and linear nature, many strike-slip faults produce a trace that is visible over a great distance (Figure 10.26). Rather than a single fracture along which movement takes place, large strike-slip faults consist of a zone of roughly parallel fractures. The zone may be up to several kilometers in width. The most recent movement, however, is often along a strand only a few meters wide, which may offset features such as stream channels (Figure 10.27). Furthermore,

FIGURE 10.25 Chief Mountain, Glacier National Park, Montana, is a klippe. (Photo by David Muench)

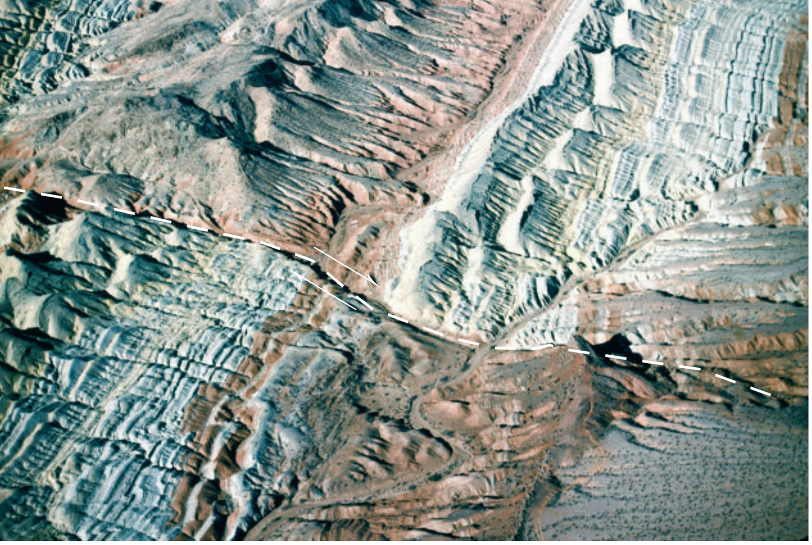


FIGURE 10.26 Aerial view of strike-slip (right-lateral) fault in southern Nevada. Notice that the ridge of white rock in the top right portion of the photo was offset to the right relative to the portion of the same ridge that appears on the bottom left of the image. (Photo by Marli Miller)

Students Sometimes Ask . . .

Has anyone ever seen a fault scarp forming?

Amazingly, yes. There have been several instances where people have been at the fortuitously appropriate place and time to observe the creation of a fault scarp—and have lived to tell about it. In Idaho a large earthquake in 1983 created a 3-meter (10-foot) fault scarp that was witnessed by several people, many of whom were knocked off their feet. More often, though, fault scarps are noticed *after* they form. For example, a 1999 earthquake in Taiwan created a fault scarp that formed a new waterfall and destroyed a nearby bridge.

crushed and broken rocks produced during faulting are more easily eroded, often producing linear valleys or troughs that mark the locations of strike-slip faults.

The earliest scientific records of strike-slip faulting were made following surface ruptures that produced large earthquakes. One of the most noteworthy of these was the great San Francisco earthquake of 1906. During this strong earthquake, structures such as fences that were built across the San Andreas Fault were displaced as much as 4.7 meters (15 feet). Because the movement along the San Andreas causes the crustal block on the opposite side of the fault to move to the right as you face the fault, it is called a *right-lateral* strike-slip fault. The Great Glen fault in Scotland is a well-known example of a *left-lateral* strike-slip fault, which exhibits the opposite sense of displacement. The total displacement along the Great Glen fault is estimated to exceed 100 kilometers (60 miles). Also associated with this fault trace are numerous lakes, including Loch Ness, the home of the legendary monster.

Many major strike-slip faults cut through the lithosphere and accommodate motion between two large crustal plates. Recall that this special kind of strike-slip fault is called a **transform** (*trans* = across, *forma* = form) **fault.** Numerous transform faults cut the oceanic lithosphere and link spreading oceanic ridges. Others accommodate displacement between continental plates that move horizontally with respect to each other. One of the best-known transform faults is California's San Andreas Fault (see Box 10.2). This plate-bounding fault can be traced for about 950 kilometers (600 miles) from the Gulf of California to a point along the Pacific Coast north of San Francisco, where it heads out to sea. Ever since its formation, about 29 million years ago, displacement along the San Andreas Fault has exceeded 560 kilometers. This movement has accommodated the northward displacement of southwestern California and the Baja Peninsula of Mexico in relation to the remainder of North America.

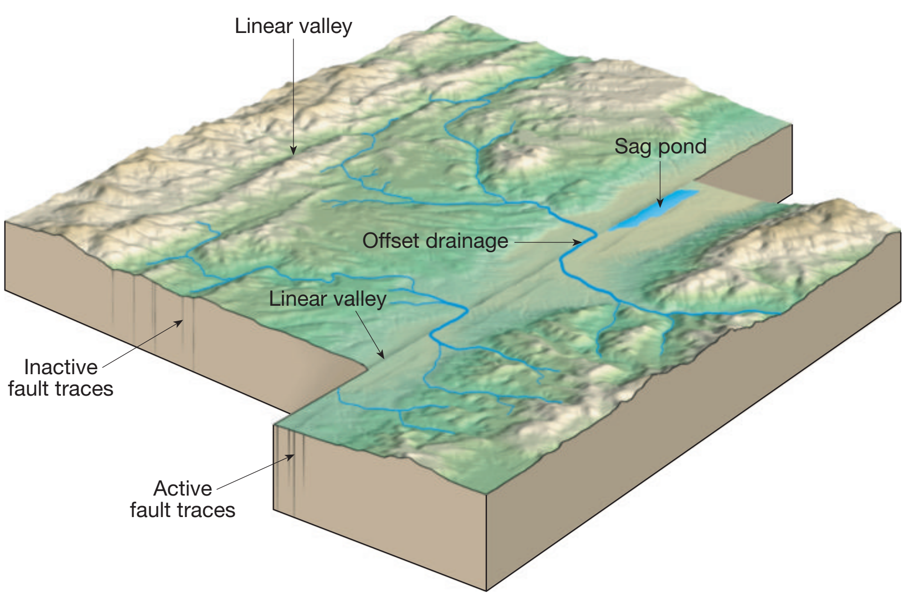
Joints

Crustal Deformation
▸ Faults and Fractures

Among the most common rock structures are fractures called joints. Unlike faults, **joints** are fractures along which no appreciable displacement has occurred. Although some joints have a random orientation, most occur in roughly parallel groups (see Figure 6.6).

We have already considered two types of joints. Earlier we learned that *columnar joints* form when igneous rocks cool and develop shrinkage fractures that produce elongated, pillarlike columns (Figure 10.28). Also recall that sheeting produces a

FIGURE 10.27 Block diagram illustrating the features associated with strike-slip faults. Note how the stream channels have been offset by fault movement. The faults in this diagram are right-lateral strike-slip faults. (Modified after R. L. Wesson and others)



Students Sometimes Ask . . .

Do faults exhibit only strike-slip or dip-slip motion?

No. Strike-slip faults and dip-slip faults are on the opposite ends of a spectrum of fault structures. Faults that exhibit a combination of dip-slip and strike-slip movements are called *oblique-slip faults.* Although most faults could technically be classified as oblique-slip, they predominately exhibit either strike-slip or dip-slip motion.

pattern of gently curved joints that develop more or less parallel to the surface of large exposed igneous bodies such as batholiths. Here the jointing results from the gradual expansion that occurs when erosion removes the overlying load.

In contrast to the situations just described, most joints are produced when rocks in the outermost crust are deformed. Here tensional and shearing stresses associated with crustal movements cause the rock to fail by brittle fracture. For example, when folding occurs, rocks situated at the axes of the folds are elongated and pulled apart to produce tensional joints. Extensive joint patterns can also develop in response to relatively subtle and often barely perceptible regional upwarping and downwarping of the crust. In many cases, the cause for jointing at a particular locale is not readily apparent.

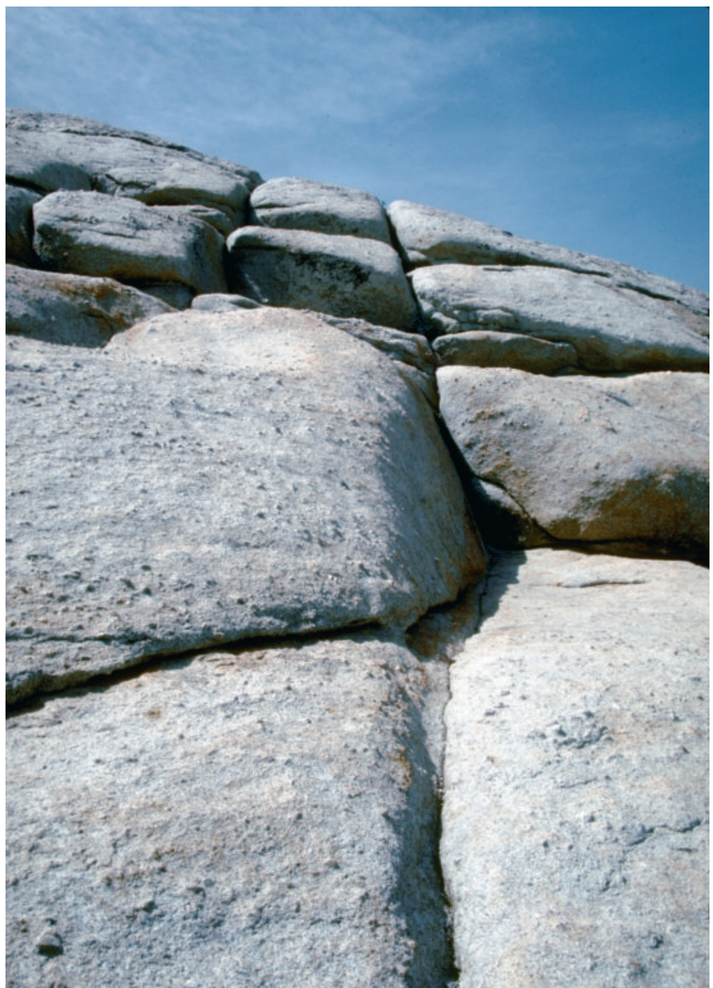
Many rocks are broken by two or even three sets of intersecting joints that slice the rock into numerous regularly shaped blocks. These joint sets often exert a strong influence on other geologic processes. For example, chemical weathering tends to be concentrated along joints, and in many areas groundwater movement and the resulting dissolution in soluble rocks is controlled by the joint pattern (Figure 10.29). Moreover, a system of joints can influence the direction that stream courses follow. The rectangular drainage pattern described in Chapter 16 is such a case.

Joints may also be significant from an economic standpoint. Some of the world's largest and most important mineral deposits were emplaced along joint systems. Hydrothermal solutions, which are basically mineralized fluids, can migrate into fractured host rocks and precipitate economically important amounts of copper, silver, gold, zinc, lead, and uranium.

Further, highly jointed rocks present a risk to the construction of engineering projects, including highways and dams. On June 5, 1976, 14 lives were lost and nearly 1 billion dollars in property damage occurred when the Teton Dam in Idaho

FIGURE 10.28 Devil's Tower National Monument is in eastern Wyoming. Established in September 1906, it was our nation's first national monument. This nearly vertical monolith rises more than 380 meters (1265 feet) above the surrounding grasslands and pine forests. This imposing structure exhibits columnar joints and the columns that result. Columnar joints form as igneous rocks cool and develop shrinkage fractures that produce five- to seven-sided elongated, pillarlike structures. (Photo by Bob Thomason/Tony Stone Images)

FIGURE 10.29 Chemical weathering is enhanced along joints in granitic rocks near the top of Lembert Dome in Yosemite National Park, California. (Photo by E. J. Tarbuck)



BOX 10.2 ▶ PEOPLE AND THE ENVIRONMENT

The San Andreas Fault System

The San Andreas, the best-known and largest fault system in North America, first attracted wide attention after the great 1906 San Francisco earthquake and fire. Following this devastating event, geologic studies demonstrated that a displacement of as much as 5 meters along the fault had been responsible for the earthquake. It is now known that this dramatic event is just one of many thousands of earthquakes that have resulted from repeated movements along the San Andreas throughout its 29-million-year history.

Where is the San Andreas fault system located? As shown in Figure 10.B, it trends in a northwesterly direction for nearly 1300 kilometers (780 miles) through much of western California. At its southern end, the San Andreas connects with a spreading center located in the Gulf of California. In the north, the fault enters the Pacific Ocean at Point Arena, where it is thought to continue its northwesterly trend, eventually joining the Mendocino fracture zone. In the central section, the San Andreas is relatively simple and straight. However, at its two extremities, several branches spread from the main trace, so that in some areas the fault zone exceeds 100 kilometers (60 miles) in width.

Over much of its extent, a linear trough reveals the presence of the San Andreas Fault. When the system is viewed from the air, linear scars, offset stream channels, and elongated ponds mark the trace in a striking manner. On the ground, however, surface expressions of the faults are much more difficult to detect. Some of the most distinctive landforms include long, straight escarpments, narrow ridges, and sag ponds

FIGURE 10.B Map showing the extent of the San Andreas Fault system. Inset shows only a few of the many splinter faults that are part of this great fault system.

formed by settling of blocks within the fault zone. Further, many stream channels characteristically bend sharply to the right where they cross the fault (Figure 10.C).

With the advent of the theory of plate tectonics, geologists began to realize the significance of this great fault system. The San Andreas Fault is a transform boundary separating two crustal plates that move very slowly. The Pacific plate, located to the west, moves northwestward relative to the North American plate, causing earthquakes along the fault (Table 10.A).

The San Andreas is undoubtedly the most studied of any fault system in the world. Although many questions remain unanswered, geologists have learned that each fault segment exhibits somewhat different behavior. Some portions of the San Andreas exhibit a slow creep with little noticeable seismic activity. Other segments regularly slip, producing small earthquakes, while still other segments may store elastic energy for 200 years or more before rupturing to generate a great earthquake. This knowledge is useful when assigning earthquake hazard potential to a given segment of the fault zone.

Because of the great length and complexity of the San Andreas Fault, it is more appropriately referred to as a "fault system." This major fault system consists primarily of the San Andreas Fault and several major branches, including the Hayward and Calaveras faults of central California and the San Jacinto and Elsinore faults of southern California (Figure 10.B). These major segments, plus a vast number of smaller faults that include the Imperial Fault, San Fernando Fault, and the Santa Monica Fault, collectively accommodate the relative motion between the North American and Pacific plates.

Blocks on opposite sides of the San Andreas Fault move horizontally in opposite directions, such that if a person stood on one side of the fault, the block on the opposite side would appear to move to the right when slippage occurred. This type of displacement is known as *right-lateral strike-slip* by geologists (Figure 10.C).

Ever since the great San Francisco earthquake of 1906, when as much as 5 meters of displacement occurred, geologists have attempted to establish the cumulative displacement along this fault over its 29-million-year history. By matching rock units across the fault, geologists have determined that the total accumulated displacement from earthquakes and creep exceeds 560 kilometers (340 miles).

FIGURE 10.C Aerial view showing offset stream channel across the San Andreas Fault on the Carrizo Plain west of Taft, California. (Photo by Michael Collier/DRK Photo)

TABLE 10.A Major Earthquakes on the San Andreas Fault System

Date	Location	Magnitude	Remarks
1812	Wrightwood, CA	7	Church of San Juan Capistrano collapsed, killing 40 worshippers.
1812	Santa Barbara channel	7	Churches and other buildings wrecked in and around Santa Barbara.
1838	San Francisco peninsula	7	At one time thought to have been comparable to the great earthquake of 1906.
1857	Fort Tejon, CA	8.25	One of the greatest U.S. earthquakes. Occurred near Los Angeles, then a city of 4000.
1868	Hayward, CA	7	Rupture of the Hayward fault caused extensive damage in San Francisco Bay area.
1906	San Francisco, CA	8.25	The great San Francisco earthquake. As much as 80 percent of the damage caused by fire.
1940	Imperial Valley	7.1	Displacement on the newly discovered Imperial fault.
1952	Kern County	7.7	Rupture of the White Wolf fault. Largest earthquake in California since 1906. Sixty million dollars in damages and 12 people killed.
1971	San Fernando Valley	6.5	One-half billion dollars in damages and 58 lives claimed.
1989	Santa Cruz Mountains	7.1	Loma Prieta earthquake. Six billion dollars in damages, 62 lives lost, and 3757 people injured.
1994	Northridge (Los Angeles area)	6.9	Over 15 billion dollars in damages, 51 lives lost, and over 5000 injured.

failed. This earthen dam was constructed of very erodible clays and silts and was situated on highly fractured volcanic rocks. Although attempts were made to fill the voids in the jointed rock, water gradually penetrated the subsurface fractures and undermined the dam's foundation. Eventually the moving water cut a tunnel into the easily erodible clays and silts. Within minutes the dam failed, sending a 20-meter-high wall of water down the Teton and Snake rivers.

Summary

- *Deformation* refers to changes in the shape and/or volume of a rock body and is most pronounced along plate margins. To describe the forces that deform rocks, geologists use the term *stress,* which is the amount of force applied to a given area. Stress that is uniform in all directions is called *confining pressure,* whereas *differential stresses* are applied unequally in different directions. Differential stresses that shorten a rock body are *compressional stresses;* those that elongate a rock unit are *tensional stresses. Strain* is the change in size and shape of a rock unit caused by stress.
- Rocks deform differently depending on the environment (temperature and confining pressure), the composition and texture of the rock, and the length of time stress is maintained. Rocks first respond by deforming *elastically* and will return to their original shape when the stress is removed. Once their elastic limit (strength) is surpassed, rocks either deform by ductile flow or they fracture. *Ductile deformation* is a solid-state flow that results in a change in size and shape of an object without fracturing. Ductile flow may be accomplished by gradual slippage and recrystallization along planes of weakness within the crystal lattice of mineral grains. Ductile deformation occurs in a high-temperature/high-pressure environment. In a near-surface environment, most rocks deform by *brittle failure.*
- The orientation of rock units or fault surfaces is established with measurements called strike and dip. *Strike* is the compass direction of a line produced by the intersection of an inclined rock layer or fault with a horizontal plane. *Dip* is the angle of inclination of the surface of a rock unit or fault measured from a horizontal plane.
- The most basic geologic structures associated with rock deformation are *folds* (flat-lying sedimentary and volcanic rocks bent into a series of wavelike undulations) and *faults.* The two most common types of folds are *anticlines,* formed by the upfolding, or arching, of rock layers, and *synclines,* which are downfolds. Most folds are the result of horizontal *compressional stresses.* Folds can be *symmetrical, asymmetrical,* or, if one limb has been tilted beyond the vertical, *overturned. Domes* (upwarped structures) and *basins* (downwarped structures) are circular or somewhat elongated folds formed by vertical displacements of strata.
- Faults are fractures in the crust along which appreciable displacement has occurred. Faults in which the movement is primarily vertical are called *dip-slip faults.* Dip-slip faults include both *normal* and *reverse faults.* Low-angle reverse faults are called *thrust faults.* Normal faults indicate *tensional stresses* that pull the crust apart. Along the spreading centers of plates, divergence can cause a central block called a *graben,* bounded by normal faults, to drop as the plates separate.
- Reverse and thrust faulting indicate that *compressional forces* are at work. Large *thrust faults* are found along subduction zones and other convergent boundaries where plates are colliding. In mountainous regions such as the Alps, Northern Rockies, Himalayas, and Appalachians, thrust faults have displaced strata as far as 50 kilometers over adjacent rock units.
- *Strike-slip faults* exhibit mainly horizontal displacement parallel to the strike of the fault surface. Large strike-slip faults, called *transform faults,* accommodate displacement between plate boundaries. Most transform faults cut the oceanic lithosphere and link spreading centers. The San Andreas Fault cuts the continental lithosphere and accommodates the northward displacement of southwestern California.
- *Joints* are fractures along which no appreciable displacement has occurred. Joints generally occur in groups with roughly parallel orientations and are the result of brittle failure of rock units located in the outermost crust.

Review Questions

1. What is rock deformation? How does a rock body change during deformation?
2. List five (5) geologic structures that are associated with deformation.
3. How is *stress* related to *force?*
4. Contrast compressional and tensional stresses.
5. Describe how shearing can deform a rock in a near-surface environment.
6. Compare strain and stress.
7. How is brittle deformation different from ductile deformation?
8. List three factors that determine how rocks will behave when exposed to stresses that exceed their strength. Briefly explain the role of each.
9. What is an outcrop?
10. What two measurements are used to establish the orientation of deformed strata? Distinguish between them.

11. Distinguish between anticlines and synclines, domes and basins, anticlines and domes.
12. How is a monocline different from an anticline?
13. The Black Hills of South Dakota are a good example of what type of structural feature?
14. Contrast the movements that occur along normal and reverse faults. What type of stress is indicated by each fault?
15. Is the fault shown in Figure 10.18 a normal or a reverse fault?
16. Describe a horst and a graben. Explain how a graben valley forms and name one.
17. What type of faults are associated with fault-block mountains?
18. How are reverse faults different from thrust faults? In what way are they the same?
19. The San Andreas Fault is an excellent example of a _____ fault.
20. With which of the three types of plate boundaries does normal faulting predominate? Reverse faulting? Strike-slip faulting?
21. How are joints different from faults?

Key Terms

anticline (p. 280)
basin (p. 282)
brittle failure (p. 277)
brittle deformation (p. 277)
compressional stress (p. 275)
deformation (p. 275)
detachment fault (p. 286)
differential stress (p. 275)
dip (p. 278)
dip-slip fault (p. 293)
dome (p. 292)
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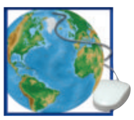

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