CHAPTER EIGHT

Faults and Faulting

8.1 INTRODUCTION

Imagine a miner in a cramped tunnel crunching forward through a thick seam of coal. Suddenly, his pick hits hard rock. The miner chips away a bit more only to find that the seam that he's been following for the past three weeks abruptly terminates against a wall of sandstone. He says, "@%$&, my seam's cut off—there's a fault with it!" From previous experience, the miner knows that because of a fault, he will have to spend precious time digging a shaft up or down to intersect the coal seam again.

Geologists adopted the term fault, but they use the term in different ways in different contexts. In a general sense, a **fault** is any surface or zone in the Earth across which measurable **slip** (shear displacement) develops. In a more restricted sense, faults are fractures on which slip develops primarily by brittle deformation processes (Figure 8.1a). This second definition serves to distinguish a "fault" (*sensu stricto*) from a fault zone and shear zone. We use the term **fault zone** for brittle structures in which loss of cohesion and slip occurs on several faults within a band of definable width (Figure 8.1b). Displacement in fault zones can

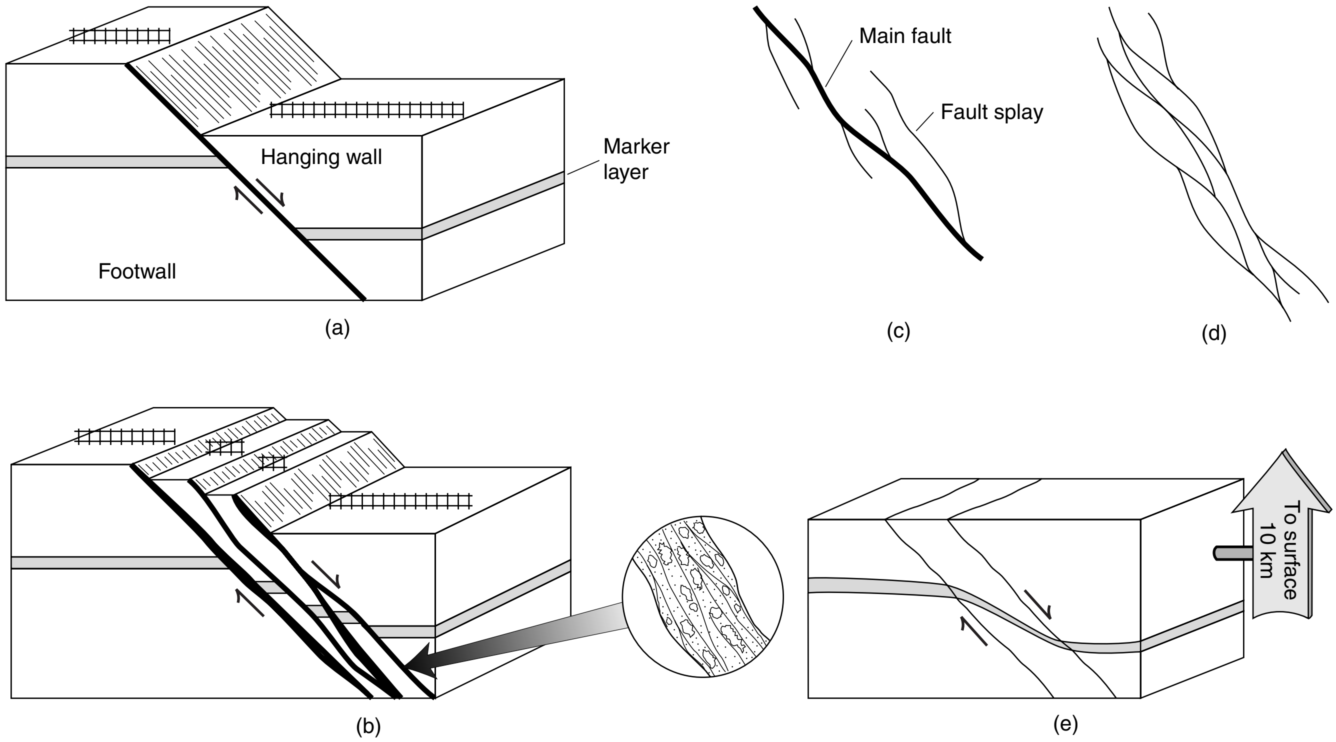


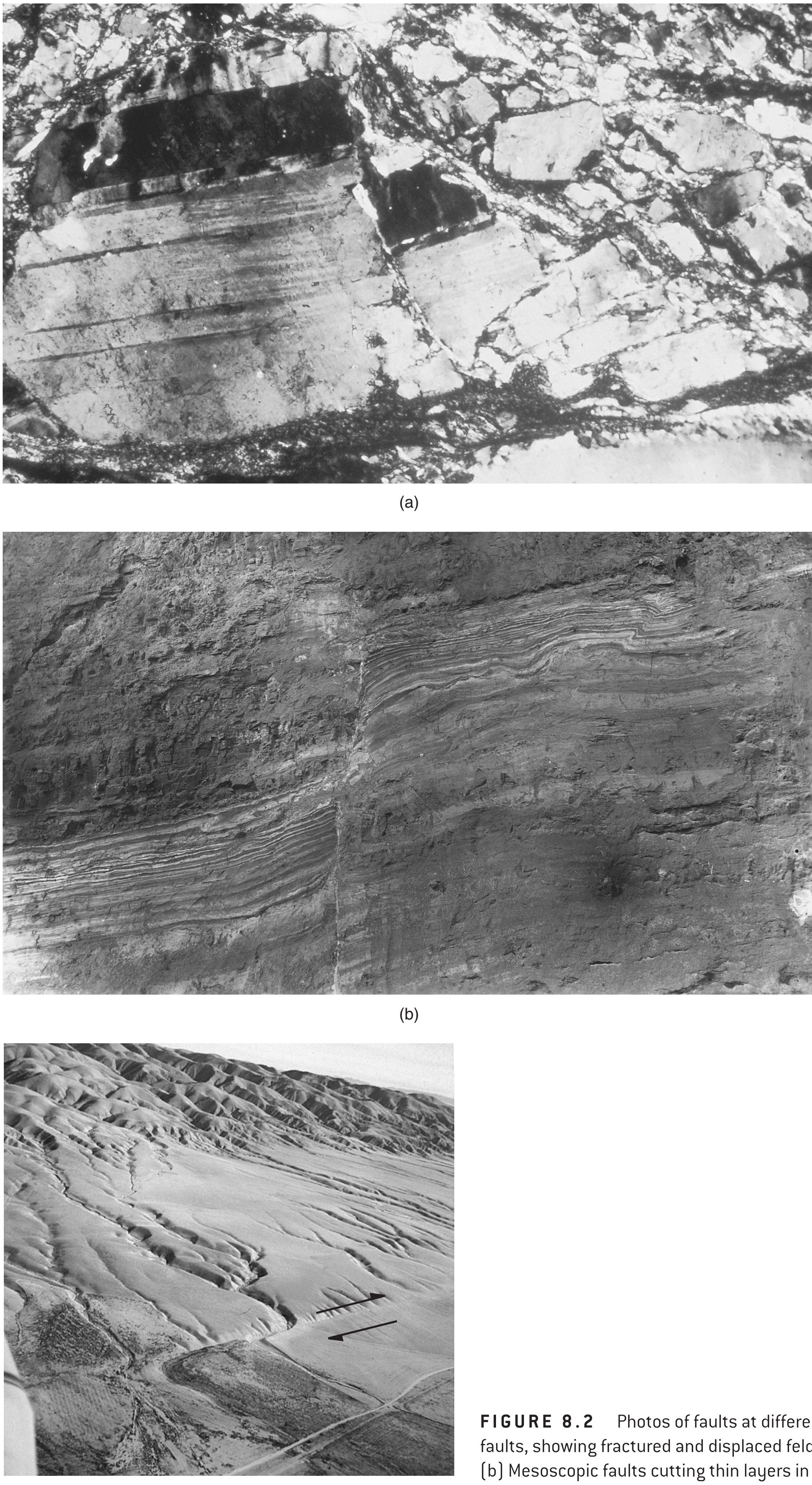
FIGURE 8.1 Sketches illustrating differences between faults, fault zones, and shear zones. (a) Fault. (b) Fault zone, with inset showing cataclastic deformation adjacent to the fault surface. (c) Sketch illustrating the relation between a principal fault and fault splays. (d) Anastomosing faults in a fault zone. (e) A shear zone, showing rock continuity across the zone. The displacements are shown to intersect the ground surface, whereas the shear zone occurs at depth in the crust.

involve formation and slip on many small, subparallel brittle faults, or slip on a principal fault from which many smaller faults diverge **(fault splays),** or slip on an anastomosing[1] array of faults (Figure 8.1c and d). **Shear zones** are ductile structures, across which a rock body does not lose mesoscopic cohesion, so that strain is distributed across a band of definable width. In ductile shear zones, rocks deform by **cataclasis,** a process involving fracturing, crushing, and frictional sliding of grains or rock fragments, or, more commonly, by **crystal plastic deformation** mechanisms (Figure 8.1e). We describe cataclastic shear zones in this chapter, but delay discussion of processes in ductile shear zones until Chapters 9 and 12.

Faults occur on all scales in the lithosphere (Figure 8.2), and geologists study them for several reasons. They control the spatial arrangement of rock units, so their presence creates puzzles that challenge even the most experienced geologic mappers. Faults affect topography and modify the landscape. Faults affect the distribution of economic resources (e.g., oil fields and ore bodies). They control the permeability of rocks and sediments, properties which, in turn, control fluid migration. Faulting creates deformation (strain ± rotation ± translation) in the lithosphere during plate interactions and intraplate movements. And, faulting may cause devastating earthquakes.

Fault analysis, therefore, plays a major role in diverse aspects of both academic and applied geology. In order to provide a basis for working with faults, this chapter introduces the terminology that is used to describe fault geometry and displacement, it discusses how to represent faults on maps and cross sections, and shows you how to recognize and interpret faults at the surface and in the subsurface. We conclude by introducing fault-system tectonics, the relation between faulting and resources, and the relation between faulting and earthquakes. Much of our discussion in this chapter focuses on the properties of mesoscopic faults; we treat the large-scale properties

[1]Anastomosing refers to a geometry of a group of wavy, subparallel surfaces that merge and diverge, resembling a braid of hair.



(a)

(b)

(c)

FIGURE 8.2 Photos of faults at different scales. (a) Microscopic faults, showing fractured and displaced feldspar grains. (b) Mesoscopic faults cutting thin layers in an outcrop. (c) The trace of the San Andreas (strike-slip) Fault across the countryside.

of fault systems more fully in Part D, where the basic concepts of plate tectonics are discussed.

8.2 FAULT GEOMETRY AND DISPLACEMENT

8.2.1 Basic Vocabulary

In order to discuss faults, we first need to introduce some fault vocabulary. To simplify our discussion, we treat a fault as a geometric surface in a body of rock. Rock adjacent to a fault surface is the **wall** of the fault, and the body of rock that moved as a consequence of slip on the fault is a **fault block.** If the fault is not vertical, you can distinguish between the **hanging-wall block,** which is the rock body above the fault plane, and the **footwall block,** which is the rock body below the fault plane (Figure 8.1a). This terminology is adopted from mining geology. Note that you cannot distinguish between a hanging wall and a footwall for a vertical fault.

To describe the attitude of a fault precisely, we measure the strike and dip (or dip and dip direction) of the fault (see Chapter 1). Commonly, geologists use adjectives such as steep, shallow, vertical, and so on, to convey an approximate image of fault dip (Table 8.1). Keep in mind that a fault is not necessarily a perfectly planar surface; it may curve and change attitude along strike and/or up and down dip. Where such changes occur, a single strike and dip is not sufficient to describe the attitude of the whole fault, and you should provide separate measurements for distinct segments of the fault. Faults whose dip decreases progressively with depth have been given the special name **listric faults.**

When fault movement occurs, one fault block slides relative to the other, which is described by the **net slip.** You can completely describe displacement by specifying the **net-slip vector,** which connects two formerly adjacent points that are now on opposite walls of the fault (Figure 8.3). To describe a net-slip vector, you must specify its magnitude and orientation (plunge and bearing, or rake on a plane), and the **sense of slip** (or shear sense). **Shear sense** defines the relative displacement of one wall of the fault with respect to the other wall; that is, whether one wall went up or down, and/or to the left or right of the other wall.

Like any vector, the net-slip vector can be divided into components. Generally we use the strike and dip of the fault as a reference frame for defining these components. Specifically, you measure the **dip-slip component** of net slip in the direction parallel to the dip direction, and the **strike-slip component** of net slip in the direction parallel to the strike. If the net-slip vector parallels the dip direction of the fault (within ~10°), the fault is called a **dip-slip fault;** if the vector roughly parallels the strike of the fault, the fault is called a **strike-slip fault.** If the vector is not parallel to either dip direction of the strike, we call the fault an

TABLE 8.1	DESCRIPTION OF FAULT DIP
Horizontal faults	Faults with a dip of about 0°; if the fault dip is between about 10° and 0°, it is called **subhorizontal.**
Listric faults	Faults that have a steep dip close to the Earth's surface and have a shallow dip at depth. Because of the progressive decrease in dip with depth, listric faults have a curved profile that is concave up.
Moderately dipping faults	Faults with dips between about 30° and 60°.
Shallowly dipping faults	Faults with dips between about 10° and 30°; these faults are also called *low-angle faults.*
Steeply dipping faults	Faults with dips between about 60° and 80°; these faults are also called *high-angle faults.*
Vertical faults	Faults that have a dip of about 90°; if the fault dip is close to 90° (e.g., is between about 80° and 90°), the fault can be called **subvertical.**

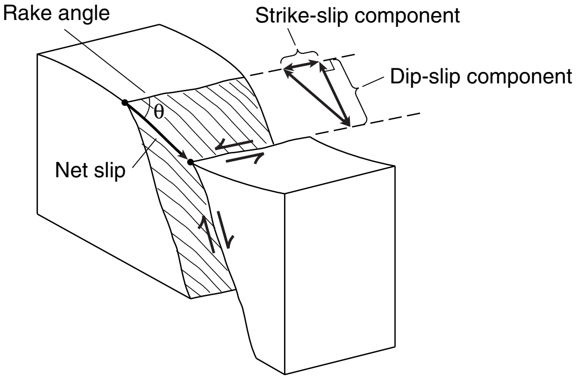


FIGURE 8.3 Block diagram sketch showing the net-slip vector with its strike-slip and dip-slip components, as well as the rake and rake angle.

oblique-slip fault. As you can see in Figure 8.3, oblique-slip faults have both a strike-slip and a dip-slip component of movement.

We describe the shear sense on a dip-slip fault with reference to a horizontal line on the fault, by saying that the movement is **hanging-wall up** or **hanging-wall down** relative to the footwall. Hanging-wall down faults are called **normal faults,** and hanging-wall up faults are called **reverse faults** (Figure 8.4a and b). To define sense of slip on a strike-slip fault, imagine that you are standing on one side of the fault and are looking across the fault to the other side. If the opposite wall of the fault moves to your right, the fault is **right-lateral** (or **dextral**), and if the opposite wall of the fault moves to your left, the fault is **left-lateral** (or **sinistral;** Figure 8.4c and d). Note that this displacement does not depend on which side of the fault you are standing on. Finally, we define shear sense on an oblique-slip fault by specifying whether the dip-slip component of movement is hanging-wall up or down, and whether the strike-slip component is right-lateral or left-lateral (Figure 8.4e–h). Commonly, an additional distinction among fault types is made by adding reference to the dip angle of the fault surface; we recognize high-angle (>60° dip), intermediate-angle (30° to 60° dip), and low-angle faults (<30° dip). We provide descriptions of the basic fault types in Table 8.2, along with descriptions of other commonly used names (such as thrust and detachment).

You may be wondering where the terms "normal" and "reverse" come from. Perhaps normal faults were thought to be "normal" because the hanging-wall block appeared to have slipped down the fault plane, just like a person slips down a slide. It is a safe guess that geologists came up with the name "reverse fault" to describe faults that are the opposite of normal. Now you know!

We also distinguish among faults on the basis of whether they cause shortening or lengthening of the layers that are cut. Imagine that a fault cuts and displaces a horizontal bed marked with points X and Y (Figure 8.5a). Before movement, X and Y project to points A and B on an imaginary plane above the bedding plane. If the hanging wall moves down, then points X and Y project to A and B'. The length AB' is greater than the length AB (Figure 8.5b). In other words, movement on this fault effectively lengthens the layer. We call a fault which results in lengthening of a layer an **extensional fault.** By contrast, the faulting shown in Figure 8.5c resulted in a decrease in the distance between points X and Y ($AB > AB''$). We call a fault which results in shortening of a body of rock a **contractional fault.** Contractional faults result in

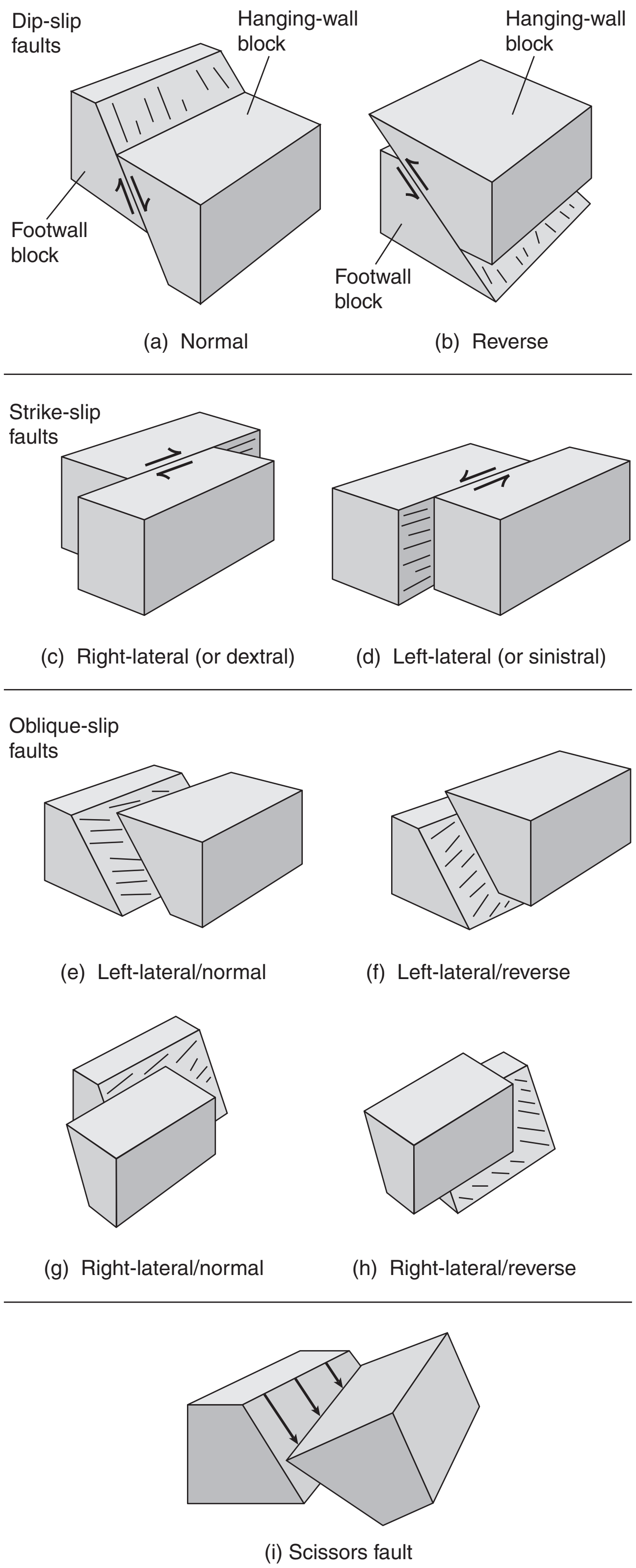


FIGURE 8.4 Block diagram sketches showing the different types of faults.

TABLE 8.2	TYPES OF FAULTS
Allochthon	The thrust sheet above a detachment is the allochthon (meaning that it is composed of *allochthonous* rock; i.e., rock that has moved substantially from its place of origin).
Autochthon	The footwall below a detachment is the autochthon; it is composed of *autochthonous* rock, or rock that is still in its place of origin.
Contractional fault	A contractional fault is one whose displacement results in shortening of the layers that the fault cuts, regardless of the orientation of the fault with respect to horizontal.
Décollement	The French word for detachment.
Detachment fault	This term is used for faults that initiate as a horizontal or subhorizontal surface along which the hanging-wall sheet of rock moved relative to the footwall. An older term “overthrust” is a regional detachment fault on which there has been a thrust sense of movement. Some detachments are listric, and on some detachments, regional normal-sense displacement occurs.
Dip-slip fault	The slip direction on a dip-slip fault is approximately parallel to the dip of the fault (i.e., has a rake between ~80° and 90°).
Extensional fault	An extensional fault is one whose displacement results in extension of the layers that the fault cuts, regardless of the orientation of the fault with respect to horizontal.
Normal fault	A normal fault is a dip-slip fault on which the hanging wall has slipped down relative to the footwall.
Oblique-slip fault	The slip direction on an oblique-slip fault has a rake that is not parallel to the strike or dip of the fault. In the field, faults with a slip direction between ~10° and ~80° are generally called oblique-slip.
Overthrust fault	This is an older term that you may find in older papers on faults, but is no longer used much today. The term is used for thrust faults of regional extent. In this context, “regional extent” means that the thrust sheet has an area measured in tens to hundreds of square km, and the amount of slip on the fault is measured in km or tens of km. Today, such faults are generally called regional detachments.
Par-autochthonous	If a fault block has only moved a small distance from its original position, the sheet is par-autochthonous (literally, relatively in place).
Reverse fault	A reverse fault is a dip-slip fault on which the hanging wall has slipped up relative to the footwall.
Scissors fault	On a scissors fault, the amount of slip changes along strike so that the hanging-wall block rotates around an axis that is perpendicular to the fault surface (Figure 8.4i).
Strike-slip fault	The slip direction on a strike-slip fault is approximately parallel to the fault strike (i.e., the line representing slip direction has a rake [pitch] in the fault plane of less than ~10°). Strike-slip faults are generally steeply dipping to vertical.
Transfer fault	A transfer fault accommodates the relative motion between blocks of rock that move because of the displacement on other faults.
Transform fault	In the preferred sense, transform faults are plate boundaries at which lithosphere is neither created nor destroyed. In a general sense, a transform fault links two other faults and accommodates the relative motion between the blocks of rock that move because of the displacement on the other two faults. However, we reserve the term *transfer fault* for this general type of displacement, independent of scale.

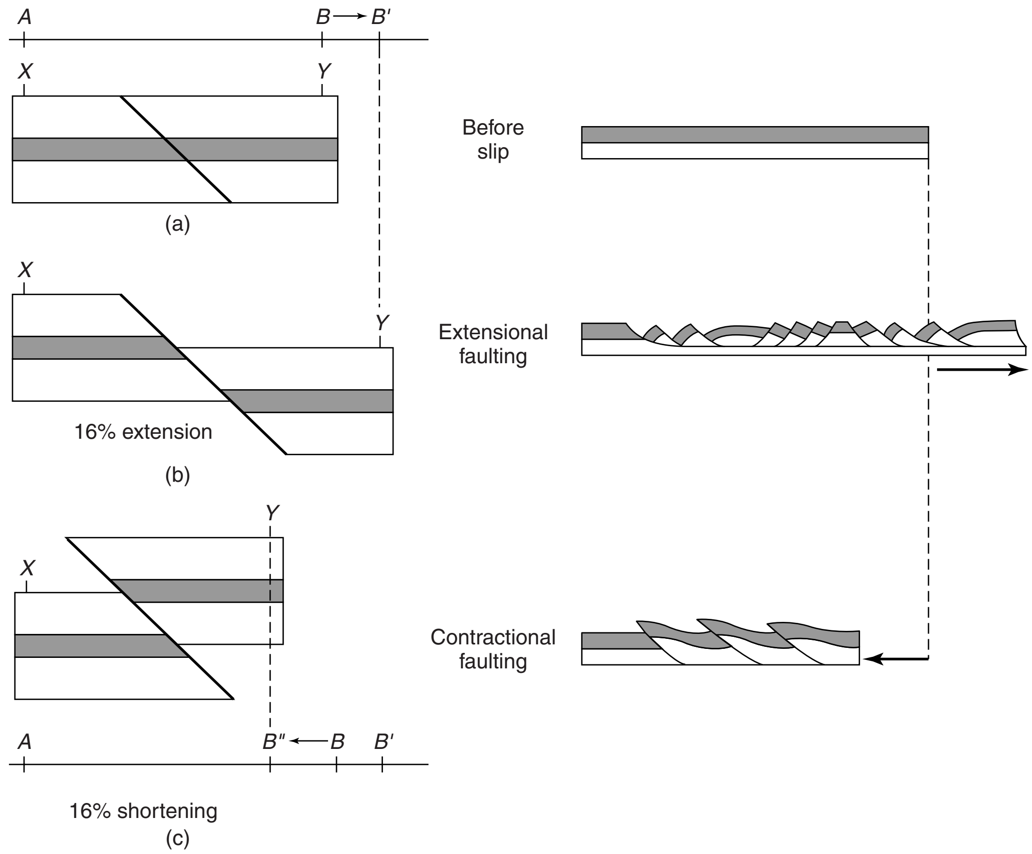


FIGURE 8.5 Extensional and contractional faulting. (a) Starting condition, (b) extension, and (c) contraction. Note the respective horizontal length changes.

duplication of section, as measured along a line that crosses the fault and is perpendicular to stratigraphic boundaries, whereas extensional faults result in loss of section. Generally, one can use the term "normal fault" as a synonym for an extensional fault, and the term "reverse fault" as synonym for contractional faults. But such usage is not always correct. Consider a normal fault that rotates during later deformation. In outcrop, this fault may have the orientation and sense of slip you would expect on a reverse fault, but, in fact, its displacement produced extensional strain parallel to layering.

8.2.2 Representation of Faults on Maps and Cross Sections

Because a fault is a type of geologic contact, meaning that it forms the boundary between two bodies of rock, faults are portrayed as a (heavy) line on geologic maps, like other contacts. We distinguish among different types of faults on maps through the use of symbols (Figure 8.6). For example, thrust faults are decorated with triangular teeth placed on the hanging-wall side of the trace. (Note that the teeth do not indicate the direction of movement!) Normal faults, regardless of dip, are commonly portrayed by placing barbs on the hanging-wall block. We represent strike-slip faults on a map by placing arrows that indicate the sense of slip on either side of the fault (Figure 8.6c).

In cross sections, faults are also represented by a thick line (Figure 8.7). If the slip direction on the fault roughly lies in the plane of the cross section, then you indicate the sense of slip on the fault by oppositely facing half-arrows drawn on either side of the fault. If the movement on the fault is into or out of the plane of the section for a strike-slip fault, you indicate the sense of slip by drawing the head of an arrow (a circle with a dot in it) on the block moving toward you, and tail of an arrow (a circle with an X in it) on the block moving into the plane. If the movement is into or out of the page for a dip-slip fault, you place the map symbol (teeth for thrust faults and barbs for normal faults) for the fault on the hanging-wall block.

If a fault cuts across the contact between two geologic units, it must displace this contact unless the net-slip vector happens to be exactly parallel to the intersection line between the fault and the contact. The point on a map or cross section where a fault intersects

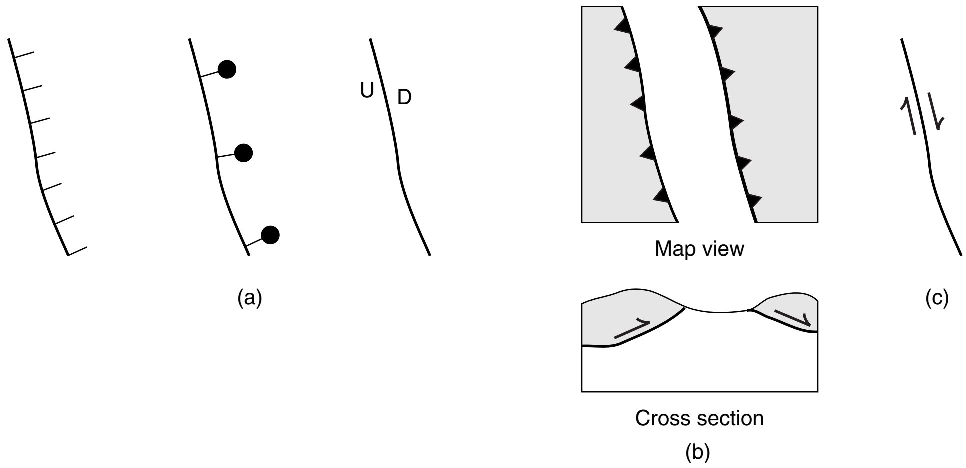


FIGURE 8.6 Basic map symbols for (a) normal fault, (b) thrust fault, and (c) strike-slip fault.

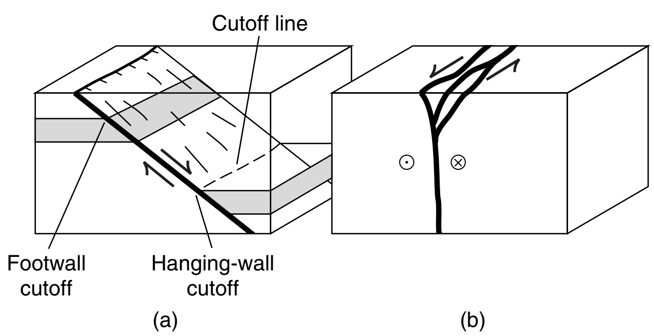


FIGURE 8.7 Block diagrams showing the different symbols for representing (a) dip-slip faults and (b) strike-slip faults (here, left-lateral). In (a) we also mark footwall and hanging-wall cutoffs.

a preexisting contact is called a **cutoff,** and in three dimensions (Figure 8.7), the intersection between a fault and a preexisting contact is a **cutoff line.** If the truncated contact lies in the hanging-wall block, the truncation is a **hanging-wall cutoff,** and if the truncated contact lies in the footwall, it is a **footwall cutoff.**

When combining map and cross-sectional surfaces with topography, we create a more realistic block diagram, giving us a three-dimensional representation of a region's geology. Consider an area that is characterized by a low-angle reverse faulting (a thrust). Where erosion cuts a hole through a thrust sheet, exposing rocks of the footwall, the hole is a **window** and the teeth are drawn outwards from the hole (Figure 8.8). An

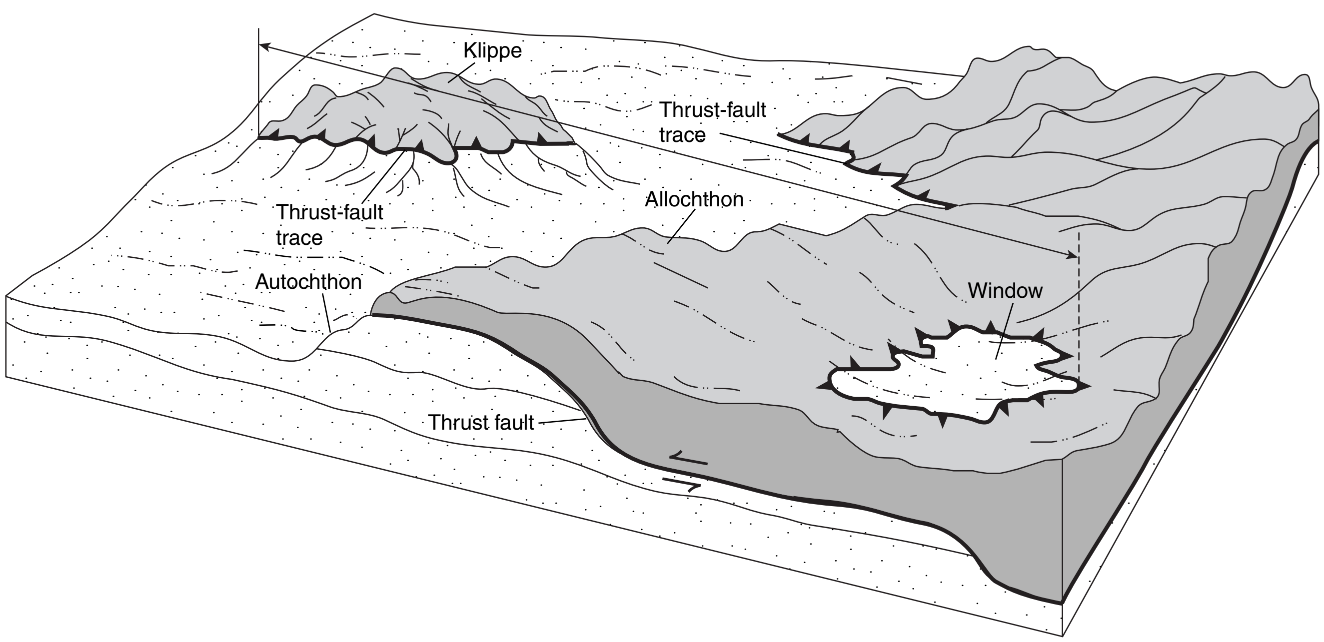


FIGURE 8.8 Block diagram illustrating klippe, window (or fenster), allochthon (gray), and autochthon (stippled) in a thrust-faulted region. Note that the minimum fault displacement is defined by the farthest distance between thrust outcrops in klippe and window.

FIGURE 8.9 Chief Mountain in Glacier National Park (USA) is an example of a klippe. The Lewis Thrust marks the contact between resistant Precambrian rock in the hanging wall and Cretaceous shale and sandstone in the footwall of this erosional remnant.

isolated remnant of a thrust sheet surrounded by exposures of the footwall is a **klippe;** this is marked by a thrust-fault symbol with the teeth pointing inwards. An imposing example of this structural geometry is Chief Mountain in Glacier Park, Montana (Figure 8.9), where old basement rocks are emplaced on young, low-grade sediments.

8.2.3 Fault Separation and Determination of Net Slip

Imagine a **marker horizon** (a distinctive surface or layer in a body of rock, such as a bed) that has been cut and offset by slip on a fault (Figure 8.10a). We define **fault separation** as the distance between the displaced parts of the marker horizon, as measured along a specified line. Separation and net slip are not synonymous, unless the line along which we measure the separation parallels the net-slip vector. The separation for a given fault along a specified line depends on the attitude of the offset marker horizon. Therefore, separation along a specified line is not the same for two nonparallel marker horizons (Figure 8.10b). Fault separation is a little difficult to visualize, so we will describe different types of fault separation with reference to Figure 8.10c, which shows an oblique-slip fault that cuts a steeply dipping bed. We define the types of separation illustrated in this figure in Table 8.3.

With the terms of Table 8.3 in mind, note that horizontal beds cut by a strike-slip fault have no strike separation and vertical beds cut by a dip-slip fault have no dip separation. If the fault cuts ground surface, this surface itself is a marker horizon for defining vertical separation, and linear features on the ground (e.g., fences, rows of trees, roads, railroads, river beds) serve as markers for defining horizontal separation. Note that Table 8.3 also defines **heave** and **throw,** which are old terms describing components of dip separation (Figure 8.10a).

In order to completely define the net-slip vector, you must specify its absolute magnitude, the direction of displacement (as a plunge and bearing) and the sense of slip. If you are lucky enough to recognize two points now on opposite walls of the fault that were adjacent prior to displacement, sometimes referred to as **piercing points,** then you can measure net slip

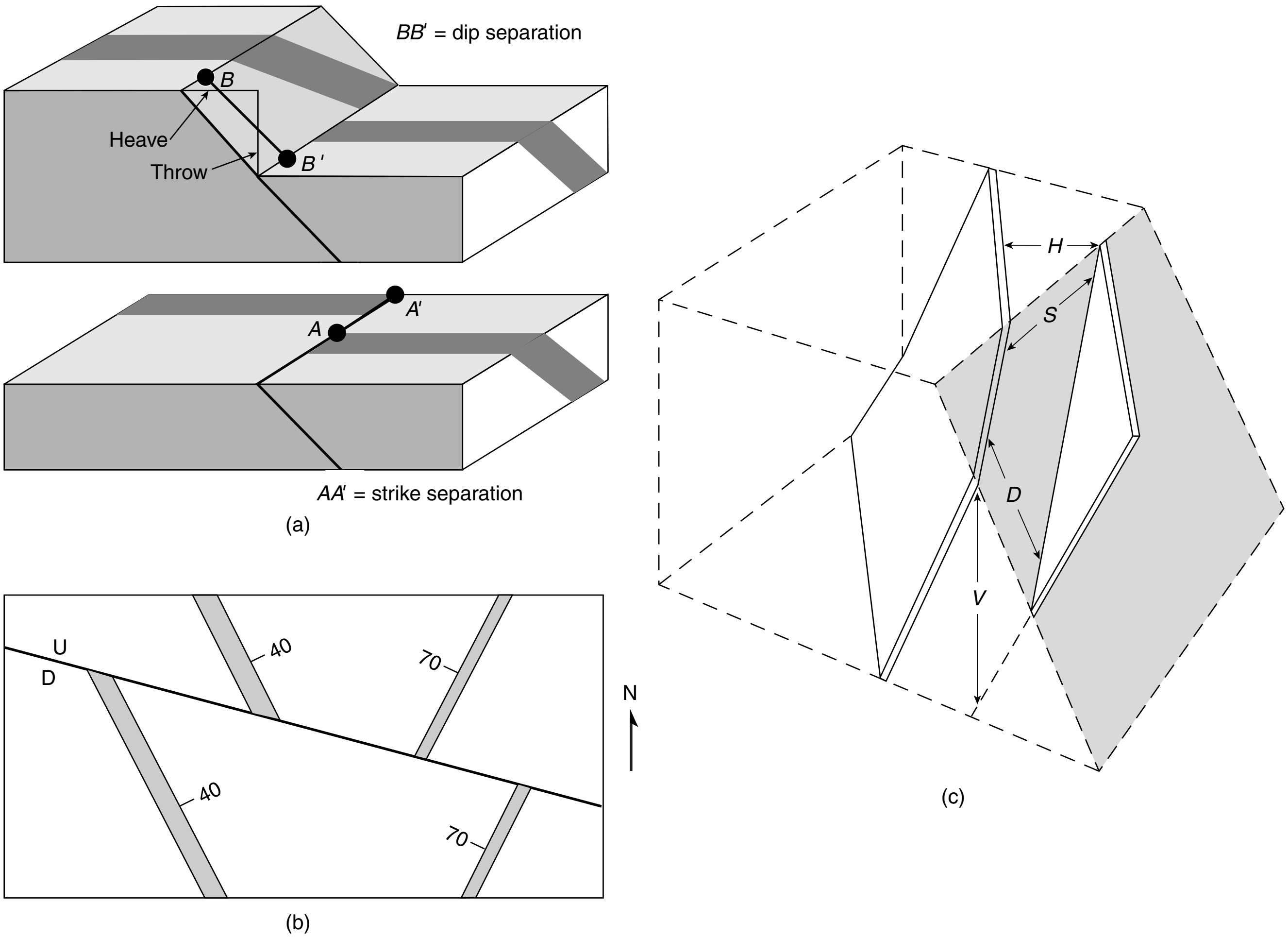


FIGURE 8.10 (a) Block diagrams showing dip separation, strike separation, heave, and throw. (b) Map view showing how separation depends on the orientation of the offset layer. The two dikes shown here dip in different directions and have, therefore, different strike separations. (c) Block diagram illustrating horizontal (*H*) and vertical (*V*) separation, as well as dip (*D*) and strike (*S*) separation.

TABLE 8.3 FAULT SEPARATION AND FAULT-SEPARATION COMPONENTS (FIGURE 8.10)

Dip separation (*D*)	The distance between the two bed/fault intersection points as measured along a line parallel to the dip direction.
Heave	The horizontal component of dip separation.
Horizontal separation (*H*)	The offset measured in the horizontal direction along a line perpendicular to the offset surface.
Stratigraphic separation	The offset measured in a line perpendicular to bedding.
Strike separation (*S*)	The distance between the two bed/fault intersection points as measured along the strike of the fault.
Throw	The vertical component of dip separation.
Vertical separation (*V*)	The distance between two points on the offset bed as measured in the vertical direction. Vertical separation is the separation measured in vertical boreholes that penetrate through a fault.

directly in the field. For example, if you observe a fence on the ground surface that has been offset by a fault, then you can define net slip, because the intersection of the fence with the ground defines a line, and the intersection of this line with the walls of the fault defines two previously adjacent points.

More commonly, however, you won't be lucky enough to observe an offset linear feature, and you must calculate the net-slip vector from other information. This can be done by measurement of (a) separation, along a specified line, of the intersection between a single marker horizon and the fault, plus information on the direction of slip; (b) separation, along two nonparallel lines, of the intersection between a single plane and the fault; or (c) separation, along a specified line, of two nonparallel marker horizons. Look at any standard structural geology methods book for an explanation of how to carry out such calculations. In the relatively rare cases where an earthquake-generating fault cuts ground surface, you can directly measure the increment of displacement accompanying a single earthquake. Generally, however, the displacement that you measure when studying ancient faults in outcrop is a **cumulative displacement** representing the sum of many incremental offsets that occurred over a long period of time.

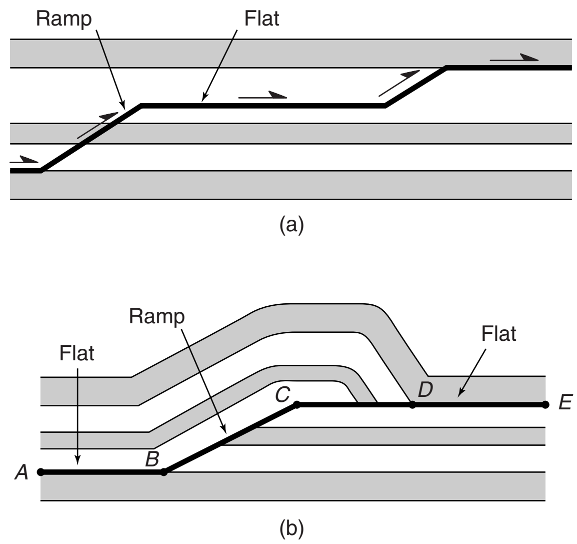


FIGURE 8.11 (a) Cross section showing the geometry of ramps and flats along a thrust fault. The fault geometry is shown prior to displacement on the fault. (b) Cross section illustrating hanging-wall and footwall flats and ramps. Segment *AB* is a hanging-wall flat on a footwall flat. Segment *BC* is a hanging-wall flat on a footwall ramp. Segment *CD* is a hanging-wall ramp on a footwall flat, and segment *DE* is a hanging-wall flat on a footwall flat.

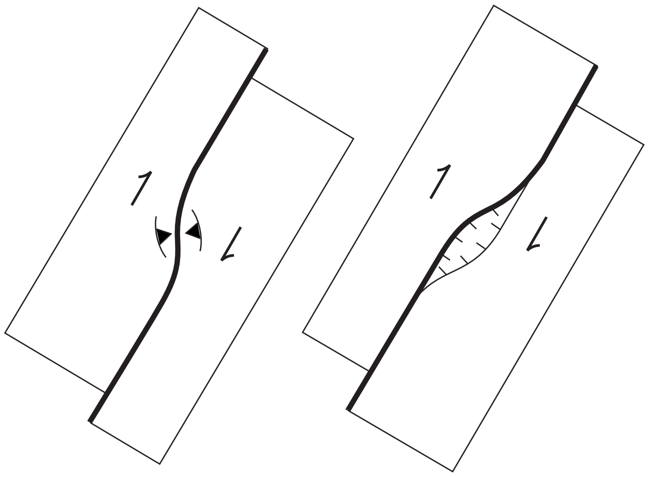
If you do not have sufficient information to determine the net slip, you can obtain valuable information about fault displacement by searching for slip lineations and shear-sense indicators. **Slip lineations** are structures on the fault that form parallel to the net-slip vector for at least the last increment of movement on the fault and, possibly, for accumulated movement during progressive deformation. **Shear-sense indicators** are structures on the fault surface or adjacent to the fault surface that define the direction in which one block of the fault moved with respect to the other. Slip lineations alone define the plunge and bearing of the net-slip vector, and with shear-sense indicators they define the direction in which the vector points. Such information can help you interpret the tectonic significance of a fault, even if you don't know the magnitude of displacement across it. We'll discuss types of slip lineations and shear-sense indicators, and how to interpret them, later in this chapter, after we have reviewed the process that leads to their formation.

The magnitude of the net-slip vector (i.e., fault displacement) on natural faults ranges from millimeters to thousands of kilometers. For example, about 600 km of net slip occurred on the oldest part of the San Andreas fault in California. In discussion, geologists refer to faults with large net slip as **major faults** and faults with small net slip as **minor faults.** Keep in mind that such adjectives are relative, and depend on context; a major fault on the scale of an outcrop may be a minor fault on the scale of a continent.

8.2.4 Fault Bends

As we mentioned earlier, fault surfaces are not necessarily planar. It is quite common, in fact, for the attitude of a fault to change down dip or along strike. In some cases, the change is gradual. For example, the dip of "down-to-Gulf" faults (see Chapter 2) along the coastal plain of Texas typically decreases with depth, so that the fault overall has a concave-up shape making these structures **listric faults.** Other faults have wavy traces because their attitude changes back and forth.

If the dip and/or strike of a fault abruptly changes, the location of the change is called a **fault bend.** Dip-slip faults that cut across a stratigraphic sequence in which layers have different mechanical properties typically contain numerous stratigraphically controlled bends that make the trace of the fault in cross section resemble a staircase. Some fault segments run parallel to bedding, called **flats,** and some cut across bedding, called **ramps** (Figure 8.11a). If the fault has not been folded subsequent to its formation, flats are (sub)horizontal, whereas ramps have dips of about 30° to 45°. Note that, as shown in Figure 8.11b, a segment of a



(a) (b)

FIGURE 8.12 Map-view illustrations of (a) a restraining bend and (b) a releasing bend along a right-lateral strike-slip fault.

fault may parallel bedding in the footwall, but cut across bedding in the hanging wall. Thus, when describing stairstep faults in a stratified sequence you need to specify whether a fault segment is a ramp or flat with respect to the strata of the hanging wall, footwall, or both.

Fault bends (or steps) along strike-slip faults cause changes in the strike of the fault. To describe the orientation of such fault bends, imagine that you are straddling the fault and are looking along its strike; if the bend moves the fault plane to the left, you say the fault steps to the left, and if the bend moves the fault plane to the right, you say that the fault steps to the right. Note that the presence of bends along a strike-slip fault results in either contraction or extension across the step, depending on its geometry. Locations where the bend is oriented such that blocks on opposite sides of the fault are squeezed together are **restraining bends,** whereas locations where the bend is oriented such that blocks on opposite sides of the fault pull away from each other are **releasing bends** (Figure 8.12).[2] Where movement across a segment of a strike-slip fault results in some compression, we say that **transpression** is occurring across the fault, and where movement results in some extension, we say that **transtension** is occurring across the fault. Note that a step to the left on a right-lateral fault yields a restraining bend, whereas a step to the right on a right-lateral fault yields a releasing bend. Try to make up the rules for a left-lateral fault yourself.

[2]The terms "restraining bends" and "releasing bends" can also be applied to steps along dip-slip faults that connect two parallel fault segments.

Natural examples of these structures, such as along the San Andreas Fault of California, will be discussed in Chapter 18.

8.2.5 Fault Terminations and Fault Length

Faulting develops at all scales, from microscopic faults that offset the boundaries of a single grain, to megascopic faults that cut laterally across thousands of kilometers of crust. But even the biggest faults do not extend infinitely in all directions. They terminate in several ways.

Faults terminate where cut by younger structures, such as another fault, an unconformity, or an intrusion (Figure 8.13a). Application of the principle of cross-cutting relationships allows you to determine the relative age of faults with respect to the structures that cut them. Some faults link to other faults while both are active (Figure 8.13b). For example, fault splays diverge from a larger fault, and faults in an anastomosing array merge and diverge along their length. Where a fault does not terminate against another structure, it must die out, meaning that the magnitude of displacement decreases along the trace of the fault, becoming zero at its tips. In some cases, a fault splits into numerous splays near its end, thereby creating a fan of small fractures called a **horsetail** (point *B* in Figure 8.13b), or it may die in an array of pinnate fractures. Alternatively, the deformation associated with the fault dies out in a zone of ductile deformation (e.g., folding or penetrative strain; point *C* in Figure 8.13b). The boundary between the slipped and unslipped region at the end of a fault is the **tip line** of the fault (Figure 8.14a). Faults that you can map today in the field terminate at the ground surface either because the fault intersected the ground when it moved or because it has been subsequently exposed by erosion. If the fault intersected the ground surface while it was still active, it is an **emergent fault** (Figure 8.14a), but if it intersects the ground surface only because the present surface of erosion has exposed an ancient, inactive fault, it is an **exhumed fault.** Exhumed and emergent faults must die out along their strike, unless they terminate at another structure; their tip line intersects the ground surface at a point. A fault that dies out in the subsurface, and thus does not intersect the ground surface, is called a **blind fault** (Figure 8.14b).

Whereas the length of some faults is limited by their intersection with, or truncation by, other structures, for many faults the trace length changes with time as the fault evolves. To visualize this process, imagine a fault that grows outward. At a given instant of time, slip has

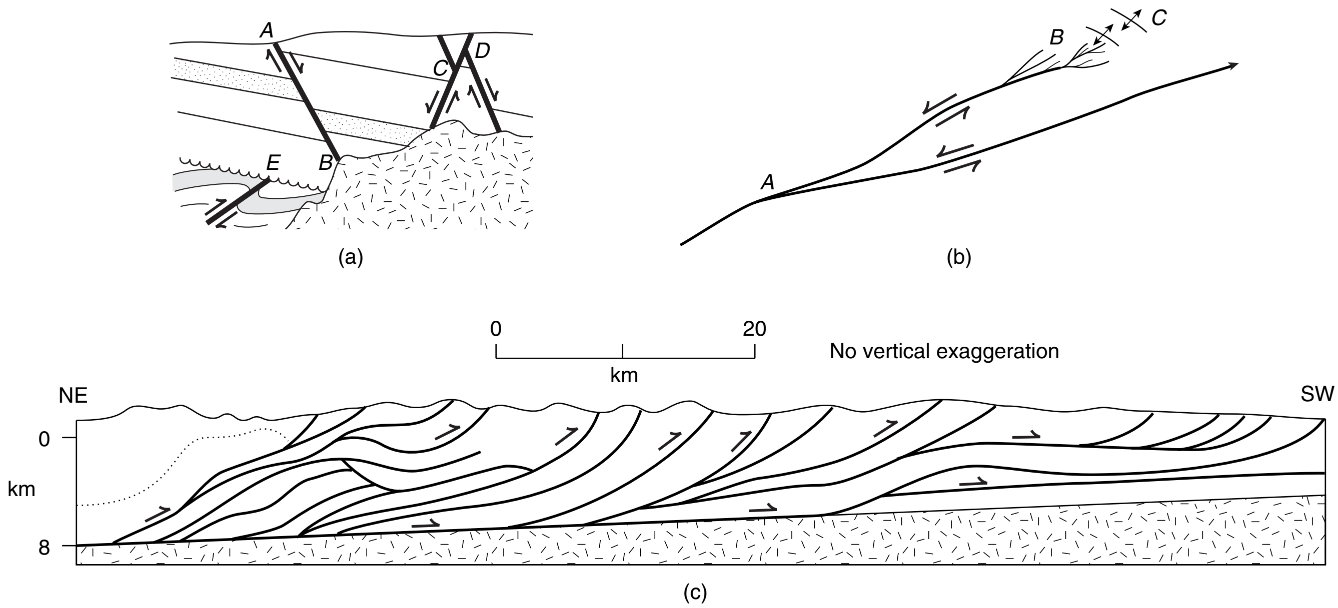


FIGURE 8.13 (a) Cross-sectional sketch showing various types of fault terminations. The fault terminates at the ground surface at point *A;* at point *B*, the fault has been cut by a pluton; at *C* and *D*, one fault cuts another; at *E*, the fault was eroded at an unconformity. (b) Termination of a fault by merging with another fault (at point *A*), or by horsetailing (at point *B*) and dying out into a zone of ductile deformation (at point *C*). (c) A series of ramps merging at depth with a basal detachment.

occurred where the fault surface already exists, but there is no slip beyond the tip of the fault (Figure 8.15a). A little later, after more fault-tip propagation, there is increased slip in the center of the fault (Figure 8.15b). As a consequence, the displacement changes along the length of the fault and the magnitude of displacement must be less than the length of the fault. Considering this relation, we might expect a general relationship between fault length and displacement: the longer the fault trace, the greater the displacement. Indeed, recent work supports this idea, though the details remain controversial. Faults that are meters long display offsets typically on the order of centimeters or less, whereas faults with lengths on the order of tens of kilometers have typical offsets on the order of several hundreds of meters. Figure 8.15c shows a plot of fault length versus offset, based on examination of thousands of faults that occur in a variety of lithologies and range in length

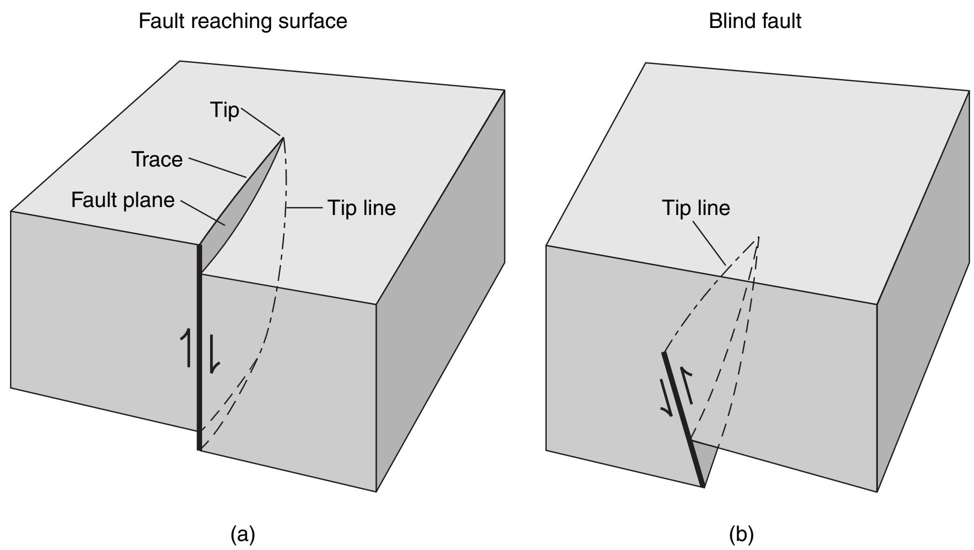


FIGURE 8.14 Tip lines for (a) an emergent fault and (b) a blind fault.

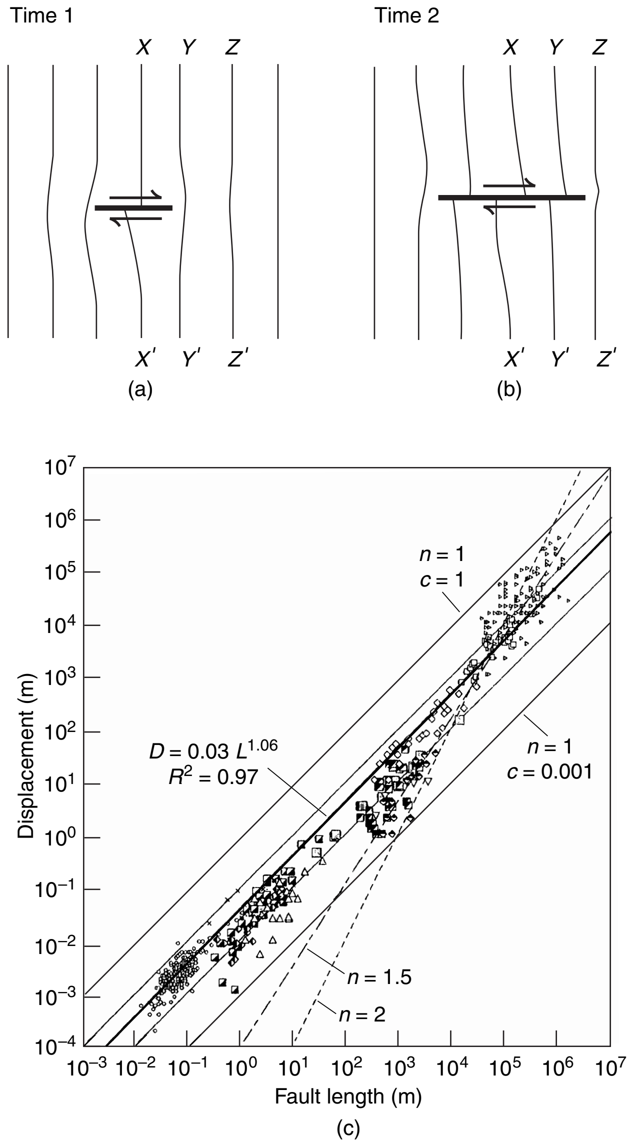


FIGURE 8.15 (a, b) Map view illustrating that displacement on a fault grows as the fault length increases. At time 1, the short fault only offsets marker line XX' by a small amount. At time 2, the fault has grown in length, and marker line XX' has been offset by a greater amount. Note that the displacement decreases toward the end of the fault. (c) Log–log plot showing the apparent relationship between fault length (L) and fault displacement (D): $D = c \cdot L^n$. The exponent, n, is called the *fractal dimension*. Various fits are possible, but a general relationship is $D = 0.03\ L^{1.06}$, suggesting an approximate displacement–length ratio of about 0.03.

from centimeters to hundreds of kilometers. The points can be fitted to a ~45° sloping band, suggesting that, given a knowledge of fault length, we can predict displacement (or vice versa) independent of the properties of the material. Debate continues as to whether a single relationship is appropriate for both regional and mesoscopic faults, as the best-fitting straight line seemingly underestimates displacement for shorter faults and overestimates for longer faults. Nevertheless, a convenient rule of thumb emerges—that fault displacement is about 3% of fault length—based on the values in Figure 8.15. Other studies indicate similar scaling properties for fault gouge/cataclasite width; Table 8.4 lists approximate relationships for fault-zone characteristics.

8.3 CHARACTERISTICS OF FAULTS AND FAULT ZONES

8.3.1 Brittle Fault Rocks

Faulting involves either shear fracturing of a previously intact rock, in which case a multitude of cracks coalesce, or slip on a preexisting fracture, which may lead to formation of new off-plane fractures and fault splays. Thus, the process of brittle faulting tends to break up rock into fragments, producing **brittle fault rock.** We classify brittle fault rock based on whether it is **cohesive** or **noncohesive** (i.e., whether or not the fragments comprising the fault rock remain stuck together to form a coherent mass without subsequent cementation or alteration) and on the *size* of the fragments that make up the fault rock. Table 8.5 summarizes the principal terms that we use to describe brittle fault rocks.

Creation of a random array of nonsystematic mesoscopic fractures that surround angular blocks of rock creates **fault breccia** (Figure 8.16a). In general, breccias have *random fabrics,* meaning they do not contain a distinctive foliation. Continued displacement across

TABLE 8.4 FIRST-ORDER RELATIONSHIPS BETWEEN FAULT PARAMETERS

	Length	Displacement	Fault Zone Width
Length	—	10^{2}	10^{4}
Displacement	10^{-2}	—	10^{2}
Fault Zone Width	10^{-4}	10^{-2}	—

Row fault property equals value times column property; for example, width = 0.01 × displacement ($W = 0.01 \times D$). *(From Scholz, 2002)*

TABLE 8.5 CLASSIFICATION OF BRITTLE FAULT ROCK

Noncohesive Brittle Fault Rocks	
Fault gouge	Rock composed of material whose grain size has been mechanically reduced by pulverization. Grains in fault gouge are less than about 1 mm in diameter. Like breccia, gouge is noncohesive. Shearing of gouge along a fault surface during progressive movement may create foliation within the gouge. Clay formed by alteration of silicate minerals in fault zones may be difficult to distinguish from true gouge.
Indurated gouge	Fault gouge that has been cemented together by minerals precipitated from circulating groundwater.
Fault breccia	Rock composed of angular fragments of rock greater than about 1 mm, and as much as several m across; fault breccia is noncohesive.
Vein-filled breccia	Fault-breccia blocks that are cemented together by vein material. Another term, *indurated breccia,* is synonymous.
Cohesive Brittle Fault Rocks	
Pseudotachylyte	A glass or microcrystalline material that forms when frictional heating melts rock during slip on a fault. Pseudotachylyte commonly flows into cracks between breccia fragments or into cracks penetrating the walls of the fault. In special cases, pseudotachylyte may be several m thick (e.g., impact sites), but generally it is mm to cm in thickness.
Argille scagliose	A fault rock that forms in very fine-grained clay- or mica-rich rock (e.g., shale or slate) and is characterized by the presence of a very strong wavy anastamosing foliation. As a consequence, the rock breaks into little scales or platy flakes.
Cataclasite	A cohesive fault rock composed of broken, crushed, or rolled grains. Unlike breccia, it is a solid rock that does not disintegrate when struck with a hammer.

(a)

(b)

FIGURE 8.16 (a) Fault breccia from the Buckskin detachment (Battleship Peak, Arizona, USA). (b) Banded clay gouge from the Lewis Thrust (Alberta, Canada).

the fault zone may crush and further fragment breccia, and/or may break off microscopic asperities protruding from slip surfaces in the fault zone, thereby creating a fine-grained rock flour that we call **fault gouge** (Figure 8.16b). Gouge and (micro)breccia are noncohesive fault rocks, meaning that they easily fall apart when collected at a fault zone or hit with a hammer.

The network of fractures between fragments in breccia and gouge allows groundwater to pass through the fault zone. Minerals like quartz or calcite may precipitate out of the groundwater, thereby cementing together rock fragments in the fault zone. As a result, breccia and gouge become **indurated,** meaning that the fragments are cemented together. In coarse breccia, the cement typically also fills veins of euhedral or blocky crystals in the open spaces between rock fragments, resulting in formation of a **vein-filled breccia** (Figure 8.16b). Circulating groundwater may also have the effect of causing intense alteration and new growth of minerals in the gouge or breccia zone. Alteration rates in fault zones tend to be greater than in intact rock, because fragmentation of rock body increases the net area of reactive surfaces. As a consequence, some minerals (e.g., feldspar) transform into clay (Figure 8.16b). The layer of clay that develops in some fault zones can act as an impermeable barrier, or **seal,** to further fluid movement. In olivine-rich rocks, such as basalt and peridotite, reaction of fault rock with water yields the mineral serpentine. As we mentioned earlier, clay and serpentine are relatively weak minerals, so their presence along a fault may allow it to slip at lower frictional stresses than it would if the original minerals were present.

Cataclasite is a cohesive brittle fault rock that differs from gouge or breccia in that the fragments interlock, allowing the fragmented rock to remain coherent even without cementation. Cataclasites generally have random fabrics (i.e., no strong foliation or lineation). Some geologists use subcategories of cataclasite, based on the proportion of matrix in the rock. In **protocataclasite,** 10–50% of the rock is matrix; in **cataclasite** (*sensu stricto*) 50–90% is matrix; and in **ultracataclasite** 90–100% is matrix.

Table 8.5 also lists two less common types of fault rock, pseudotachylyte and argille scagliose. **Pseudotachylyte** (from the prefix *pseudo-*, which means "like," and the noun tachylyte, which is a type of volcanic glass) is glass or very finely crystalline material that forms when frictional sliding generates enough heat to melt the rock adjacent to the fault (Figure 8.17). Such conditions occur during earthquakes. Because rock is not a good conductor, the heat generated by

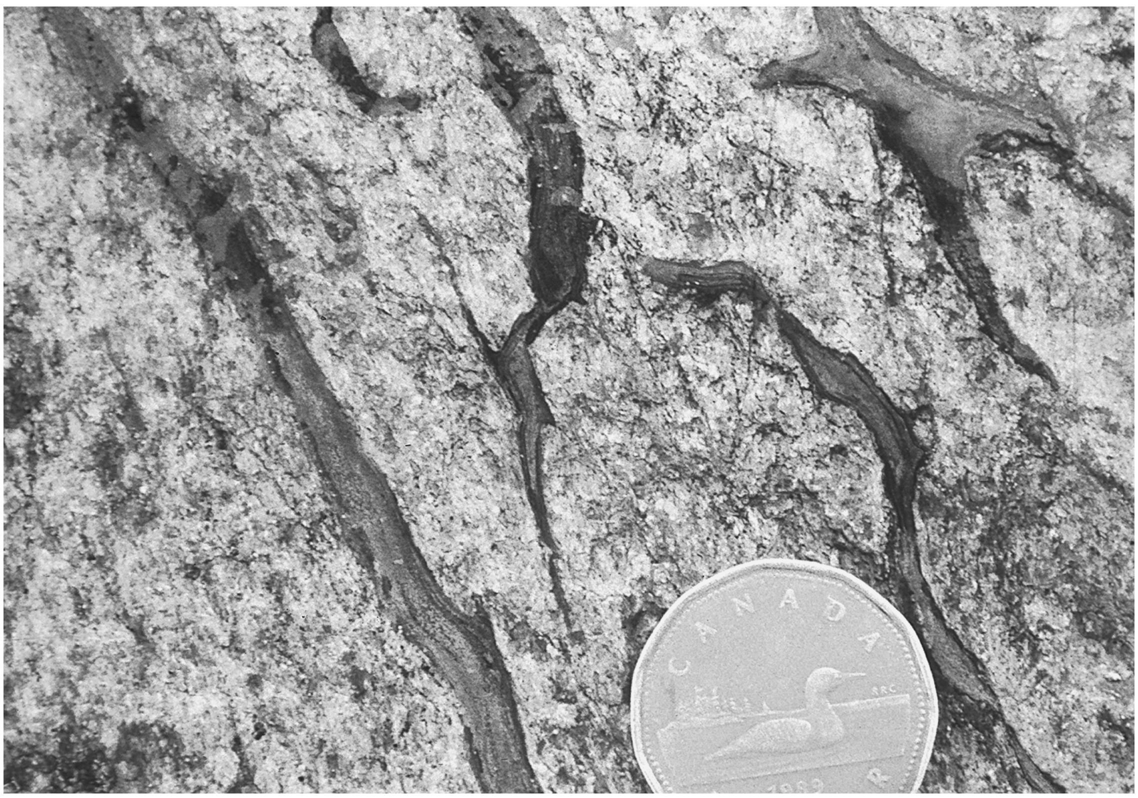
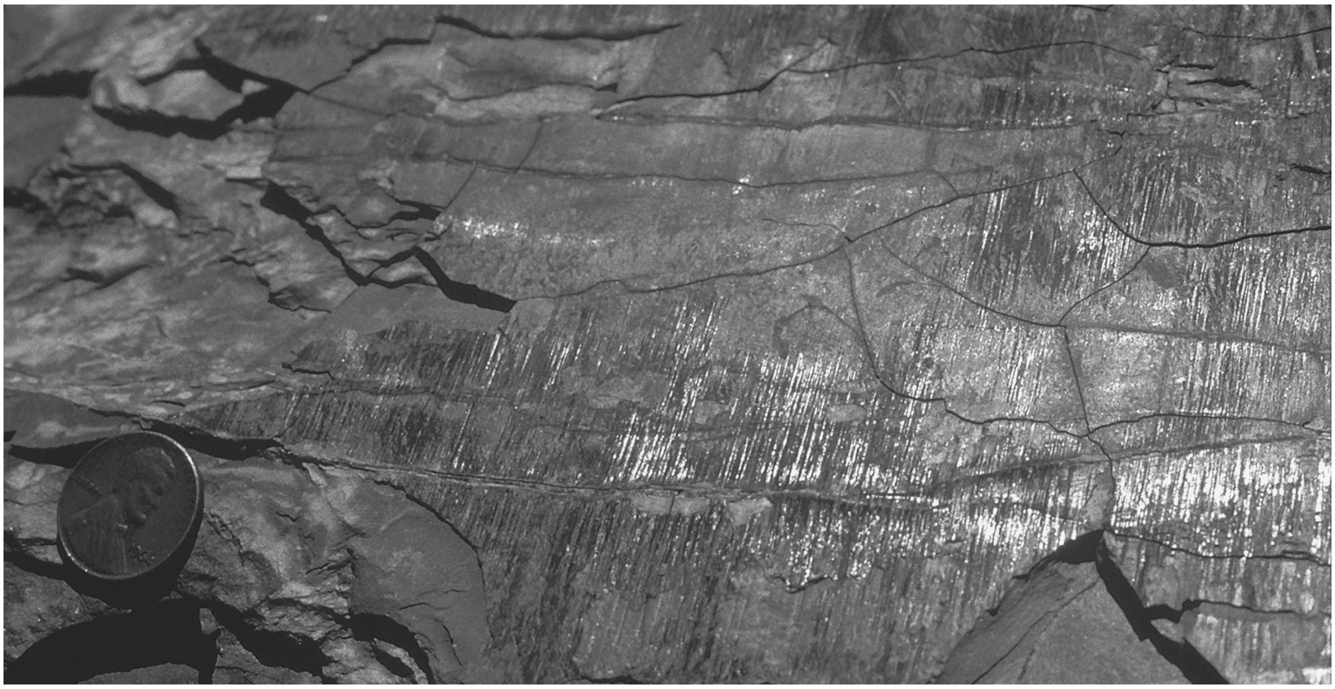


FIGURE 8.17 Pseudotachylyte (dark, wispy bands) near the Grenville Front (Ontario, Canada); looney for scale. *(Courtesy of J. Magloughlin)*



(a)

frictional sliding cannot flow away from the fault, and temperatures in the fault zone quickly become very high (>1000°C). Melt formed in such a setting squirts into cracks and pores in the fault walls where it cools so quickly that it solidifies into a glass. **Argille scagliose** refers to a strongly foliated fault rock formed by pervasive shearing of a clay-rich or very fine-grained mica-rich lithology such as shale or slate. In argille scagliose ("scaly clay" in Italian) foliation planes (microscopic shear surfaces) are anastomosing and very shiny, yielding a rock that has the overall appearance of a pile of oyster shells. Argille scagliose can develop under nonmetamorphic conditions (e.g., at the base of a sedimentary mélange) or in low-grade metamorphic conditions. Scaly fabrics also form in other fine-grained lithologies, such as coal and serpentinite.

8.3.2 Slickensides and Slip Lineations

Displacement on a fault in the brittle field involves frictional sliding and/or pressure solution slip. Each process yields distinctive structures (slickensides and slip lineations) on the fault surface, which may provide information about the direction of net slip and, in some cases, the shear sense of slip during faulting.

If slip on a fault takes place by frictional sliding, asperities on the walls of the fault break off and/or plow into the opposing surface and wear down. As a result, the two walls of the fault may become smoother and, in some cases, attain a high polish. Fault surfaces that have been polished by the process of frictional sliding are called **slickensides** (Figure 8.18). Slickensides form either on the original wall rock of the fault, or on the surface of a thin layer of gouge/cataclasite.

(b)

FIGURE 8.18 (a) Shiny slickensided surface in Paleozoic strata of the Appalachians (Maryland, USA); coin for scale. (b) Slip fibers on a fault surface, showing steps that indicate sense of shear; compass for scale.

Some asperities on one wall of the fault plow into the surface of the other wall, thereby creating **groove lineations** on the slickenside, resulting in formation of a **lineated slickenside,** also called **slickenlines.** These lineations resemble the glacial striations created when rocks entrained in the base of a moving glacier scratch across bedrock, though they are much smaller.

Groove lineations are not the only type of lineation that forms on brittle faults. On fault surfaces coated with fine-grained material, fault slip may mold gouge into microscopic linear ridges that, along with grooves, create a lineation visible on the fault surface. Also, some fault surfaces initiate with small lateral steps, whose presence gives the fault a corrugated appearance. These **corrugations** resemble grooves, but can be longer than the total displacement on the fault. The origin of corrugations is not well understood.

Some fault surfaces are coated by elongate **fibers** of vein minerals (typically quartz, calcite, or chlorite), whose long axes lie subparallel to the fault surface (Figure 8.19a). These fibers grow incrementally by the "crack-seal" deformation mechanism, or by solution mass transfer through a fluid film along the fault surface. Thus, fiber formation does not always involve brittle rupture along the fault surface. Whether frictional sliding is accompanied by fiber growth depends on the strain rate and fluid conditions during faulting. Fibers form at smaller strain rates and require the presence of water films, and typically form in imbricate sheets (Figure 8.19b). As movement continues, multiple sheets of fibers may develop on top of one another, so that a vein up to several centimeters thick eventually develops along the fault plane. In relatively thick veins (> 2 or 3 cm), the internal portion of the vein may consist of blocky growth, which forms either by recrystallization of earlier-formed fibers, or by precipitation of euhedral crystals in gaps along the fault surface.

Slip lineations trend parallel to an increment of displacement on a fault, and thus allow you to determine whether the increment resulted in strike-slip, dip-slip, or oblique-slip offset. If all movement on the fault has been in the same direction, then the slip lineation defines the orientation of the net-slip vector. You must use caution when interpreting slip lineations, however, because they may only define the last few increments of displacement on a fault surface on which slip in a range of directions occurred. Moreover, the last increment of frictional sliding sometimes erases grooves formed during earlier increments, and formation of one sheet of fiber lineations may cover and obscure preexisting sheets of fiber lineations. If more than one set of lineations are preserved on a fault surface, you can study them to distinguish among multiple nonparallel slip increments.

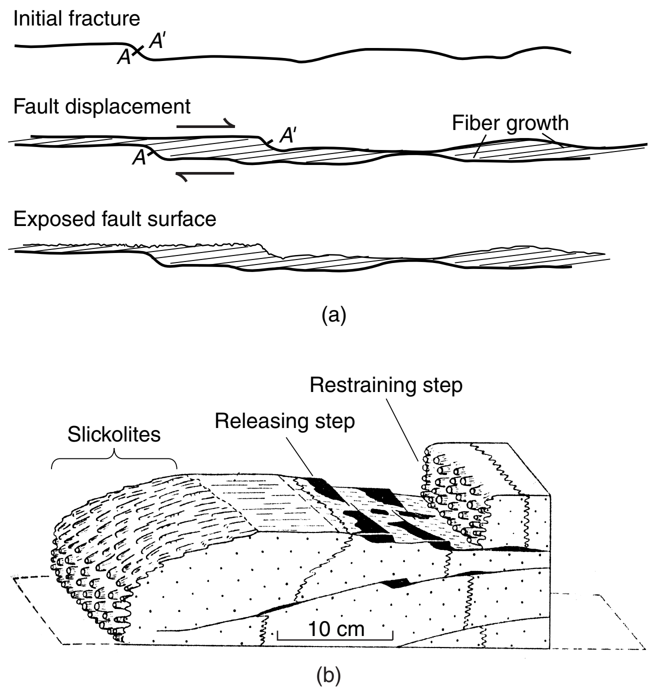


FIGURE 8.19 (a) Illustration of the growth of slip fibers along a fault, and (b) block diagram illustrating steps along a fiber-coated fault surface. Restraining steps become pitted by pressure solution, releasing steps become the locus of vein growth, and oblique restraining steps become slickolites.

Structures on a fault surface may provide constraints on the shear sense. For example, as frictional sliding occurs, an original irregularity on a slickenside becomes polished more smoothly on the upslip side, and remains rougher on the downslip side, and subtle steps may develop on the fault surface. These features create an anisotropy on a slickenside; the surface feels smoother as you slide your hand in the shear direction as opposed to sliding your hand in the direction opposite to the shear direction. However, anisotropy on a slickenside is subtle and may be ambiguous because the intersection of pinnate fractures with the fault surface may create steps that face in the opposite direction to the steps formed by differential polishing.

Sheets of mineral fibers formed during slip on a fault provide a more reliable indication of shear sense. Because the fibers composing the sheets form at a low angle to the fault surface, fiber sheets tend to overlap one another like shingles, and tilt away from the direction of shear. Features developed on mesoscopic steps along a fault that moved by pressure-solution slip may also define shear sense. Restraining steps oppose movement on the fault, and therefore become pitted by pressure solution. Pit axes on the steps are roughly parallel to the net-slip vector, and thus are subparallel to the fault surface. If the restraining-step face is not perpendicular to the fault surface, its pit axes are oblique to the step face. Surfaces containing such oblique pits look like a cross between a stylolite and a slip lineation, and thus some authors call them **slickolites** (Figure 8.19b). Releasing steps, at which the opposing walls of the fault pull away from one another, typically become coated with fibrous veins. The long axes of the fibers of these veins parallel the pit axes on releasing steps along the same fault.

8.3.3 Subsidiary Fault and Fracture Geometries

Fault zones consist of one or more major faults, along with an array of subsidiary faults, including both discrete smaller faults that occur within a larger fault zone (and may anastomose with one another) and fault splays that branch off. Such subsidiary faults may initiate when the primary rupture splits into more than one surface during its formation in intact rock, when numerous subparallel faults initiate simultaneously in a fault zone, when conjugate shear fractures develop at an angle to the principal fault, or when numerous preexisting surfaces in the fault zone reactivate during a deformation event.

In the case of emergent strike-slip faults, like the San Andreas Fault, a particularly interesting array of subsidiary faults, known as R- (or Riedel) and P-shears, develops. To picture how these develop, imagine an experiment in which you place a thin layer of clay over two wooden blocks, and then shear one block horizontally past the other (Figure 8.20a). The clay develops ductile strain before any fractures appear. When fracturing in the clay begins to develop, the first fractures are short shear fractures that are inclined at an angle to the trace of the throughgoing fault that eventually forms (Figure 8.20b). These short fractures are called *R*- or **Riedel shears.**[3] Generally, you will find two distinct sets of Riedel shears (R and R′) that together define a conjugate pair. The bisector of the acute angle between conjugate R- and R′-shears reflects the local orientation of σ_1 adjacent to the future fault. As shear continues, a third set of fractures, called P-shears, develops. P-shears link together the previously formed Riedel shears, and, eventually, a throughgoing fault zone consisting of linked R-, R′-, and P-shears develops. Whether subsidiary shears also form at depth in the earth remains debatable. In fact, some geologists suggest that they only form when a weak layer is sheared by relative displacement of two stronger blocks on either side or below, much as in our experiment.

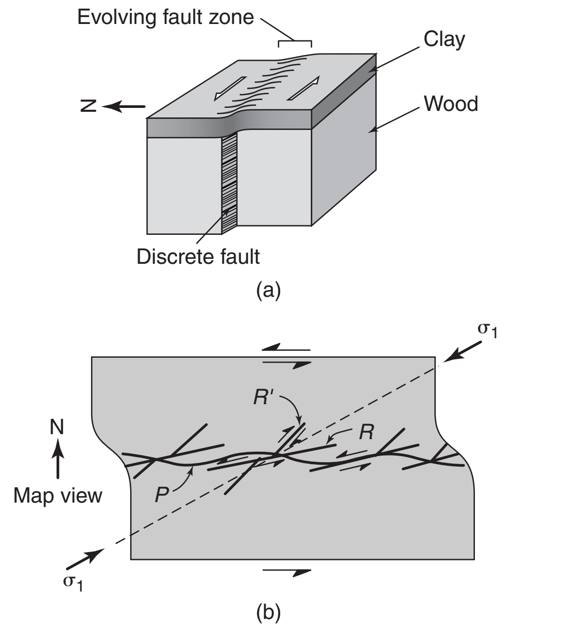


FIGURE 8.20 Growth of R-, R′-, and P-shears. (a) Schematic diagram illustrating a layer of clay that deforms when underlying blocks of wood slide past one another. (b) Map view of the top surface of the clay layer, illustrating the orientation of Riedel (R), conjugate Riedel (R′), and P-shears. Note that the acute bisector of the R- and R′-shears is parallel to the remote σ_1 direction.

Tensile fractures can also develop in association with faulting, which may occur in the wall rock, or in the fault zone itself. A series of parallel tensile fractures that forms within a fault zone defines an **en echelon array** or **stepped array** that tends to dilate and become veins (see Chapter 7). Typically, stepped veins in a fault zone initiate at an angle of about 45° to the zone boundary and rotate with the direction of shear. With progressive displacement of the fault-zone walls, the earlier-formed parts of the veins rotate, but new vein increments initiate at an angle of about 45° to the walls, producing a sigmoidal shape whose sense of rotation defines the sense of shear on the fault (see Figure 7.23).

8.3.4 Fault-Related Folding

Faults and folds commonly occur in the same outcrop. The spatial juxtaposition of these structures, one brittle and one ductile, may seem paradoxical at first. How can rock break at the same time that it distorts ductilely? One explanation for the juxtaposition of faults and folds in an outcrop is that these structures formed at different times under different pressure and temperature conditions. For example, imagine that the folds

[3]After the geologist who first discussed this relationship.

formed 500 million years ago at a depth of 15 km in the crust, but, once formed, the rock body containing the folds was unroofed and moved to shallow depths, where a later, and totally separate, deformation event created brittle faults in the body. Typically, in such examples, the faults cut across the preexisting folds and are not geometrically related to the folds. But in many instances it is clear from the spatial and geometric relation between folds and faults that the two structures formed together during the same deformation event. Folds that form in association with faults are called **fault-related folds.**

From earlier experiments we learned that the transition between brittle and ductile deformation can depend on strain rate (Chapter 5; recall the behavior of Silly Putty®). If ductile strain (e.g., folding) in a rock body develops fast enough to accommodate regional deformation, then differential stress does not get very high in the rock and faulting need not occur. But if regional deformation cannot be accommodated sufficiently fast by ductile strain, then differential stress builds until it exceeds the failure strength of the rock or the frictional strength of an existing fracture, and faulting occurs. Alternatively, strain rate may vary with position in a rock body, so that while one region in a rock body faults, another region in the rock body folds. Finally, strain rate and differential stress magnitude can vary with time at a given location during the same overall deformation event, so that episodes of faulting and folding may alternate at the location.

In this section, we introduce several types of folds that develop in association with faults, but we again return to the subject of fault-related folding when we discuss aspects of regional deformation (Chapters 17 and 18). Imagine a horizontally stratified rock that undergoes shortening due to regional compression. Initially, the ends of the body move toward one another very slowly. If, during this stage, differential stress in the body is lower than the shear failure strength of the rock, the layers will fold. Such folding may yield merely a gentle flexure of adjacent beds, or a pronounced asymmetric anticline-syncline pair (Figure 8.21). If, at a later time, folding can no longer accommodate the internal displacement of the block, the differential stress magnitude in the block increases and faulting initiates. Stress buildup leading to faulting may reflect a change in the strain rate or may reflect locking up of the folds. By locking up, we mean that the change in the geometry of the beds resulting from fold formation makes continued folding more difficult (i.e., like a stick, the layers can be bent only so far before they break). Note that the sense

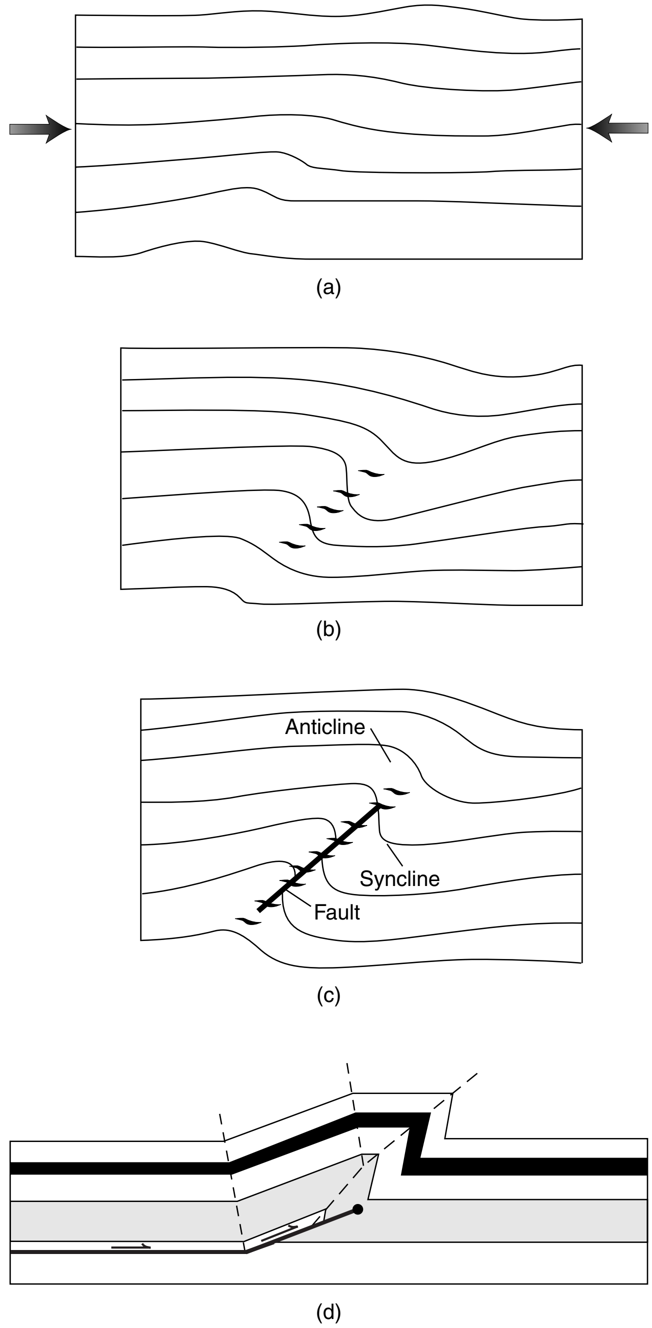


FIGURE 8.21 Progressive development of fault-related folding in a stratified sequence. (a) A small flexure develops during shortening of the layers, and a pronounced anticline-syncline pair develops. (b) *En echelon* (or stepped) gashes form in the fold. (c) A fault breaks through the fold, cutting through a gentle flexure. (d) Geometry of a fault-propagation fold.

of asymmetry of these folds[4] reflects the sense of shear on the fault. In some cases, a fault cuts through the fold along the hinge of the fold, whereas in other cases, the fault breaks through the limbs of the folds between the adjacent anticline and syncline (at the

[4] Sometimes called *fault-inception folds.*

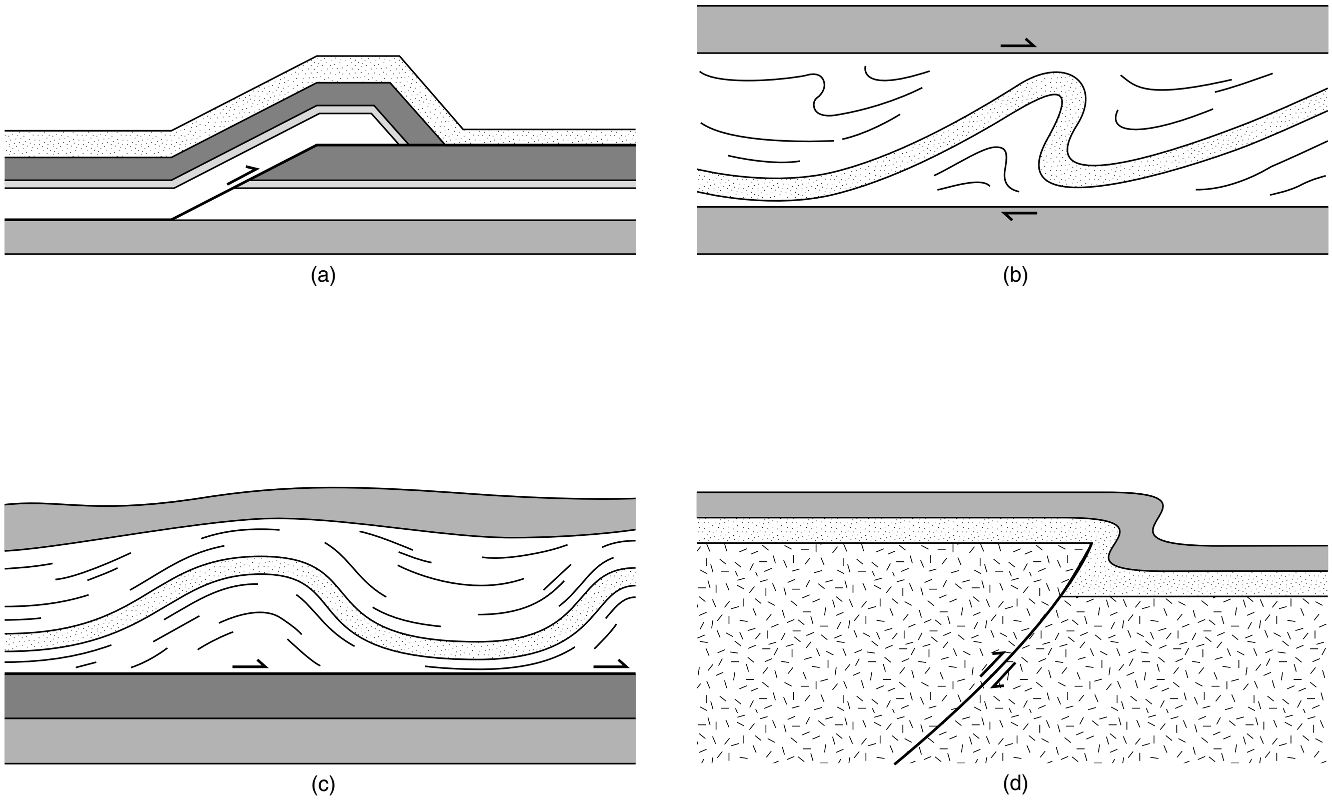


FIGURE 8.22 Fault-related folds. (a) Fault-bend fold on a thrust. (b) Folding in a fault zone. (c) Detachment fold. (d) Drape fold over faulted basement.

inflection surface of the fold); these are called **fault-propagation folds** (Figure 8.21d).

Bends in a fault cause folding of strata that move past the bend during displacement, because the moving layers must accommodate the bends without gaps. Folds that form in this manner are called **fault-bend folds.** Fault-bend folds form in association with all types of faults, but most of the literature concerning them pertains to examples developed along dip-slip faults (Figure 8.22a).

A third situation in which folding accompanies faulting occurs where a sequence of interlayered weak and strong rock layers (e.g., interbedded shale and sandstone) is caught between the opposing walls of a fault zone (Figure 8.22b). In such a circumstance, displacement of the rigid fault-zone walls with respect to one another causes the intervening rock to fold, much like a carpet that is caught between the floor and a sliding piece of furniture. Folds in fault zones formed by such a process tend to be very asymmetric. Typically, the hinges of folds within the fault zone are initially perpendicular to the overall shear direction, but with continued displacement, they may be bent, eventually curving into parallelism with the shear direction. Fault zones in which such folding occurs are effectively ductile shear zones (see Chapter 12), but we mention them here because this deformation can occur in association with brittle sliding.

The asymmetry of fault-zone folds provides a clue to the sense of shear on associated faults. However, because the hinges of fault-zone folds change during progressive shear, the geometry of a single fold may not necessarily provide the net-slip vector. You can define its direction if you measure a slip lineation on the fault surface, but if such lineations are not present, you may be able to obtain this information by measuring numerous folds with a range of hinge trends using a technique called the **Hanson slip-line method.**[5] We leave a description of this method for laboratory books on structural geology.

The sheet of rock above a detachment fault may deform independently of the rock below. The resulting folds are called **detachment folds** (Figure 8.22c). Shear on the detachment fault accommodates the contrast in strain between the folded hanging-wall and the unfolded footwall.

[5]Named after its originator.

TABLE 8.6 SHEAR-SENSE INDICATORS FOR BRITTLE FAULTS AND FAULT ZONES

Offset markers	You can define shear sense if you are able to define the relative displacement of two piercing points on opposite walls of the fault, or can calculate the net-slip vector based on field study of the separation of marker horizons.
Fault-related folds	The sense of asymmetry of fault-related folds defines the shear sense. Typically, fault-inception folds verge in the direction of shear (see Chapter 11 for a definition of fold vergence). If the hinges of folds in a fault zone occur in a range of orientations, you may need to use the Hanson slip-line method to determine shear sense. Note that the asymmetry of rollover folds relative to shear sense is opposite to that of other fault-related folds.
Fiber-sheet imbrication	The imbrication of slip-fiber sheets on a fault provides a clear indication of shear sense. Fiber sheets tilt away from the direction of shear.
Steps on slickensides	Microscopic steps develop along slickensided surfaces. Typically the face of the step is rougher than the flat surface. However, slickenside steps may be confused with the intersection between pinnate fractures and the fault, giving an opposite shear sense.
En echelon veins	*En echelon* veins tilt toward the direction of shear. If the veins are sigmoidal, the sense of rotation defines the shear sense.
Carrot-shaped grooves	Grooves on slickensides tend to be deeper and wider at one end and taper to a point at the other, thus resembling half a carrot. The direction in which that "carrot" points defines the direction of shear.
Chatter marks	As one fault block moves past another, small wedge-shaped blocks may be plucked out of the opposing surface. The resulting indentations on the fault surface are known as chatter marks.
Pinnate fractures	The inclination of pinnate fractures with respect to the fault surface defines the shear sense.

In continental-interior platform regions (e.g., the Midcontinent of the United States), a thin and relatively weak veneer of Phanerozoic sedimentary rock was deposited over relatively rigid Precambrian crystalline basement. At various times during the Phanerozoic, steeply dipping faults in the basement reactivated, causing differential movement of basement blocks. This movement forces the overlying layer of sedimentary rocks to passively bend into a fold that drapes over the edge of the basement block. Fault-related folds that form in this way are called **drape folds** or **forced folds** (Figure 8.22d). Kinematically, they are fault-propagation folds, but they are given a separate name to emphasize their unique tectonic setting.

We close with a comment on the term **drag fold,** which is used to refer to all fault-related folds in older literature. This general application of the term is misleading, because "drag" implies that the fold formed by shear resistance on the fault that retarded movement of the hanging wall with respect to the footwall. Rather, as we've seen, fault-related folds form in a number of different ways, most of which do not involve such "drag." Thus we discourage and avoid the use of this term.

8.3.5 Shear-Sense Indicators of Brittle Faults—A Summary

So far we have explored how various features of fault zones can be used to determine the shear sense. Given the lengthy descriptions that this required, we provide a concise summary of the various brittle shear-sense indicators in Table 8.6.

8.4 RECOGNIZING AND INTERPRETING FAULTS

Displacement on a fault can alter the landscape. The nature of the disruption depends on the type of fault, on how much displacement occurred, on the rate of displacement, on how recently faulting occurred, on the climate in the area, and on whether the fault is emergent or blind. We include climate and age of faulting in our list of factors because they determine the extent to which erosion can erase the effects of displacements. For example, in a humid climate with abundant rainfall, a fault's geomorphologic manifestation may disappear

FIGURE 8.23 Fault scarp in the Basin and Range Province (Nevada, USA). Note the normal sense of offset, with person for scale.

within months or years, whereas in a desert climate, a fault's geomorphologic manifestation may last for millennia.

Seismic faulting along an emergent strike-slip fault typically creates a **surface rupture,** which is manifested by broken ground and fissures. Displacement on an emergent dip-slip fault creates a step in the ground surface, called a **fault scarp** (Figure 8.23). Both dip-slip and strike-slip faulting can offset topographic features, such as ridges or river beds, and a recent fault may offset roads, fences, and sections of buildings. You can identify the trace of the San Andreas Fault in the wine district of California, for example, by offset rows of casks in a wine cellar, and historic faults by offsets in ancient buildings in Greece. Erosion typically destroys surface ruptures and fault scarps within tens to thousands of years, depending on climate. However, the abrupt change in elevation of the ground surface caused by the faulting can be maintained even after the original scarp has eroded away. Recent movement on a blind fault, by definition, does not result in a surface rupture, but the trace of the fault may be evident from differential movement of the ground surface, which can be detected by ground-based surveying equipment or by modern GPS (global positioning system) equipment.

Displacement across regional strike-slip faults (such as the Alpine Fault of New Zealand and California's San Andreas Fault) is not necessarily exactly parallel to the fault surface, in part due to the existence of restraining and releasing bends along the fault, and in part due to the relative motion of lithospheric plates during or after faulting. Transpression across a strike-slip fault may result in the formation of **fault-parallel ridges** that may be tens to hundreds of meters high (Figure 8.24), and transtension along a large strike-slip fault results in the development of **sag ponds.** Because of these features, the traces of major strike-slip faults are so obvious that they can be seen from space. If transpression occurs along a fault for millions of years, a mountain range develops adjacent to the fault (e.g., the New Zealand Alps), and if transtension occurs for millions of years, a sizable sedimentary basin develops adjacent to the fault. As we will see in Chapter 19, displacement in a regional-scale strike-slip fault zone may result in the formation of en echelon or stepped structures near the ground surface, which may have surface manifestations in the form of local ridges or troughs oriented at an angle to the main fault trace.

Inactive emergent or exhumed faults also have a topographic manifestation, for two reasons. First, the fault zone itself may have a different resistance to erosion than the surrounding intact rock. If the fault zone consists of noncohesive breccia or gouge, it tends to be weaker than the surrounding intact rock and thus more easily erodes. As a consequence, the fault trace on the ground surface evolves into a linear trough that may control surface drainage. Alternatively, if the fault becomes indurated, it may become more resistant to erosion than the surrounding region and will stand out in relief. Second, if the fault juxtaposes two rock units with different resistance to erosion, then a topographic scarp develops along the trace of the fault because the weaker unit erodes more rapidly, and the land surface underlain by the weaker unit becomes topographically lower (Figure 8.25). Such a **fault-line scarp** differs from a fault scarp, in that it is not the plane of the fault itself. In fact, the slope direction of a fault-line scarp may even be opposite to the dip direction of the underlying fault.

Fault zones are sometimes manifested by subtle features of the landscape. For example, fault zones containing abundant fractures may control local groundwater movement and may preferentially drain the surface, an effect that can cause changes in vegetation that appear as lineaments in remote-sensing images. Similarly, faults may offset structurally controlled topographic ridges, and thus may cause an alignment of ridge terminations. Ancient and inactive faults can be identified in the field, even if they do not disrupt the landscape or if fault-related structures have been covered or removed. Features indicative of faulting include the following: juxtaposition of rock units that were not in contact when first formed, offset of marker horizons, loss or duplication of section; and the presence of slickensided surfaces, slip lineations, fault rocks, and fault-related folds. Doing some of your own fieldwork in faulted regions is simply the best way to obtain an appreciation of the many manifestations of faulting.

FIGURE 8.24 Ridges formed along a transpressional segment of the San Andreas Fault. The striped white and gray rocks are basement that is pushed up relative to the (dark) sedimentary cover, in response to shortening along this strike-slip fault.

8.4.1 Recognition of Faults from Subsurface Data

To many geologists, particularly those working in economic geology, petroleum geology, and hydrogeology, the "field" consists of the subsurface region of the crust, and "field data" includes drill-hole logs (including downhole records of lithology based on cores or cuttings, electrical-conductivity measurements, or gamma-ray measurements), geophysical measurements (e.g., regional variation in the Earth's gravity and magnetic fields), and seismic-reflection data. Analysis of drill-hole logs and geophysical data provides an important basis for identification of subsurface faults.

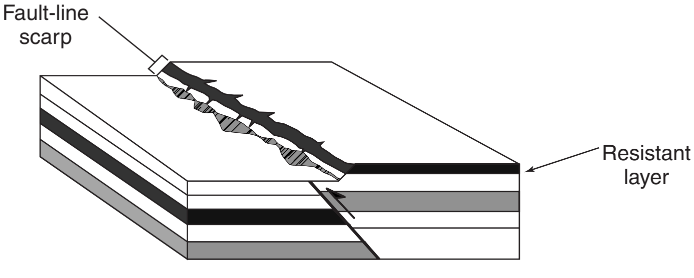


FIGURE 8.25 (a) Block diagram illustrating a fault-line scarp caused by the occurrence of a resistant stratigraphic layer (in black) that has been uplifted on one side.

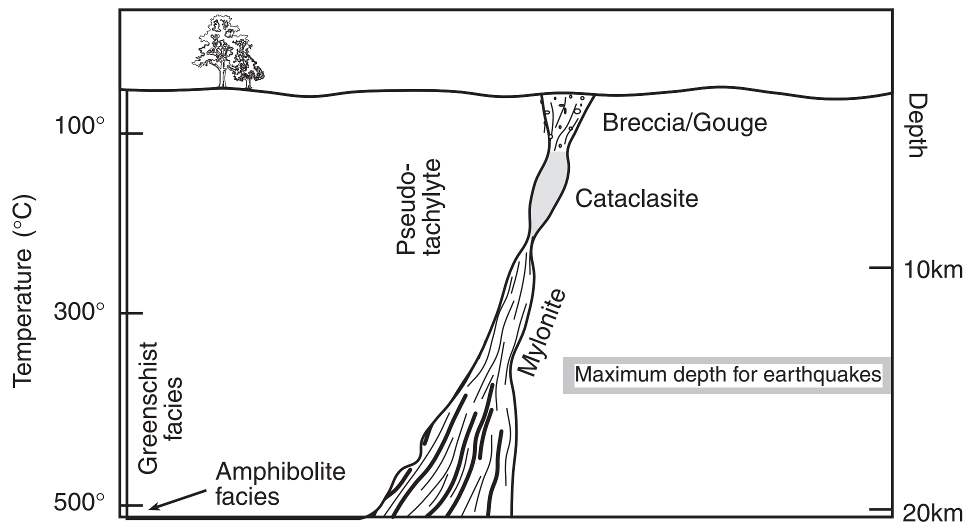


FIGURE 8.26 Change in fault character with depth for a steeply dipping fault. Note the change in fault zone width and types of structures with depth.

Evidence for subsurface faulting includes the following: (1) abrupt steps on structure-contour maps; (2) excess section (i.e., repetition of stratigraphy) or loss of section in a drill core; (3) zones of brecciated rock in a drill core, although weak fault rocks typically do not survive drilling intact (and thus fault zones may appear as gaps in a core); (4) seismic-reflection profiles, on which faults appear either as reflectors themselves, or as zones which offset known reflectors; and (5) linear anomalies or an abrupt change in the wavelength of gravity and/or magnetic anomalies, suggesting the occurrence of an abrupt change in depth to a particular horizon. We explore the use of geophysical methods in (regional) structural analysis in Chapter 15.

8.4.2 Changes in Fault Character with Depth

The characteristics of a fault depend on the magnitude of displacement on the fault, on whether or not faulting ruptures a previously intact rock or activates a preexisting surface, and on the pressure and temperature conditions (i.e., burial depth) at which faulting occurs. Figure 8.26 is a synoptic diagram of the various expressions of a major crustal fault zone. At very shallow depths in the earth (less than ~5 km), mesoscopic faults that form by reactivation of a preexisting joint or bedding surface typically result in discrete slickensided or fiber-coated surfaces. Mesoscopic shallow-level faults that break through previously intact rock tend to be bordered by thin **breccia** or **gouge zones,** and macroscopic faults, which inevitably break through a variety of rock units and across contacts, tend to be bordered by wider breccia and gouge zones, and subsidiary fault splays.

As we discussed earlier, rocks become progressively more ductile with depth in the crust, because of the increase in temperature and pressure that occurs with depth. Consequently, at depths between ~5 km and 10–15 km, faulting tends to yield a fault zone composed of **cataclasite.** Whereas cataclasite forms by brittle deformation on a grain scale, movement in the fault zone resembles viscous flow and strain is distributed across the zone (i.e., we have ductile behavior; see Chapter 9). The **brittle-plastic transition** for typical crustal rocks lies at a depth of 10–15 km in the crust. We purposely specify the transition as a range, because rocks consist of different minerals, each of which behaves plastically under different conditions, and because the depth of transition depends on the local geothermal gradient. Temperature conditions at a depth of around 10–15 km are in the range of 250°C to 350°C (i.e., lower greenschist facies of metamorphism), where plastic deformation mechanisms become the dominant contributor to strain in (quartz-rich) crustal rocks. The activity of plastic deformation mechanisms below this brittle-plastic transition yields a fine-grained and foliated fault-zone rock, called **mylonite.**

The degree of ductile deformation that accompanies faulting also depends on the strain rate and on the fluid pressure. At slower strain rates, a given rock type is weaker and tends to behave more ductilely. Thus, rocks can deform ductilely even at shallow crustal levels if strain rates are slow, whereas rocks below the brittle-plastic transition can deform brittlely if strain rates are high. A given shear-zone interval may therefore contain both mylonite and cataclasite, either as a consequence of variations in strain rate at the same depth or because progressive displacement eventually transported the interval across the brittle-plastic boundary. At very rapid strain rates (seismic slip), fault displacement in dry rock generates **pseudotachylyte.**

The width of a given fault zone typically varies as a function of rock strength; fault zones tend to be nar-

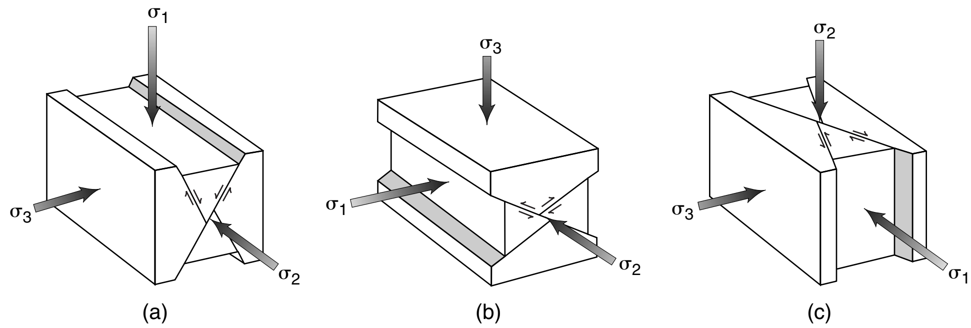


FIGURE 8.27 Anderson's theory of faulting predicts (a) (high-angle) normal faults, (b) (low-angle) reverse faults (or thrusts), and (c) (vertical) strike-slip faults.

rower in stronger rock. Thus, the width of a transcrustal fault zone may vary with depth. Very near the surface (within a few kilometers), the fault diverges into numerous splays, because the near surface rock is weakened by jointing and by formation of alteration minerals (e.g., clay). At somewhat greater depths, rock is stronger, and the fault zone may be narrower. At still greater depths, where cataclastic flow dominates, the fault zone widens. Similarly, we expect to find widening of the mylonitic segment of the fault zone with depth, where the rocks as a whole become weaker.

8.5 RELATION OF FAULTING TO STRESS

Faulting represents a response of rock to shear stress, so it only occurs when the differential stress ($\sigma_d = \sigma_1 - \sigma_3 = 2\sigma_s$) does not equal zero. Because the shear-stress magnitude on a plane changes as a function of the orientation of the plane with respect to the principal stresses, we should expect a relationship between the orientation of faults formed during a tectonic event and the trajectories of principal stresses during that event. Indeed, faults that initiate as Coulomb shear fractures will form at an angle of about 30° to the σ_1 direction and contain the σ_2 direction. This relationship is called **Anderson's theory of faulting.**[6] Why isn't σ_1 at 45° to the fault planes, where the shear stress is maximum? Recall the role of the normal stress, where the ratio of shear stress to normal stress on planes orientated at about 30° to σ_1 is at a maximum (see Chapter 6).

The Earth's surface is a "free surface" (the contact between ground and air/fluid) that cannot, therefore, transmit a shear stress. Therefore, regional principal stresses are parallel or perpendicular to the surface of the Earth in the upper crust. Considering that gravitational body force is a major contributor to the stress state, and that this force acts vertically, stress trajectories in homogeneous, isotropic crust can maintain this geometry at depth. Anderson's theory of faulting states that in the Earth-surface reference frame, normal faulting occurs where σ_2 and σ_3 are horizontal and σ_1 is vertical, thrust faulting occurs where σ_1 and σ_2 are horizontal and σ_3 is vertical, and strike-slip faulting occurs where σ_1 and σ_3 are horizontal and σ_2 is vertical (Figure 8.27). Moreover, the dip of thrust faults should be ~30°, the dip of normal faults should be ~60°, and the dip of strike-slip faults should be about vertical. For example, if the σ_1 orientation at convergent margins is horizontal, Anderson's theory predicts that thrust faults should form in this environment, and indeed belts of thrust faults form in collisional mountain belts.

Anderson's theory is a powerful tool for regional analysis, but we cannot use this theory to predict all fault geometries in the Earth's crust for several reasons. First, faults do not necessarily initiate in intact rock. The frictional sliding strength of a preexisting surface is less than the shear failure strength of intact rock; thus, preexisting joint surfaces or faults may be reactivated before new faults initiate, even if the preexisting surfaces are not inclined at 30° to σ_1 and do not contain the σ_2 trajectory. Preexisting fractures that are not ideally oriented with respect to the principal stresses become oblique-slip faults. Second, a fault surface is a material feature in a rock body whose orientation may change as the rock body containing the fault undergoes progressive deformation. Thus, the fault may rotate into an orientation not predicted by Anderson's theory. Local stress trajectories may be different from regional stress trajectories because of local heterogeneities and weaknesses (e.g., contacts between contrasting lithologies, preexisting faults) in the Earth's crust. As a consequence, local fault geometry might not be geometrically related to regional stresses. Third, systematic changes in stress trajectories

[6]After the British geologist E. M. Anderson.

are likely to occur with depth in mountain belts (e.g., along a regional detachment), but Anderson's theory assumes that the stress field is homogeneous and that the principal stresses are either horizontal or vertical, regardless of depth.

8.5.1 Formation of Listric Faults

Let us consider changes in stress field with depth in more detail, because they explain the occurrence of **listric faults,** which are faults that have decreasing dip with depth. Deformation intensity, as manifested by strain magnitude, decreases from the interior to the margin of an orogenic belt. We can infer from this observation that the magnitude of horizontal tectonic stress similarly decreases from the interior to the margin of the belt. Now, envision a sheet of rock above a detachment that is subjected to a greater horizontal σ_1 at the hinterland than at the foreland (Figure 8.28). You might expect that such an apparent violation of equilibrium would cause the block to translate toward the foreland, but friction inhibits such movement in our model, generating shear stress at the base. Because the top of the sheet is a free surface (the contact between rock and atmosphere), principal stresses must be parallel or perpendicular to the top surface of the sheet. However, σ_1 near the bottom of the sheet cannot parallel σ_1 at the top surface, because there is shear stress at the bottom of the sheet. Calculating stress trajectories given these conditions, we find that they curve into the bottom of the sheet. If we accept Anderson's premise that faults in the sheet initiate as Coulomb shears at about 30° to the σ_1 trajectory, then the faults must also curve; that is, their dips decrease with depth. This scenario provides a good explanation for the occurrence of listric faulting in the brittle regime, but note that the stress field is more complicated when the detachment is a ductile shear zone.

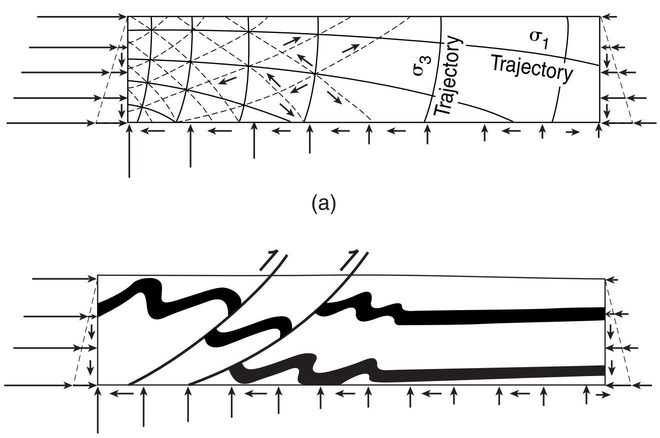


FIGURE 8.28 Curved principal stress trajectories and listric faults in a sheet of rock that is pushed from the (left) side. (a) Cross section of stress trajectories in a block bounded on the top by a free surface and on the bottom by a frictional sliding surface. (b) Predicted pattern of reverse faulting, assuming that faults form at ~30° to the σ_1 trajectory (note that only one set of reverse faults is illustrated).

8.5.2 Fluids and Faulting

There is no doubt that fluids can play a major role in fault zones, as you will commonly find that fault rocks are altered by reaction with a fluid phase (e.g., clay forms in fault zones from the reaction of feldspar with water), and that fault zones contain abundant veins composed of minerals that precipitated from a fluid (e.g., quartz, calcite, chlorite, economic minerals). The fracturing that accompanies fault displacement creates open space within the fault zone for fluid to enter. Because of the increase in open space, fluid pressure in the fault zone temporarily drops relative to the surrounding rock. The resulting fluid-pressure gradient can actually drive groundwater into the fault zone until a new equilibrium is established. Such faulting-triggered fluid motion is known as **fault valving** or **seismic pumping.**

The presence of water in fault zones affects the stress at which faulting occurs in three ways. First, alteration minerals formed by reaction with water in the fault zone tend to have lower shear strength than minerals in the unaltered rock, and thus their presence may permit the fault to slip at a lower frictional stress than it would otherwise. Second, the presence of water in a rock may cause hydrolytic weakening of silicate minerals, and therefore allow deformation to occur at lower stresses. Third, the pore pressure of water (P_{fluid}) in the fault zone decreases the effective normal stress in a rock body, and thus decreases the magnitude of the shear stress necessary to initiate a shear rupture in intact rock or initiate frictional sliding on a preexisting surface. See Chapter 6 for more details on the mechanics of these effects.

The observation that increasing fluid pressure leads to faulting at a lower regional σ_d is illustrated by the history of earthquake activity in the Rocky Mountain Arsenal near Denver, Colorado. In the early 1960s, the U.S. military chose to dispose of large quantities (sometimes as much as 30 million liters per month) of liquid toxic waste by pumping it into the groundwater reservoir via a 4-km-deep well at the arsenal. Geophysicists noticed that when the waste was injected, dozens of small earthquakes occurred near the bottom of the well. Evidently, injection of the waste increased P_{fluid}, thereby decreasing the effective normal stress on faults such that the local stress was sufficient to cause preexisting faults near the well to slip until the elevated fluid pressure dissipated by fluid flow out of the well.

The concept that an increase in fluid pressure decreases effective stress across a fault helps to resolve one of the great paradoxes of structural geology, namely the movement of thrust sheets on regional-scale detachments. Look at the schematic image of a thrust sheet depicted in Figure 8.29a. The sheet is represented by a large rectangular block slipping on a detachment at its base. If you assume that the shear resistance at the base of the sheet is comparable to the frictional sliding strength of rock observed in the laboratory, then in order to move the sheet, the magnitude of the horizontal stress applied at the end of the sheet must be very large. Herein lies the paradox: the horizontal stress must be so large that it would exceed the strength of the sheet, so that the thrust sheet would deform internally (by faulting and folding) close to where the stress was applied before the whole sheet would move (Figure 8.29b). As an analogy of this paradox, picture a large Persian rug lying on a floor. If you push at one end of the rug, it simply wrinkles at that end, but it does not slip across the floor, because the shear resistance to sliding is too great. So, how do large thrust sheets form and move great distances on detachment faults?

Fluid pressure offered the first reasonable solution to this paradox. If fluid pressure (P_{fluid}) in the detachment zone approaches lithostatic pressure (i.e., the magnitude of fluid pressure approaches the weight of the overlying rock), then the effective normal stress across the fault plane approaches zero (see Chapter 6). Therefore, the shear stress necessary to induce sliding on the detachment would also become very small, so that thrust sheets can move before deforming internally (Figure 8.29c). This powerful idea, known as the **Hubbert-Rubey hypothesis,**[7] emphasizes the role of elevated fluid pressure during movement on detachments, and has wide applicability. Indeed, modern measurements confirm that in regions where detachment faults move, P_{fluid} near the detachment interval often exceeds hydrostatic conditions and may even approach lithostatic values. Note that fluid pressure cannot exceed lithostatic pressure for long timescales, as this would forcefully push the rocks upward.

Initially, geologists thought that all large thrust sheets slide down gently foreland-dipping slopes in response to gravity, effectively gliding on a cushion of fluid. But **gravity sliding** does not explain most examples of thrusting in orogenic belts, because thrust sheets typically dip away from the direction in which they move. Recall, for example, the thrust-ramp geometry that characterizes large thrust faults, leading to a regionally hinterland-dipping surface. We return to this problem in Chapter 18, where we discuss newer concepts of thrust-sheet movement and the role of the Hubbert-Rubey hypothesis in these models.

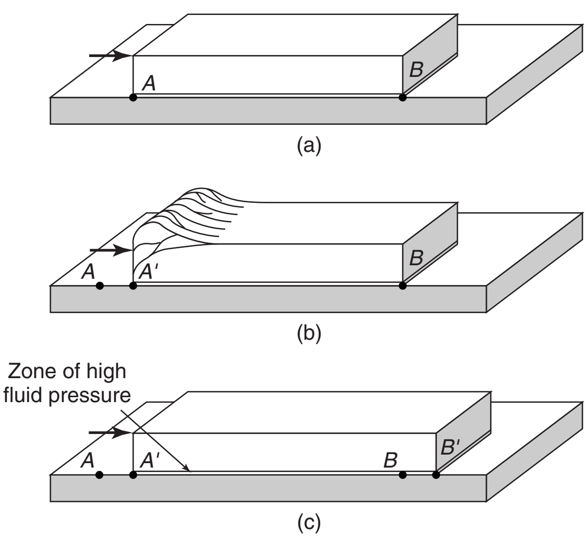


FIGURE 8.29 The thrust sheet paradox. (a) Block and sectional view of a thrust sheet on a frictional surface (heavy line). (b) Because of frictional resistance, the shear stress necessary to initiate sliding exceeds the yield strength of the frontal end of the thrust sheet, causing fracturing and folding. (c) Thrust sheet moves coherently if it rides on a "cushion" of fluid. *A′* and *B′* are displaced positions of points *A* and *B*.

8.5.3 Stress and Faulting—A Continuing Debate

The issue of how large the shear stress (σ_s) must be in order to initiate faults or to reactivate preexisting faults remains highly controversial. The magnitude of σ_s necessary to trigger faulting depends on fluid pressure, lithology, strain rate, temperature, and the orientation of the preexisting fault. Recall that for a given range of orientations, the stress necessary to initiate frictional sliding on a preexisting fault is less than that necessary to initiate a new fault (Chapter 6). Rock mechanics experiments provide one avenue of approach into this problem. From laboratory triaxial loading experiments, we determine that, all other factors being equal, the shear stress for failure increases as confining pressure increases, and that σ_s is greatest for contractional faulting and least for extensional faulting. Limiting conditions for each of the three fault types are shown in Figure 8.30, based on the relationship:

$$\sigma_d \geq \beta\,(\rho \cdot g \cdot z)\,(1 - \lambda) \qquad \text{Eq. 8.1}$$

[7]Named after the American geologists M. King Hubbert and William Rubey.

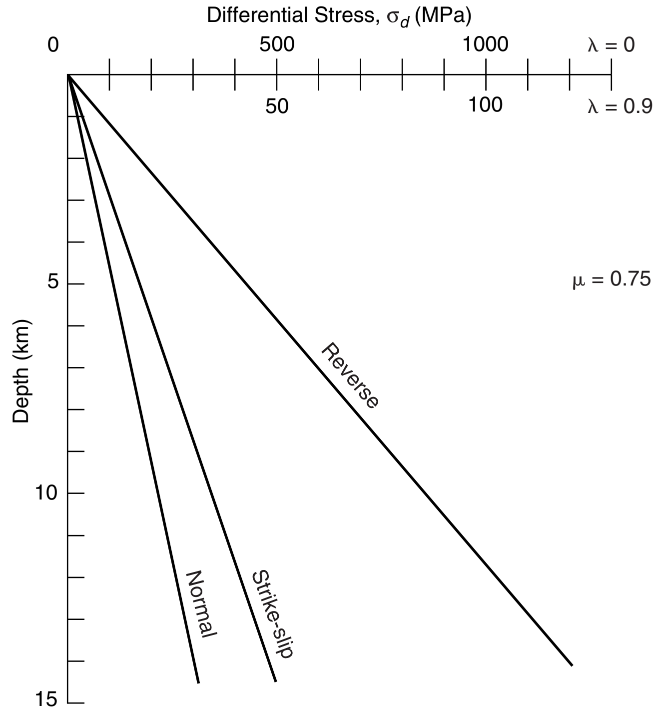


FIGURE 8.30 Graph showing variation in differential stress necessary to initiate sliding on reverse, strike-slip, and normal faults, as a function of depth. The relationship is given by Equation 8.1, assuming a friction coefficient, $\mu = 0.75$, and a fluid pressure parameter, $\lambda = 0$ (no fluid present) and $\lambda = 0.9$ (fluid pressure is 90% of lithostatic pressure).

where σ_d is differential stress (i.e., $2\sigma_s$); β is 3, 1.2, and 0.75 for reverse, strike-slip, and normal faulting, respectively; and λ is a fluid pressure parameter, defined as the ratio of pore-fluid pressure and lithostatic pressure (λ ranges from ~0.4, for hydrostatic fluid pressure, to 1 for lithostatic fluid pressure).

Geologists have questioned the validity of stress estimates based on laboratory studies, because of uncertainty about how such estimates scale up to crustal dimensions and about time-dependent changes. Thus, alternative approaches have been used to determine stress magnitudes needed to cause crustal-scale faulting, including analysis of heat generation during faulting, direct measurement of stresses near faults, and borehole data.

If movement on a fault surface in the brittle regime of the crust involves frictional sliding, some of the work done during fault movement is transformed into heat. This process, called **shear heating,** obeys the equation

$$\sigma_s \cdot u = E_e + E_s + Q \qquad \text{Eq. 8.2}$$

where σ_s is the shear stress across the fault, u is the amount of slip on the fault, E_e is the energy radiated by earthquakes, E_s is the energy used to create new surfaces (breaking chemical bonds), and Q is the heat generated. This equation suggests that $\sigma_s \cdot u > Q$, and thus if u is known, the value of Q can provide a minimum estimate of σ_s. With this concept in mind, geologists have studied metamorphism near faults in order to calculate the amount of heat (Q) needed to cause the metamorphism and thereby provide an estimate of σ_s during faulting. Assuming that metamorphism is entirely a consequence of shear heating, these studies conclude that shear stresses across faults must be quite large. These studies, however, have been criticized, because many geologists reject the assumption that the observed metamorphism is primarily due to shear heating, and because observed heat flow adjacent to active faults is not greater than heat flow at a distance from the fault. For example, while the San Andreas Fault is a huge active fault, it is not bordered by a zone of high heat flow as predicted by the shear-heating model.[8]

In recent years, geologists have attempted to define the stress state during crustal faulting by in situ measurements of stress fields in the vicinity of faults. Recall from Chapter 3 that hydrofracture measurements in drill holes, strain-release measurements, and borehole breakouts all provide an estimate of stress orientation, and in some cases stress magnitude, in the shallow crust. In general, these measurements suggest that stresses during faulting are relatively low. For example, direct measurement of stress in the crust around the San Andreas Fault suggests that the σ_1 trajectory bends so that it is nearly perpendicular in the immediate vicinity of the fault. Such a change in stress orientation suggests that the San Andreas Fault is behaving like a surface of very low friction, and therefore could not support a large σ_s. Based on these observations, some geologists have concluded that the σ_s needed to cause movement on the San Andreas Fault, and by inference other faults, is relatively low (the **weak-fault hypothesis**).

Another way to estimate stresses during faulting comes from studies of seismicity. When an earthquake occurs, energy is released and the value of σ_s across the fault decreases. This decrease, called the **stress drop,** provides a minimum estimate of the value of σ_s that triggered the earthquake. Stress drops for earthquakes estimated in this manner range between about 0.1 MPa and 150 MPa, but are typically in the range of 1–10 MPa.

In summary, geologists do not yet agree about the stress state necessary to cause crustal faulting, because estimates derived from different approaches do not

[8]This observation is key to the strength debate on the San Andreas Fault, where low heat flow supports the suggestion of low shear stresses, but also high fluid activity.

TABLE 8.7 GEOMETRIC CLASSIFICATION OF FAULT ARRAYS

Parallel fault array	As the name suggests, a parallel fault array includes a number of fault surfaces that roughly parallel one another.
Anastomosing array	A group of wavy faults that merge and diverge along strike, thereby creating a braided pattern in map view or cross section.
En echelon array	A group of parallel fault segments that lie between two enveloping surfaces and are inclined at an angle to the enveloping surfaces.
Relay array	In map view, a relay array is a group of parallel or subparallel non-coplanar faults that are spaced at a distance from one another across strike, but whose traces overlap with one another along strike. As displacement dies out along the strike of one fault in the array, displacement increases along an adjacent fault. Thus, displacement is effectively "relayed" (transferred) from fault to fault. In a thrust belt containing a relay array of faults, regional shortening can be constant along the strike of the belt, even though the magnitude of displacement along individual faults dies out along strike.
Conjugate fault array	An array composed of two sets of faults that are inclined to one another at an angle of about 60°. Conjugate fault arrays can consist of dip-slip faults or strike-slip faults. If the faults in the array are strike-slip, then one set must be dextral and the other sinistral.
Nonsystematic fault array	In some locations, faulting occurs on preexisting fractures. If the fracture array initially had a wide range of orientations, then slip on the fractures will yield faults in a wide range of orientations. Such an array is called a nonsystematic array.

agree with one another. Most likely, there is a range of stress conditions that can cause faulting, reflecting parameters such as the geometry of the fault, the nature of the faulted material, properties of the fault rocks, and the fluid pressure in the fault zone.

8.6 FAULT SYSTEMS

Faults typically do not occur in isolation, but rather are part of a group of associated faults that develop during the same interval of deformation and in response to the same regional stress field. We classify groups of related faults either by their geometric arrangement or their tectonic significance (i.e., the type of regional deformation resulting from their movement). A group of related faults is called a *fault system* or *fault array.* Although the terms "system" and "array" can be used interchangeably, geologists commonly use array when talking about geometric classifications and system in the context of tectonic classifications. As we continue, we'll first describe a geometric classification for fault arrays, and then we'll introduce the three tectonically defined types of fault systems: normal, thrust, and strike-slip systems. In many areas, faults in normal and thrust systems merge with a detachment in sedimentary rocks at shallow depth (10–15 km), in which case the system may be referred to as a **thin-skinned system.** If faulting involves deeper crustal rocks (i.e., basement), we call this a **thick-skinned** system. Most mountain belts show a transition between thin- and thick-skinned systems, so these are not adequate indicators of the tectonic setting.

Our description of fault systems in this chapter is meant only to be a brief introduction. We provide detailed descriptions of fault systems later in this book (Chapters 16–19), where you will also find ample illustration of associated structures and descriptions of tectonic settings.

8.6.1 Geometric Classification of Fault Arrays

Groups of faults may be classified as parallel, anastomosing, *en echelon,* relay, conjugate, or random arrays, depending on the relationships among faults in the array, and on how the faults link with one another along strike. We define these terms in Table 8.7 and illustrate them in Figure 8.31.

Typically, faults in parallel arrays of normal or thrust faults dip in the same direction over a broad region. Subsidiary faults in the array that parallel the major faults are called **synthetic faults,** whereas subsidiary faults whose dip is opposite to that of the major faults are called **antithetic faults.**

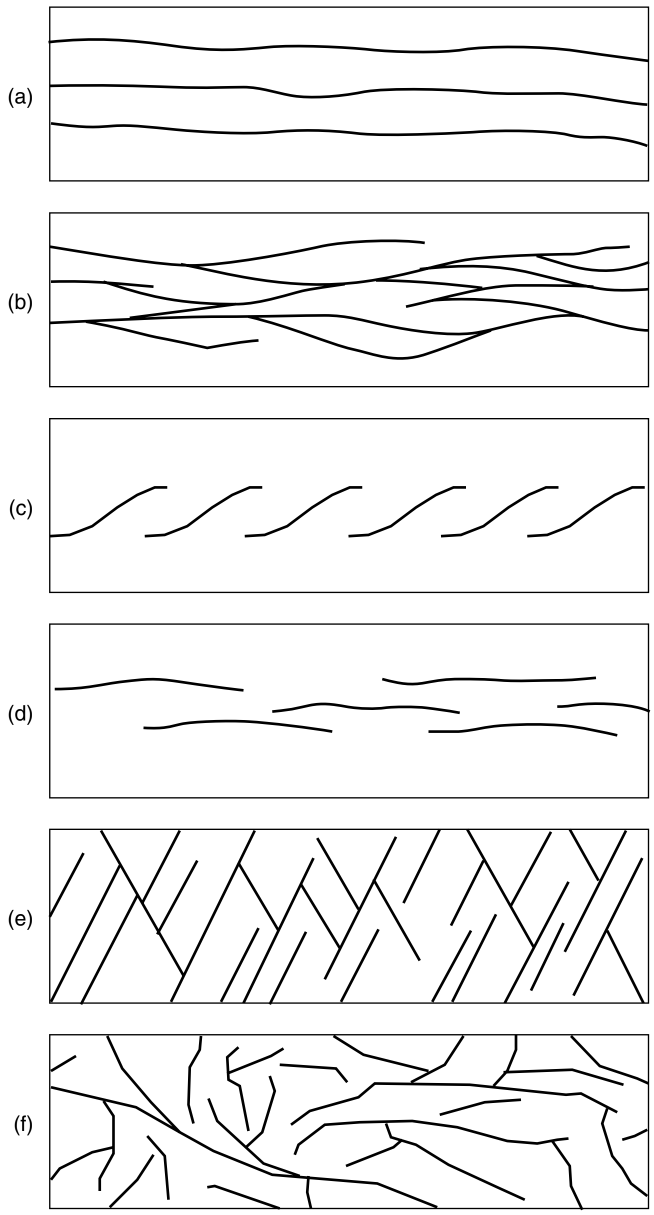


FIGURE 8.31 Map-view sketches of various types of fault arrays. (a) Parallel array, (b) anastomosing array, (c) *en echelon* or stepped array, (d) relay array, (e) conjugate array, and (f) random array. See descriptions in Table 8.7.

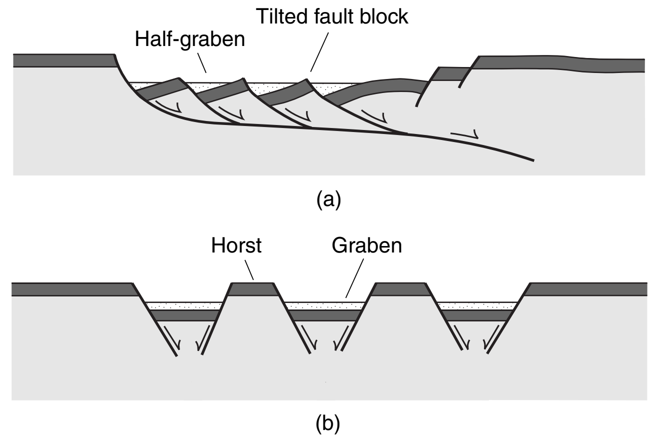


FIGURE 8.32 Normal fault systems. (a) Half-graben system. (b) Horst-and-graben system.

8.6.2 Normal Fault Systems

Regional normal fault systems form in **rifts,** which are belts in which the lithosphere is undergoing extension; along **passive margins,** which are continental margins that are not currently plate margins; and along **mid-ocean ridges.** Typically, faults in a normal fault system comprise relay or parallel arrays; they can be listric or planar, or contain distinct fault bends. Movement on both planar and listric normal fault systems generally results in rotation of hanging-wall blocks around a horizontal axis, and therefore causes tilting of overlying fault blocks and/or formation of rollover anticlines and synclines. The geometry of tilted blocks and rollover folds developed over normal faults depends on the shape of the fault and on whether or not synthetic or antithetic faults cut the hanging-wall block.

As a consequence of the rotation accompanying displacement on a normal fault, the original top surface of the hanging-wall block tilts toward the fault to create a depression called a **half graben** (Figure 8.32a). Note that a half graben (from the German word for "trough") is bounded by a fault on only one side. Most of the basins in the Basin and Range Province of the western United States are half grabens, and the ranges consist of the exposed tips of tilted fault blocks. In places where two adjacent normal faults dip toward one another, the fault-bounded block between them drops down, creating a **graben** (Figure 8.32b). Where two adjacent normal faults dip away from one another, the relatively high footwall block between the faults is called a **horst.** Horsts and grabens commonly form because of the interaction between synthetic and antithetic faults in rift systems. We further explore normal fault systems and their tectonic settings in Chapter 16.

8.6.3 Reverse Fault Systems

Reverse fault systems are commonly arrays of thrust faults that form to accommodate large regional shortening. Not surprisingly, thrust systems are common along the margins of convergent plate boundaries and in collisional orogens. In such tectonic settings, thrusting occurs in conjunction with formation of folds, resulting in tectonic provinces called **fold-thrust belts.**

To a first approximation, fold-thrust belts resemble the wedge of snow or sand that is scraped off by a plow (Figure 8.33a). Typically, the numerous thrusts in a fold-thrust belt merge at depth with a shallowly dipping detachment. At a crustal scale, major thrust faults are

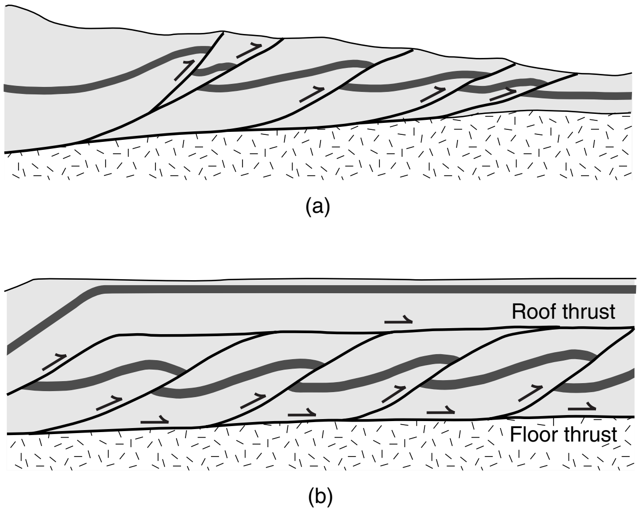


FIGURE 8.33 Reverse fault systems. (a) Imbricate fan in thrust system. (b) Duplex system with horses in between the floor and roof thrusts.

listric; but in detail, where thrusts cut upwards through sequences of contrasting strata, they have a stair-step profile, with **flats** following weak horizons and **ramps** cutting across beds. Ramp-flat geometries generally develop best in sequences of well-stratified sedimentary rock, such as at passive margins caught in between colliding continents, or in the sequence of sediment on the **craton** side of the orogen that is derived by erosion of the evolving orogen, called a **foreland basin.**

As is the case with normal fault systems, faults in a thrust-fault system tend to comprise relay or parallel arrays. An **imbricate fan** of thrust faults (Figure 8.33a) consists of thrusts that either intersect the ground surface or die out up dip, whereas a **duplex** (Figure 8.33b) consists of thrusts that span the interval of rock between a higher-level detachment called a **roof thrust** and a lower level detachment called a **floor thrust.** We can't say much more about thrust systems without introducing a lot of new terminology, so we'll delay further discussion of the subject until Chapter 18.

8.6.4 Strike-Slip Fault Systems

Strike-slip fault systems occur at transform boundaries, which are boundaries where two plates slide past one another without the creation or subduction of lithosphere; they can also occur within plates and as components of convergent orogens. Major continental strike-slip fault systems are complicated structures. We have already explored several associated structures of regional strike-slip faults. Typically, they splay into many separate faults in the near surface, which, in cross section, resembles the head of a flower. Because of this geometry, such arrays are called **flower structures** (Figure 8.34). We present further description of

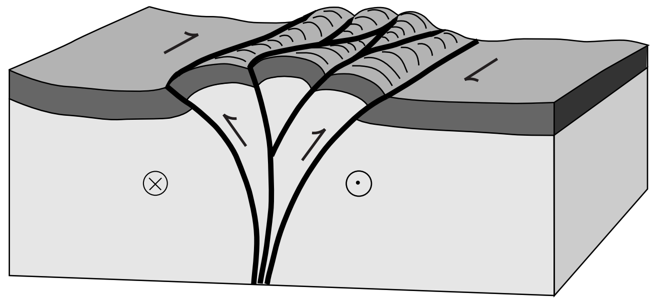


FIGURE 8.34 The formation of a (positive) flower structure from strike-slip faulting. The symbols $\otimes$ and $\odot$ indicate motion away and toward an observer, respectively.

these complexities and of the tectonic settings in which strike-slip faults occur in Chapter 19.

8.6.5 Inversion of Fault Systems

Once formed, a fault is a material discontinuity that may remain weaker than surrounding regions for long periods of geologic time. Thus, faults can be reactivated during successive pulses of deformation at different times during an area's history. If the stress field during successive pulses is different, the kinematics of movement on the fault may not be the same, and the resulting displacement from one event may be opposite to the displacement resulting from another event. For example, a normal fault formed during rifting of a continental margin may be reactivated as a thrust fault if that margin is later caught in the vice of continental collision. Likewise, the border faults of a half graben or failed rift may be reactivated later as thrust, or as strike-slip faults, if the region is later subject to compression. This reversal of displacement on a fault or fault system is called **fault inversion.** When inversion results in contraction of a previously formed basin, the process is called **basin inversion.**

8.6.6 Fault Systems and Paleostress

With the Andersonian concept of faulting in mind, the geometry of a fault system is a clue to the regional stress conditions that caused the faulting. A thrust system reflects conditions where regional σ_1 is horizontal and at a high angle to the trace of the system. A normal fault system reflects conditions where regional σ_3 is horizontal and trends at a high angle to the trend of the system. There is no single rule defining the relationship between strike-slip fault systems and stress trajectories. Oceanic transforms, for example, typically parallel the σ_3 direction, whereas continental strike-slip faults commonly trend oblique to the σ_1 direction.

Geologists study slip on variably oriented faults in a region to determine the stress field, assuming that all the faults moved in response to the same stress. Most

FIGURE 8.35 Damage from the Kobe earthquake in 1995 (Japan).

places in Earth's crust are fractured, and though there may be dominant systematic arrays of fractures in the area, there are likely to be many nonsystematic fractures as well. If a body of rock containing abundant preexisting fractures in a range of orientations is subjected to a regional homogeneous stress field, a shear stress will exist on all fractures that are not principal planes (i.e., not perpendicular to one of the principal stresses). The orientation of the resolved shear stress on each fracture is determined by the orientation and relative magnitudes of the principal stresses defining the regional stress state. Fractures whose frictional resistance is exceeded will move, and the direction of movement will be approximately parallel to the maximum resolved shear stress on the fracture surface. Fractures whose dip direction happens to be parallel to the shear stress become dip-slip faults, and fractures whose strike direction happens to be parallel to the shear stress become strike-slip faults. All other fractures that slip become oblique-slip faults.

During the past few decades, methods have been developed that permit the principal stress directions to be derived from measurements of slip trajectories on a nonsystematic array of faults, assuming that they moved in response to a regionally homogeneous stress. In principle, you need measurements of the shear sense and the trend of the net-slip vector on only four nonparallel faults to complete this **paleostress analysis,** but in practice, geologists use measurements from numerous faults and employ statistics to obtain a best-fit solution. It is beyond the scope of this book to provide the details of paleostress analysis from slip data on fault arrays, but we provide some references on the subject in Chapter 3.

8.7 FAULTING AND SOCIETY

All this terminology and theory may make us forget that the study of faulting is not just an academic avocation. Faults must be studied carefully by oil and mineral exploration geologists, because faulting controls the distribution of valuable materials. Similarly, faults and fractures play an important role in groundwater mobility and the general availability of water. Regional fault analysis is required for the localization and building of large human-made structures, such as dams and nuclear power plants. Such structures, which may have devastating impact when they fail, should clearly not be built near potentially active faults. But perhaps the most dramatic effect of fault activity on society comes in the form of earthquakes (Figure 8.35), which can be responsible for great loss of life (in some cases measured in the hundreds of thousands) and can destroy the economic stability of industrialized countries (damages may be hundreds of billions of dollars). We close this chapter by briefly looking at these more immediate consequences of faulting.

8.7.1 Faulting and Resources

Faults contribute to the development of **oil** traps by juxtaposing an impermeable seal composed of packed gouge or fault-parallel veining against a permeable reservoir rock, or by juxtaposing an impermeable unit, like shale, against a reservoir bed. Faulting may also affect oil migration by providing a highly fractured zone that serves as a fluid conduit through which oil migrates. Finally, syndepositional faulting (growth faults) affects the distribution of oil reservoir units.

Valuable **ore** minerals (e.g., gold) commonly occur in veins or are precipitated from hydrothermal fluids that were focused along fault zones, because fracturing provides enhanced permeability. Thus, fault breccias are commonly targets for mineral exploration. As we pointed out earlier, displacement on faults may control the distribution of ore-bearing horizons, and thus mining geologists map faults in a mining area in great detail.

Hydrogeologists also are cognizant of faults, because of their effect on the migration of **groundwater.** A fault zone may act as a permeable zone through which fluids migrate if it contains unfilled fractures, or the fault zone may act as a seal if it has been filled with vein material or includes impermeable gouge. In addition, faults may truncate aquifers, and/or juxtapose an aquitard against an aquifer, thereby blocking fluid migration paths.

8.7.2 Faulting and Earthquakes

Non-geologists often panic when they hear that a fault has been discovered near their home, because of the common perception that all faults eventually slip and cause earthquakes. Faulting is widespread in the crust, but fortunately, most faults are **inactive faults,** meaning that they haven't slipped in a long time, and are probably permanently stuck. Relatively few faults are **active faults,** meaning that they have slipped recently or have the potential to slip in the near future. Even when slip occurs, not all movement on active faults results in seismicity. If an increment of faulting causes an earthquake, we say that the fault is **seismic,** but if the offset occurs without generating an earthquake, we call the slip **aseismic.** Aseismic faulting is also called **fault creep** by seismologists.

Why do earthquakes occur during movement on faults? Earthquakes represent the sudden release of elastic strain energy that is stored in a rock, and can be generated when an intact rock ruptures, or when asperities on a preexisting fault snap off or suddenly plow. Rubbing two bricks while applying some pressure is a good analogy of this process. The bricks move until the indentation of asperities once again anchors them.

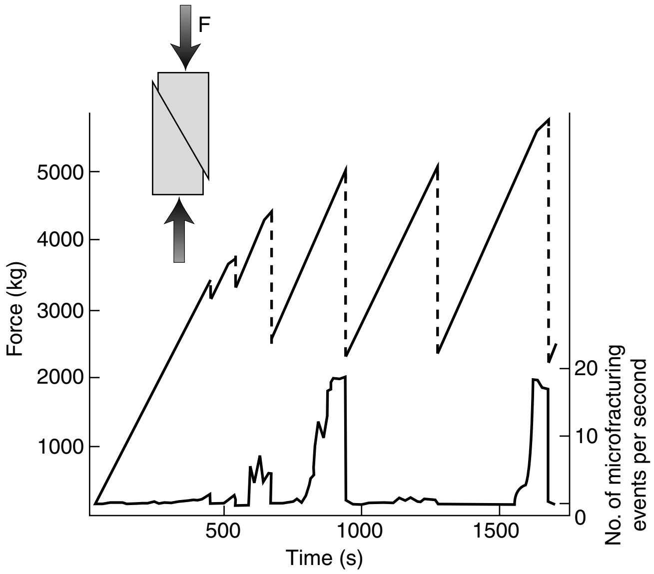


FIGURE 8.36 Laboratory frictional sliding experiment on granite, showing stick-slip behavior. The stress drops (dashed lines) correspond to slip events. Associated microfracturing activity is also indicated.

This start-stop behavior of faults is called **stick-slip behavior** (Figure 8.36). During the stick phase, stress builds up (as illustrated by the solid line), while during the slip phase the fault moves and stress at the site of faulting drops (dashed line). Typically, the stress drop is not complete, meaning that the differential stress does not decrease to zero. If we accept this model of faulting, **fault creep** occurs where a fault zone is very weak, perhaps due to the presence of weak material, hydrolytic weakening in the fault zone, or high fluid pressures in the fault zone.

Geoscientists have struggled for decades to delineate regions that have the potential to be seismic. This work involves the study of faults to determine if they are active or inactive, and whether active faults are seismic or aseismic. But fault studies alone do not provide a complete image of seismicity, because not all earthquakes occur on recognized faults. Some represent the development of new faults, some represent slip on blind faults, and some represent non-fault-related seismicity (e.g., volcanic explosions). Delineation of seismically active regions plays a major role in land-use planning. Obviously, the potential for **seismicity** must be taken into account when designing building codes for homes and when situating large public facilities like nuclear power plants, schools, hospitals, dams, and pipelines.

The primary criterion for delineating a seismically active region comes from direct measurements of seismicity. The underlying idea is that places with a potential for earthquakes in the near future probably have

FIGURE 8.37 Nearly horizontal layers of light-colored rock and gravel, and dark-colored peat have been broken and distorted as a result of earthquake activity near Banning, California, along the San Andrea Fault.

suffered earthquakes in the past. Networks of seismographs record earthquakes and provide the data needed to pinpoint the **focus** of each earthquake; that is, the region in the Earth where the seismic energy was released. Maps of earthquake **epicenters** (the point on the Earth's surface that lies directly above the focus) emphasize that most earthquakes occur along plate boundaries, but that dangerous earthquakes may also occur within plate interiors. Thus, we recognize **plate-boundary seismicity** and **intraplate seismicity.** Cross sections showing the distribution of earthquake foci show that, with the exception of convergent-margin seismicity, most earthquakes occur at depths shallower than ~15 km, which defines the lower boundary of the brittle upper crust.[9] Convergent-margin earthquakes occur along subducted slabs down to depths of about 650 km, defining the **Wadati-Benioff zone** (see Chapter 14). The deep earthquakes in a Wadati-Benioff zone occur well below the expected depth for brittle faulting. So why do these deep-focus earthquakes occur? One suggestion is that deep-focus earthquakes represent the stress release associated with sudden mineral phase changes in the downgoing slab (e.g., olivine to spinel). Since different mineral phases occupy different volumes, a sudden phase change causes a movement in the rock body that could result in the generation of an earthquake and perhaps even in the formation of pseudotachylyte. The question is not resolved.

A reliable and detailed record of seismicity is only a few decades old, because a worldwide network of seismograph stations was not installed until after World War II. The information from these stations proved critical for the formulation of plate tectonic theory in the 1960s (Chapter 14), but governments actually funded this network to monitor underground nuclear testing in the Cold War era of the 1950s. Seismic studies cannot delineate the potential for seismicity in areas that have only infrequent earthquake activity. To identify such cryptic seismic zones, geologists rely on field data (Figure 8.37). We search for features such as cross-cutting relations with very young stratigraphic units or landforms. If the fault cuts a very young sequence of sediment or a very young volcanic flow or ash, then the fault must itself be very young. Similarly, if the fault cuts a young landform, like an alluvial fan or a glacial moraine, then the fault must be very young. Fault

[9]This seismically active region of the crust is also called the schizosphere, whose lower boundary is sometimes (incorrectly) called the brittle-ductile transition; we use the term brittle-plastic transition for this boundary.

scarps and triangular facets suggest that the faulting occurred so recently that erosion has not had time to erase its surface manifestation. The presence of uncemented gouge suggests that a fault was active while the rock was fairly close to the surface, a situation that implies movement on the fault occurred subsequent to uplift and exhumation of rock. The presence of pseudotachylyte may indicate that the fault was seismic. Changes in base level due to faulting at the face of a mountain range can cause a stream to cut down through alluvium that it previously deposited, creating matched terraces on opposite sides of the valley. The development of such paired terraces may indicate the occurrence of seismicity. Finally, accurate surveying of the landscape may indicate otherwise undetectable ground movements that could be a precursor to seismicity.

In special cases, it may be possible to determine the **recurrence interval** on a fault, meaning the average time between successive faulting events. This is done by studying the detailed stratigraphy of sediments deposited in marshes or ponds along the fault trace. Seismic events are recorded by layers in which sediment has been liquefied by shaking (these layers are sometimes called *seismites*), which causes disruption of bedding and sand volcanoes. By collecting organic material from the liquefied interval (e.g., wood), it may be possible to date the timing of liquefaction. Once we know the recurrence interval and the size of the earthquakes, we can estimate the seismic risk for an area.

The past few decades have seen intense study of earthquakes and related processes. As a result we have become reasonably successful in predicting *where* earthquakes will occur (if not precisely, at least the general area), but *when* they occur remains an imprecise science at best. Error margins of 50–100 years are inadequate for modern society, but being able to improve significantly on this with our current understanding of faulting seems unlikely. Geology operates on timescales much larger than our human ones. Perhaps preparation is our best bet when it comes to earthquake hazards.

8.8 CLOSING REMARKS

Chapters 6, 7, and 8 have provided an overview of brittle deformation processes and structures. At the end of Chapter 8, we also discussed briefly society's need for better understanding of fault processes. Brittle deformation is only a small part of what contributes to the development of structures and deformation of the lithosphere. We have alluded, for example, to the existence of ductile shear zones, in which displacement is not accommodated by brittle failure, and we have mentioned folding, which also deforms rocks without loss of cohesion. In the next section of this book, we turn our attention to the complementary topic of ductile deformation and resulting structures. After we have introduced these structures, we will be able to describe the relationship between faults and other geologic structures in greater detail, and describe the tectonic settings in which they form.

ADDITIONAL READING

Anderson, E. M., 1951. *Dynamics of faulting and dyke formation.* Oliver and Boyd: Edinburgh.

Bonnet, E., Bour, O., Odling, N. E., Davy, P., Main, I., Cowie, P., and Berkowitz., B., 2001. Scaling of fracture systems in geological media. *Reviews of Geophysics,* 39, 347–383.

Boyer, S. E., and Elliot, D., 1982. Thrust systems. *American Association of Petroleum Geologists Bulletin,* 66, 1196–1230.

Chester, F. M., and Logan, J. M., 1987. Composite planar fabric of gouge from the Punchbowl Fault, California. *Journal of Structural Geology,* 9, 621–634.

Hubbert, M. K., and Rubey, W. W., 1954. Role of fluid pressure in mechanics of overthrust faulting. *Bulletin of the Geological Society of America,* 70, 115–205.

Keller, E., and Pinter, N., 1996. *Active tectonics: earthquakes, uplift, and landscape.* Prentice Hall: Englewood Cliffs, 338 pp.

Kirby, S. H., 1983. Rheology of the lithosphere. *Reviews of Geophysics and Space Physics,* 21, 1458–1487.

Mandl, G., 1988. *Mechanics of tectonic faulting, models and basic concepts.* Elsevier: Amsterdam.

Petit, J. P., 1987. Criteria for the sense of movement on fault surfaces in brittle rocks. *Journal of Structural Geology,* 9, 597–608.

Scholz, C. H., 2002. *The mechanics of earthquakes and faulting* (2nd edition). Cambridge University Press: Cambridge.

Sibson, R. H., 1974. Frictional constraints on thrust, wrench and normal faults. *Nature,* 249, 542–544.

Sibson, R. H., 1977. Fault rocks and fault mechanisms. *Journal of the Geological Society of London,* 133, 190–213.

Sylvester, A. G., 1988. Strike-slip faults. *Geological Society of America Bulletin,* 100, 1666–1703.

Wise, D. U., Dunn, D. E., Engelder, J. T., Geiser, P. A., Hatcher, R. D., Kish, S. A., Odom, A. L., and Schamel., S., 1984. Fault-related rocks: suggestions for terminology. *Geology,* 12, 391–394.