CHAPTER TWO

Primary and Nontectonic Structures

2.1 INTRODUCTION

In the first chapter, "Overview," we introduced a general definition for a **geologic structure** as any definable shape or fabric in a rock body. Most of the scope of structural geology is focused on **tectonic structures,** meaning those structures that form in response to forces generated by plate interactions (such as convergence, collision, rifting, strike-slip movement, subduction, buoyancy—concepts that will be discussed later). But before we get to tectonic structures, we have to examine structures that form during or shortly after the deposition of rocks, and we want to mention structures that are not an immediate consequence of plate interactions. Collectively, these structures may be called **nontectonic structures,** because it is inferred that they form in the absence of tectonic stresses, but strictly speaking this usage is misleading. Most geologic structures are indirectly, if not directly, a consequence of tectonic activity. Examples are the creation of slopes down which sediments slide and the occurrence of volcanic activity leading to the flow of basalt, both of which are manifestations of movements in the Earth. All of these phenomena are ultimately a manifestation of movement in the Earth. Under the broad masthead of nontectonic structures we discuss depositional, penecontemporaneous, intrusive, and gravity-slide structures for both sedimentary and igneous rocks. Impact structures, which result from the collision of extraterrestrial objects with Earth's surface, are briefly discussed at the end of the chapter.

2.2 SEDIMENTARY STRUCTURES

When you look at an outcrop of sedimentary rock, the most obvious fabric to catch your eye is the primary layering or stratification (Figure 2.1), which is called **bedding.** Some of the more commonly used terms that are related to stratification are listed in Table 2.1.

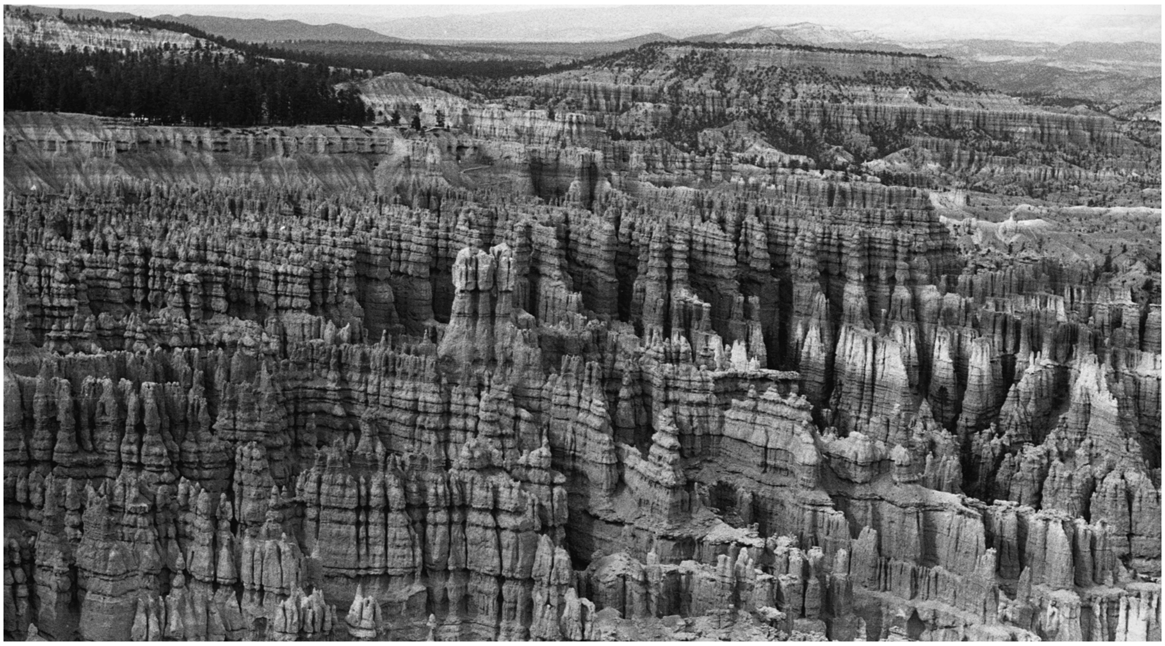


FIGURE 2.1 Differential erosion of bedding surfaces in the Wasatch Formation, Bryce Canyon (Utah).

TABLE 2.1 SOME TERMINOLOGY OF STRATIFICATION

Bedding	Primary layering in a sedimentary rock, formed during deposition, manifested by changes in texture, color, and/or composition; may be emphasized in outcrop by the presence of parting
Compaction	Squeezing unlithified sediment in response to pressure exerted by the weight of overlying layers
Overturned beds	Beds that have been rotated past vertical in an Earth–surface frame of reference; as a consequence, facing is down
Parting	The tendency of sedimentary layers to split or fracture along planes parallel to bedding; parting may be due to weak bonds between beds of different composition, or may be due to a preference for bed-parallel orientation of clay
Strata	A sequence composed of layers of sedimentary rock
Stratigraphic facing	The direction to younger strata, or, in other words, the direction to the depositional top of beds
Younging direction	Same as stratigraphic facing

What defines bedding in an outcrop? In the Painted Desert of Arizona, beds are defined by spectacular variations in colors, and outcrops display garish stripes of maroon, red, green, and white. In the Grand Canyon and the Rocky Mountains, beds are emphasized by contrasts in resistance to erosion; sandstone and limestone beds form vertical cliff faces while shale layers form shallow slopes. On the cliffs that form the east edge of the Catskill Mountains in New York State, there are abrupt contrasts in grain size between adjacent beds, with a coarse conglomerate juxtaposed against siltstone or shale. Strictly speaking, a bed is the smallest subdivision of a sedimentary unit.[1] It has a definable top or bottom and can be distinguished from adjacent beds by differences in grain size, composition, color, sorting, and/or by a physical parting surface.

[1]Stratigraphers divide sequences of strata into the following units: supergroup, group, formation, member, bed. Several beds make a member, several members make a formation, and so on. Criteria for defining units are somewhat subjective. Basically, a unit is defined as a sequence of strata that can be identified and mapped at the surface or in the subsurface over a substantial region. The basis for recognizing a unit can be its age, its component sequence of lithologies, and the character of its bedding.

All of the features defining bedding, with the exception of parting, are a consequence of changes in the source of the sediment (or provenance) or the depositional environment.

In some outcrops, bedding is enhanced by the occurrence of **bedding-parallel parting.** Parting forms when beds are unroofed (i.e., overlying strata are eroded away) and uplifted to shallower depths in the crust. As a consequence, the load pushing down on the strata decreases and the strata expand slightly. During this expansion, fractures form along weak bedding planes and define the parting. This fracturing reflects the weaker bonds between contrasting lithologies of adjacent beds, or the occurrence of a preferred orientation of sedimentary grains (e.g., mica). If a sedimentary rock has a tendency to have closely spaced partings, it is said to display **fissility.** Shale, which typically has a weak bedding-parallel fabric due to the alignment of constituent clay or mica flakes, is commonly fissile.

There are three reasons why platy grains like mica have a preferred orientation in a sedimentary rock. First, the alignment of grains can reflect settling of asymmetric bodies in Earth's gravity field. Platy grains tend to lie down flat. To understand why, throw a handful of coins or cards into the air: after the coins fall they lie flat against the floor. You will rarely see a coin stand on its edge. Second, the alignment of grains can reflect flow of the fluid in which the grains were deposited. In a moving fluid, grains are reoriented so that they are hydrodynamically stable, meaning that the traction caused by the moving fluid is minimized (as is the case if the broad face of the grain parallels the flow direction). Typically, grains end up being imbricated, meaning that they overlap one another like roof shingles. **Imbrication** is a useful primary sedimentary feature that can be used to define **paleocurrent direction,** which is the direction of the current when the sediment was deposited. Third, the alignment of grains can form as a consequence of compaction subsequent to deposition. As younger sediment is piled on top, water is progressively squeezed out of the older sediment below and the grains mechanically rotate into an orientation with their flat surfaces roughly perpendicular to the applied load.

Bedding in outcrops is often highlighted by differential weathering and erosion (Figure 2.1). For example, chemical weathering (like dissolution of carbonate) of a sequence of strata containing alternating limestones and quartz sandstones will result in an outcrop face on which the quartz-rich layers stand out in relief. Fresh limestone and dolomite are almost identical in color, but weathering of a sequence of alternating limestones and dolostones will result in a color-banded outcrop, because the dolostones tend to weather to a buff-tan color and the limestones become grayish. Erosion of a sequence of alternating sandstones and shales may result in a stair-step outcrop face, because the relatively resistant sandstone beds become vertical cliffs and the relatively weak shale beds create slopes. Geologists have developed a jargon for describing specific types of bedding, as defined in Table 2.2.

TABLE 2.2 TERMS TO DESCRIBE TYPES OF BEDDING

Massive beds	Beds that are relatively thick (typically several m) and show no internal layering. Massive bedding develops in sedimentary environments where large quantities of sediment are deposited very rapidly or in environments where bioturbation (churning of the sediments by worms and other organisms) occurred.
Medium beds	Beds that are 10–30 cm thick
Rhythmic beds	A sequence of beds in which the contrast between adjacent beds is repeated periodically for a substantial thickness of strata
Thick beds	Beds that are 30–100 cm thick. Very thick beds are tens of m thick.
Thin beds	Beds that are less than 3 cm thick
Thinly laminated beds	Beds that are less than 0.3 cm thick

2.2.1 The Use of Bedding in Structural Analysis

Recognition of bedding is critical in structural analysis. Bedding provides a reference frame for describing deformation of sedimentary rocks, because when sediments are initially deposited, they form horizontal or nearly horizontal layers, a concept referred to as the **Law of Original Horizontality.** Thus, if we look at an outcrop and see tilting or folding, what we are noticing are deviations in bedding attitude from original horizontality.

In complexly deformed and metamorphosed sedimentary rocks, geologists have to search long and hard to find subtle preserved manifestations of original bedding (e.g., variations in grain size, color, composition). Only by finding bedding can a geologist unravel the folding history of a region. When found, bedding is labeled S_0 (pronounced ess-zero), where the S is an abbreviation for surface. Later we will discuss surfaces in rocks that are formed by deformation, like rock cleavage, which are labeled S_1, S_2, and so on (Chapter 11).

The study of certain depositional structures within beds and on bedding surfaces is useful in tectonic analysis because they may provide important information on **depositional environment** (the setting in which the sediment was originally deposited), on **stratigraphic facing** or **younging direction** (the direction in which strata in a sequence are progressively younger), and on **current direction** (the direction in which fluid was flowing during deposition). Facing indicators allow you to determine whether a bed is right-side-up (facing up) or overturned (facing down), with respect to the Earth's surface. Recognition of facing is powerful both for stratigraphic studies and for structural studies. For example, the structural interpretation of a series of parallel beds in two adjacent outcrops depends on the facing—if the facing is the same in both outcrops, then the strata are probably **homoclinal,** meaning that they have a uniform dip. However, if the facing is opposite, then the two outcrops are likely on different limbs of a fold whose hinge area is not exposed. We'll return to the use of facing in structural analysis in Section 10.2.

2.2.2 Graded Beds and Cross Beds

Patterns within beds may contain information about stratigraphic facing and current directions that are often critical for tectonic interpretations. **Graded beds** display progressive fining of clast/grain size from the base to the top (Figure 2.2), and are a consequence of deposition from turbidity flows. A **turbidity flow** is a cloud of sediment that moves down a slope under water because the density of the sediment-water mixture is greater than that of clean water, and denser liquids sink through less dense liquids. Turbidite flows are triggered by major storms or earthquakes (because of their association with seismicity, the occurrence of turbidites may indicate that the sediment source region was tectonically active), and move down gentle slopes at considerable speed. Typically, a flow is confined to a submarine channel or canyon; when a broadening of the channel or a decrease in slope slows the speed of a turbidity current, the sediment cloud settles. During settling, the largest grains fall first, and the finest grains last. Each turbidity flow produces a separate graded sequence or a **turbidite,** which is often capped by pelagic sediment, meaning deep-marine sediments like

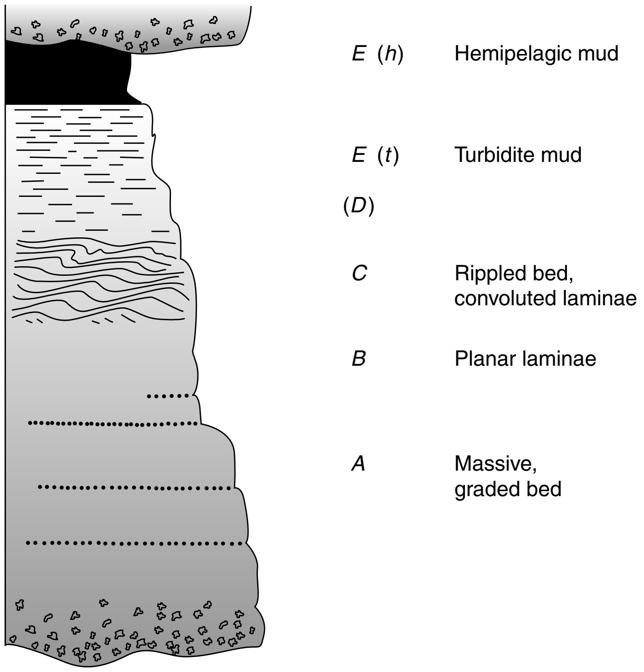


FIGURE 2.2 Graded bedding in a turbidite (or Bouma) sequence.

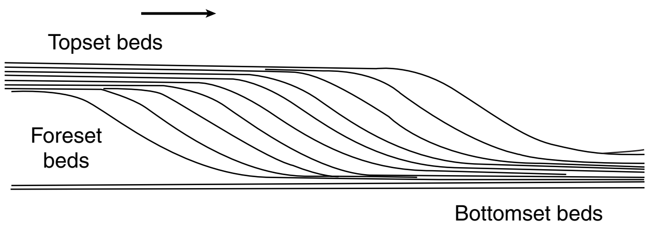
clay and plankton shells. Turbidites display an internal order, called a **Bouma sequence**[2] (Figure 2.2), which reflects changing hydrodynamic conditions as the turbidity current slows down.

In pre–plate tectonics geological literature (i.e., pre-Beatles), thick sequences of turbidites were referred to as **flysch,** a term originating from Alpine geology (see Part E). Flysch was considered to be an orogenic deposit, meaning a sequence of strata that was deposited just prior to and during the formation of a mountain range. Exactly why such strata were deposited, however, was not understood. Modern geologists now realize that the classical flysch sequences are actually turbidites laid down in a deep trench marking an active plate boundary (like a subduction zone). Turbidite flows are common in ocean trenches, which represent seismically active, convergent plate margins. During and after deposition, the trench turbidites are scraped up and deformed by continued convergence between plates, and may eventually be caught in a continental collision zone (see Chapter 16).

Cross beds are surfaces within a bed that are oblique to the overall bounding surfaces of the bed (Figure 2.3). Cross beds, which are defined by subtle partings or concentrations of grains, form when sediment moves from the windward or upstream side of a dune, ripple, or delta to a face on the leeward or downstream side, where the current velocity is lower and the

[2]Named after the sedimentologist Arnold Bouma.

sediment settles out. Thin beds parallel to the upper bounding surface are called **topset beds,** the inclined layers deposited parallel to the slip face are called **foreset beds,** and the thin beds parallel to the lower bounding surface are called **bottomset beds** (Figure 2.3a). The foreset beds, which typically are curved (concave up) and merge with the topset and bottomset beds, are the cross beds. If the topset beds and the upper part of the foreset beds are removed by local erosion, the bottomset beds of the next higher layer of sediment are juxtaposed against the foreset beds of the layer below. Thus, cross beds tend to be truncated at the upper bedding surface, whereas they are asymptotic to the lower bedding surface (Figure 2.3b). This geometry provides a clear stratigraphic facing indicator. The current direction in a cross-bedded layer is taken to be approxi-



(a)

(b)

FIGURE 2.3 (a) Terminology of cross bedding, and (b) cross beds in the Coconino Sandstone, Oak Creek Canyon (Arizona).

TABLE 2.3	**COMMON SURFACE MARKINGS**
Animal tracks	Patterns formed when critters like trilobites, worms, and lizards tromp over and indent the surface (the characteristic trails of these organisms are a type of *trace fossil*).
Clast imbrication	The shingle-like overlapping arrangement of tabular clasts on the surface of a bed in response to a current. Imbrication develops because tabular clasts tend to become oriented so that the pressure exerted on them by the moving fluid is minimized.
Flute casts	Asymmetric troughs formed by vortices (mini tornadoes) within the fluid that dig into the unconsolidated substrate. The troughs are deeper at the upstream end, where the vortex was stronger. They get shallower and wider at the downstream end, because the vortex dies out. Flute casts can be used as facing indicators.
Mudcracks	Desiccation of mud causes the mud to crack into an array of polygons and intervening mudcracks. Each polygon curls upwards along its margins, so that the mudcracks taper downwards and the polygons resemble shallow bowls. Mudcracks can be used as facing indicators, because an individual crack tends to taper downwards (Figure 2.4).
Raindrop impressions	Circular indentations on the bed-surface of mudstone, formed by raindrops striking the surface while it was still soft.
Ripple marks	Ridges and valleys on the surface of a bed formed as a consequence of fluid flow. If the current flows back and forth, as along a beach, the ripples are *symmetric,* but if they form in a uniformly flowing current, they are *asymmetric* (Figure 2.5). The crests of symmetric ripples tend to be pointed, whereas the troughs tend to be smooth curves. Thus, symmetric ripples are good facing indicators. Asymmetric ripples are not good facing indicators, but do provide current directions.
Traction lineation	Subtle lines on the surface of a bed formed either by trails of sediment that collect in the lee of larger grains, or by alignment of inequant grains in the direction of the current to diminish hydraulic drag.
Worm burrows	The traces of worms or other burrowing organisms that live in unconsolidated sediment. They stand out because of slight textural and color contrasts with the unburrowed rock.

mately perpendicular to the intersection between the truncated foresets and overlying bed.

2.2.3 Surface Markings

Local environmental phenomena, such as rain, desiccation (i.e., drying), current traction, and the movement of organisms, affect the surface of a bed of sediment. If the sediment is unlithified, these phenomena leave an imprint that is known as a **surface marking.** Some of the more common surface markings are listed in Table 2.3.

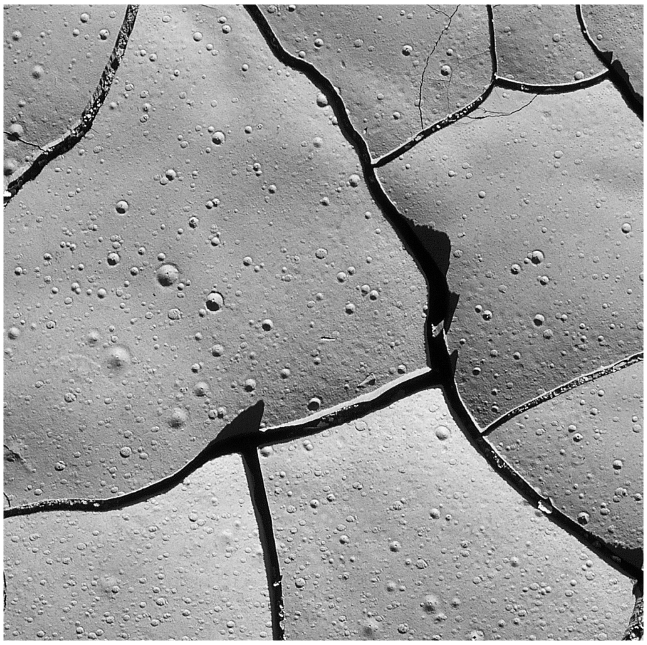


FIGURE 2.4 Mudcracks, separating a mud layer into platelets. The circular indentations are raindrop imprints.

FIGURE 2.5 Asymmetric ripple marks. The arrow indicates approximate current direction during deposition.

2.2.4 Disrupted Bedding

Load casts, which are also called ball-and-pillow structures, are bulbous protrusions extending downward from a sand layer into an underlying mud or very fine sand layer (Figure 2.6). They form prior to lithification where a denser sand lies on top of less dense mud and a disturbance by a storm or an earthquake causes blobs of sand to sink into the underlying mud. Load casts are useful stratigraphic facing indicators when they retain some connection to the host layer.

Where sand and mud layers are progressively buried, it is typical for the mud layers to compact and consolidate before the sand layers do. As a consequence, the water in the sand layer is under pressure. If an earthquake, storm, or slump suddenly cracks the permeability barrier surrounding the sand, water and sand are released and forced into the mud along cracks. When this happens near the Earth's surface, little mounds of sand, called **sand volcanoes,** erupt at the ground surface. The resulting wall-like intrusions of sand (or in some localities, even conglomerate) are called **clastic dikes** (Figure 2.7). At depth, partially consolidated beds of sand and mud break into pieces, resulting in a chaotic layering that is known, simply, as **disrupted bedding** (Figure 2.8).

Studies of disrupted bedding, sedimentary dikes, and sand volcanoes in lake and marsh deposits provide an important basis for determining the recurrence interval of large earthquakes. In these studies, investigators dig a trench across the deposit and then look for disrupted intervals within the sequence. Radiocarbon dating of organic matter in the disrupted layers defines the absolute age of disruption events and allows us to estimate the recurrence of earthquakes.

2.2.5 Conformable and Unconformable Contacts

Earlier we defined a contact as any surface between two geologic units. There are three basic types of contacts: (1) **depositional contacts,** where a sediment layer is deposited over preexisting rock; (2) **fault contacts,** where two units are juxtaposed by a fracture on which sliding has occurred; and (3) **intrusive contacts,** where one rock body cuts across another rock body. In this section, we consider in more detail the nature and interpretation of depositional contacts.

Relatively continuous sedimentation in a region leads to the deposition of a sequence of roughly parallel

FIGURE 2.6 Load cast or ball-and-pillow structure; wine stakes for scale (Eifel, Germany).

FIGURE 2.7 Clastic dike in Proterozoic sandstones. Note that the dike sharply cuts across bedding and that very coarse clasts are preserved in its center (Sudbury, Ontario).

FIGURE 2.8 Disrupted bedding in turbidite; hammer for scale (Cantabria, Spain).

sedimentary units in which the contacts between adjacent beds do not represent substantial gaps in time. Gaps in this context can be identified from gaps in the fossil succession. The boundary between adjacent beds or units in such a sequence is called a **conformable contact.** For example, we say, "In eastern New York, the Becraft Limestone was deposited conformably over the New Scotland Formation." The New Scotland Formation is an argillaceous limestone representing marine deposition below wave base, whereas the Becraft Limestone is a pure, coarse-grained limestone representing deposition in a shallow-marine beach environment. Bedding in the two units is parallel, and the contact between these two units is gradational.

If there is an interruption in sedimentation, such that there is a measurable gap in time between the base of the sedimentary unit and what lies beneath it, then we say that the contact is **unconformable.** For example, we say, "In eastern New York, the Upper Silurian Rondout Formation is deposited unconformably on the Middle Ordovician Austin Glen Formation," because Upper Ordovician and Lower Silurian strata are absent. Unconformable contacts are generally referred to as **unconformities,** and the gap in time represented by the unconformity (that is, the difference in age between the base of the strata above the unconformity and the top of the unit below the unconformity) is called a **hiatus.** In order to convey a meaningful description of a specific unconformity, geologists distinguish among four types of unconformities that are schematically shown in Figure 2.9 and defined in Table 2.4.

Unconformities represent gaps in the rock record that can range in duration from thousands of years to billions of years. Examples of great unconformities, representing millions or billions of years, occur in the Canadian shield, where Pleistocene till buries

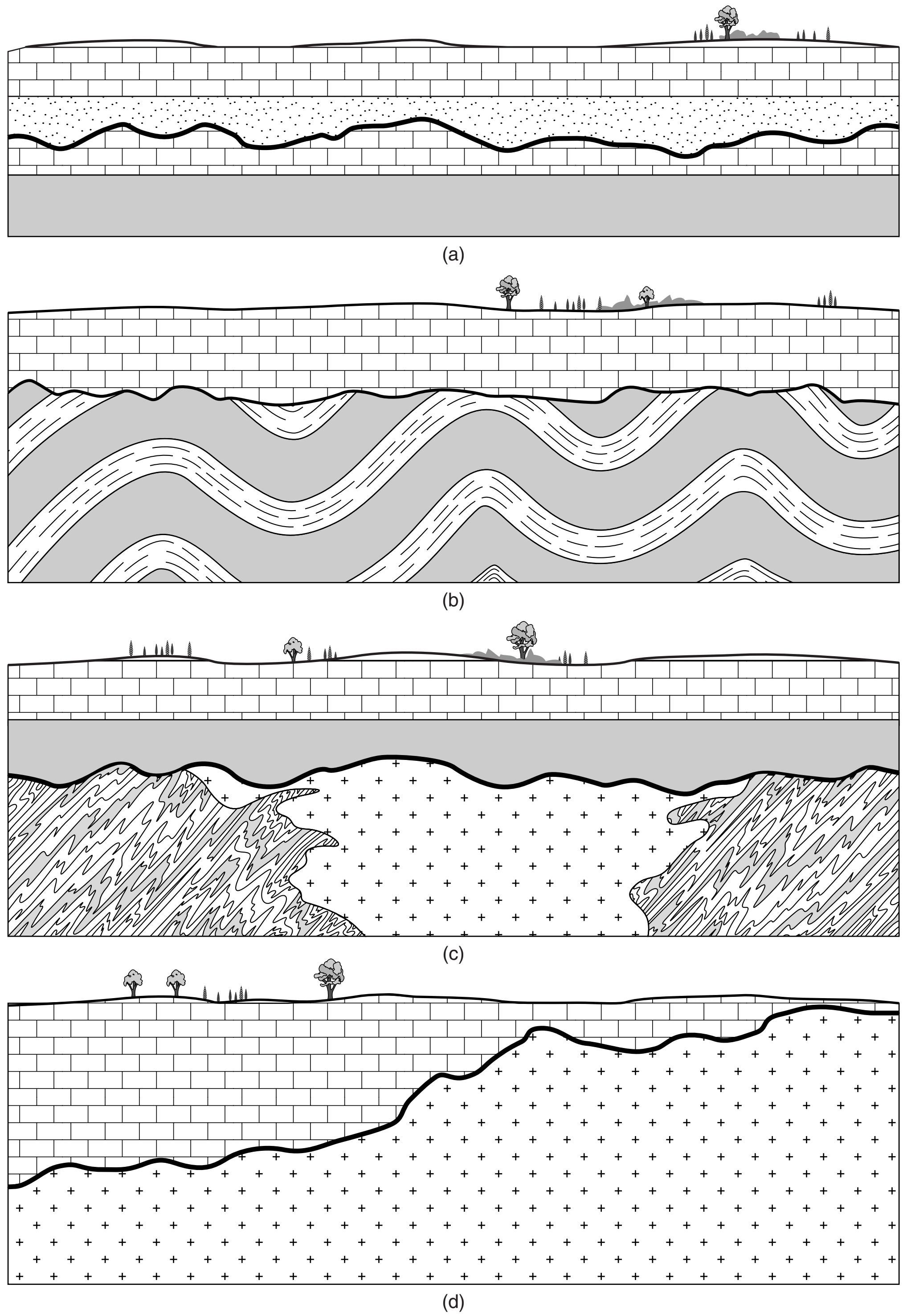


FIGURE 2.9 The principal types of unconformities: (a) disconformity, (b) angular unconformity, (c) nonconformity, (d) buttress unconformity.

TABLE 2.4	TYPES OF UNCONFORMITIES
Disconformity	At a disconformity, beds of the rock sequence above and below the unconformity are parallel to one another, but there is a measurable age difference between the two sequences. The disconformity surface represents a period of nondeposition and/or erosion (Figure 2.9a).
Angular unconformity	At an angular unconformity, strata below the unconformity have a different attitude than strata above the unconformity. Beds below the unconformity are truncated at the unconformity, while beds above the unconformity roughly parallel the unconformity surface. Therefore, if the unconformity is tilted, the overlying strata are tilted by the same amount. Because of the angular discordance at angular unconformities, they are quite easy to recognize in the field. Their occurrence means that the sub-unconformity strata were deformed (tilted or folded) and then were truncated by erosion prior to deposition of the rocks above the unconformity. Therefore, angular unconformities are indicative of a period of active tectonism. If the beds below the unconformity are folded, then the angle of discordance between the super- and sub-unconformity strata will change with location, and there may be outcrops at which the two sequences are coincidentally parallel (Figure 2.9b).
Nonconformity	Nonconformity is used for unconformities at which strata were deposited on a basement of older crystalline rocks. The crystalline rocks may be either plutonic or metamorphic. For example, the unconformity between Cambrian strata and Precambrian basement in the Grand Canyon is a nonconformity (Figure 2.9c).
Buttress unconformity	A buttress unconformity (also called *onlap unconformity*) occurs where beds of the younger sequence were deposited in a region of significant predepositional topography. Imagine a shallow sea in which there are islands composed of older bedrock. When sedimentation occurs in this sea, the new horizontal layers of strata terminate at the margins of the island. Eventually, as the sea rises, the islands are buried by sediment. But along the margins of the island, the sedimentary layers appear to be truncated by the unconformity. Rocks below the unconformity may or may not parallel the unconformity, depending on the pre-unconformity structure. Note that a buttress unconformity differs from an angular unconformity in that the younger layers are truncated at the unconformity surface (Figure 2.9d).

Proterozoic and Archean gneisses. In Figure 2.10 the classic unconformity between Paleozoic sedimentary rocks and Precambrian gneisses is shown and many introductory geology books show this contact in the Grand Canyon.

It is a special experience to put your finger on a major unconformity and to think about how much of Earth's history is missing at the contact. Imagine how James Hutton felt when, in the late eighteenth century, he stood at Siccar Point along the coast of Scotland (Figure 2.11), and stared at the Caledonian unconformity between shallowly dipping Devonian Red Sandstone and vertically dipping Silurian strata and, as the present-day waves lapped on and off the outcrop and deposited new sand, suddenly realized what the contact meant. His discovery is one of the most fundamental in field geology.

FIGURE 2.10 Unconformable contact between mid-Proterozoic Grenville gneiss (dark gray) and Cambrian sandstone and Pleistocene soils (southern Ontario, Canada).

How do you recognize an unconformity (Figure 2.12) in the field today? Well, if it is an angular unconformity or a buttress unconformity, there is an angular discordance between bedding above and below the unconformity. A nonconformity is obvious, because crystalline rocks occur below the contact. Disconformities, however, can be more of a challenge to recognize. If strata in the sequence are fossiliferous, and you can recognize the fossil species and know their age, then you can recognize a gap in the fossil succession. Commonly, an unconformity may be marked by a surface of erosion, as indicated by scour features, or by a **paleosol,** which is a soil horizon that formed from weathering prior to deposition of the overlying sequence. Some unconformities are marked by the occurrence of a **basal conglomerate,** which contains clasts of the rocks under the unconformity. Recognition of a basal conglomerate is also helpful in determining whether the contact between strata and a plutonic rock is intrusive or whether it represents a nonconformity.

FIGURE 2.11 Angular unconformity in the Caledonides at Siccar Point (Scotland). The hammerhead rests on the unconformity, which is tilted due to later deformation.

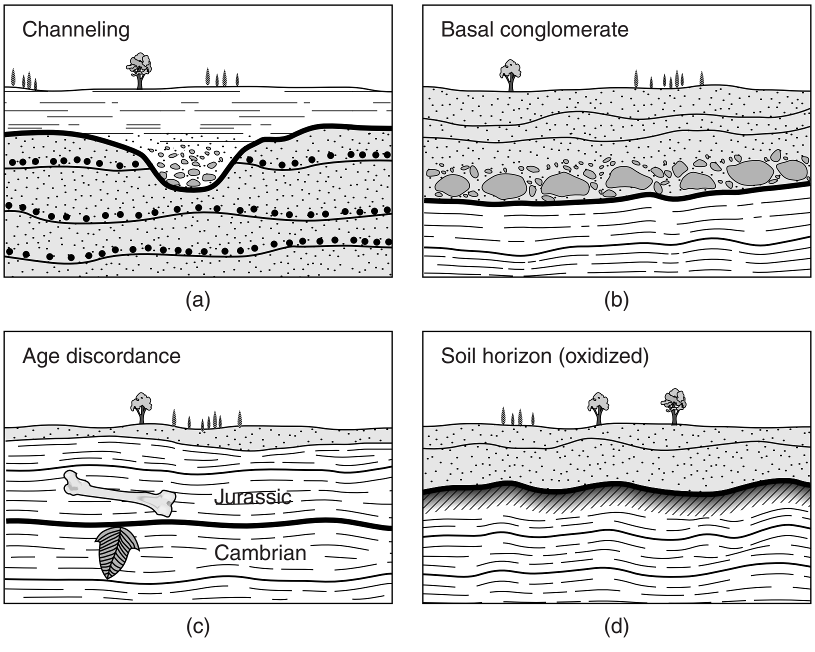


FIGURE 2.12 Some features used to identify unconformities: (a) scour channels in sediments, (b) basal conglomerate, (c) age discordance from fossil evidence, and (d) soil horizon or paleosol.

2.2.6 Compaction and Diagenetic Structures

When a clastic sediment initially settles, it is a mixture primarily of grains and water. The proportion of solid to fluid varies depending on the type of sediment. Gooey mud, which consists of clay and water, contains more water than well-packed sand. Progressive burial of sediment squeezes the water out, and the sediment compacts. **Compaction** results in a decrease in porosity (>50% in shale and >20% in sand) that results in an increase in the density of the sediment.

Lateral variation in the amount of compaction within a given layer, or contrasts in the amount of compaction in a vertical section, is a phenomenon called **differential compaction.** Differential compaction within a layer can lead to lateral variation in thickness that is called **pinch-and-swell structure.** Pinch-and-swell structure can also form as a consequence of tectonic stretching, so again, you must be careful when you see the structure to determine whether it is a depositional structure or a tectonic structure.

The compaction of mud leads to development of a preferred orientation of clay in the resulting mudstone. Clay occurs in tiny flakes shaped like playing cards. In a wet sediment the flakes are not all parallel to one another, as in a standing house of cards, but after compaction the flakes are essentially parallel to one another, as in a collapsed house of cards. The preferred orientation of clay flakes, as we have seen, leads to bedding plane fissility that produces a **shale.** For example, in the Gulf Coast sequence of the southern United States this progression is preserved in drill cores that were obtained to study the relationship between oil maturation and clay mineralogy. In contrast, the compaction of sand composed of equant grains causes the grains to pack together more tightly, but produces little rock fabric.

Deeper compaction can cause **pressure solution,** a process by which soluble grains preferentially dissolve along the faces at which stress is the greatest. In pure limestones or sandstones, this process causes grains to suture together, meaning that grain surfaces interlock with one another like jigsaw puzzle pieces. In conglomerates, the squeezing together of pebbles results in the formation of indentations on the pebble surfaces creating **pitted pebbles** (Figure 2.13). In limestones and sandstones that contain some clay, the clay enhances the pressure solution process. Specifically, pressure solution occurs faster where the initial clay concentration is higher. As a result, distinct seams of clay residue develop in the rock. These seams are called **stylolites** (Figure 2.14). In rocks with little clay (<10%), stylolites are jagged and tooth-like in cross section, like the sutures in your skull. The teeth are caused by the distribution of grains of different solubility along the stylolite. In rocks with more clay, the stylolites are wavy and the teeth are less pronounced, because the clay seams become thicker than tooth amplitude. Some of the dissolved ions that are removed at pressure-solved surfaces precipitate locally in the rock in veins or as cement in pore spaces, whereas some get transported out of the rock by moving groundwater. The proportion of reprecipitated to transported ions is highly variable, but as much as 40% of the rock can be dissolved and removed during formation of stylolites.

FIGURE 2.13 Pitted pebble; coin for scale.

FIGURE 2.14 Suture-like stylolites in limestone. Pocket knife for scale.

Some sedimentary rocks exhibit color banding that cuts across bedding. This color banding, which is called **Liesegang banding,** is the result of diffusion of impurities, or of reactions leading to alternating bands of oxidized and reduced iron. Because it can be mistaken for bedding or cross bedding, it is mentioned in the context of primary sedimentary structures. To avoid mistaken identity, search the outcrop to determine whether sets of bands cross each other (possible for Liesegang bands, but impossible for bedding), and whether the bands are disrupted at fractures or true bedding planes, because these are places where the diffusion rate changes.

2.2.7 Penecontemporaneous Structures

If sediment layers have an initial dip, meaning a gentle slope caused by deposition on a preexisting slope or tilting prior to full lithification in a tectonically active region, gravity can pull the layers down the slope. The ease with which sediments move down a slope is increased by fluid pressure in the layers, which effectively keeps the layers apart. Movement is resisted by weak electrostatic adhesion between grains, but this resistance can be overcome by the energy of an earthquake or a storm, and the sediment will move down the slope. If the sediment completely mixes with water and becomes a turbid suspension flowing into deeper water, then all of the preexisting primary structure is lost and the grains are resedimented as a new graded bed (turbidite; see Figure 2.2) farther down the slope. If the flowing mixture of sediment and water is dominantly sediment, it churns into a slurry containing chunks and clasts that are suspended in a matrix. Such slurries are called **debris flows,** and where preserved in a stratigraphic sequence, they become matrix-supported, poorly sorted conglomerates containing a range of clast sizes and shapes.

If the beds were lithified sufficiently prior to movement, so that they maintain cohesion, then the movement is called **slumping.** During slumping, the sedimentary layers tend to be folded and pulled apart and are thrust over one another. The folds and faults formed during this slumping are called **penecontemporaneous structures,** because they formed almost (Greek prefix *pene*) at the same time as the original deposition of the layers. Penecontemporaneous folds and faults are characteristically chaotic. The folds display little symmetry, and folds in one layer are of a different size and orientation than the structures in adjacent layers. Penecontemporaneous faults are not associated with pronounced zones of brittle fracturing (we turn to the characteristics of brittle behavior in Chapter 6). One key to the recognition of slump structures in a sedimentary sequence is that the deformed interval is *intraformational,* meaning that it is bounded both above and below by relatively undeformed strata (Figure 2.15). Commonly, intervals of penecomtemporaneous structures occur in a sequence that also includes debris flows and turbidites, all indicative of an unstable depositional environment. While slump structures can be mistaken for local folding adjacent to a tectonic detachment fault, the opposite, tectonic folds mistakenly interpreted as slump structures, may also occur. Not a simple matter to distinguish between the two!

We tend to think of debris flows and landslides as being relatively small structures, capable of disrupting a hillslope and perhaps moving a cottage or two, but generally not much more. However, the geologic record shows that catastrophic landslides of enormous dimension have occurred on occasion. In northern Wyoming, for example, a giant Eocene slide in association with volcanic eruption displaced dozens of mountain-sized blocks and hundreds of smaller blocks. One such large block, Heart Mountain, moved intact for several tens of kilometers, apparently riding on a cushion of compressed air above a nearly planar subhorizontal (detachment) fault (Figure 2.16).

FIGURE 2.15 Penecontemporaneous folds in the Maranosa Arenaci (Italian Apennines).

FIGURE 2.16 Heart Mountain detachment; with Paleozoic carbonates on Eocene deposits (Wyoming).

2.3 SALT STRUCTURES

Salt is a sedimentary rock that forms by the precipitation of evaporite minerals (typically halite [NaCl] and gypsum or anhydrite calcium sulfates) from saline water. Salt deposits accumulate in any **sedimentary basin,** meaning a low region that is the site of deposition, where saline water, such as seawater, evaporates sufficiently for salt to precipitate. Particularly thick salt deposits lie at the base of **passive-margin basins,** so named because they occur along tectonically inactive edges of continents. To see how these basins form, imagine a supercontinent that is being pulled apart, like Pangea in the early Mesozoic. This process, called **rifting,** involves brittle and ductile faulting, the net result of which is to thin the continental lithosphere until it breaks and an oceanic ridge is formed. During the early stages of rifting, the rift basin is dry or contains freshwater lakes. Eventually, the floor of the rift drops below sea level and a shallow sea forms. If evaporation rates are high, various salts (typically, halite and gypsum/anhydrite) precipitate out of the seawater and are deposited on the floor of the rift. When the rift evolves into an open ocean, the continental margins become passive margins that gradually subside. With continued subsidence, the layer of evaporite (salt) is buried by clastic sediments and carbonates typical of continental-shelf environments. We'll discuss this tectonic environment in more detail later (Extension Tectonics, Chapter 16), but for now we leave you a picture of a thick pile of sediment with a layer of salt near its base. This is the starting condition for the formation of **salt intrusions.**

Salt differs from other sedimentary rocks in that it is much weaker and, as a consequence, is able to flow like a viscous fluid under conditions in which other sedimentary rocks behave in a brittle fashion.[3] In some cases, deformation of salt is due to tectonic faulting or folding, but because salt is so weak, it may deform solely in response to gravity, and thereby cause deformation of surrounding sedimentary rock. If gravity is the only reason for salt movement, the deformation resulting from its movement is called **halokinesis** (combining the Greek words for salt and movement, respectively) and the resulting body of salt is called a **salt structure.**

[3]In fact, dry salt moves crystal plastically, whereas movement of damp salt involves pressure solution. We discuss these deformation mechanisms in Chapter 9.

2.3.1 Why Halokinesis Occurs

Halokinesis begins in response to three factors: (1) the development of a density inversion, (2) differential loading, and (3) the existence of a slope at the base of a salt layer. All three of these factors occur in a passive-margin basin setting. Salt is a nonporous and essentially incompressible material. So when it gets buried deeply in a sedimentary pile, it does not become denser. In fact, salt actually gets less dense with depth, because at greater depths it becomes warmer and expands. Other sedimentary rocks (like sandstone and shale), in contrast, form from sediments that originally had high porosity and thus become denser with depth because the pressure caused by overburden makes them compact. This contrast in behavior, in which the density of other sedimentary rocks exceeds the density of salt at depths greater than about 6 km, results in a **density inversion,** meaning a situation where denser rock lies over less dense rock. Salt density is about 2200 kg/m^3, whereas the density of the sedimentary rocks is about 2500 kg/m^3. A density inversion is an unstable condition because the salt has positive buoyancy. **Positive buoyancy** means that forces in a gravity field cause lower density materials to try to rise above higher density materials, thereby decreasing the overall gravitational potential energy of the system. **Negative buoyancy,** the reverse, is a force that causes a denser material to sink through a less dense material (see Section 2.2.4). A familiar example of positive buoyancy forces is the push that your hand feels when you try to hold an air-filled balloon under water. Holding a brick under water illustrates negative buoyancy. When the positive buoyancy force exceeds the strength of the salt and is sufficient to upwarp strata that lie over the salt structure, then it will contribute to the formation of the salt structure.

Differential loading of a salt layer takes place when the downward force on the salt layer caused by the weight of overlying strata varies laterally. This may occur where there are primary variations in the thickness or composition of the overlying strata, primary variations in the original surface topography of the salt layer, or changes in the thickness of the overlying strata due to faulting. Regardless of its cause, differential loading creates a situation in which some parts of the salt layer are subjected to a greater vertical load than other parts, and the salt is squeezed from areas of higher pressure to areas of lower pressure. For example, imagine a layer of salt whose upper surface initially bulges upward to form a small "dome." The weight of a column of rock and water from sea level down to a horizontal surface in the salt layer on either side of the dome is greater than the weight of the column on the top of the dome, because salt is less dense than other sedimentary rocks. As a consequence, salt is squeezed into the dome, making it grow upwards. A salt layer that has provided salt for the production of a salt structure, and thus has itself been changed by halokinesis, is called the **source layer.**

The combination of differential loading and buoyancy force drives salt upward through the overlying strata until it reaches a level of **neutral buoyancy,** meaning the depth at which it is no longer buoyant. At this level, salt has the same density as surrounding strata. The density of mildly compacted clastic strata equals that of salt at depths of around 500–1500 m below the surface of the basin, depending on composition. At the level of neutral buoyancy, salt may begin to flow laterally, much like a thick pile of maple syrup flows laterally over your pancakes. This process, which is also driven by gravity (above the level of neutral buoyancy, the salt is subjected to a negative buoyancy force), is known as **gravity spreading.** Where salt is extruded at the land surface, it becomes a salt glacier (Figure 2.17). At the seafloor, salt also spreads like a salt glacier, except that during movement it continues to be buried by new sediment.

2.3.2 Geometry of Salt Structures and Associated Processes

In response to positive buoyancy force and to differential loading, salt will flow upward from the source bed, which thins as a consequence. If the source bed thins to the point of disappearing and the strata above the source bed and below the source bed become juxtaposed, we say that the contact between these two beds is a **primary weld.** In general, a weld is any contact between strata that were once separated by salt. At any given time, a region may contain salt structures at many stages of this development. Geologists working with salt structures have assigned a rich vocabulary to these structures based on their geometry; some are described in Table 2.5 and shown in Figure 2.18. The name assigned to a specific structure depends on its shape today, but in the context of geologic time, this shape may be only temporary.

Because salt both rises up into preexisting strata and rises during the time of deposition of overlying strata, geologists distinguish between two types of salt structure growth. If the salt rises after the overlying strata have already been deposited, then the rising salt will warp and eventually break through the overlying

FIGURE 2.17 A salt glacier that originates from a salt dome (western Iran).

TABLE 2.5	TERMINOLOGY OF SALT STRUCTURES
Detached bulb	Stems or walls connecting salt stocks to the source bed may pinch out, so that salt that was originally separated by the stem or wall becomes juxtaposed. Flow of the salt into bulbs results in folds that wrap around themselves and thus have circular profiles.
Detached canopy	Stems or walls connecting salt canopies to the source bed may pinch out, so that salt that was originally separated by the stem or wall becomes juxtaposed.
Salt anticline	An upward salt bulge relative to the source layer that is elongate in plan view. Strata overlying these structures conformably warp around the structure.
Salt canopy	In regions where the source bed was quite thick so that many diapirs form, salt walls or salt stocks may spread out and merge at a higher stratigraphic level. Salt canopies may flow dominantly in one direction in response to gravity, and if so are sometimes called *salt-tongue canopies.*
Salt diapir	A salt structure that pierces bedding in overlying strata.
Salt dome	An upward salt bulge relative to the source layer that is roughly symmetric in plan view. Strata overlying these structures conformably warp around the structure.
Salt glacier	Salt that flows out over the land when the salt diapir pierces ground surface (Figure 2.17).
Salt pillow	Same as *salt dome.*
Salt roller	An asymmetric salt dome that resembles an ocean wave.
Salt stock	Equant salt diapir in plan view. Mature salt diapirs are generally narrower at depth and broader at the top; the lower part is called the *stem* and the upper part is called the *bulb.*
Salt wall	Elongate salt diapir in plan view.

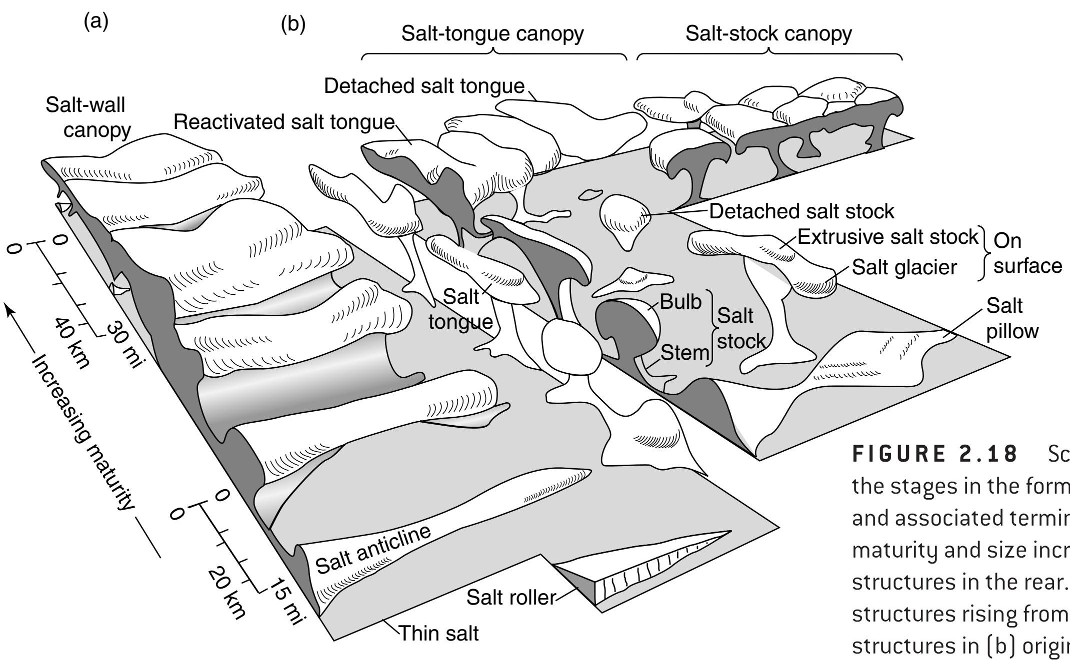


FIGURE 2.18 Schematic diagram showing the stages in the formation of salt structures and associated terminology. Structural maturity and size increase toward the structures in the rear. Sequence in (a) shows structures rising from line sources, whereas structures in (b) originate from point sources.

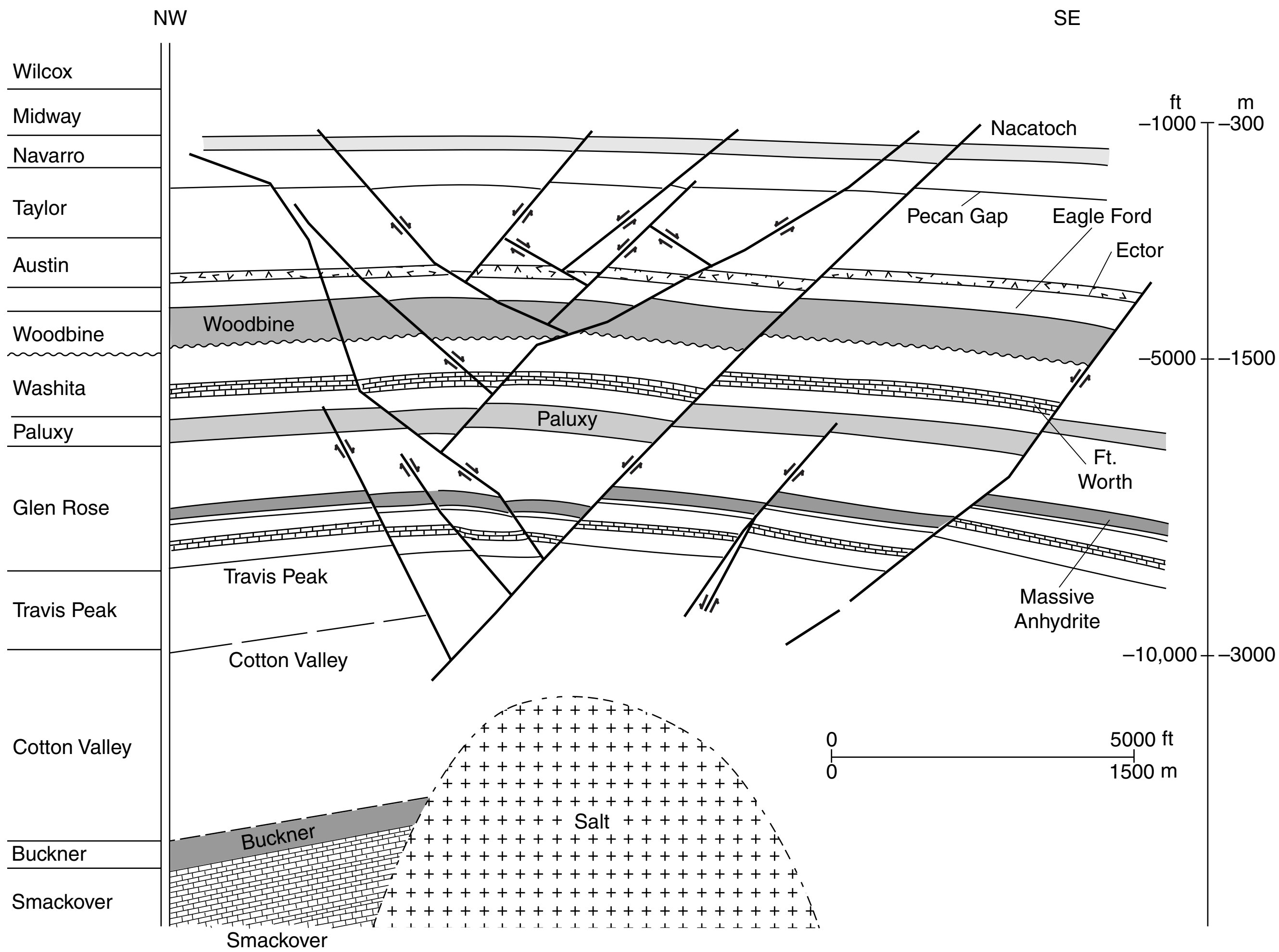


FIGURE 2.19 Cross section illustrating a normal fault array over the top of a salt dome in Texas.

strata. This process is called **upbuilding.** If, however, the rise of the salt relative to the source layer occurs coevally with further deposition, the distance between the source layer and the surface of the basin also increases. This process is called **downbuilding.** As salt moves, it deforms adjacent strata and creates complex folds and local faults. When salt diapirs approach the surface, the overlying strata are arched up and therefore are locally stretched, resulting in the development of normal faults in a complex array over the crest of the salt structure (Figure 2.19).

The structural geometry of passive-margin basins is complicated because sedimentation continues during salt movement. Sedimentary layers thicken and thin as a consequence of highs and lows in elevations caused by the salt, and the resulting differential compaction causes further salt movement. The sedimentation pattern may change in a locality when a salt structure drains out and flows into a structure at another locality. Thus, in regions where halokinesis occurs, it is common to find places where an arch evolves into a basin, or vice versa, a process we call **inversion.**

2.3.3 Gravity-Driven Faulting and Folding

The formation of salt structures is a dynamic process that is intimately linked to faulting in the overlying strata. Salt is so weak that it makes a good glide horizon on which detachment and movement of overlying strata occurs. In fact, on many passive margins, a thick package of sedimentary rocks tends to detach and slump seaward, gliding on a detachment fault in the layer of salt at its base. This movement resembles the slumping of sediment of a hillslope, though the scale of displacement it quite different. As slumping occurs, the landward portion of the basin is stretched and is therefore broken by a series of normal faults whose dip tends to decrease with depth. This change in dip with depth makes the faults concave up, which are called **listric faults.** As movement occurs on a listric fault, the strata above the fault arch into **rollover folds** (Figure 2.20). Many listric normal faults intersect the ground surface in southern Texas, because this region is part of the passive-margin basin along the Gulf

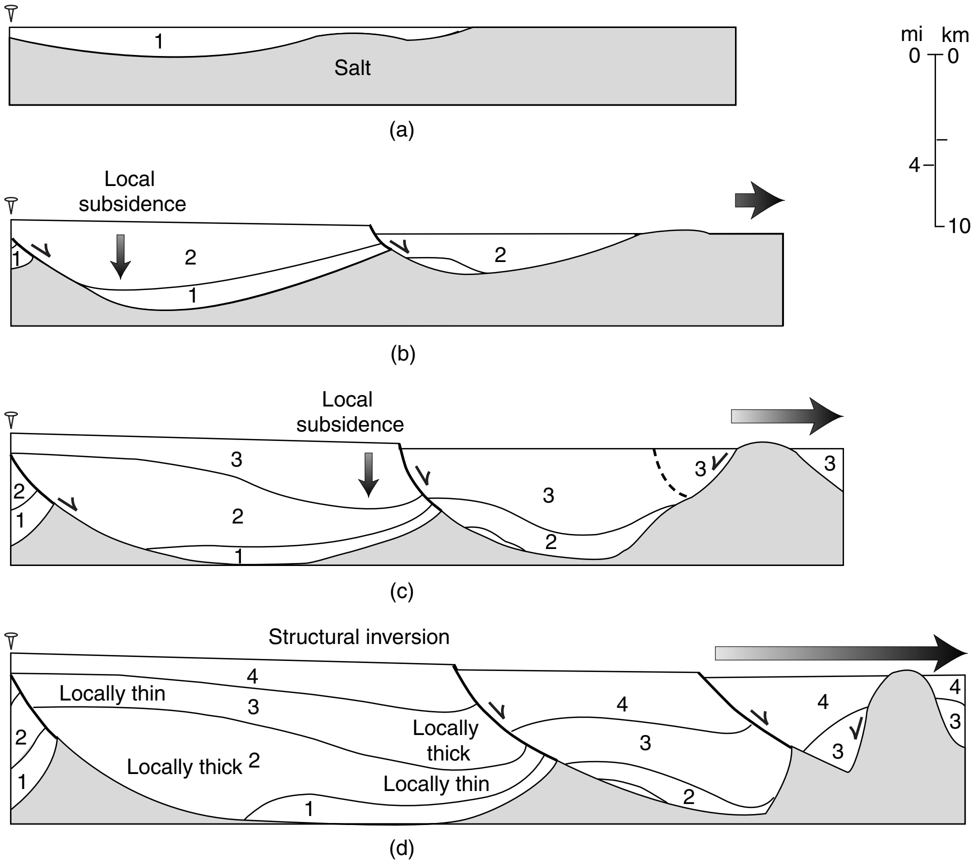


FIGURE 2.20 Cross sections showing "down-to-Gulf" type movement of a passive-margin salt wedge; the sections also show listric normal faults and salt rollers. The sequence (a) through (d) shows successive stages in the evolution of the margin, and accompanying extension.

Coast. Because the faults dip south and they transport rock toward the Gulf of Mexico, they are sometimes called **down-to-Gulf faults.** Slip on these faults thins the stratigraphic section above the salt layer and thus results in differential loading. As a result, the salt can rise beneath the fault, evolving from a salt dome to a salt roller, to a diapir that eventually cuts the overlying fault. At the toe of the passive margin wedge, a series of thrust faults develop to accommodate displacement of the seaward-moving section, just as thrusts develop at the toe of a hillside slump.

2.3.4 Practical Importance of Salt Structures

Why spend so much time dealing with salt structures and passive margins? Simply because these regions are of great economic and societal importance. Passive-margin basins are major oil reservoirs, and much of the oil in these reservoirs is trapped adjacent to salt bodies. Oil rises in the upturned layers along the margins of the salt body and is trapped against the margin of the impermeable salt. In recent years, salt bodies in passive-margin basins are being used as giant storage tanks for gas or oil, and are being considered as potential sites for the storage of nuclear waste. This is one of many examples where structural geology is central to reaching important societal decisions.

2.4 IGNEOUS STRUCTURES

You may recall from your introductory geology course that there are two principal classes of igneous rocks, and that these classes are distinguished from one another based on the environment in which the melt cools. **Extrusive rocks** are formed either from lava that flowed over the surface of the Earth and cooled under air or water, or from ash that exploded out of a volcanic vent. **Intrusive rocks** cooled beneath the surface of the Earth. During the process of intruding, flowing, settling and/or cooling, igneous rocks can develop primary structures. In this context, we use the term "primary structure" to refer to a fabric that is a consequence of igneous processes.

Where do **magmas,** the melt phase of igneous rocks, come from? If you have not had a course in igneous petrology, we'll quickly outline the nature of magmatic activity. Magma forms where conditions of heat and pressure cause existing rock (either in the crust or in the mantle) to melt. Commonly, only certain minerals within the solid rock melt (the ones that melt at a lower temperature), in which case we say that the rock has undergone **partial melting.** A magma formed by partial melting has a composition that differs from that of the rock from which it was extracted. For example, a 1–6% partial melt of ultramafic rock (peridotite) in the mantle yields a mafic magma which, when solidified, forms the gabbro and basalt that characterizes oceanic crust. Melting of an intermediate-composition crustal rock (diorite) yields a silicic magma which, when solidified, forms granite or rhyolite. Once formed, magma is less dense than the surrounding rock, and buoyancy forces cause it to rise. The density decrease is a consequence of the expansion that accompanies heating and melting, the formation of gas bubbles within the magma, and the difference in composition between magma and surrounding rock.

Magma moves by oozing up through a network of cracks and creeping along grain surfaces. The difference between the pressure within the magma and the pressure in the surrounding rock is so substantial that, as magma enters the brittle crust, it can force open new cracks. Magma continues to rise until it reaches a level of neutral buoyancy, defined as the depth where pressure in the magma equals lithostatic pressure in the surrounding rock, meaning that the buoyancy force is zero. At the level of neutral buoyancy, the magma may form a **sheet intrusion,** or may pool in a large **magma chamber** that solidifies into a bloblike intrusion called a **pluton.** We describe common types of igneous intrusions in Table 2.6. If the magma pressure is sufficiently high, the magma rises all the way to the surface of the Earth, like water in an artesian well, and is extruded at a volcano.

TABLE 2.6	TERMINOLOGY OF IGNEOUS INTRUSIONS
Batholith	A huge bloblike intrusion; usually a composite of many plutons.
Dike	A sheet intrusion that cross cuts stratification in a stratified sequence, or is roughly vertical in an unstratified sequence.
Hypabyssal	An intrusion formed in the upper few km of the Earth's crust; hypabyssal intrusions cool relatively quickly, and thus are generally fine grained.
Laccolith	A hypabyssal intrusion that is concordant with strata at its base, but bows up overlying strata into a dome or arch.
Pluton	A moderate-sized bloblike intrusion (several km in diameter). Sometimes the term is used in a general sense to refer to any intrusion, regardless of shape or size.
Sill	A sheet intrusion that parallels preexisting stratification in a stratified sequence, or is roughly subhorizontal in an unstratified sequence.
Stock	A small, bloblike intrusion (a few km in diameter).

2.4.1 Structures Associated with Sheet Intrusions

One important aspect of sheet intrusions that is of interest to structural geologists is their relationship to stress. In Chapter 3, we will introduce the concept of stress in detail, but for now, we point out that stress acting on a plane is defined as the force per unit area of the plane. Intuitively, therefore, you can picture that the Earth's crust is held together least tightly in the direction of the smallest stress, which we call the direction of least principal stress. Sheet intrusions, in general, form perpendicular to the direction of the least principal stress, assuming no preexisting planes of weakness (such as faults). For example, in regions where the greatest stress is caused by the weight of the overlying rocks, and is therefore vertical, the least principal stress is horizontal, so vertical dikes form. **Dike swarms,** which are arrays of subparallel dikes occurring over broad regions of the crust, probably represent intrusion at depth in association with horizontal extension, which causes the horizontal stress to be tensile.

Not all dikes occur in parallel arrays. In the immediate vicinity of volcanoes, the ballooning and/or collapse of a magma chamber locally modifies the stress field and causes a complex pattern of fractures. As a result, the pattern of dikes around a volcano (Figure 2.21) includes **ring dikes,** which have a circular trace in map view, and **radial dikes,** which run outward from the center of the volcano like spokes of a wheel. At a distance from the volcano, where the local effects of the volcano on the stress field are less, radial dikes may change trend to become perpendicular to the regional least principal stress.

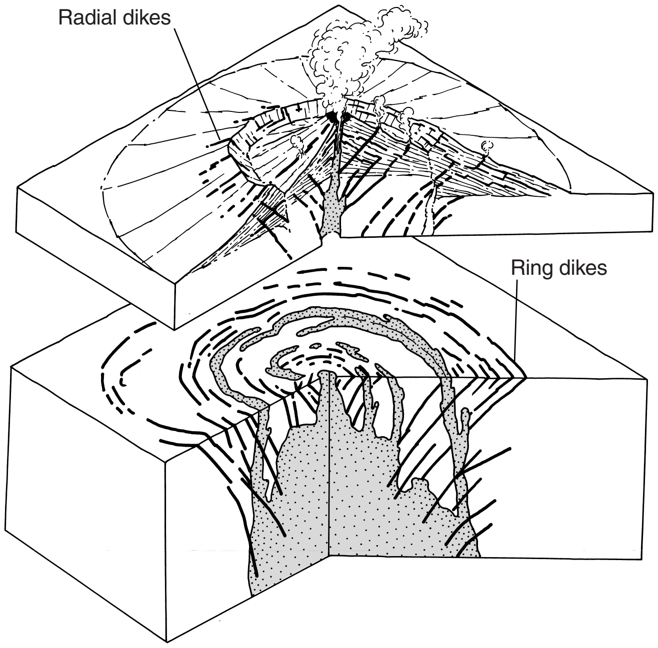


FIGURE 2.21 Types of sheet intrusions around a volcano.

Sills, or subhorizontal sheetlike intrusions, form where local stress conditions cause the least principal stress to be vertical, and/or where there are particularly weak horizontal partings in a stratified sequence. Sill intrusion can result in the development of faults. If the thickness of the intrusion changes along strike, there is differential movement of strata above the intrusion. **Laccoliths** resemble sills in that they are concordant with strata at their base, but unlike sills, they bow up the overlying strata to create a dome. For example, the laccoliths of the Henry Mountains in Utah are several kilometers in diameter.

Sheet intrusions occur in all sizes, from thin seams measured in centimeters, to the Great Dike of Zimbabwe, which is nearly 500 km long and several kilometers wide. Considering the dimensions of large intrusions (tens to hundreds of kilometers long), it is important to keep in mind that a large volume of magma can flow past a given point. The occurrence of flow may be recorded as primary igneous structures in the rock. For example, examination of dike-related structures may show the presence of drag folds, scour marks, imbricated phenocrysts, and flow foliation, particularly along the walls of an intrusion.

2.4.2 Structures Associated with Plutons

The nature of primary structures found in plutonic rocks depends on the depth in the Earth at which the intrusion solidified, because these structures reflect the temperature contrast between the intrusion and the country rock.[4] Remember that the Earth gets warmer with depth: at the surface, the average temperature is ~10°C, whereas at the center it may be as much as 4000°C. The change in temperature with depth is called the **geothermal gradient.** In the shallow crust, the geothermal gradient is in the range of 20°C/km to 40°C/km. At greater depths, however, the gradient must be less, because temperatures at the continental **Moho** (at about 40 km depth) are in the range of about 700°C (~15°C/km), and temperatures at the base of the lithosphere (at about 150 km) are in the range of about 1280°C (<10°C/km). The origin of the increased geothermal gradient near the surface is the concentration of radioactive elements in the minerals of more silicic rocks. Granitic magma begins to solidify at temperatures between 550°C to 800°C. Therefore, the temperature contrast between magma and country rock decreases with depth. In the case of shallow-level plutons, which intrude at depths of less than about 5 km, the contrast between magma and country rock is several hundred degrees. Contacts at the margins of such shallow intrusions are sharp, so that you can easily place your finger on the contact. Angular blocks of country rock float in the magma near the contact, and the country rock adjacent to the contact may be altered by fluids expelled by the magma or may be baked (a rock type called hornfels). At greater depth, the temperature contrast between magma and country rock decreases, and contacts are more gradational, until the country rock itself is likely to be undergoing partial melting. Minerals that melt at lower temperatures (like quartz and feldspar) turn to liquid, while refractory minerals (that is, minerals that melt at higher temperatures, such as amphibole and pyroxene) are still solid, though quite soft. Movement of the melt causes the soft solid layers to be contorted into irregular folds. When this mass eventually cools, the resulting rock, which is composed of a marble-cake-like mixture of light and dark contorted bands, is called a **migmatite** (Figure 2.22).

Many plutons exhibit an **intrusion foliation** that is particularly well developed near the margin of the pluton and is subparallel to pluton–host rock contact. This foliation is defined by alignment of **inequant** crystals

[4]To a geologist, "country rock" is not only a type of music, but also a casual term for the rock that was in a locality before the intrusion.

FIGURE 2.22 A migmatite from the North Cascades (Washington State, USA) showing complex folding and disruption.

and by elongation of chunks of **country rock** or early phases of pluton that were incorporated in the magma (called **xenoliths,** which means "foreign rocks"). Such fabric is a consequence of shear of the magma against the walls of the magma chamber and of the flattening of partially solidified magma along the chamber walls in response to pressure exerted as new magma pushes into the interior of the chamber. Similarly, intrusion foliation can be developed along the margins of dikes.

Because intrusion foliation forms during the formation of the rock and is not a consequence of tectonic movements, it is a nontectonic structure. However, it may be difficult to distinguish from schistosity resulting from tectonic forces. Plutons tend to act as mechanically strong blocks, so that regional deformation is deflected and concentrated along the margins of the pluton. Interpretation of a particular foliation therefore depends on regional analysis and the study of deformation microstructures (Chapter 9). For example, if the fabric remains parallel to the boundary of the intrusion, even when the boundary changes and individual grains show no evidence for solid-state deformation, the foliation is likely a primary igneous structure. The distinction between tectonic and primary structures in plutons has proven to be quite difficult and more often than not is ambiguous.

2.4.3 Structures Associated with Extrusion

As basaltic lava flows along the surface of the Earth, the surface of the flow may wrinkle into primary folds that resemble coils of rope, or may break into a jumble of jagged blocks that resembles a breccia. Lava flows with the rope-coil surface are called **Pahoehoe flows,** and lavas with the broken-block surface are called **Aa flows.** The wrinkles in a Pahoehoe lava should not be mistaken for tectonic folds, and the jumble of blocks caused by **autobrecciation** (that is, breaking up during flow) of Aa lavas should not be mistaken for a tectonic breccia related to faulting.

If basaltic lava is extruded beneath seawater, the surface of the flow cools quickly, and a glassy skin coats the surface of the flow. Eventually, the pressure in the glass-encased flow becomes so great that the skin punctures, and a squirt of lava pushes through the hole and then quickly freezes. The process repeats frequently, resulting in a flow composed of blobs (centimeters to meters in diameter) of lava. Each blob, which is called a pillow, is coated by a rind of fine-grained to glassy material. As the pillows build out into a large pile, creating a **pillow basalt,** successive pillows flow over earlier pillows and, while still soft, conform to the shape of the earlier flow surface (Figure 2.23). As a result, pillows commonly have a rounded top and a pointed bottom (the "apex") in cross section, and this shape can be used as a stratigraphic facing criterion.

In 1902, Mt. Pelee on the Caribbean island of Martinique erupted. It was a special kind of eruption, for instead of lava flows, a spine of rhyolite rose day by day from the peak of the volcano. This spine, as it turned out, was like the cork of a champagne bottle slowly being worked out. When the cork finally pops out of the champagne bottle, a froth of gas and liquid flows down the side of the bottle. Likewise, when, on the morning of May 2, the plug exploded off the top of Mt. Pelee, a froth of hot (>800°C) volcanic gas and ash floated on a cushion of air and rushed down the side of the mountain at speeds of up to 100 km/h. This ash flow engulfed the town of St. Pierre, and in an instant, almost 30,000 people were dead. When the ash stopped moving, it settled into a hot layer that welded together. Such a layer of welded tuff is called an **ignimbrite,** often displaying a foliation. Volcanic ash is composed of tiny glass shards with jagged spinelike forms that are a consequence of very rapid cooling. When the ash settles, the glass shards are still hot and soft, so the compaction pressure exerted by the weight of overlying ash causes the shards to flatten, thereby creating a primary foliation in ignimbrite that is comparable to bedding.

Rhyolitic lavas commonly display subtle color banding, called **flow foliation,** that has been attributed to flow of the lava before complete solidification. The banding forms because lavas are not perfectly homogeneous materials. Since the temperature is not perfectly uniform, there may be zones in which crystals have formed, while adjacent regions are still molten. Shear resulting from movement of the lava smears out these initial inhomogeneities into subparallel bands. To visualize this, think of a bowl of pancake batter into which you have dripped spoonfuls of chocolate batter. If you slowly stir the mixture, the blobs of chocolate smear out into sheets. Chocolate blobs that were initially nearby would smear into parallel sheets with an intervening band of pancake batter. In the flow, movement of the lava smears out blobs of contrasting texture into layers, which, when the rock finally freezes, have a slightly different texture than adjacent bands and thus are visible markers in outcrop. Commonly, continued movement causes previously formed layers to fold, so flow-banded outcrops typically display complex primary folds.

FIGURE 2.23 Pillow basalt from the Point Sal Ophiolite, California (USA). The asymmetric shape of the pillows and location of the "points" (apex) indicate that the stratigraphic top of the flow is up.

2.4.4 Cooling Fractures

As shallow intrusions and extrusive flows cool, they contract. Because of their fine grain size, these bodies are susceptible to forming natural cracks, or **joints,** in response to the thermal stress associated with cooling. When such joints are typically arranged in roughly hexagonal arrays that isolate columns of rock, the pattern is called **columnar jointing** (Figure 2.24). Popular tourist and movie director destinations like Devil's Tower in Wyoming, Giant's Causeway in Ireland, or the Massif Central in southern France offer spectacular examples. The long axes of columns are perpendicular to isotherms (surfaces of constant temperature) and thus they are typically perpendicular to the boundaries of the shallow intrusion or flow. If you look closely at unweathered columnar joints, the surfaces of individual joints are ribbed. We will learn later that this feature is a consequence of the way in which fractures propagate through rock (Chapter 6).

2.5 IMPACT STRUCTURES

Glancing at the Moon through a telescope, the most obvious landforms that you see are craters. Like early Earth, the Moon has been struck countless times by meteors, and each impact has left a scar which, because the Moon is tectonically inactive and has no atmosphere or water, has remained largely unchanged through succeeding eons. The Earth has been pummeled at least as frequently as the Moon, but many objects disintegrate and burn in the atmosphere before reaching the surface, and the scars of many that did strike the surface have been erased by erosion and particularly by tectonics. The vast majority of impacts on the Earth-Moon system occurred prior to about 3.9 Ga, when the solar system contained a multitude of fragments that were not yet incorporated into planets. Considering that 70% of today's Earth's surface is underlain by oceanic lithosphere, most of which is less than 200 million years old, and that all but a relatively small portion of the continental crust has either been covered by younger strata or has been involved in plate tectonics, it is not surprising that impact structures are so rare on Earth.

But even though **impact structures** are rare, and in some cases difficult to recognize, they do exist. For our discussion we distinguish three categories, based on the most obvious characteristic of the impact: (1) relatively recent surficial impacts that are defined by a visible crater, (2) impacts whose record at the present Earth surface is the disruption of sedimentary strata, and (3) impacts whose record is a distinctive map-view circular structure in basement.

(a) (b)

FIGURE 2.24 Columnar jointing at Devil's Postpile, California (USA); (a) side view, (b) top view.

Today there are about 150 impacts recognized on our planet but only a few dozen obvious impact craters can be seen. One of the largest and perhaps most famous is the approximately 50,000-year-old Barringer Meteor Crater in Arizona, which is 1.2 km in diameter and 180 m deep, and is surrounded by a raised rim that is about 50 m high (Figure 2.25a). The size of the impacting object is estimated only to be 30–50 m! As shown in the cross section (Figure 2.25b), the impact created a breccia that is about 200 m thick beneath the floor of the crater. The raised rims of the crater are not only composed of shattered rock ejected from the crater, but are also sites where bedrock has been upturned. Ancient impact sites that are no longer associated with a surficial crater dot the Midcontinent region of the United States. These sites are defined by relatively small (less than a few kilometers across) semicircular disruption zones, in which the generally flat-lying Paleozoic strata of the region are fractured, faulted, and tilted. They were originally called **crypto-volcanic structures** (from the Greek *crypto,* meaning "hidden"), because it was assumed that they were the result of underlying explosive volcanism. Typically, steeply dipping normal faults, whose map traces are roughly circular, define the outer limit of these structures. These faults are cross cut by other steep faults

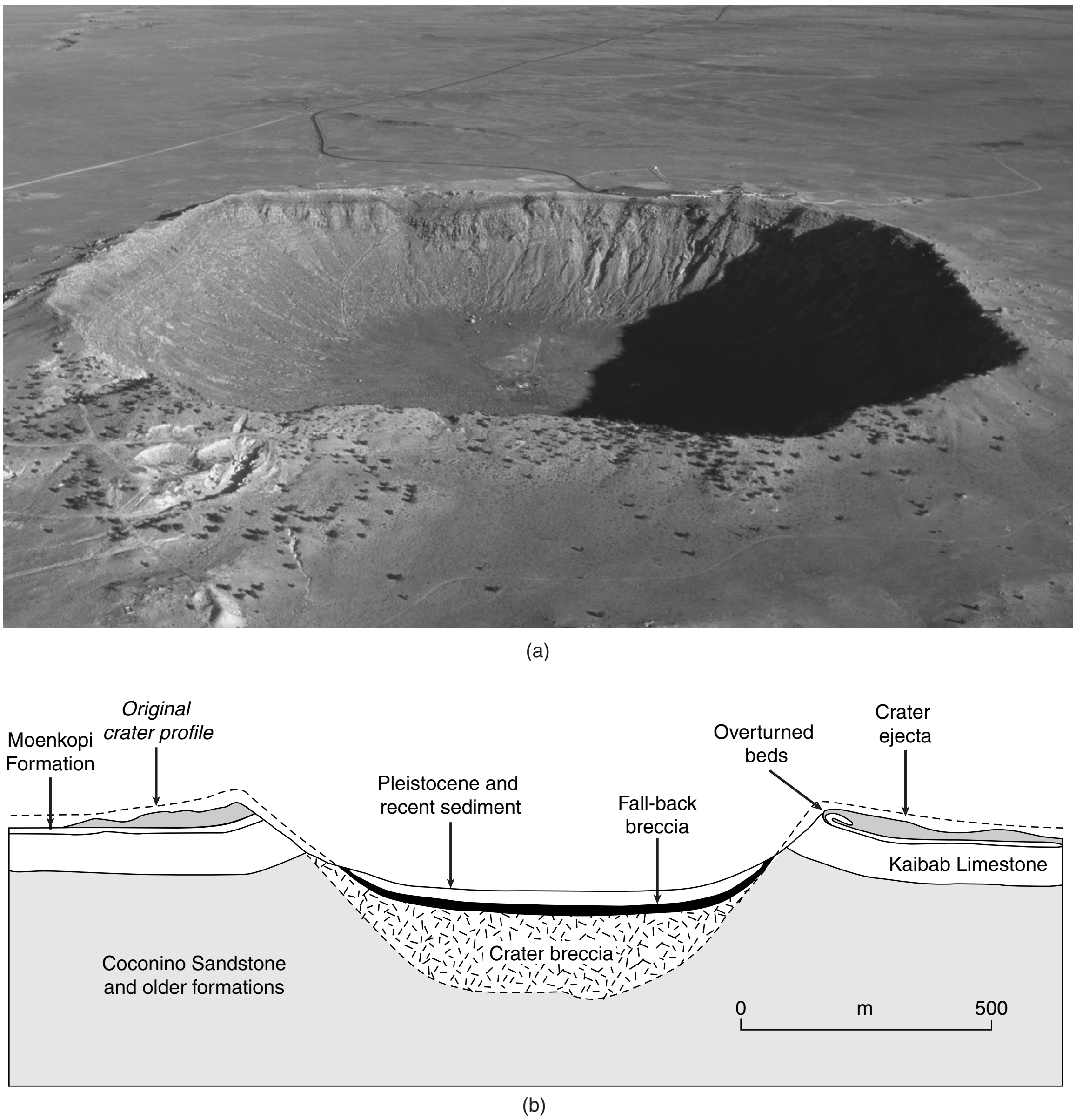


FIGURE 2.25 (a) Barringer Meteor Crater of Arizona (USA) and (b) geologic cross section showing distribution of impact structures.

that radiate from the center of the structure like spokes of a wagon wheel. This fracture geometry is similar to that around volcanoes (Figure 2.21), which appeared to support the volcanic interpretation. Near the center of the structure, bedding is steeply dipping, and faulting juxtaposes units of many different ages. Locally, the strata are broken into huge blocks that jumbled together to create an **impact breccia.** Throughout this region, rocks are broken into distinctive **shatter cones,** which are conelike arrays of fractures similar to those found next to a blast hole in rock (Figure 2.26). The apex of the cone points in the direction from which the impacting object came. In impact structures, shatter cones point up, confirming that they were caused by impact from above, as would be the case if the structure was due to an incoming meteor.

Why do impact structures have the geometry that they do? To see why, think of what happens when you drop a pebble into water. Initially, the pebble pushes down the surface of the water and creates a depression, but an instant later, the water rushes in to fill the depression, and the place that had been the center of the depression rises into a dome. In the case of meteor impact against rock, the same process takes place. The initial impact gouges out a huge crater and elastically compresses the rock around the crater. But an instant later, the rock rebounds. At the margins of the affected zone it pulls away from the walls, creating normal faults, and in the center of the zone, it flows upward, creating the steeply tilted beds. Because the rock in the near surface behaves in a brittle manner when this occurs, this movement is accompanied by faulting and brecciation.

The incredibly high pressures that develop during an impact create distinctive changes in the rocks of the impact site. The shock wave that passes through the rock momentarily subjects rocks to very high pressures, a condition that causes **shock metamorphism.** Shock metamorphism of quartz yields unusual high-density polymorphs like stishovite, and characteristic deformation microstructures. In addition, the kinetic energy of impact is suddenly transformed into heat, with the result that rocks of the impact site are momentarily heated to temperatures as high as 1700°C. At such temperatures, the whole rock melts, only to freeze quickly into glass of the same composition as the original rock. In some cases, melt mixes with impact breccia, and injects into cracks between larger breccia fragments, forming a glasslike rock called **pseudotachylyte.**

Impact structures affecting now exposed basement crystalline rocks characteristically cause distinctive circular patterns of erosion in the basement that stand out in satellite imagery or through geophysical methods. One of the best known basement impact structures is the Sudbury complex in southern Ontario, Canada. Not only are the characteristic features of impact, like shatter cones and pseudotachylyte, readily visible in the field, but the Sudbury impact, occurring about 1.85 Ga, was large enough to affect the whole crust and cause an impact melt that produced valuable economic deposits that have been mined for many years.

FIGURE 2.26 Shatter cones that were formed by the Sudbury impact that occurred around 1.85 Ga (Ontario, Canada). The apex of a shatter cone points in the direction from which the impacting object came. Pocket knife for scale.

2.6 CLOSING REMARKS

In this chapter we explored the description of various types of sedimentary, igneous, and impact structures whose formation is not an immediate consequence of plate tectonic forces, of isostatic consequences, or of thickening or thinning of the crust. A discussion of nontectonic structures is a good way to start a structural geology course, because it gets you thinking about geometries and shapes. However, these structures are the topic of entire classes in sedimentology and igneous petrology, so you will realize that we cannot do justice to the richness of these topics. Before delving into tectonic structures, we will first introduce the fundamentals of stress and deformation. Although the description of tectonic structures does not require an understanding of these concepts, they are needed when examining how and why tectonic structures form. So, stress, deformation, and rheology are the topics of the next few chapters.

ADDITIONAL READING

Collinson, J. D., and Thompson, D. B., 1989. *Sedimentary structures* (2nd edition). Unwin Hyman: London.

Jackson, M. P. A., and Talbot, C. J., 1994. Advances in salt tectonics. In Hancock, P. L., ed., *Continental deformation.* Pergamon Press, pp. 159–179.

Melosh, H. J., 1996. *Impact cratering: A geologic process.* Oxford University Press: New York.

Paterson, S. R., Vernon, R. H., and Tobisch, O. T., 1989. A review of criteria for the identification of magmatic and tectonic foliations in granitoids. *Journal of Structural Geology,* 11, 349–363.

Selley, R. C., 1988. *Applied sedimentology.* Academic Press: New York.

Shrock, R. R., 1948. *Sequence in layered rocks.* McGraw-Hill: New York.

Worrall, D. M., and Snelson, S., 1989. Evolution of the northern Gulf of Mexico, with emphasis on Cenozoic growth faulting and the role of salt. In *The geology of North America—an overview,* v. A, pp. 97–138. Geological Society of America.