# 7 Structure of Rock Bodies

Although many people think that Earth's crust is permanent and fixed, evidence of crustal movement comes in many forms and is there for all to see. In the Mediterranean area, some ancient harbors, such as Ephesus in Asia Minor, are now high and dry, several kilometers inland from the sea. Other ancient shorelines have been submerged well below low tide. Earthquakes are perhaps the most convincing evidence that the crust is moving. During earthquakes, the crust not only vibrates, but segments of it are fractured and displaced along a fault. One impressive example was the movement along the San Andreas Fault during the 1906 San Francisco earthquake, which offset fences and roads by as much as 7 m. Another is the 1899 earthquake at Yakutat Bay, Alaska, during which a beach was uplifted 15 m above sea level. The raised beach terraces along the coast of southern California also testify to crustal movement in prehistoric times. There, ancient wave-cut cliffs and terraces—containing remnants of beaches with barnacles, shells, and sand—rise in a series of steps more than 500 m above the present shore.

Since the beginning of geologic studies more than 200 years ago, geologists have shown that rock layers in certain parts of the continents are folded, fractured, and deformed on a
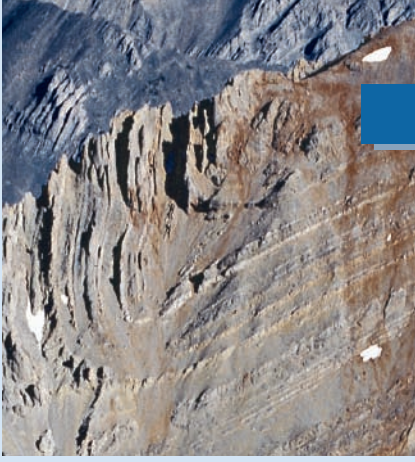
gigantic scale. The large-scale shape, geometry, and deformation of rock bodies are part of what we call their structure.

Deformation of the crust is most intense in the great mountain belts of the world, where sedimentary rocks, which were originally horizontal and below sea level, are now folded, contorted, fractured, and, in some places, completely overturned. In some mountains, large bodies of rock have been thrust several tens of kilometers over younger strata. The folded rocks in the world's major mountain belts (the Appalachians, the Rockies, the Andes, the Himalayas, the Urals, and the Alps) all exemplify this type of deformation. These folded and warped rock layers testify to the continuing motion of the lithosphere and the deformation it produces.

Spectacular examples of structural deformation are found in the large flexures of folded mountain belts. In the photo above of Borah Peak, Idaho, the sedimentary rock layers were originally deposited on the seafloor, but were uplifted more than four kilometers above sea level and deformed into anticlines and synclines by horizontal compression at a convergent plate margin. Erosion has cut through the flexures to reveal their internal structure.

The extensive deformation of rock bodies in the shields and mountain belts shows that Earth's tectonic system has operated throughout geologic time, with shifting plates constantly deforming at their margins. In this chapter we examine how and why rocks deform, how we describe deformed rocks, and then give specific examples of deformed rock bodies.

1. Deformation of Earth's crust is well documented in historical times by earthquakes along faults, by raised beach terraces, and by deformed rock bodies.
2. Rocks deform when applied stress exceeds their strength. They may deform by ductile flow or brittle fracture. Extensional stress causes rocks to stretch and thin. Contractional stress causes rocks to shorten and thicken.
3. Joints are fractures in rocks along which there is no horizontal or vertical displacement.
4. Faults are fractures, along which slippage or displacement has occurred. The three basic types are (a) normal faults, (b) thrust faults, and (c) strike-slip faults.
5. Folds in rock strata range in size from microscopic wrinkles to large structures hundreds of kilometers long. The major types of folds are (a) domes and basins, (b) plunging anticlines and synclines, and (c) complex folds.

## PRINCIPLES OF ROCK DEFORMATION

Rocks deform in response to differential stress. The resulting structure depends on the stress orientation. At high temperatures, ductile flow of rocks occurs. At low temperatures, brittle fractures form.

The folds and faults exposed in canyon walls and mountain ranges show that crustal rocks can be deformed on large scales and in dramatic ways (Figure 7.1). But why are some rocks warped into great folds and others only fractured or faulted? Why are some only gently folded, whereas others are complexly folded and faulted? In short, what factors control the type of deformation that rocks experience?

To understand this, you need to understand the forces on rocks. Force applied to an area is **stress.** Stress is the same thing as pressure and is a measure of the intensity of the force or of how concentrated the force is. Everyday experience tells us that solids will bend or break if too much stress is placed on them; that is, they deform if the stress exceeds their **strength** (their natural resistance to deformation). Rocks behave in the same way and deform in response to the forces applied to them.

All of Earth's rocks are under some type of stress, but in many situations the stress is equal in all directions and the rocks are not deformed. In many tectonic settings, however, the magnitude of stress is not the same in all directions and rocks experience **differential stress.** As a result, the rocks yield to the unequal stress and deform by changing shape or position. Geologists call the change in shape **strain.** In other words, differential stress causes strain.

Although strain proceeds by several complex phenomena, two end-member styles of deformation are recognized, with all gradations between them being possible. Under some conditions, rock bodies change shape by breaking to form continuous **fractures** and they lose cohesion; this is **brittle deformation** (Figure 7.2). We commonly see such behavior of solids in our daily experience: Chairs, baseball bats, pencils, and wooden beams break if too much force is applied to them. On the other hand, **ductile deformation** occurs when a rock body deforms permanently without fracturing or losing cohesion. The most obvious type of ductile deformation is the viscous flow of fluids, such as molten magma, but solids can also deform ductilely. At first it may seem strange that solids can flow. But bending of metal provides a familiar example. Consider sheet metal in a car fender. A minor collision commonly makes a dent, not a fracture in the fender. Likewise, rocks can flow in a solid state, under the right conditions. This type of solid state flow is usually called plastic flow and is accomplished by slow internal creep, gliding on imperfections in crystals, and recrystallization.

Depending on the temperature or pressure of the surroundings and the rate at which stress is applied, most types of rocks can deform by brittle fracture or
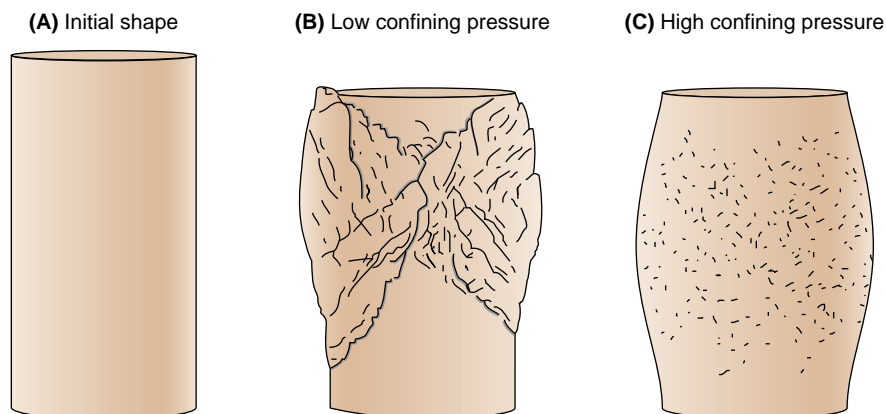
**How are solid rocks deformed?**
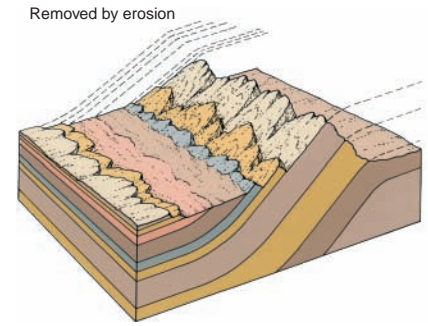
**FIGURE 7.1    A sequence of inclined beds** striking toward the background and dipping 40° to the east (left) forms the flank of the San Rafael Swell, in Utah. The diagram shows the form of the flexure and the upper beds, which have been partially removed by erosion.

ductile flow. Low pressures, low temperatures, and rapid deformation rates favor brittle deformation (Figure 7.2B). As a result, brittle structures are most common in the shallow crust. We use the term **shear** to describe slippage of one block past another on a fracture. High confining pressures, high temperatures, and low rates of deformation all favor ductile behavior (Figure 7.2C). Ductile deformation is more common in the mantle and deeper parts of the crust. Another example of this difference can be seen in the behavior of glass. When a glass rod is cold, it is strong and brittle fractures form when enough stress is applied (Figure 7.3). When the same glass rod is hot but not molten, it is weak and easy to bend. To visualize the role of the rate of deformation, consider taffy (or Silly Putty™) as an example. If warm taffy is pulled slowly and steadily, it is ductile and stretches continuously, without breaking, to form long, thin strands. On the other hand, if it is stretched rapidly, it may break and form brittle fractures.
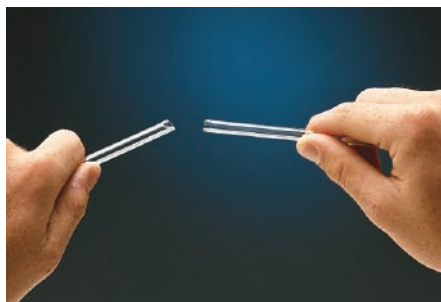
It should be clear which types of rock structures form by ductile deformation and which form by brittle behavior. The flow of rocks in a solid state to form folds in metamorphic rocks is a good example of ductile behavior (Chapter 6 Opener). The formation of fractures, joints, and faults are common expressions of deformation in the brittle regime.

**What effect does confining pressure have on the style of rock deformation?**

**(A)** Initial shape    **(B)** Low confining pressure    **(C)** High confining pressure



**FIGURE 7.2    Brittle versus ductile behavior** of rocks is controlled by external conditions. These marble cylinders were compressed in a laboratory under different confining pressures, but at the same temperature and with the same deforming stress. At low pressure, the cylinder deformed in a brittle fashion and fractured and faulted. At high pressure, the cylinder deformed in a ductile fashion as mineral grains flowed and recrystallized.

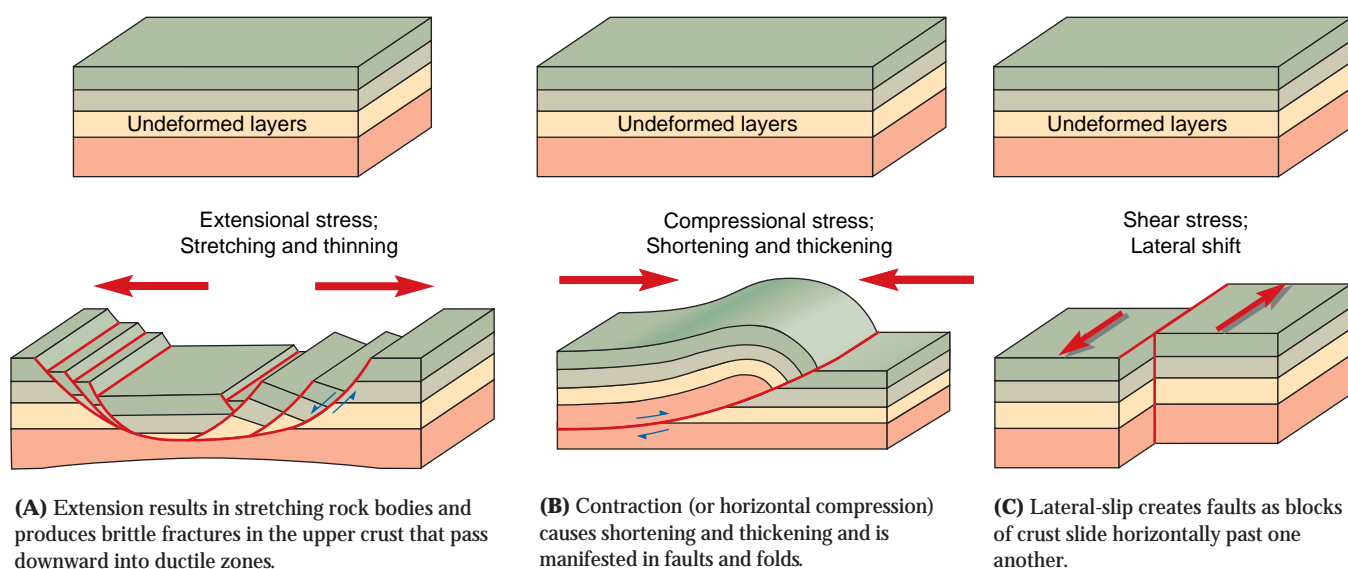**FIGURE 7.3**  **Temperature affects how solids deform.** *(Photographs by Stan Macbean)*

**(A)** A cold glass rod deforms by brittle fracture and breaks.

**(B)** A hot glass rod bends and shows ductile behavior without breaking.

We can also better understand rock structures if we consider the orientations of the stresses acting on a body (Figure 7.4). For example, **tension** occurs where the stresses point away from one another and tend to pull the rock body apart. In contrast, **compression** tends to press a body of rocks together. Three distinct types of deformation occur because of differential stresses caused by tectonic processes (Figure 7.4). To visualize this, imagine how two adjacent blocks can interact. They can move away from one another (**extension**), move toward one another (**contraction**), or slip horizontally past one another (**lateral-slip**). Obviously, the orientation of the stresses acting on the blocks determines which of the three cases is dominant. In the simplest case, the stress orientations are directly related to plate tectonic settings. Extension is caused when the differential stresses point away from one another (Figure 7.4A). This type of deformation results in lengthening and is common at divergent boundaries. In brittle rocks it is expressed by fracturing and faulting, and in ductile rocks by stretching and thinning. Contraction is caused by horizontal compression when the differential stresses are directed toward one another (Figure 7.4B). Contraction is common at convergent boundaries and causes shortening and thickening of rock bodies, expressed as faults in brittle rocks and folds in ductile rocks. Lateral-slip is the kind of shear that occurs when rocks slide horizontally past one another along nearly vertical fractures and dominates at transform plate boundaries (Figure 7.4C).



**(A)** Extension results in stretching rock bodies and produces brittle fractures in the upper crust that pass downward into ductile zones.

**(B)** Contraction (or horizontal compression) causes shortening and thickening and is manifested in faults and folds.

**(C)** Lateral-slip creates faults as blocks of crust slide horizontally past one another.

**FIGURE 7.4**  **Extension, contraction, and lateral-slip** produce fundamentally different types of structures in rocks. Moreover, each is caused by different stress orientations and dominates at different plate tectonic settings.

# GEOMETRY OF ROCK STRUCTURES

> The orientation of planar features in rocks, such as bedding planes, faults, and joints, can be defined by measurements of dip (the downward inclination of the plane) and strike (the direction or trend of the plane).

Many structural features of the crust are too large to be seen from one point on the ground. They are recognized only after the geometry of the rock bodies is determined from geologic mapping. At an outcrop, two fundamental observations—dip and strike—describe the orientations of bedding planes, fault planes, joints, and other planar features in the rock. The **dip** of a plane is the angle and direction of its inclination from the horizontal. The **strike** is the compass bearing of a horizontal line on the plane, such as a bedding plane or a fault. These two measurements together define the orientation of the planar surface in space.

The concept of dip and strike can be easily understood by referring to Figure 7.5, which shows an outcrop of tilted beds along a coast. The water provides a necessary reference to a horizontal plane. The trend of the waterline along the bedding plane is the direction of strike. The angle between the bedding plane and the water surface is the angle of dip. Figure 7.1 shows a sequence of beds striking south (to the top of the picture) and dipping 40° to the east (to the left of the picture). Another way to visualize dip and strike is to think of the roof of a building. The dip is the direction and amount of inclination of the roof, and the strike is the trend of the ridge.

In the field, dip and strike are measured with a geologic compass, which is designed to measure both direction and angle of inclination. A long crossbar shows the strike, a short line perpendicular to it shows the direction of dip, and the number represents the angle of dip (Figure 7.5). The symbols provide a way to view a map in three dimensions and simplify the construction of vertical cross sections through the crust.

A useful tool to interpret the structure of dipping layers of rock is called the **rule of Vs**. Look carefully at the sketch and photo in Figure 7.5. Erosion has cut V-shaped notches into the resistant layer of rock. Each V points in the direction the bed dips.

**How do we measure the orientation of rock bodies, faults, and joints?**
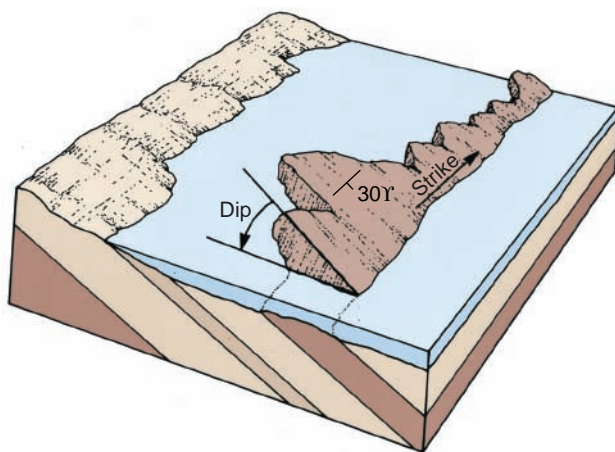


**FIGURE 7.5    The concept of dip and strike** can be understood by studying rock layers such as the ones shown in this photograph. The strike of a bed is the compass bearing of a horizontal line drawn on the bedding plane. It can readily be established by reference to the horizontal waterline in this example. The dip is the angle and direction of inclination on the bed, measured at right angles to the strike and is represented with a T-shaped symbol and accompanying dip in degrees.

## JOINTS

> Joints are tension fractures in brittle rocks along which no shear has occurred. They form at low pressure and are found in almost every exposure.

The simplest and most common structural features of rocks at Earth's surface are cracks or fractures, known as **joints**, along which little displacement (or slip) has occurred. Their most important feature is the absence of shear; no movement occurs parallel to the fracture surface. Joints form by the brittle failure of rocks at low pressure as stress accumulates and exceeds the rock's strength. They do not occur at random but are usually perpendicular to the direction of tension.

**What is the difference between a joint and a fault?**

Multiple sets of joints that intersect at angles ranging from 45° to 90° are very common. They divide rock bodies into large, roughly rectangular blocks. These joint systems can form remarkably persistent patterns extending over hundreds of square kilometers. Each set probably formed at a different time and under a different stress orientation.

The best areas to study joints are where brittle rocks, such as thick sandstones, have been fractured and their joint planes accentuated by erosion. The massive sandstones of the Colorado Plateau are excellent examples. Joints are expressed by deep, parallel cracks that have been enlarged by erosion; they are most impressive when seen from the air (Figure 7.6). In places, joints control the development of stream courses, especially secondary tributaries and areas of solution activity.

Joints result from broad regional upwarps, or from contraction or extension associated with faults and folds; some tensional joints result from erosional unloading and expansion of the rock. Columnar joints in volcanic rocks are produced by tensional stresses that are set up as the lava cools and contracts.

At first it may seem that joints are insignificant and uninteresting, but they have great economic importance. They can be paths for groundwater migration and for the movement and accumulation of petroleum. Analysis of joint patterns has been



**FIGURE 7.6**    **Joint systems in resistant sandstones** in Arches National Park, Utah, have been enlarged by weathering, forming long, narrow crevasses. The system of intersecting joints reflects the orientation of stress that deformed the rock body.

important in exploration and the development of these resources. Joints also control the deposition of copper, lead, zinc, mercury, silver, gold, and tungsten ores. Hot aqueous solutions associated with igneous intrusions migrate along joint systems and minerals crystallize along the joint walls, forming mineral **veins.** Modern prospecting techniques therefore include detailed analysis of fractures. Major construction projects are especially affected by joint systems within rocks, and allowances must be made for them in project planning. For example, dams must be designed so that the stresses caused by water storage tend to close any fractures in the bedrock foundation. Joint systems can be either an asset or an obstacle to quarrying operations. Closely spaced joints severely limit the sizes of blocks that can be removed. If a quarry follows the orientation of intersecting joints, however, the expense of removing building blocks is greatly reduced, and waste is held to a minimum.
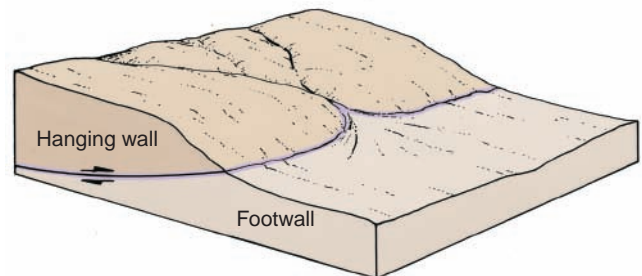
## FAULTS

Faults are fractures in Earth's crust along which displacement has occurred. Three basic types of faults are recognized: (1) normal faults, (2) reverse faults, and (3) strike-slip faults. Normal faults are usually the result of extension, thrust faults the result of horizontal compression, and strike-slip faults the result of lateral slip.

Slippage (or shear) along brittle fractures in Earth's crust creates **faults** (Figure 7.7). Like other deformation features, they form by the application of differential stress. In a road cut, or in the walls of a canyon, a fault plane may be obvious, and the displaced, or offset, beds can easily be seen. Elsewhere, the surface expression of a fault can be very subtle, and detailed geologic mapping may be necessary before the precise location of such a fault can be established. Displacement along faults ranges from a few centimeters to hundreds of kilometers.
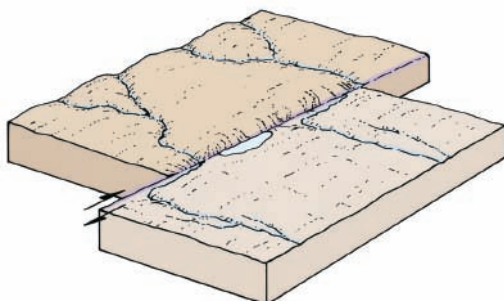
**Why are there three different types of faults?**
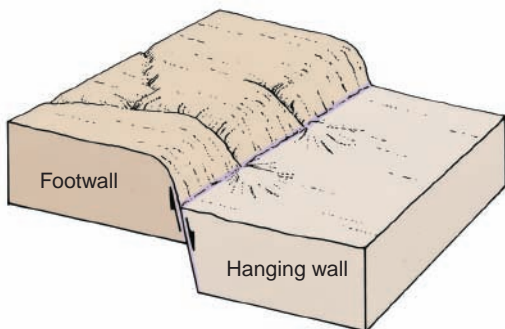
**(A)** In normal faults, the hanging wall moves downward in relation to the footwall.

**(B)** In thrust faults, the hanging wall moves upward in relation to the footwall.

**(C)** In strike-slip faults, the displacement is horizontal.

**FIGURE 7.7    The three major types of faults** are distinguished by the direction of relative displacement.

Faults grow by a series of small movements, which occur as stress built up in the crust is suddenly released in earthquakes. Displacement can also occur by an almost imperceptibly slow movement called tectonic creep.

*Normal (Extensional) Faults.*    Along **normal faults,** movement is mainly vertical, and the rocks above the fault plane (the **hanging wall**) move downward in relation to those beneath the fault plane (the **footwall**) (Figure 7.7A). Most normal faults are steeply inclined, usually between 65° and 90°, but some dip at low angles (Figure 7.4). Their predominantly vertical movement commonly produces a cliff, or **scarp**, at the surface.

Normal faults are rarely isolated features. A group of parallel normal faults may develop a series of fault-bounded blocks. A narrow block dropped down between two normal faults is a **graben** (German, "trough" or "ditch"), and an upraised block is a **horst** (see Figure 7.8B). A graben typically forms a conspicuous fault valley or basin marked by relatively straight, parallel walls. Horsts form plateaus bounded by faults. Normal faults usually juxtapose younger rocks over older rocks. In the field, this may be mistaken for normal layer-upon-layer deposition, especially where the normal fault dips at a low angle (Figure 7.4). However, careful fieldwork will also reveal that some rock layers are apparently missing along the fault. This is called omission of strata.

Large-scale normal faulting is the result of horizontal extensional stress, which stretches, thins, and pulls apart the lithosphere. Normal faults are common because rocks are weaker during extension than during compression. This type of extensional stress occurs on a global scale along divergent plate margins. Consequently, normal faults are the dominant structures along the oceanic ridge, in continental rift systems, and along rifted continental margins.

In the Basin and Range Province of western North America, normal faulting produced a series of grabens and horsts trending north from central Mexico to Oregon and Idaho. The horsts commonly form mountain ranges from 2000 to 4000 m high, which are considerably dissected by erosion. Grabens form topographic basins, which are partly filled with erosional debris from the adjacent ranges. Between the Wasatch Range, in central Utah, and the Sierra Nevada, on the Nevada-California border, faulting has extended Earth's crust some 100 km during the last 15 million years. Modern earthquakes and displacement of young alluvial fans and surface soils show that many of the faults are still active. The great rift valley of east Africa is another example of large-scale normal faulting produced by a zone of extension in Earth's crust.

*Reverse (Contractional) Faults.*    Faults in which the hanging wall has moved up and over the footwall are **reverse faults** (Figure 7.7B). **Thrust faults** are low-angle reverse faults and dip at angles less than 45°. Movement on a thrust fault is predominantly horizontal, and displacement can be more than 50 km.

Thrust faults result from horizontal compression with the maximum stress perpendicular to the trend of the fault. This shortens and thickens the crust. In contrast to normal faults, thrust faults usually place older over younger strata and instead of omitting layers, units are repeated in a vertical section (Figure 7.9). Where resistant rocks are thrust over nonresistant strata, a scarp is eroded on the upper plate. The scarp is not straight or smooth, as are cliffs produced by normal faulting. Rather, the outcrop of the fault surface typically is irregular in map view due to the low angle of dip (Figure 7.9B).

Most contractional faults form at convergent plate margins. Thrust faults are typically associated with folds and are prominent in all of the world's major folded mountain belts. They commonly evolve from folds in the manner diagrammed in Figure 7.9.

*Strike-Slip Faults.*    **Strike-slip faults** are high-angle fractures in which slip is horizontal, parallel to the strike of the fault plane. Ideally, there is little or no
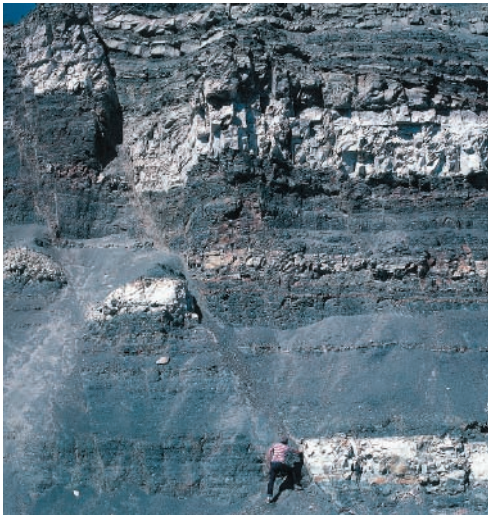
**What tectonic settling commonly produces normal faults?**

**Faults**

**Why are thrust faults and folds found in the same regions?**

**(A)** Faults exposed on the vertical walls of a valley or road cut are easily recognized by the displaced strata.



**(B)** Grabens and horsts are commonly produced where the crust is extended. This area shows alternating horsts (ridges) and grabens (valleys).



**(C)** Normal faults commonly form cliffs on the upthrown block. This is the Hurricane fault in southern Utah where the red Mesozoic strata are displaced downward about 1.5 km. The trace of the fault is along the base of the gray cliff.



**(D)** Thrust faults form where the crust is contracted as in a folded mountain belt. Here, a series of thrust faults (shown in white) juxtapose Precambrian rocks on Paleozoic sandstones.



**(E)** Strike-slip faults are commonly expressed by a series of straight linear ridges and troughs that can be traced for long distances. Here the San Andreas Fault in southern California offsets a drainage system.

**FIGURE 7.8    The surface expression of faults** is highly variable and influenced by the character of the rock units that have been displaced, the type and degree of erosion, and the nature of the exposed surface.
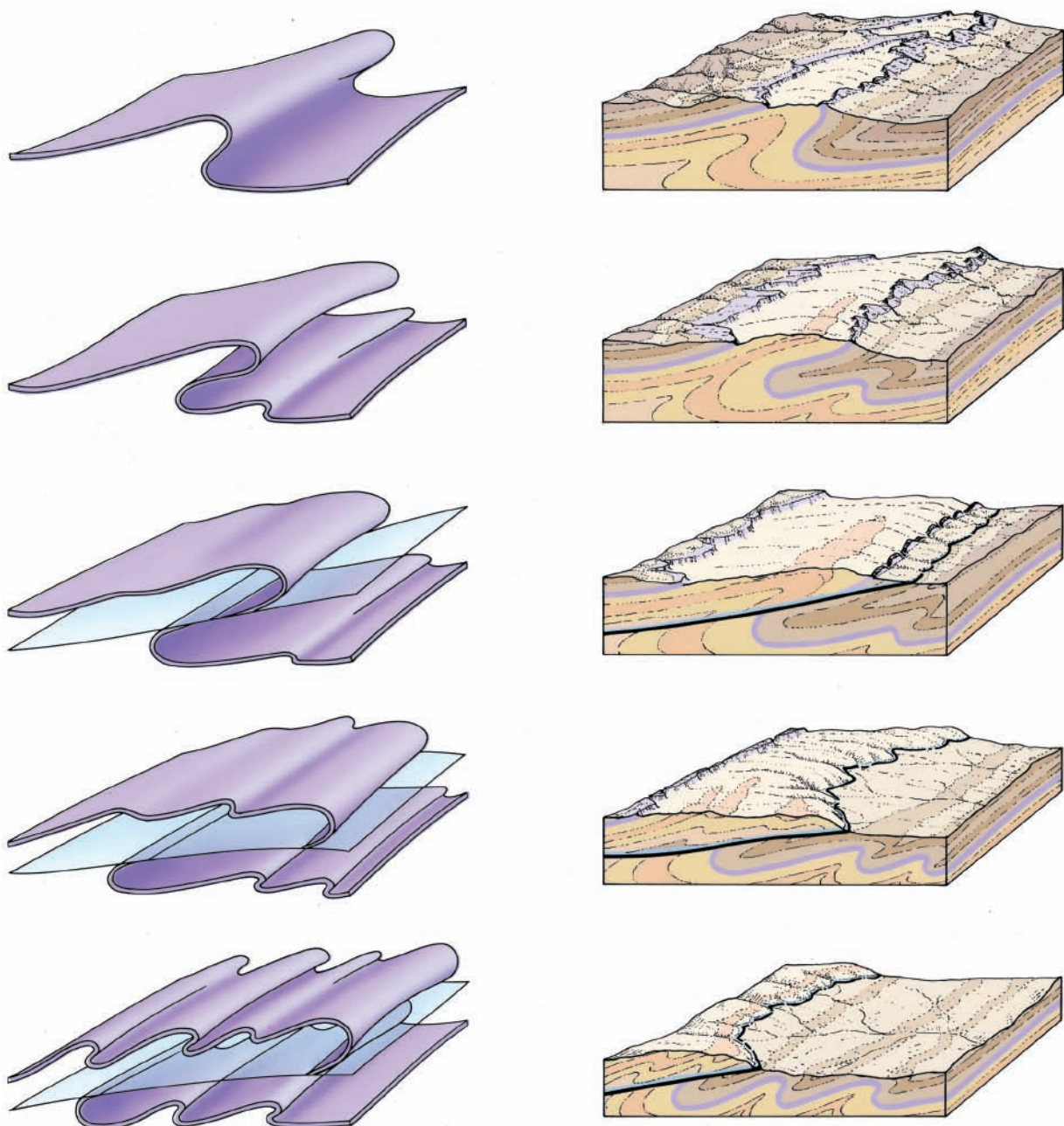
**(A)** Progressive development of a fold in a colored layer into a thrust fault (nearly horizontal plane).

**(B)** Thrust faults form sinuous scarps that are modified by contemporaneous erosion.

**FIGURE 7.9** **The evolution of thrust faults from folds** is depicted in this sequence of diagrams. A fault forms when the strength of the folding rock layers is exceeded.

vertical movement, so high cliffs do not usually form along strike-slip faults. Instead, these faults are expressed topographically by a straight valley or by a series of low ridges and commonly mark discontinuities in the drainage and types of landscape. No crustal thinning or thickening are produced, except at bends in the fault where extension or contraction can occur.

Some topographic features produced by strike-slip faulting and subsequent erosion are shown in Figure 7.7C. One of the more obvious is the offset of the drainage pattern. The relative movement is often shown by abrupt right-angle bends in streams at the fault line. A stream follows the fault for a short distance and then turns abruptly and continues down the regional slope (Figure 7.8E). As the blocks

**FIGURE 7.10**    **Recent displacement along a fault,** at the base of the Lost River Range in southern Idaho, has produced the fresh cliff at the base of the mountain front. Cumulative movement on the fault during a vast period of time produced the mountain range. *(Courtesy of Glenn Embree)*

move, some parts may be depressed to form sag ponds. Others buckle into low, linear ridges. Faults also disrupt patterns of groundwater movement, as is reflected by contrasts in vegetation and soils, and by the occurrence of springs along the fault trace.

Strike-slip faults result from horizontal shear along nearly vertical faults. They commonly are produced by lateral-slip where one tectonic plate slides past another at a transform fault boundary. The most famous is probably California's seismically active San Andreas Fault. Strike-slip faults also join adjacent segments of the oceanic ridge. Others form where two segments of continental crust are stretched or shortened at different rates.

## Observed Movement on Faults

The movement along faults during earthquakes rarely exceeds a few meters. In the great San Francisco earthquake of 1906, the crust slipped horizontally as much as 7 m along the San Andreas Fault, so roads, fence lines, and orchards were offset. Recent offsets along vertical faults in Nevada and Idaho have produced fresh scarps from 3 to 6 m high (Figure 7.10). The Good Friday earthquake in Alaska, in 1964, was accompanied by a 13-m uplift near Montague Island. The largest well-authenticated displacement during an earthquake appears to have occurred in 1899 near Yakutat Bay in Alaska, where beaches were raised as much as 15 m above sea level.

Movement along faults is not restricted to uplift during earthquakes, however. Precise surveys along the San Andreas Fault show slow shifting along the fault plane at an average rate of 4 cm/yr. Such slow movements, known as tectonic creep, break buildings constructed across the fault line and eventually result in considerable displacement.

The important point is that the total displacement on a fault that may amount to many kilometers does not occur in a single violent event. Rather, it is the result of numerous periods of displacement. On most faults, there are hundreds to thousands of years between major earthquakes. Slow tectonic creep may continue between these major events.

**What kind of stress makes strike-slip faults?**

**How long does it take for thousands of meters of displacement to occur on a normal fault?**
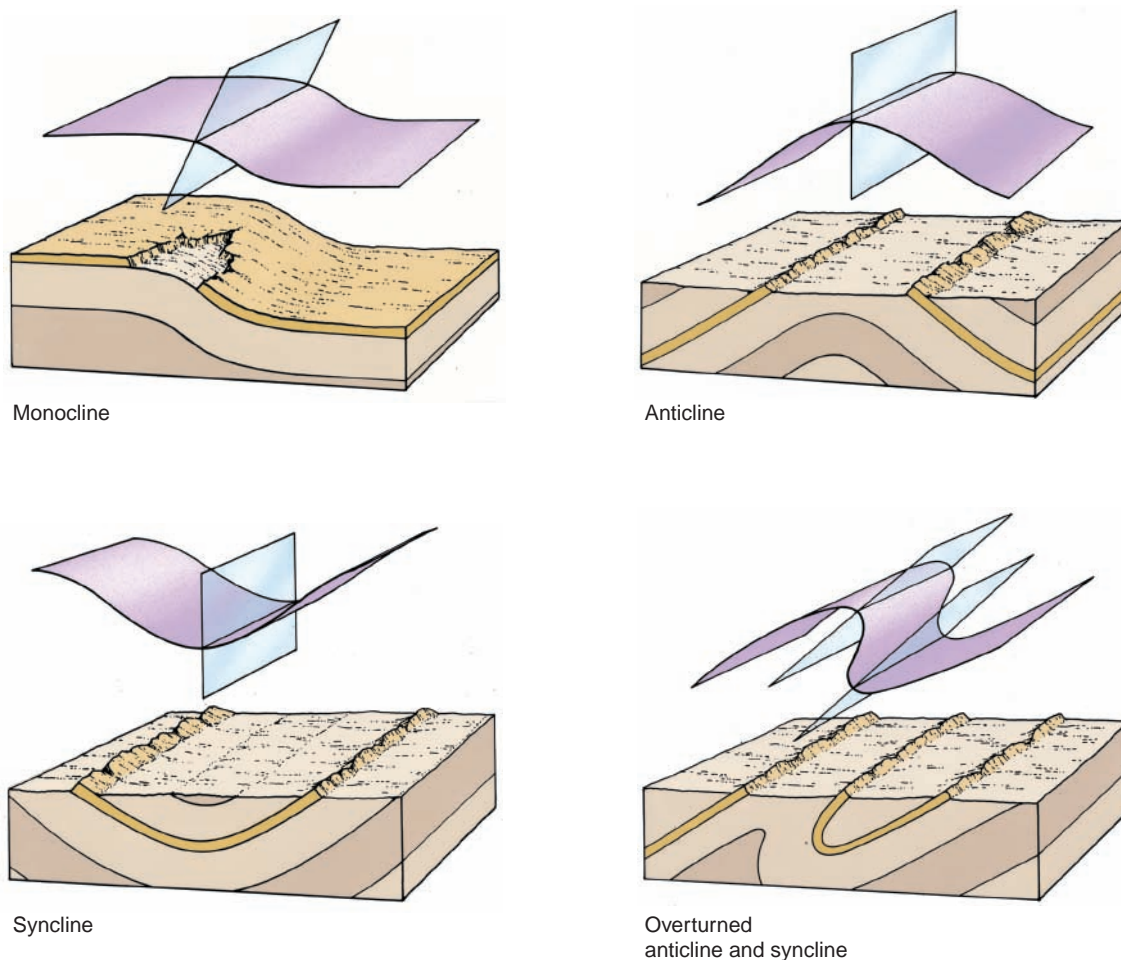
## FOLDS

Folds are warps in rock strata during ductile deformation. They are three-dimensional structures ranging in size from microscopic crinkles to large domes and basins that are hundreds of kilometers across. Most folds develop by horizontal compression at convergent plate boundaries where the crust is shortened and thickened. Broad, open folds form in the stable interiors of continents, where the rocks are only mildly warped.

**How do we describe the geometry of folded rocks?**

Almost every exposure of sedimentary rock shows some evidence that the strata have been deformed. In some areas, the rocks are slightly tilted; in others the strata are folded like wrinkles in a rug. Small flexures are abundant in sedimentary rocks and can be seen in mountainsides and road cuts and even in hand specimens. These warps in the strata are called **folds** and are a manifestation of ductile deformation in response to horizontal compression. This kind of deformation is also called contraction. Large folds cover thousands of square kilometers, and they can best be recognized from aerial or space photographs or from geologic mapping. Like faults, folds form slowly over millions of years, as rock layers gradually yield to differential stress and bend.

Folds are of great economic importance because they commonly form traps for oil and gas and may control localization of ore deposits. Consequently, it is of more than academic interest to understand folds.

Monocline

Anticline

Syncline

Overturned
anticline and syncline

**FIGURE 7.11**    **The nomenclature of folds** is based on the three-dimensional geometry of the structure, although most exposures show only a cross section or map view.

# Geologic Maps: Models of the Earth

The key tool used in almost all geologic investigations is a geologic map. A geologic map shows more than just locations and distances; it also shows more than elevation. Instead, a geologic map shows the distribution of rock types, faults, folds, and other geologic features, as well as the ages of the rocks. It is a scale model of the rock bodies, revealing much about their origin, internal structure, and deformation.

Geologic maps are constructed by careful fieldwork. A geologist working alone or as part of a small team collects critical data about rock types during traverses on foot across a field area. The locations and rock type of each outcrop are found and plotted on a map. In arid regions where vegetation is sparse, large areas can be mapped on an aerial photograph as shown here. By plotting the distribution of rock types along with strikes and dips of planar features, we reveal the three-dimensional structure of the rocks.

The goal of field geologists is to first recognize major rock bodies (formations) that can be mapped and then to determine the location and nature of the contacts between the rock units. They may do this by walking along the con-

tacts in the field and marking their locations on a map or a photo. If the contacts are obscured by vegetation or by younger rocks, contacts may need to be inferred from the evidence gathered. Soils, selective growth of vegetation, stream patterns, and other geologic features can help. A geologist working on a map must also decide the relative ages of the rock units using the principles of superposition and crosscutting relationships.

Observations are recorded on maps and in field notebooks. A brunton compass (pocket transit) may be used to measure the strike and dip of deformed layers of sedimentary rock. A geologic hammer is used to collect samples and break off weathered surfaces. A stereoscope is used with a pair of aerial photographs to view the landscape in three dimensions. The data are then plotted on a topographic map. In some areas, where locations are difficult to establish precisely because of vegetation cover, a GPS unit (global positioning system; see page 592) and computer may be used to record locations or contacts. Regardless of how the data are collected and recorded, the end result is a scale model of Earth.
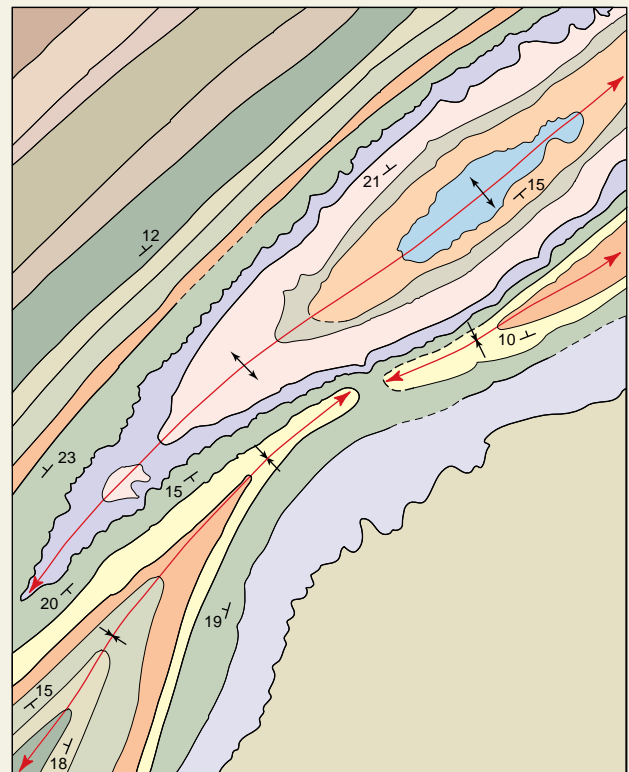


*(Courtesy U.S. Department of Agriculture)*

**FIGURE 7.12    The hinge plane of a fold** is an imaginary plane that divides the fold in equal parts. The line formed by the intersection of the hinge plane and a bedding plane is called the hinge. The downward inclination of the hinge is called the plunge.

**Folds**
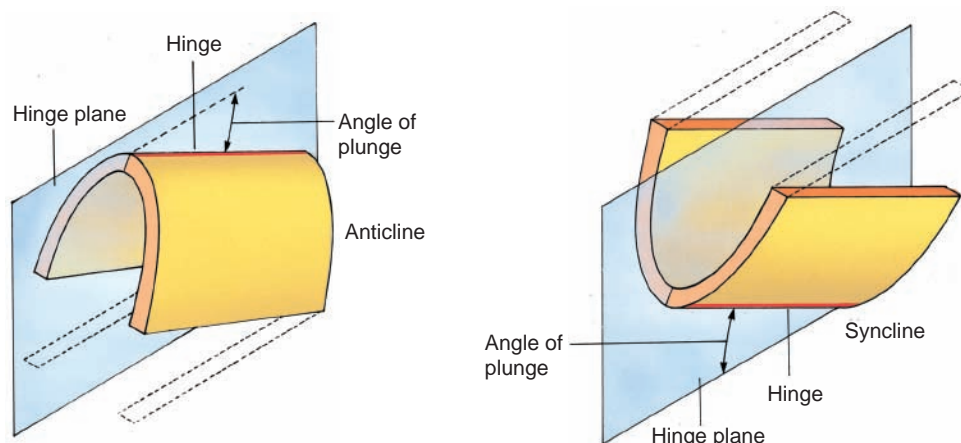
**What is the outcrop pattern of a series of plunging anticlines and synclines?**

*Fold Nomenclature.*    Three general types of folds are illustrated in Figure 7.11. An **anticline,** in its simplest form, is uparched strata, with the two **limbs** (sides) of the fold dipping away from the crest. Rocks in an eroded anticline are progressively *older* toward the interior of the fold. **Synclines,** in their simplest form, are downfolds, or troughs, with the limbs dipping toward the center (Figure 7.11). Rocks in an eroded syncline are progressively *younger* toward the center of the fold. **Monoclines** are folds that have only one limb; horizontal or gently dipping beds are modified by simple steplike bends.

For purposes of description and analysis, it is useful to divide a simple fold into two more-or-less equal parts by an imaginary plane known as the **hinge plane.** The hinge marks the region of maximum curvature in the fold. The line formed by the intersection of the hinge plane and a bedding plane is the **hinge line,** and the downward inclination of the hinge line is called the **plunge** (Figure 7.12). A **plunging fold,** therefore, is a fold in which the hinge line is inclined.

In most folds, the hinge plane is not vertical but is inclined, and the fold is overturned and one limb is steeper than the other (7.11D). The direction the hinge plane is rotated from vertical indicates the direction the rocks were displaced. In other words, movement is toward the steep limb. In Figure 7.11D, you can see that the folded rocks were transported from left to right, just as they were in Figure 7.9.

*Fold Belts.*    Where contraction is intense (typically in **orogenic belts** at convergent plate boundaries), the rock layers are deformed into a series of tight folds in a long linear belt. The internal geometry of many fold belts is not exceedingly complex. In many ways, the folds resemble the wrinkles in a rug. However, complexity in the outcrop patterns of fold belts results from erosion, so folds may be difficult to recognize on aerial photographs without some experience in geologic observation and interpretation. The diagrams in Figure 7.13 illustrate a fold belt with plunging folds and its surface expression after the upper part has been removed by erosion. The outcrop of the eroded plunging anticlines and synclines forms a characteristic zigzag pattern. The nose of an anticline forms a V that points in the direction of plunge, and the oldest rocks are in the center of the fold (see Figure 7.15D). The nose of a syncline forms a V that opens in the direction of plunge, and the youngest rocks are in the center of the fold. Together, the outcrop pattern, the strike and dip of the beds, and the relative ages of the rocks in the center of the fold make it possible to determine the structure's subsurface configuration. Thrust faults commonly form in association with these contractional folds.

Intense deformation in the cores of some mountain ranges produces complex folds like those in Figure 7.14. Some folds are refolded. Such structures commonly exceed 100 km across and can extend through a large part of a mountain belt. Details of such intensely deformed structures are extremely difficult to work out because of the complexity of the outcrop patterns. These complexities arise from
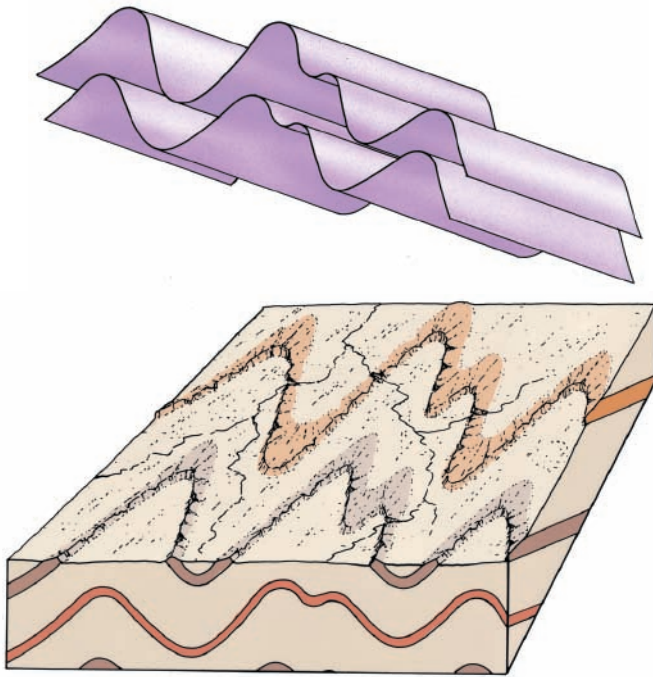
**(A)** The basic form of folded strata is similar to that of a wrinkled rug. In this diagram, the strata are compressed and plunge toward the background.

**(B)** If the tops of the folded strata are eroded away, a map of the individual layers shows a zigzag pattern at the surface. Rock units that are resistant to erosion form ridges, and nonresistant layers are eroded into linear valleys. In a plunging anticline, the surface map pattern of an anticline forms a V pointing in the direction of plunge.

**FIGURE 7.13**    **A series of plunging folds** forms a zigzag outcrop pattern.

multiple episodes of folding at elevated temperatures and pressures in lower crust. Under these conditions, rocks are much more likely to deform ductilely.

Figure 7.14 illustrates the geometry and surface expression of a large complex fold. Figure 7.14A is a perspective drawing of a single bed in a typical complex fold. This **overturned fold** is a huge anticlinal structure with numerous minor anticlines and synclines forming indentations on the larger fold; Figure 7.14B shows the fold after it has been subjected to considerable erosion, which has removed most of the upper limb. Note the cross section of the structure on the mountain front and the outcrop pattern compared with that in Figure 7.14A. The topographic expression of complex folds is variable. They usually are expressed in a series of
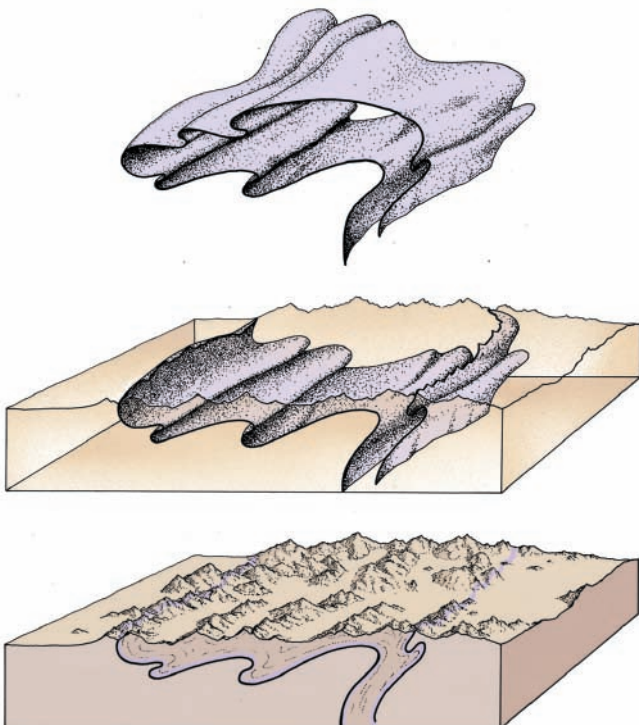
**(A)** Rocks that have been intensely deformed commonly consist of large overturned folds, with minor folds on the limbs. Some of these are refolded folds.

**(B)** The surface outcrop of the fold, after erosion has removed the upper surface, shows great complexity, so that its details and overall structure are difficult to recognize.

**(C)** The topographic expression of complex folds can be a series of linear mountain ridges.

**FIGURE 7.14**    **Complexly deformed folds** produce intricate outcrop patterns.

**(A)** Anticlines and synclines are easily recognized when erosion cuts across the structures and exposes them in a vertical cross section, such as these folds in the Calico Hills of southern California.



**(B)** A small structural dome in west Texas has been truncated by erosion and is expressed as a series of circular ridges and valleys. The dome is only about 5 km across. *(Courtesy of U.S. Geological Survey)*
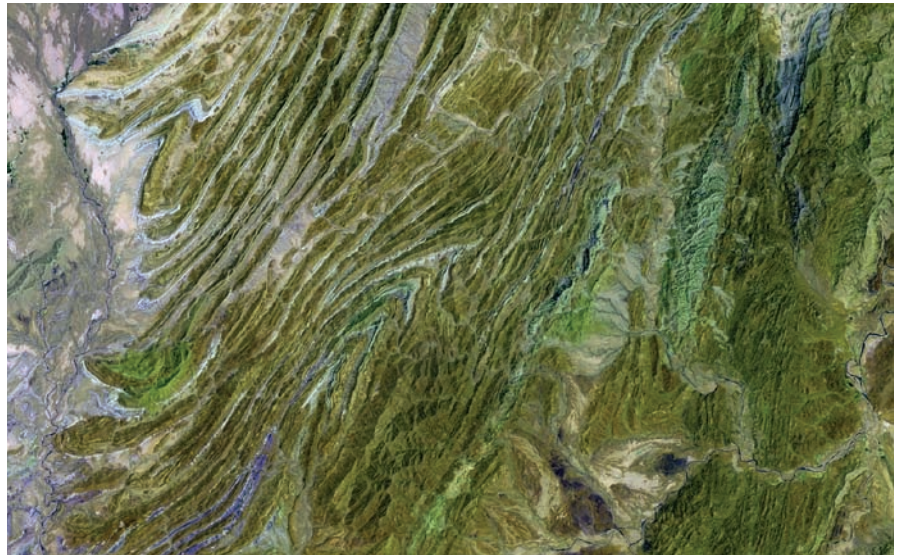


**(C)** Folds in the Sulaiman Ranges of western Pakistan are plunging anticlines and synclines formed when India collided with Asia. This area is about 30 km across. *(Courtesy of U.S. Geological Survey and EROS Data Center)*



**(D)** A plunging anticline forms a typical V-shape pattern pointing in the direction of plunge. This is an exceptionally clear example near St. George, Utah, where colorful beds form alternating ridges and valleys.

**FIGURE 7.15** **The surface expression of folds** is extremely variable because of the great range in the size and shape of the structures and because of the variety of ways in which folds may be modified by erosion. Common types of folds are shown above.
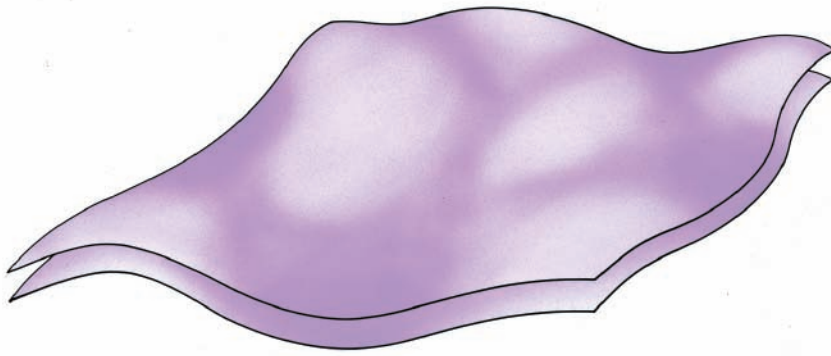
**(A)** A single folded bed is warped into broad domes and basins.
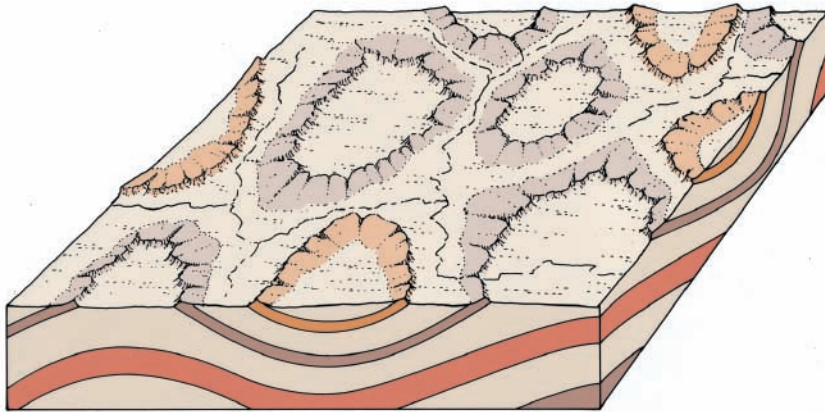
**(B)** As erosion proceeds, the tops of the domes are eroded first. The outcrop pattern of eroded domes and basins typically is circular or elliptical.

**FIGURE 7.16    The geometry and topographic expression of domes and basins** involve broad upwarps and downwarps of layered rocks. When eroded, the exposed rock forms circular or elliptical outcrop patterns.

mountains (Figure 7.14C). Complex folds are common in the Swiss Alps (see Figure 21.14C), but they were recognized only after more than half a century of detailed geologic studies. They are also common in the roots of ancient mountain systems and thus are exposed in many areas of the shields.

An excellent example of an orogenic belt is the Zagros Mountains of southern Iran. The fold belt is only a small part of the Alpine Himalaya chain that extends from southern Europe and across southern Asia (see Figure 7.17). The Landsat satellite photograph of the region shows that the sedimentary beds have been folded above a series of major thrust faults. Anticlines form the hills and are separated by intervening synclines. The Valley and Ridge Province of the Appalachian Mountains in the eastern United States is another good example. The fold belt extends from New England to central Alabama and consists of enormous parallel folds hundreds of kilometers long. Thrust faults slice through many of the folds. The region has been deeply eroded so that resistant formations form long, high ridges, and the weak formations form intervening valleys (see Figures 1.10 and 23.17). Other folded mountain belts include the Sulaiman Range of Pakistan (Figure 7.15C) and the Alps in France and Switzerland (see Figure 21.14c). Folds are the obvious expression of the ductile deformation of rocks in Earth's crust. All of these fold belts formed at convergent plate margins where an oceanic plate was thrust underneath a continental plate or where two continents collided.

***Domes and Basins.***    In contrast to fold belts at convergent margins, the sedimentary rocks covering much of the continental interiors have been only gently warped into broad **domes** and **basins** many kilometers in diameter. One large basin covers practically all of the state of Michigan. Another underlies the state of Illinois. An elongate dome underlies central Tennessee, central Kentucky, and southwestern Ohio. Although these flexures in the sedimentary strata are extremely large, the configuration of the folds is known from geologic mapping and from information gained through drilling. The nature of these flexures and their topographic expression are illustrated in Figures 7.15 and 7.16.

**FIGURE 7.17** **A vast orogenic belt** is being created by the collision of Arabia with southern Asia. This portion of the fold and thrust belt is part of the Zagros Mountains of Iran. The deformation accompanies the underthrusting of the Arabian subplate beneath Asia. Doubly plunging anticlines and synclines as well as elongate domes lie above major thrust faults. In the lower right, a light-colored salt dome has pierced through an anticline. The salt is not resistant and a valley has formed. *(Courtesy of U.S. Geological Survey EROS Data Center. Cross section modified from Nadine MacQuarrie.)*

Deformed strata

Salt layer
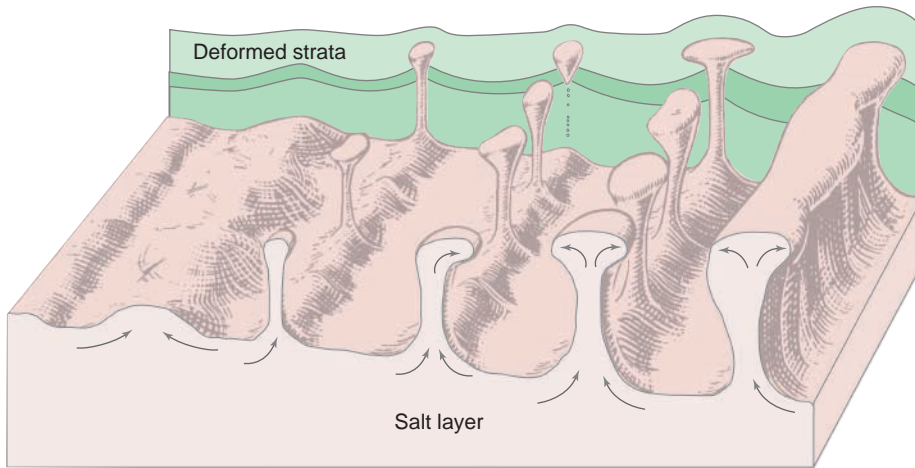
**FIGURE 7.18    Diapirs** form by the rise of low-density ductile materials such as salt and gypsum taking on a variety of forms. The overlying strata may be faulted and folded as they are pierced or domed upward. As salt domes form, the layer of salt is thinned *(Modified after F. Trusheim)*.

The form of a single bed warped into broad domes and basins is shown in perspective in Figure 7.16A. If erosion cuts off the tops of the domes, the surface exposure of the layer looks like the one shown in Figure 7.16B. The deformed layers in both domes and basins typically have circular or elliptical outcrop patterns. There is a major difference, however. The rocks exposed in the central parts of eroded domes are the oldest rocks, whereas the rocks exposed in the centers of basins are the youngest. The rule of Vs is also a useful tool to interpret the structure of domes and basins. Look carefully at the aerial photos of the dome in Figure 7.15B and the model in Figure 7.16. Note how stream erosion has cut V-shaped valleys into the resistant layers of rock. The "V" points in the direction the bed dips. Thus, for domes the Vs point away from the center and for basins they point inward. A classic example of a broad fold in the continental interior is the large dome that forms the Black Hills of South Dakota (see Figure 23.11). Resistant rock units form ridges that can be traced completely around the core of the dome, and nonresistant formations make up the intervening valleys.

How these broad domes form is still enigmatic. Their circular shapes and distance from convergent margins are puzzling. Many could have formed by multiple periods of deformation. For example, gentle east-west contraction far from an active convergent margin, followed by north-south contraction related to a different plate margin, could produce a series of broad domes and intervening basins.

*Diapirs.*    Some domes and basins also could form by vertical adjustments caused by density differences in the crust or upper mantle. Many small domes, like the one in Figure 7.15B, are associated with buoyant rise of material that is less dense than the overlying rock. For example, in thick sequences of sedimentary rocks, beds of salt may deform and rise as a **diapir,** a streamlined body shaped somewhat like an inverted teardrop (Figure 7.18). Plugs of salt may rise and pierce overlying sedimentary strata to form salt domes. The deformed sedimentary beds are faulted and typically dip away from the center of the structure. The white mass in the lower right part of Figure 7.18 is a salt dome that has reached the surface to flow like a glacier. The movement of salt has modified the seafloor south of the Mississippi delta on a grand scale, with subsidence basins and domes pockmarking the seafloor (see Figure 11.17). Other small domes are formed by the intrusion of magma (see Figure 4.18A).

Thrust fault plane

Eroded strata

Paleozoic limestones

Mesozoic sandstones
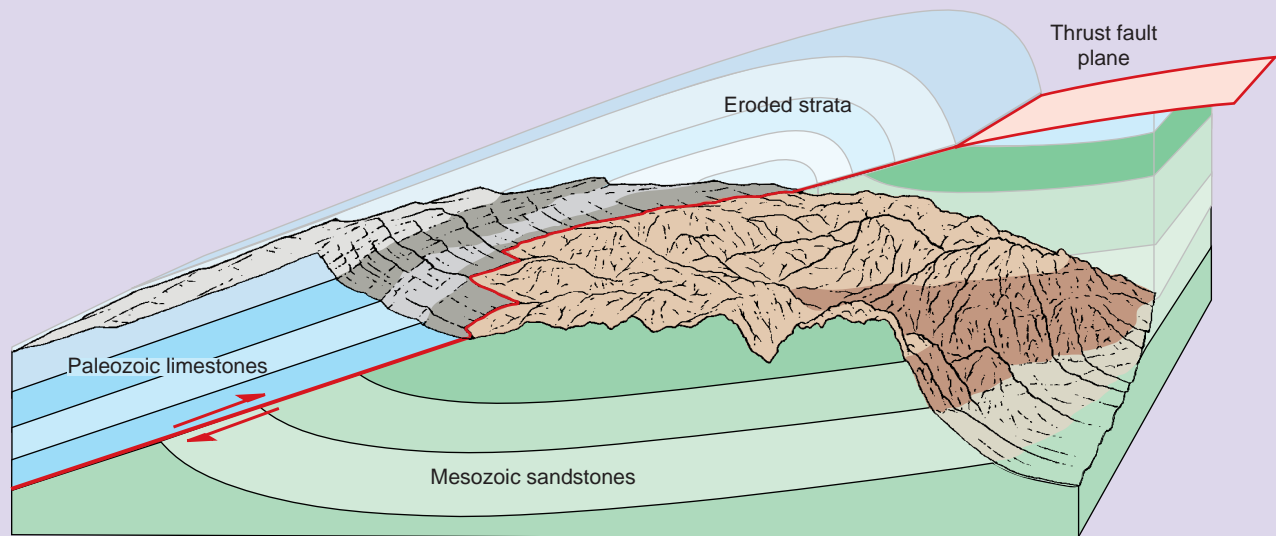
The Keystone Thrust in Nevada is one of the most dramatic expressions of thrust faulting in the world. Here a sequence of dark bluish-gray limestones and dolomites are thrust over a sequence of reddish sandstone. The important questions are: (1) What was this area like before erosion, and (2) How did this structure form?

## Observations

1. Older beds lie above younger beds! The fossils in the gray limestones show that they are about 350 million years *older* than the layers of red sandstone beneath them.
2. The contact between the two rock layers is not a sedimentary contact, but a zone of broken and crushed rock—a typical fault zone.
3. By carefully mapping the dip of the beds in the thrust sheet, a large fold is revealed.
4. The landscape has been eroded and huge volumes of rock have been removed from the area.

## Interpretations

The geologic interpretation is shown in the lower diagram. The three major structural elements are (1) the older sequence of rocks, (2) the younger sequence of rocks, and (3) the thrust fault. If we project these major geologic features to their position prior to erosion, the magnitude of the structure becomes much more apparent. One hundred million years ago, a convergent plate margin existed along the western margin of North America. As a result of intense compression at the plate boundary, a thrust fault broke through these layers of rock. Gradually, over the course of the next few millions of years, the great slab of gray limestones was pushed up and over the reddish sandstone layers. The great folded mountain belt that once dominated this region is now dismembered by extensive erosion.

## KEY TERMS

| | | | |
|---|---|---|---|
| anticline (p. 180) | fault (p. 173) | limb (p. 180) | strain (p. 168) |
| basin (p. 183) | fold (p. 178) | monocline (p. 180) | strength (p. 168) |
| brittle deformation (p. 168) | footwall (p. 174) | normal fault (p. 174) | stress (p. 168) |
| compression (p. 170) | fracture (p. 168) | orogenic belt (p. 180) | strike (p. 171) |
| contraction (p. 170) | graben (p. 174) | overturned fold (p. 181) | strike-slip fault (p. 174) |
| diapir (p. 185) | hanging wall (p. 174) | plunge (p. 180) | syncline (p. 180) |
| differential stress (p. 168) | hinge line (p. 180) | plunging fold (p. 180) | tension (p. 170) |
| dip (p. 171) | hinge plane (p. 180) | reverse fault (p. 174) | thrust fault (p. 174) |
| dome (p. 183) | horst (p. 174) | rule of Vs (p. 171) | vein ( p. 173) |
| ductile deformation (p. 168) | joint (p. 172) | scarp (p. 174) | |
| extension (p. 170) | lateral-slip (p. 170) | shear (p. 169) | |

## REVIEW QUESTIONS

1. List evidence that Earth's crust is in motion and has moved throughout geologic time.
2. What is the difference between a fault and a joint?
3. Compare the kinds of deformation produced by extension and those produced by contraction.
4. Contrast brittle and ductile behavior in rocks.
5. Do you think the fold shown in Figure 7.1 could have formed when the rocks were at the surface as they are now?
6. Explain the terms dip and strike.
7. Sketch a cross section of the structure of the rocks shown in Figure 7.15B.
8. List some of the surface features that commonly are produced by strike-slip faults.
9. Draw a simple block diagram of a normal fault, a thrust fault, and a strike-slip fault, and show the relative movement of the rock bodies along each. List the defining characteristics of each type of fault.
10. What are horsts and grabens? What global tectonic features are found where horsts and grabens most commonly are formed?
11. What global tectonic features are found where thrust faults commonly are formed?
12. What are the major features of an orogenic belt?
13. Make a perspective sketch of an anticline and the adjacent syncline, and label the following features: (a) hinge plane, (b) hinge, (c) angle of plunge, and (d) limbs.
14. Sketch the outcrop pattern of a plunging fold.
15. Describe the development of complex alpine-type folds.
16. What kinds of folds or other structures might form above a salt diapir?

## ADDITIONAL READINGS

Davis, G. H., and S. J. Reynolds. 1996. *Structural Geology of Rocks and Regions.* New York: Wiley.

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Van der Pluijm, B.A., and S. Marshak. 1997. *Earth Structure: An Introduction to Structural Geology and Tectonics.* New York: WCB/McGraw-Hill.

## MULTIMEDIA TOOLS

***Earth's Dynamic Systems Website***
The Companion Website at **www.prenhall.com/hamblin** provides you with an on-line study guide and additional resources for each chapter, including:

- On-line Quizzes (Chapter Review, Visualizing Geology, Quick Review, Vocabulary Flash Cards) with instant feedback
- Quantitative Problems
- Critical Thinking Exercises
- Web Resources

***Earth's Dynamic Systems CD***
Examine the CD that came with your text. It is designed to help you visualize and thus understand the concepts in this chapter. It includes:

- Animations showing the creation of folds and faults
- Slide shows with examples of different kinds of structures
- A direct link to the Companion Website