CHAPTER

4 Introduction to Faults

A fault is a surface or narrow zone along which one side has moved relative to the other in a direction parallel to the surface or zone. Most faults are brittle shear fractures (Figure 4.1*A*) or zones of closely spaced shear fractures (Figure 4.1*B*), but some are narrow shear zones of ductile deformation where movement took place without loss of cohesion at the outcrop scale (Figure 4.1*C*). We generally use the term *fault* for shear fractures or zones that extend over distances of meters or larger. Features at the scale of centimeters or less are called **shear fractures**, and shear fractures at the scale of a millimeter or less are **microfaults** that may be visible only under a microscope. Faults are often structural features of first-order importance on the Earth's surface and in its interior. They affect blocks of the Earth's crust thousands or millions of square kilometers in area, and they include major plate boundaries hundreds or even thousands of kilometers long.

The word *fault* is derived from an eighteenth and nineteenth-century mining term for a surface across which coal layers were offset. Many such mining terms were transferred to geology in its early days, despite the fact that the mining lexicon was often complex and ambiguous. This century has seen a number of attempts to rationalize and systematize this terminology, although there is still no agreement on precise definitions for some words. We try to employ only those terms that are the most prevalent and the most useful in describing faults.

4.1 Types of Faults

A fault divides the rocks it cuts into two **fault blocks**. For an inclined fault, geologists have adopted the miners' terms **hanging wall** for the bottom surface of the upper fault block and **footwall** for the top surface of the lower fault block (Figures 4.1, 4.2). In a tunnel, these are the surfaces that literally *hang* overhead or lie under *foot*. The fault block above the fault is the **hanging wall block**, and the block below the fault is the **footwall block**. For a vertical fault, of course, these distinctions do not apply, and the sides of the fault are named in accordance with geographic directions: the northwest side and the southeast side, for instance.

Faults are classified in terms of the attitude of the fault surface. If a fault dip is more than 45°, it is a **high-angle fault**; if less than 45°, it is a **low-angle fault**.

We also divide faults into three categories depending on the orientation of the relative **displacement**, or **slip**, which is the net distance and direction that the hanging wall block has moved with respect to the footwall block (Figure 4.2). On **dip-slip faults**, the slip is approximately parallel to the dip of the fault surface; on **strike-slip faults**, the slip is approximately horizontal, parallel to the strike of the fault surface; and on **oblique-slip faults**, the slip is inclined obliquely on the fault surface. An oblique-slip vector can always be described as the sum of a strike-slip component and a dip-slip

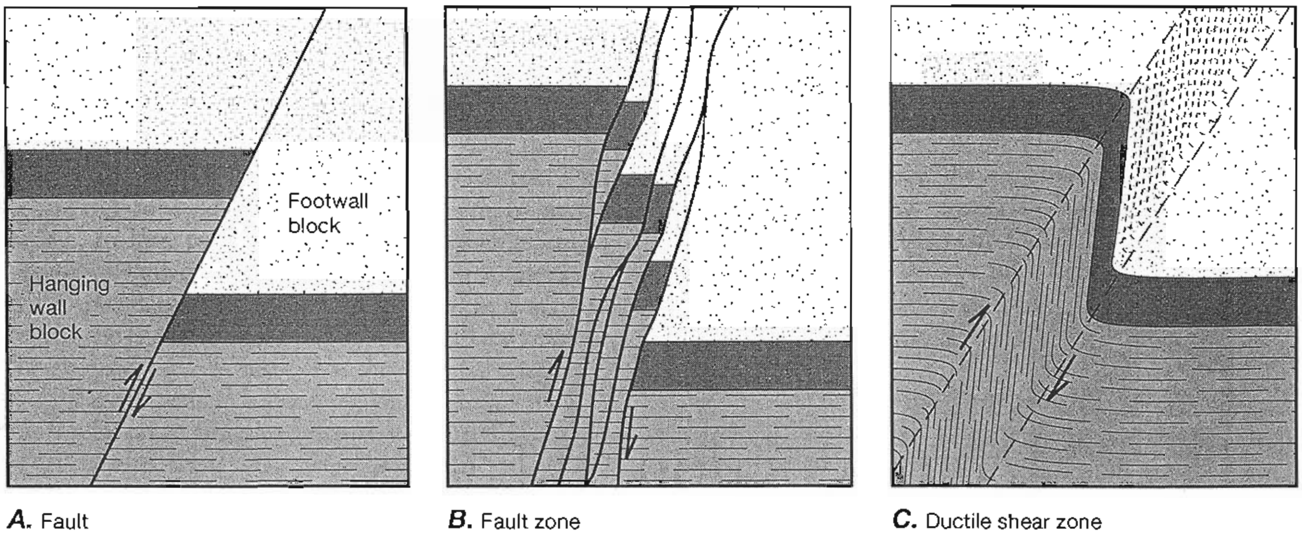


Figure 4.1 Three styles of faulting. *A.* A single fault consists of a single shear fracture. A fault zone comprises *B.* a set of associated shear fractures or *C.* a zone of ductile shear.

component, or as the sum of a horizontal component and a vertical component. The dip-slip component may in turn be described as the sum of a vertical component and a horizontal component, which are sometimes called the **throw** and the **heave**, respectively.

We subdivide faults further in terms of the relative movement along them. Inclined dip-slip faults on which the hanging wall block moves down relative to the footwall block are **normal faults** (Figure 4.3*A*). Those on which the hanging wall block moves up relative to the footwall block are **thrust faults** (Figure 4.3*B*).[1] Vertical faults characterized by dip-slip motion cannot, of course, be classified as either normal or reverse faults, so we simply specify which side of the fault has moved up or down. Strike-slip faults are **right-lateral**, or **dextral**, if the fault block across the fault from the observer moved to the right (Figure 4.3*C*); they are **left-lateral**, or **sinistral**, if that block moved to the left (Figure 4.3*D*). Oblique-slip faults may be described according to the nature of the strike-slip and dip-slip components. Figure 4.3*E*, for example, shows sinistral normal slip, and Figure 4.3*F* shows sinistral reverse slip. For **rotational faults** the slip changes rapidly with horizontal distance along the fault (Figure 4.3*G*).

[1] Thrust faults that dip more steeply than 45° are sometimes called reverse faults.

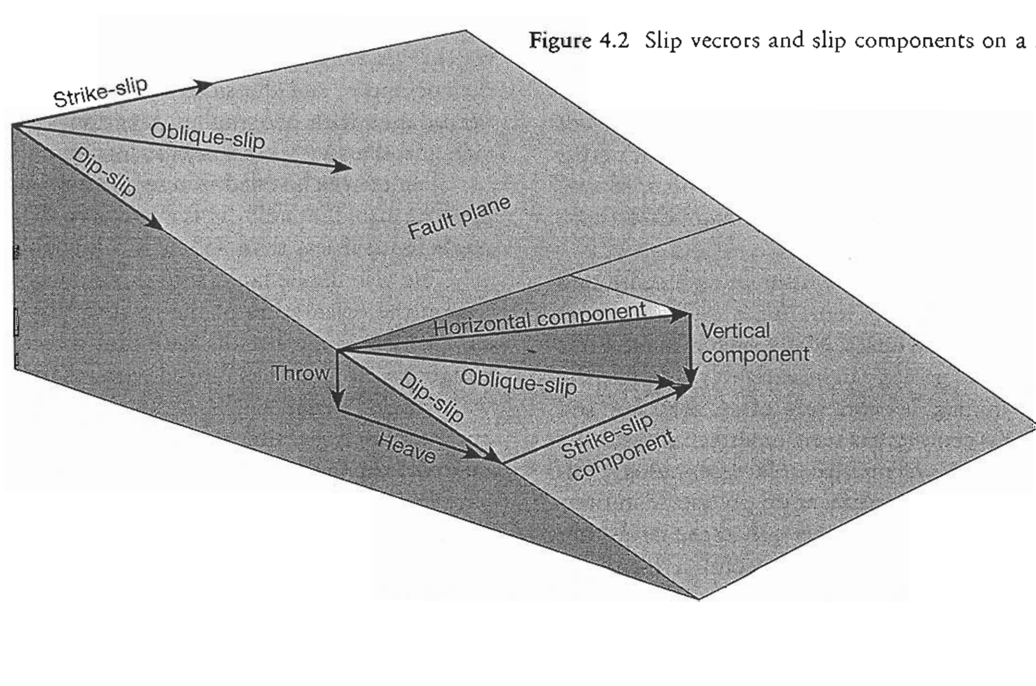


Figure 4.2 Slip vectors and slip components on a fault surface.

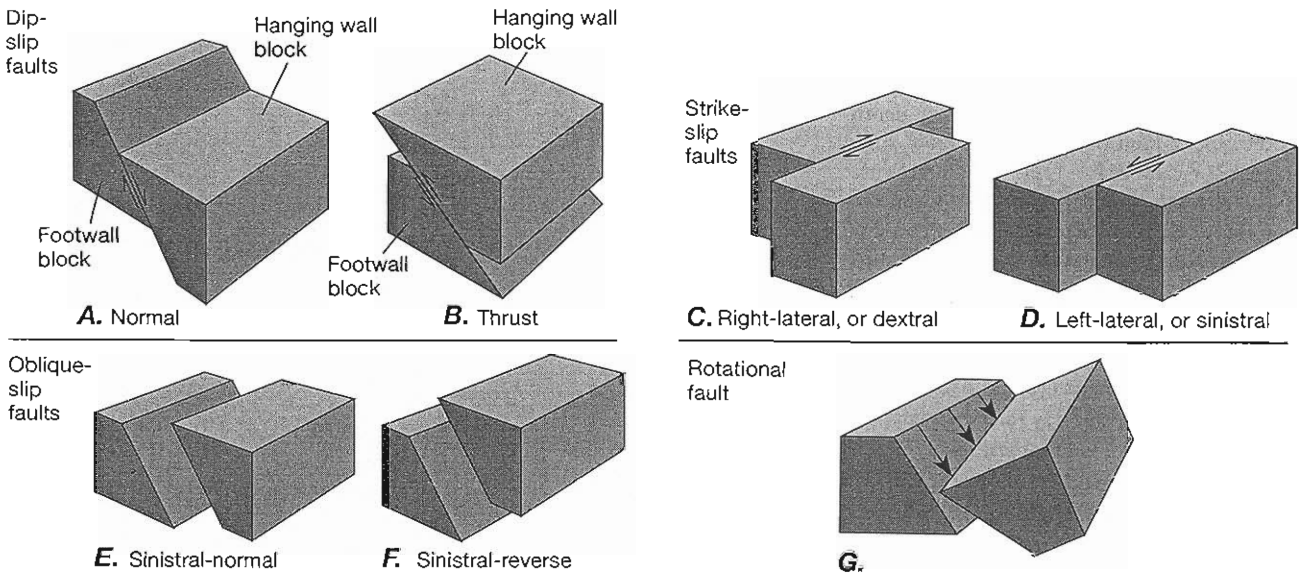


Figure 4.3 Faulted blocks showing the characteristic displacement for the different classes of faults.

4.2 Recognition of Faults

The criteria for recognizing faults can be divided into three broad categories: features intrinsic to faults themselves, effects on geologic or stratigraphic units, and effects on physiographic features. We briefly consider each of these categories.

Features Intrinsic to Faults

Faults can often be recognized by the characteristic textures and structures developed in rocks as a result of shearing (Table 4.1). These textures and structures vary with the amount and rate of shear and with the physical conditions at which the faulting occurred, including

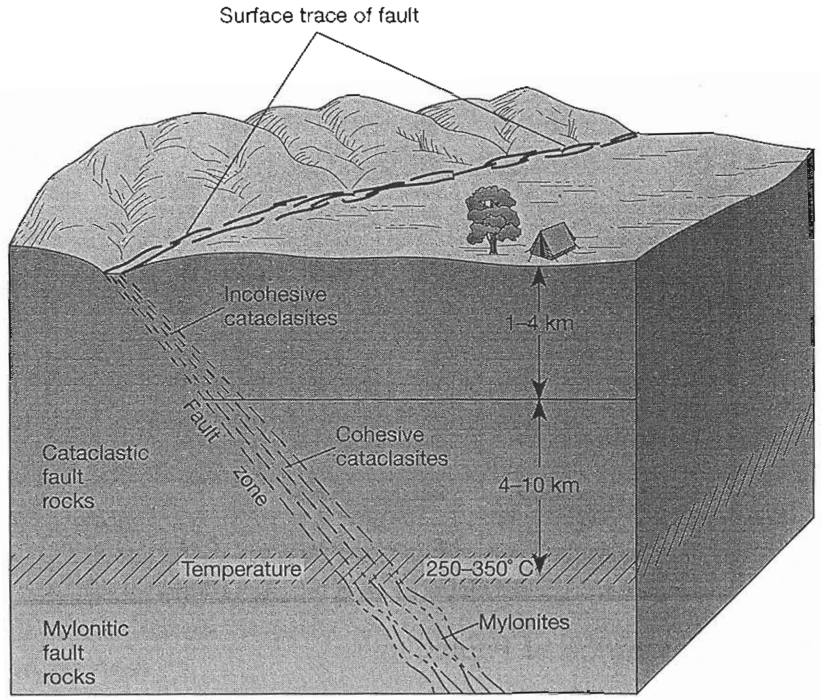
Table 4.1 Fault Rock Terminology[a]

Cataclastic rocks					
Fabric	Texture	Name		Clasts	Matrix
Generally no preferred orientations	Cataclastic: sharp, angular fragments	Breccia series	Megabreccia	> 0.5 m	<30%
			Breccia	1–500 mm	<30%
			Microbreccia	<1 mm	<30%
		Gouge		<0.1 mm	<30%
		Cataclasite		Generally ≤ ~10 mm	>30%
		Pseudotachylite			Glass, or grain size ≤1 μm

Mylonitic rocks					
Fabric	Texture	Name		Matrix grain size	Matrix
Foliated and lineated	Metamorphic: Interlocking grain boundaries, sutured to polygonal	Mylonitic gneiss		>50 μm	
		Mylonite series	Protomylonite	<50 μm	<50%
			Mylonite	<50 μm	50%–90%
			Ultramylonite	<10 μm	>90%

[a] The terminology applied to fault rocks is by no means generally agreed upon. The definitions of the different categories, and the quantitative boundaries we have placed on them, should therefore be understood as guidelines to present usage, which can vary from one geologist to another. We believe, for example, that what we have defined as mylonite would fit anyone's definition, but other geologists use *mylonite* in a broader sense, even to include what we call mylonitic gneiss.

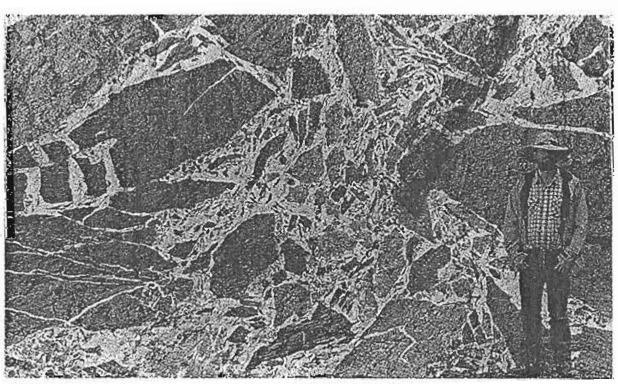
Figure 4.4 Schematic block diagram of a portion of the earth's crust, showing the surface trace of a fault zone (i.e., its exposure on the Earth's surface) and the variation with depth of the type of fault rock within the fault zone. Incoherent cataclasites (plus pseudotachylite if dry) characterize depths above 1–4 km. Below that, coherent cataclasites (plus pseudotachylite if dry) are present at depths of up to 15 km. Mylonites are present at depths greater than 10–15 km and temperatures greater than 250–350° C.



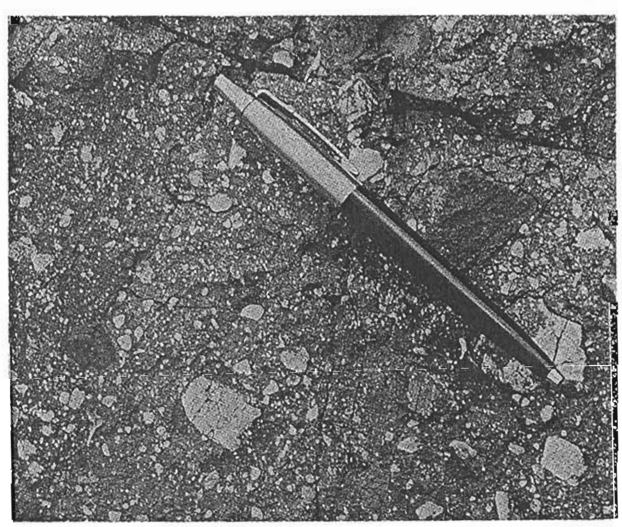
temperature and pressure which typically are a function of the depth of faulting (Figure 4.4).

Faults formed at depths less than about 10 to 15 km typically have **cataclastic rocks** present in the fault zone. This term refers in general to rocks that have been fractured into clasts or ground into powder during brittle deformation. Individual fragments are generally sharp, angular, and internally fractured. Cataclastic rocks usually lack any internal planar or linear structure. Friable cataclastic rocks are typical of faulting above depths of 1 to 4 km. Cohesive cataclasites may form at depths up to 10 to 15 km.

The terminology and classification are not universally agreed upon (see Table 4.1). We divide cataclastic rocks into four main categories. The breccia series, gouge, and cataclasite are distinguished on the basis of clast size and the percentage of matrix; whereas pseudotachylite forms a separate category based on the character of the matrix. The percentage of fine-grained matrix distinguishes rocks in the breccia series (less than about 30 percent; Figure 4.5*A*) from cataclasites (more than about 30 percent; Figure 4.5*B*). We subdivide the breccia series into megabreccia (Figure 4.5*A*), breccia, and microbreccia. In megabreccia and breccia, the clasts



A.



B.

Figure 4.5 Cataclastic rocks. *A*. Megabreccia composed of very large fragments of limestone (Titus Canyon, Death Valley National Monument). *B*. Cataclasite from the Whipple Mountain detachment, southeast California.

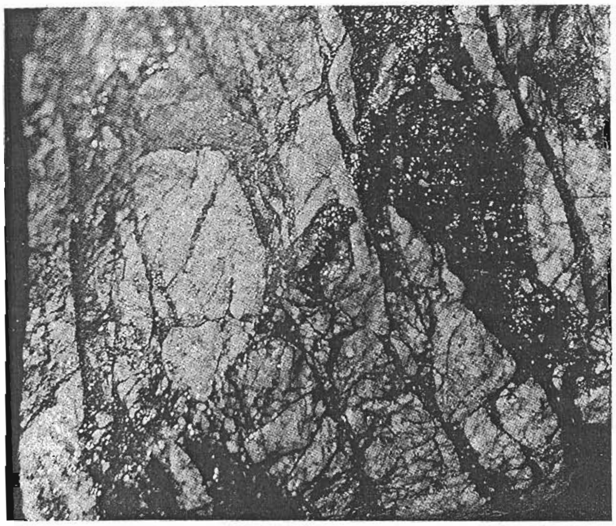


Figure 4.6 Veins of dark pseudotachylite cutting a light-colored gneiss. Actual width of view shown is about 6 cm.

are predominantly rock fragments. In microbreccia the clasts are principally mineral grain fragments. Gouge is essentially a continuation of the breccia series to finer clast size. In outcrop, it appears as a finely ground, whitish rock powder. Remarkably, all cataclastic rocks look very much the same over a wide range of scales (compare Figure 4.5*A*, *B*). Although breccia, microbreccia, and gouge are generally noncohesive, deposition of silica during or subsequent to formation can turn them into hard, cohesive, silicified fault rock.

Cataclasites include a range of clast sizes and vary from 30 percent fine-grained matrix up to 100 percent matrix (Figure 4.5*B*). They are generally cohesive rocks.

Pseudotachylite (Figure 4.6) is a massive rock that frequently appears in microbreccias or surrounding rocks as dark veins of glassy or cryptocrystalline material. It characteristically contains a matrix of crystals less than 1 μm in diameter and/or small amounts of glass or devitrified glass cementing a mass of fractured material together. Under a petrographic microscope, the matrix appears isotropic; that is, between crossed polarizers no light is transmitted. This behavior is characteristic of glass and of extremely fine-grained material. During an earthquake under dry conditions at depths generally less than 10 to 15 km, frictional heating can be sufficient to melt small portions of the rock. The resulting material may intrude through fractures in the adjacent rock before quenching to form veins of pseudotachylite.

Cataclastic rocks occur in zones ranging from a few millimeters in thickness up to extensive zones one or more kilometers thick. In general, the greater the thickness and the smaller the grain size, the greater the amount of displacement that has accumulated on the fault.

Fault zones formed at depths exceeding about 10 to 15 km are characterized by another type of very fine-grained rock called **mylonitic rocks** (Figure 4.7). These

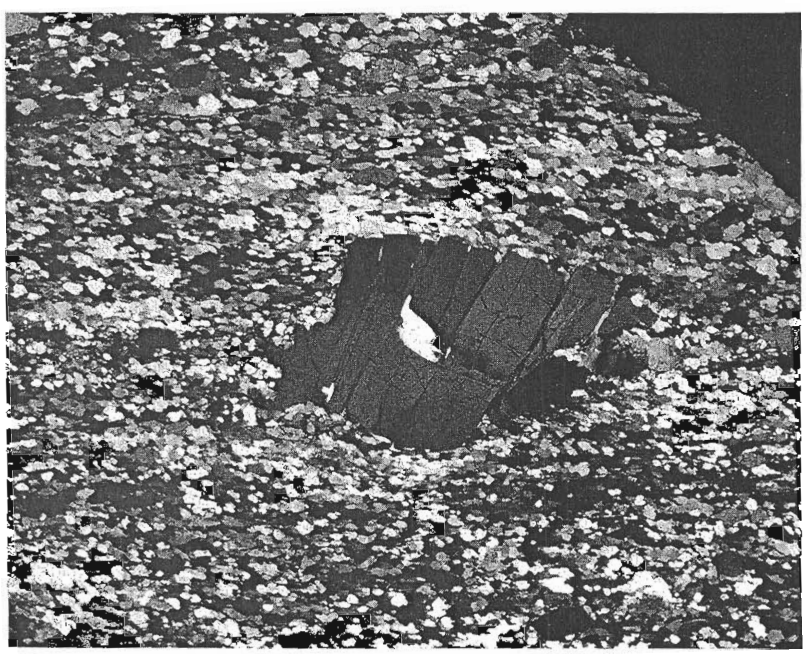


Figure 4.7 Quartz mylonite showing large feldspar porphyroclast in a much finer-grained matrix of strongly recrystallized quartz grains.

rocks form only as a resulr of ductile deformation, which occurs in crustal rocks at temperatures generally in excess of 250°C to 350°C.[2] Mylonitic rocks have a matrix of very fine grains that are derived by reducrion of grain size from the original rock. Variable amounts of relict coarse mineral grains, called porphyroclasts, may be present, surrounded by the fine-grained matrix. The fine grains show an interlocking grain boundary texture characteristic of metamorphic rocks; the grain boundaries thcmselves may be polygonal, forming 120° triple junctions, or they may be highly sutured. Mylonitic rocks exhibit a strong planar and linear intcrnal structure, called foliation and lineation, rcspectively (sec Chapter 13). These structures are subparallel to the fault zone.

Mylonitic rocks form as a result of the recrystallization of mineral grains during rapid ductile deformation. Their polygonal to sutured grain boundaries differ from fine-grained cataclasites, in which the grains have the sharp, angular shape characteristic of brittle fracturing.

If the grain size is reduced from the original grain size but is coarser than about 50 μm, the rock is a mylonitic gneiss. If the matrix grain size is less than 50 μm, the rock belongs to the mylonite series, which we subdividc on the basis of increasing percentage of fine-grained matrix (Table 4.1) into protomylonite, mylonite, and ultramylonite. In ultramylonites, the characteristic grain size of less than 10 μm causes the matrix to appcar glassy in a hand sample.

Mylonites are generally found in ductile shear zones ranging in thickness from a fraction of a meter to several meters. Some mylonites, however, are much thicker bodies that apparently define wide shear zones. All transitional stages, from original country rock through mylonitic gneisses to ultramylonites, may be present in such a zone.

Where exposed, fault planes commonly are smooth, polished surfaces called slickensides,[3] which form in response to shearing on the fault planes or in the fault gouge. Fault surfaces, including slickensides, typically contain strongly oriented linear features, known as slickenlines, slickenside lineations, or striations, that are parallel to the direction of slip. These lineations are of three types: ridges and grooves, mineral streaks, and mineral fibers or slickenfibers. Ridges and grooves may tesult from scratching and gouging of the fault surface (Figure 4.8*A*), from the accumulation of gouge behind hard protrusions or asperities, from the development of irregularities in the fracture surface itself forming ridge-in-groove lineations or fault mullions (Figure 4.8*B*), or from growth of slickenfibers (Figure 4.8*C*). Slickenfibers are long, single-crystal mineral fibers that grow parallel to the direction of fault displacement. They fill gaps that develop along the fault during gradual shearing (see Figure 4.15 and Sections 4.3, 13.8, and 14.6). Mineral streaks are streaks on slickensides that develop as a result of the pulverization and shearing out of mineral grains within the gouge.

Faults that develop at relatively shallow depths are dilatant—that is, they dcvelop open spaces that increase the rock's volume. Such fault zones providc pathways for the flow of ground water and hydrothermal fluids. As a result, many fault zones contain secondary deposits of minerals, including calcite (see Figure 4.5*A*) and silica (quartz, opal, or chalcedony) as vein deposits or as cement for the preexisting fault gouge or breccia. Many ore deposits form by precipitation of ore minerals from hydrothermal fluids flowing along fault zones.

Effects of Faulting on Geologic or Stratigraphic Units

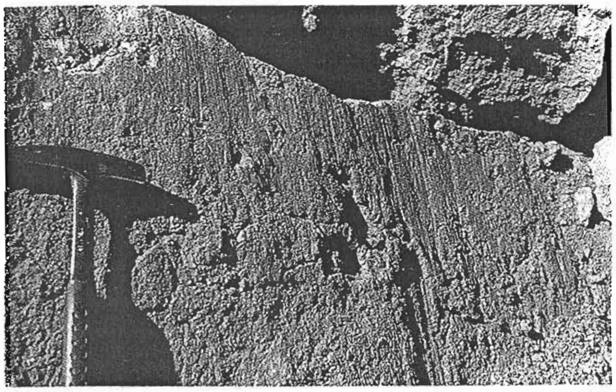
Displaccment along faults generally places adjacent to one another rocks that do not belong together in ordinary geologic sequences. The resulting discontinuity provides some of the best evidence for the presence of the fault.

A bteak in an otherwise continuous geologic fcature, such as sedimentary bedding, may indicate the presence of a fault. A stratigraphic discontinuity, however, may also result from an unconformity or an intrusive contact, and it is important to distinguish such features from faults. Characteristic features of unconformities include fossil soil horizons, erosional channels, basal conglomerates, depositional contacts, and the parallelism or near parallelism of the strata that overlie the unconformity (see Figure 2.11 and the discussion in Section 2.4). Distinctive features of intrusive contacts (Figure 4.9) include metamorphism in the adjacent country rocks, fragments of country rock suspended in the intrusion (xenoliths), and dikes or veins of igneous rock cutting the country rock adjacent to the intrusion.

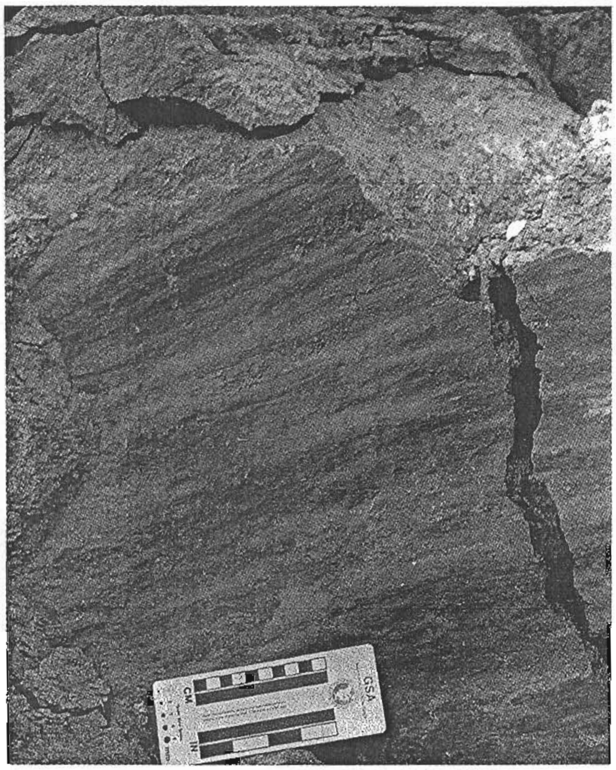
The presence of horses, or fault slices, along a discontinuity is clear evidence of a fault. Horses are volumes of rock that are bounded on all sides by faults (see Figure 4.24). They are sliced from either the footwall or the hanging wall block by a branch of the fault and are displaced a significant distance from their original position. Thus they may appear markedly out of place stratigraphically. If the local stratigraphy is known, identification of the original stratigraphic position of

[2] We discuss in detail the structures and processes of ductile deformation in Parts III and IV of this book. We include here a brief description of some of these features because they characterize many fault zones.

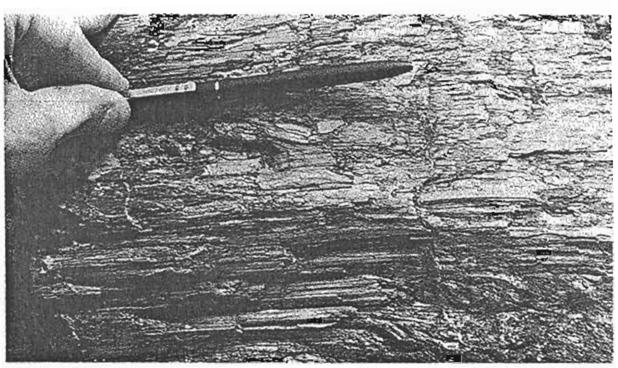
[3] Confusion exists about the exact meaning of this term. Many authors use *slickenside* to refer to the lineations that occur on the fault surfaces. This usage is not consistent with the original definirion, however.



A.



B.



C.

Figure 4.8 (*Left*) Lineations on fault surfaces formed during fault slip. *A*. Lineations formed by scratching and gouging of the fault surface. *B*. Ridge-in-groove lineations, or fault mullions. C. Serpentine slickenfiber lineations.

the rocks in a horse provides a constraint on the direction and amount of movement. In areas where horses separate two similar rock types, a horse of a different lithology may be the only observable evidence of a fault.

Repetition of strata or omission of strata in a known stratigraphic sequence is another possible indication of a fault. This criterion is especially important in subsurface geology, where often only drill-hole data are available. Figure 4.10 shows a diagrammatic cross section of a region of horizontal bedding with drill-hole information indicating the repetition (Figure 4.10*A*) and omission (Figure 4.10*B*) of strata. Such data are referred to as showing repeated section and missing section, respectively. If enough information is present, it is possible to map a fault in the subsurface solely on the basis of information obtained by drilling.

As in the truncation of structures, it is important to make sure that the omission of strata does not result from an unconformity and that the repetition of strata does not result from a facies change associated with alternating transgressions and regressions. The distinction between faults and facies changes can be subtle, and failure to distinguish them correctly has resulted in some spectacular geologic errors.

Bedding surfaces near faults may curve in the direction of motion of the opposite fault block. These folds are called drag folds. They are most likely to develop where the traces of the sedimentary layers on

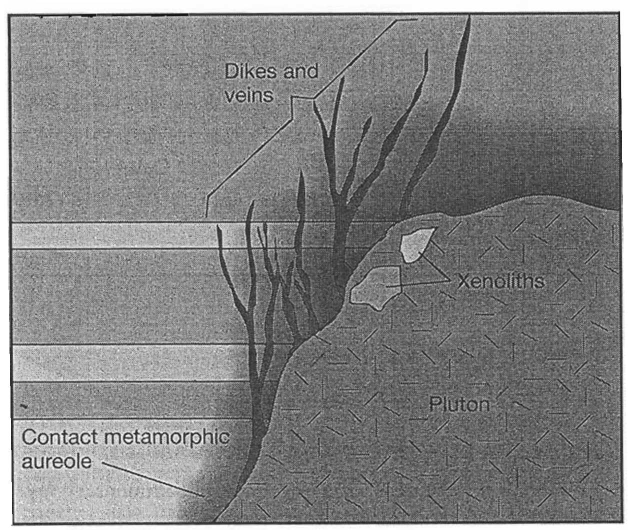


Figure 4.9 Characteristics of a plutonic intrusive contact in sedimentary rocks.

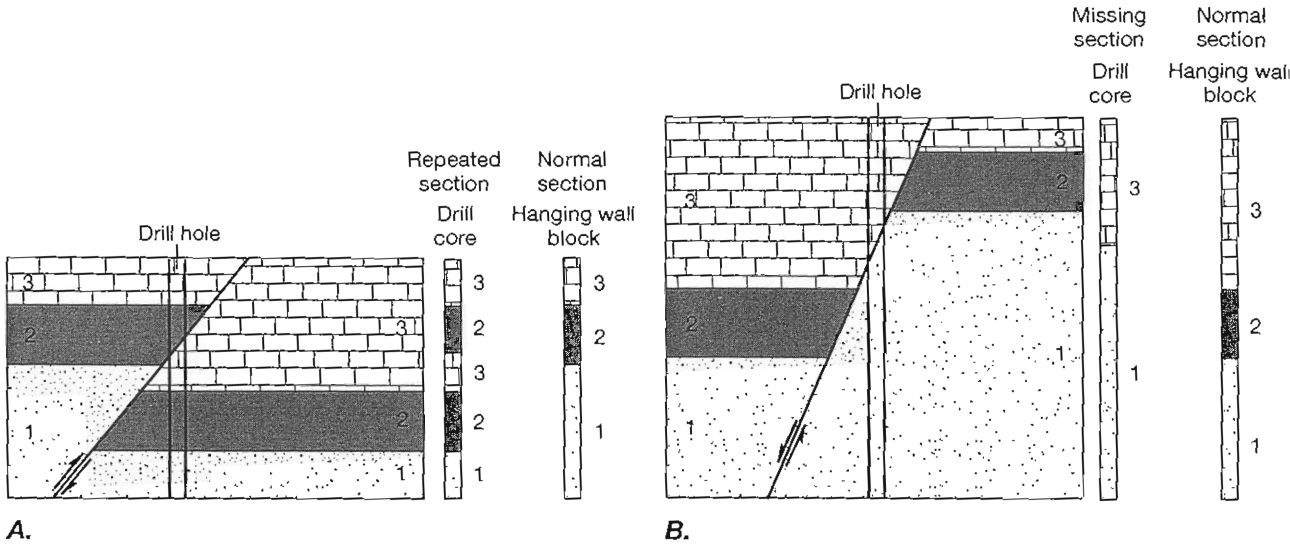


Figure 4.10 *A.* Thrust fault resulting in repeated section in a vertical drill hole. *B.* Normal fault resulting in missing section in a vertical drill hole.

the fault plane—that is, the cutoff lines—are at a high angle to the slip direction on the fault (Figure 4.11*A*, *B*). Drag folds are less likely to form if the cutoff lines are nearly parallel to the slip direction (Figure 4.11*C*).

Drag folds are especially well developed along many thrust faults (Figure 4.11*A*). Along normal faults, rollover anticlines, which develop in the hanging wall block, are more common (Figure 4.11*D*). The direction of bending in these folds is opposite to that found in drag folds, and they reflect the deformation necessary to accommodate the hanging wall block to a curved fault surface (see Sections 5.1 and 5.4).

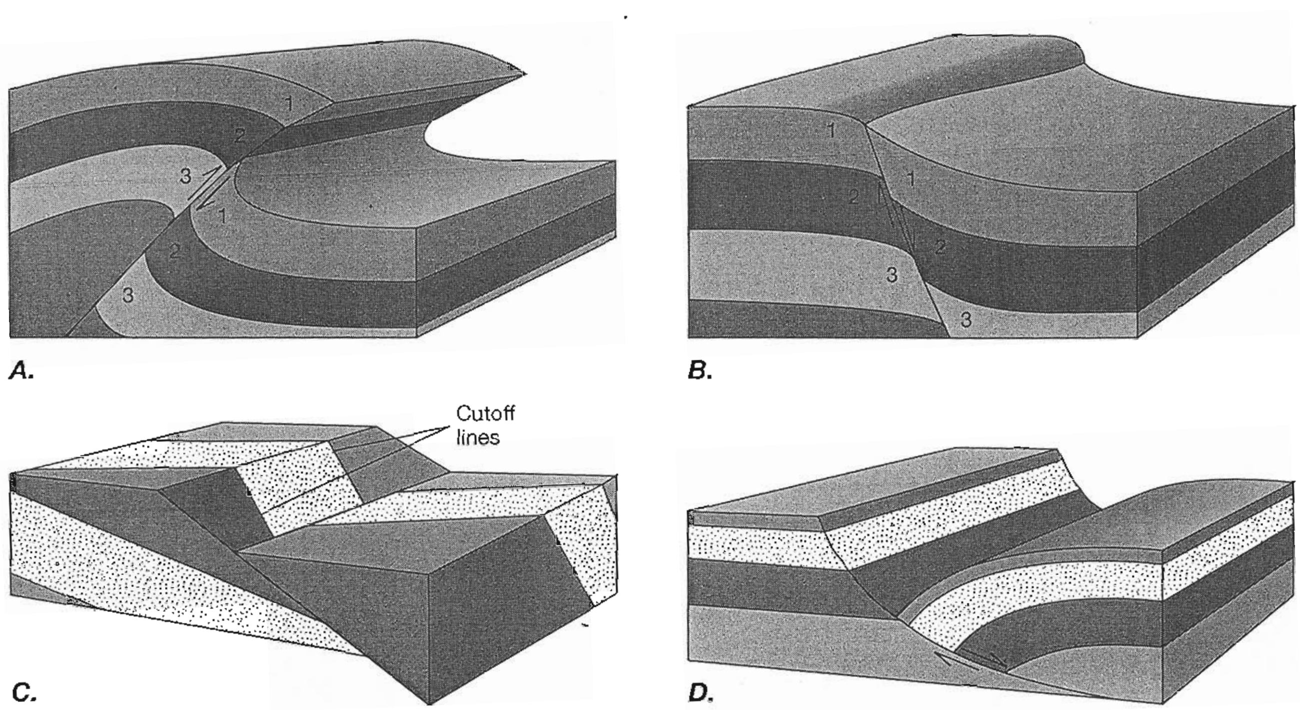


Figure 4.11 Drag folds in sedimentary layers along faults. *A.* A thrust fault, and *B.* a normal fault. *C.* If the cutoff line of the bedding makes a small angle with the displacement direction on the fault surface, the formation of drag folds is less likely. *D.* Rollover anticline in the hanging wall of a normal fault.

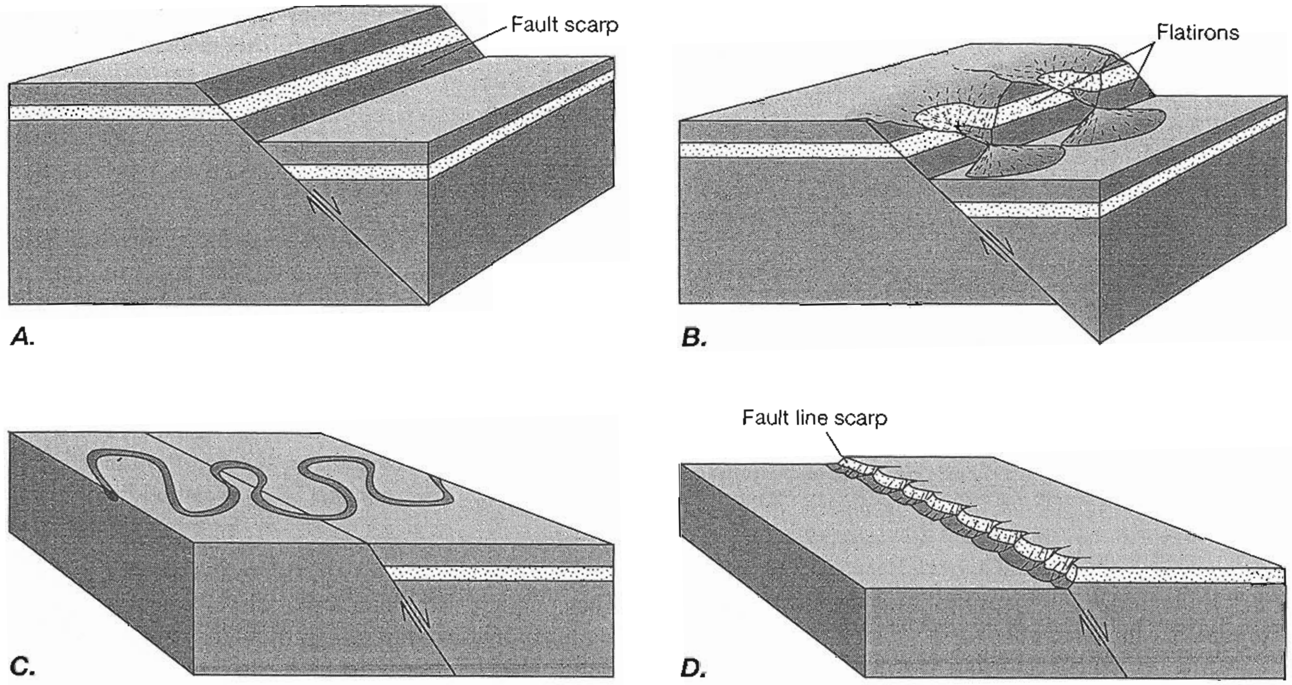


Figure 4.12 Erosion of fault scarps. *A*. Faulting produces a fault scarp. *B*. Erosion of valleys in the fault scarp produces flatirons. *C*. Erosion wears away the thin resistant layer in the topographically high footwall block and levels the topography. *D*. Erosion reaches the level of the resistant layer in the hanging wall block. More rapid erosion in the less resistant layers in the footwall block leaves a topographic step, a fault line scarp.

Physiographic Criteria for Faulting

Many active and inactive faults have pronounced effects on topography, stream channels, and ground water flow. Because these effects frequently suggest the existence of a fault, they are useful in geologic mapping.

Scarps are linear features characterized by sharp increases in the topographic slope; they suggest the presence of faults. There are three types of fault-related scarps: Fault scarps are continuous linear breaks in slope that result directly from displacement of topography by a fault (Figure 4.12*A*). Fault line scarps are erosional features that are characteristic of both active and inactive faults. Figure 4.12 illustrates three steps in the progressive erosion at a fault. Initially, the upthrown footwall block (Figure 4.12*A*) forms a fault scarp. Erosion carves valleys in the fault scarp, leading to formation of flatirons along the mountain front (Figure 4.12*B*). Eventually the upthrown block is eroded down to the same level as the downthrown hanging wall block (Figure 4.12*C*). Subsequent erosion exposes the thin layer in the hanging wall block, which is more resistant than the surrounding layers. Further erosion occurs most rapidly in the least resistant rocks (Figure 4.12*D*), leaving a scarp that, as in this case, does not necessarily indicate the sense of displacement on the fault.

Fault benches are linear topographic features characterized by an anomalous decrease in slope. They form where a fault displaces an originally smooth slope so that a strip of shallower slope results, or where erosion of the less resistant rocks in a fault zone produces a shallower slope than is supported by the surrounding, more resistant rocks. Fault benches may be associated with any of the different fault types.

Ridges, valleys, or streams may be offset along a fault. Figure 4.13 shows two stream channels that have been offset by strike-slip motion on a fault. The deflection of the stream channels may indicate the sense of slip on the fault, but if the fault displacement is sufficiently large, original stream channels may be abandoned. In this case, the "dog leg" in the channel may not correspond to the sense of displacement.

A fault surface or fault zone may act as either a conduit or a barrier for ground water, depending on the permeability of the material both in the fault and on either side of the fault. A breccia zone forms an excellent conduit for water, but a thick gouge zone containing abundant clay minerals may act as a barrier to flow. If faulting offsets an aquifer, or juxtaposes an impermeable rock, such as a plutonic or metamorphic rock, against a good aquifer, it may also significantly alter the flow of ground water. Thus fault traces are

Figure 4.13 Photograph of San Andreas fault, central California, showing streams offset along the trace of the fault.

often characterized by springs and by water-filled depressions called sag ponds.

A stream tends to form a consistent equilibrium profile characterized by a slope that gradually steepens toward the headwaters. Such a profile can change because of fault movements or because of the variable erodability of the bedrock. Any sharp changes in the profile or valley shape of a stream that cannot be ascribed to the change in erosional resistance of the bedrock may betray a fault, but further evidence should be sought.

4.3 Determination of Fault Displacement

Complete determination of the total displacement on a fault requires knowledge of the magnitude and direction of its displacement. Some features indicate the total displacement; others permit a partial or approximate determination; still others only place a constraint on the possible displacements.

Complete Determination of Displacement

The complete determination of displacement on a fault requires identification of a particular preexisting linear feature that intersects the fault surface and is displaced by it. The piercing points of a linear feature are the points where it intersects the fault surface (Figure 4.14). The vector connecting the two piercing points uniquely determines the direction and magnitude of fault displacement, and the relative positions of the linear feature

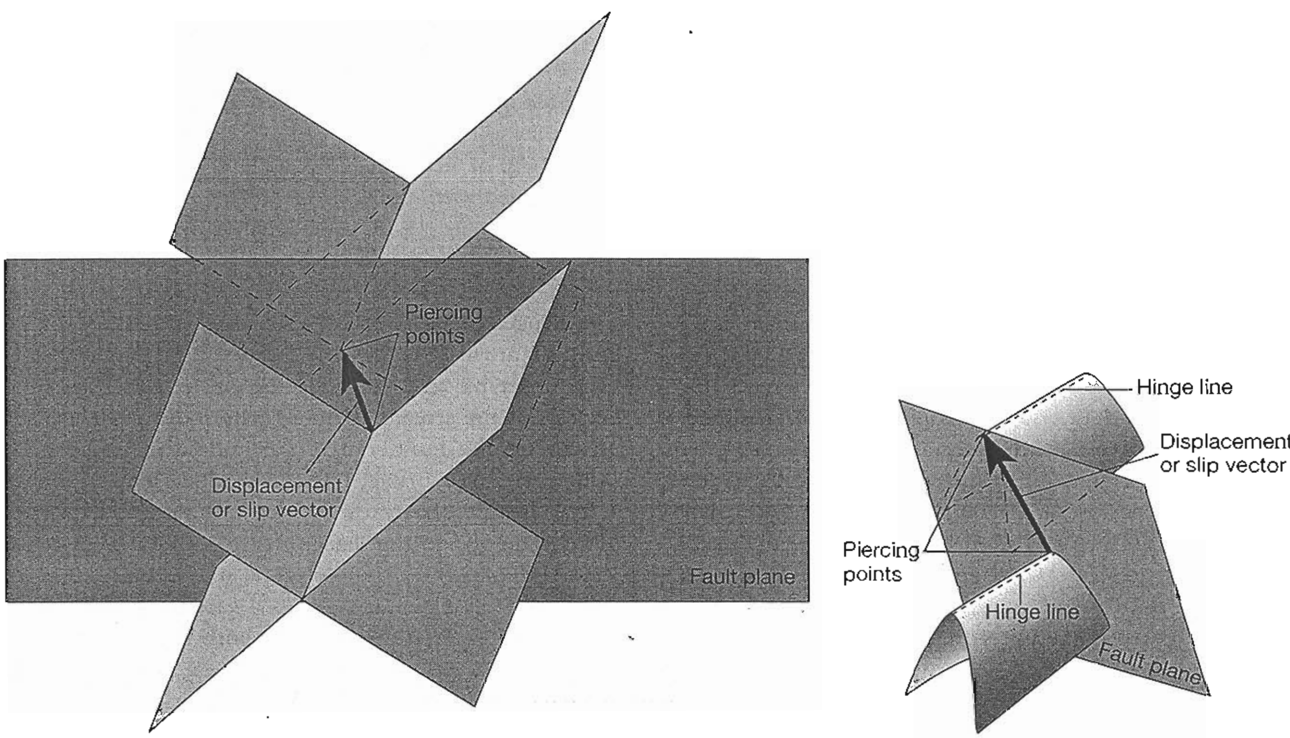


Figure 4.14 Complete determination of the displacement or slip vector from the offset of a unique linear feature cut by a fault. *A.* The intersection of two planar features. *B.* The hinge line of a fold.

on opposite sides of the fault give the sense of shear of the fault.

Several linear geologic features provide piercing points on fault surfaces. The intersection of two distinct planes always defines a unique line that, when cut by a fault, can be used to determine the fault displacement (Figure 4.14*A*). Examples include two other faults; two differently oriented veins or dikes; a bedding plane and a fault, vein, or dike; and an unconformity and a geologic contact such as a bedding plane or intrusive contact. The point of maximum curvature, or hinge line, of a fold also provides a unique line by which to determine the displacement (Figure 4.14*B*). Buried river channels and linear sandstone bodies (shoestring sands) are linear stratigraphic features that can be used, and cylindrical bodies such as volcanic necks and some ore deposits can serve in the same manner.

Partial Determination of Displacement from Small-Scale Structures

In many cases where a fault or ductile shear zone is identified, it is possible to determine the orientation of the displacement vector and the sense of shear, but not the magnitude of the displacement. This type of information can be obtained by examining features at the microscopic to hand sample scale.

As we have noted, slickenside lineations form parallel to the direction of displacement on a fault (see Figure 4.8), but the magnitude of the displacement is more difficult to obtain. Ridge-in-groove lineations that form during propagation of the fracture may be longer than the displacement vector on the fault. Mineral streaks may result from comminution and smearing out of mineral grains and may give a minimum estimate of the displacement magnitude, although this correlation has never been proved.

Slickenfiber lineations grow at a small angle to the shear fracture boundary such that an arrow pointing along a fiber from its point of attachment to one fault block indicates the direction of relative motion of the opposite block (Figure 4.15*A*). The fibers probably grow during slow aseismic movement on a fault, and the fiber growth keeps up with the displacement. Opposite ends of the same mineral fiber should therefore join points that were adjacent when the fiber started growing. Thus, for the period of time of fiber growth, the length of the fiber is a measure of the displacement magnitude on a particular fracture. Shear displacement that might have occurred before the onset of fiber growth would not, of course, be recorded. The maximum displacement magnitudes that are recorded on individual fractures by these fibers are rather small—on the order of 10 to 20 cm. Fibers much longer than this are either not formed or not preserved. In principle, the minimum total

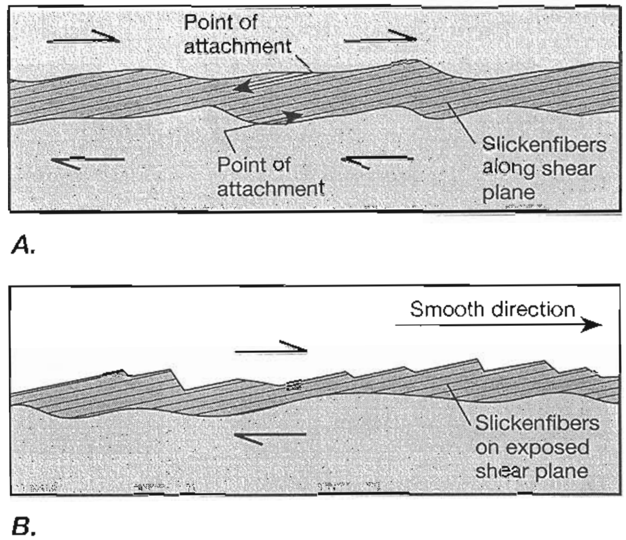


Figure 4.15 Slickenfibers as indicators of shear sense and minimum displacement. *A.* An arrow along a slickenfiber with its base at the point of attachment to one wall of the fault points in the direction of relative slip of the opposite wall of the fracture. The length of the fiber from one wall to the other is a measure of the minimum displacement on that fault. *B.* The smooth, or "downstairs," direction on the stepped surface of an exposed set of slickenfibers defines the direction of relative slip of the missing block.

displacement across a fault zone should be the sum of the fiber lengths on all shear fractures in a cross section of the zone, but where displacements on the order of meters or more are involved, this is an impractical measurement to make.

Generally, slickenfibers are found on fault surfaces that have been exposed by erosion of one of the fault blocks. Because the crystals grow at a low angle to the fault surface and tend to break off either along the fibers or at a high angle to them, such a fault surface tends to have a stepped texture (Figure 4.15*B*). The surface feels smoothest to the hand when rubbed in the direction of relative motion of the missing block.

During brittle faulting, minor secondary fractures can develop along the fault surface at low to moderate angles to the fault. These secondary fractures may be either extension or shear fractures. In general, the extension fractures are not striated and may be filled with a secondary mineral; secondary shear fractures are striated.

Secondary fractures provide four criteria that are useful for determining the sense of shear on the fault surface.

1. As viewed on an exposed fault surface, extension fractures cut below the fault surface in the direction of movement of the missing fault block (Figure 4.16*A*). These extension fractures are essentially the

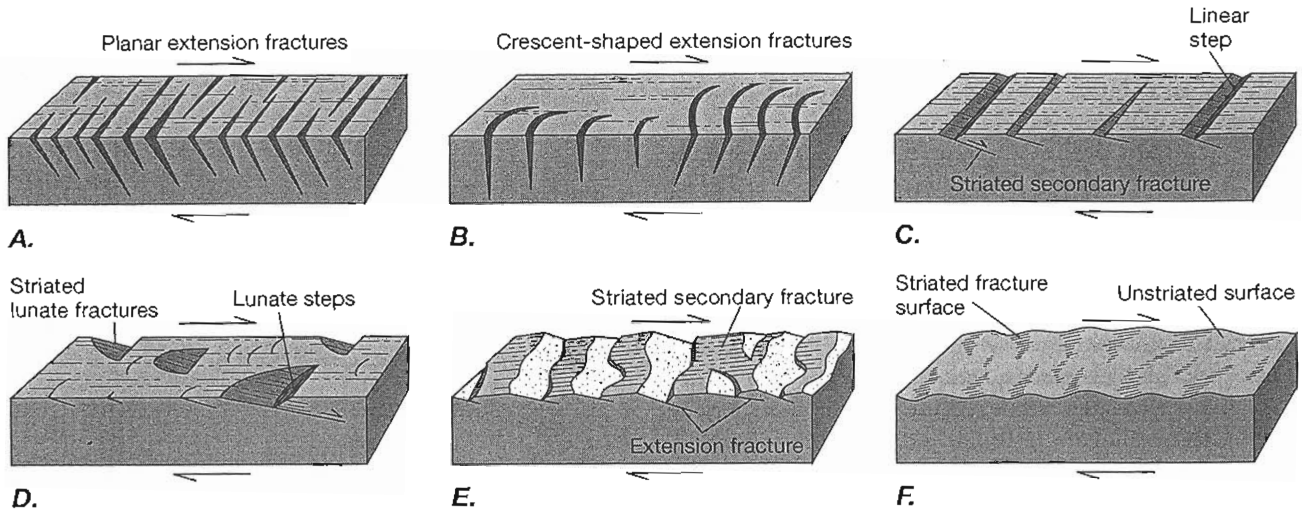


Figure 4.16 Shear sense criteria on brittle faults. Block diagrams show the relationship between secondary fractures and the sense of shear on a brittle fault. The top plane is the shear plane; relative motion is indicated by arrows. Extension fractures are unstriated and may be filled with secondary minerals. Striated fracture surfaces are shear fractures.

same as the pinnate fractures shown in a section normal to the fault in Figure 3.7 (Section 3.1). In cross section, sets of these fractures may form *en echelon* arrays of gash fractures.

2. If the extension fractures are crescent-shaped as exposed on the fault surface, they are concave in the direction of motion of the missing fault block (Figure 4.16*B*).
3. If striated secondary fractures extend beneath the main fault plane, then the fractures cut down into the surface in the direction of motion of the missing fault block (Figure 4.16C). Fracturing of the acute wedge of rock between the secondary shear and the fault surface produces steps in the surface that face in a direction opposite to the motion of the missing fault block. The steps may be predominantly linear (Figure 4.16C), or they may have a lunate morphology (Figure 4.16*D*).
4. Some striated secondary shears do not cut below the fault surface. They may alternate with unstriated secondary extension fractures that cut below the fault surface (Figure 4.16*E*), or they may be simply the faces of irregularities in the fault surface (Figure 4.16*F*). In these cases, the striared surfaces face opposite to the direction of motion of the missing fault block, and sreps formed by breaking across the acute angle between the secondary shear and extension ftactures also face in that direction (Figure 4.16*E*).

Because of the step faces formed in association with these secondary fractures, the fault surface generally feels smoothest to the hand when rubbed in a direction *opposite* to the sense of motion of the missing block. This criterion is opposite to that for determining shear sense on a fault surface dominated by slickenfibers. Thus one must be careful in using the smoothest direction on a fault surface to determine shear sense.

Ductile sheat zones may contain a number of small-scale structures that indicate the shear sense. Platy minerals may become aligned to form a foliation (labeled S in the figure) that makes an angle of about 45° to the shear zone at its boundary and becomes roughly parallel to the zone at its center (Figure 4.17*A*). The sigmoidal pattern of the foliation defines the sense of shear as indicated in the figure.

Ductile faults also characteristically exhibit extended tube-shaped folds in layering that are called sheath folds (Figure 4.17*B*). The long dimension of these folds is approximately parallel to the direction of slip on the ductile fault.

Many rocks in ductile shear zones contain large crystals. Some are relict crystals, or porphyroclasts (after the Latin word *porphyry,* which means purple, and the Greek word *klastos* which means broken), that survived the shearing and reduction in grain size from the original rock. Others are porphyroblasts (after *porphyry* and the Greek word *blastos* which means growth), which are mineral grains that grow to a relatively large size in a rock during metamorphism and deformation. (The use of *porphyry* for a rock with large crystals comes from the fact that statues of Roman emperors were carved from purple volcanic rock containing large feldspar phenocrysts.)

Porphyroclasts found in mylonites may have asymmetric "tails" of very fine grains that are recrystallized from the edges of the porphyroclast itself. The sense of asymmetry of the tails defines the sense of shear in the deformed rock. We distinguish two different tail mor-

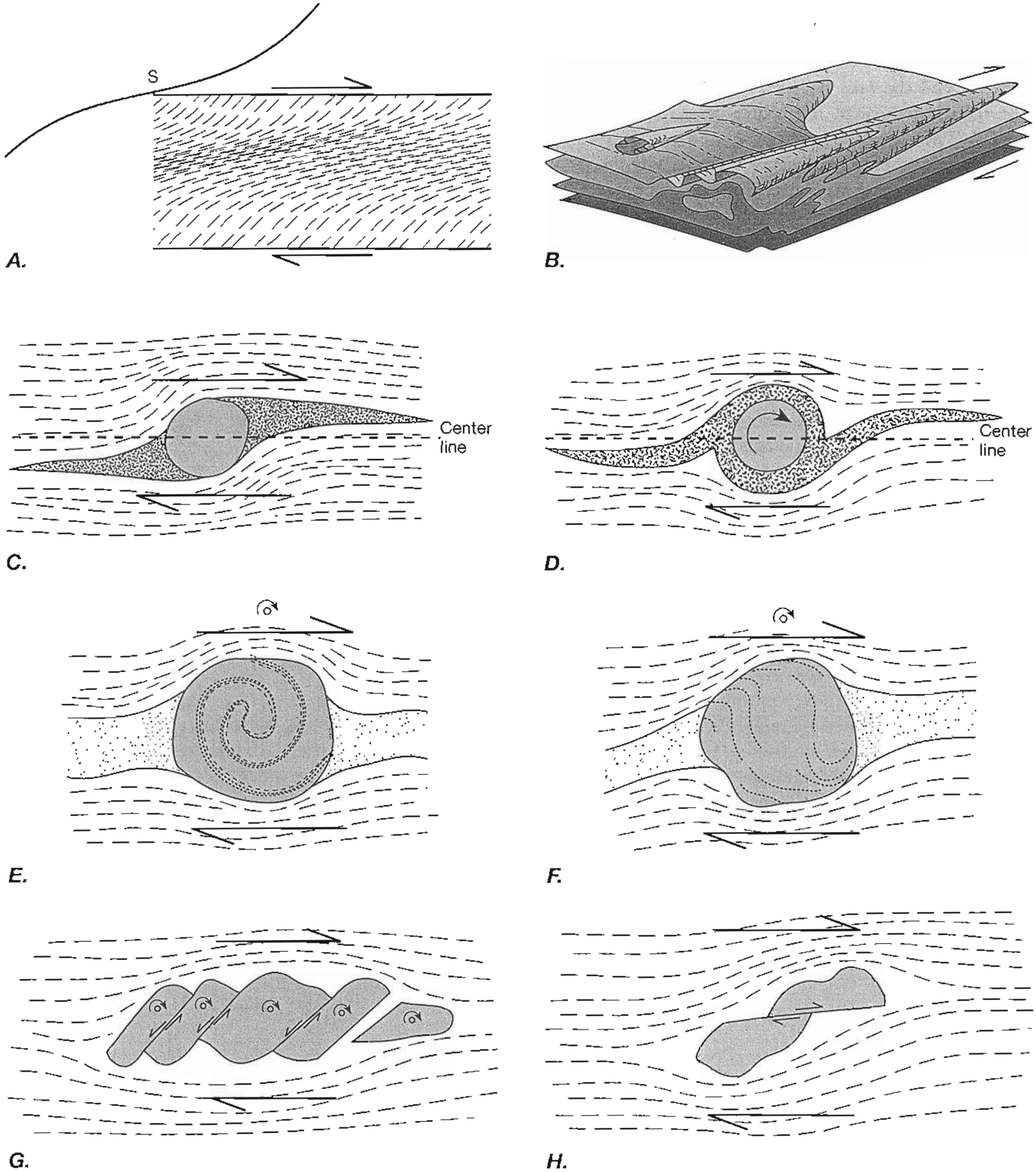


Figure 4.17 Shear sense criteria in ductile shear zones. *A*. Sense of curvature of a foliation defined by the parallel alignment of platy mineral grains. *B*. The orientation and asymmetry of sheath folds shown with the surrounding layers of rock removed. *C*. The sense of asymmetry of σ-type "tails" of recrystallized porphyroclastic material relative to the porphyroclast. *D*. δ-type "tails." *E*. The sense of rotation of a helical train of mineral grains accumulated within a porphyroblastic crystal grain such as garnet or staurolite. *F*. An inclusion train that indicates only a small amount of rotation is not, by itself, adequate evidence of shear sense. *G* and *H*. The sense of shear on fractures in mineral grains such as feldspar or mica depends on the shear sense of the fault and the angle between the fracture and the shear plane.

phologies, the σ-type and the δ-type. On σ-porphyroclasts, the tails extend from each side of the grain in the "downstream" direction of the relative shear in the matrix, and the tails do not cross the line parallel to the foliation through the center of the grain (Figure 4.17*C*). The δ-type is derived from the σ-type by rotation of the porphyroclast in a sense consistent with the shear, and the tails do cross the center line (Figure 4.17*D*). Similarly shaped asymmetric tails may also develop from the crystallization of a mineral different from the porphyroclast in asymmetric zones called pressure shadows. These tails, however, are difficult to interpret unambiguously and should not be confused with the tails composed of recrystallized porphyroclast mineral.

Porphyroblasts do not deform with the rest of the rock but rotate as rigid grains during ductile deformation of the matrix. Minerals that form porphyroblasts include garnet and staurolite. As they grow during deformation, they enclose adjacent minerals, such as micas, from the matrix. Continued rotation and growth of a porphyroblast result in a helical train of inclusions that defines the sense of rotation of the grain and thereby the sense of shear in the rocks (Figure 4.17*E*, *F*). Interpretation of shear sense from inclusion trains that indicate only small amounts of rotation, however, is unreliable because such inclusion trains may actually preserve crenulations in the original foliation rather than a record of porphyroblast rotation (Figure 4.17*F*).

Some porphyroclastic minerals, such as mica and feldspar, tend to shear on discrete fractures or crystallographic planes to accommodate ductile deformation in the surrounding matrix. If the fractures initially make a high angle with the shear plane, then the shear sense on these planes is opposite to that in the surrounding matrix (Figure 4.17*G*). If, on the other hand, the fractures make a relatively low angle with the shear plane, the shear sense on the fractures is the same as it is in the matrix (Figures 4.17*H* and 4.7).

Partial Determination of Displacement from Large-Scale Structures

In regions where fault displacement measures tens to hundreds of kilometers, large-scale geologic features that have been offset can be used to determine the displacement direction and shear sense and to estimate the displacement magnitude of the fault. Such features include shorelines, sides of sedimentary basins, and the source and depositional site of distinctive sediments. Figure 4.18 shows a paleogeographic map of an Oligocene sedimentary basin in western California that has been cut by the San Andreas fault. The different map patterns distinguish marine from nonmarine deposits. Offset of the shoreline along the fault from A to B suggests a right-lateral component of displacement of approximately 300 km since Oligocene time.

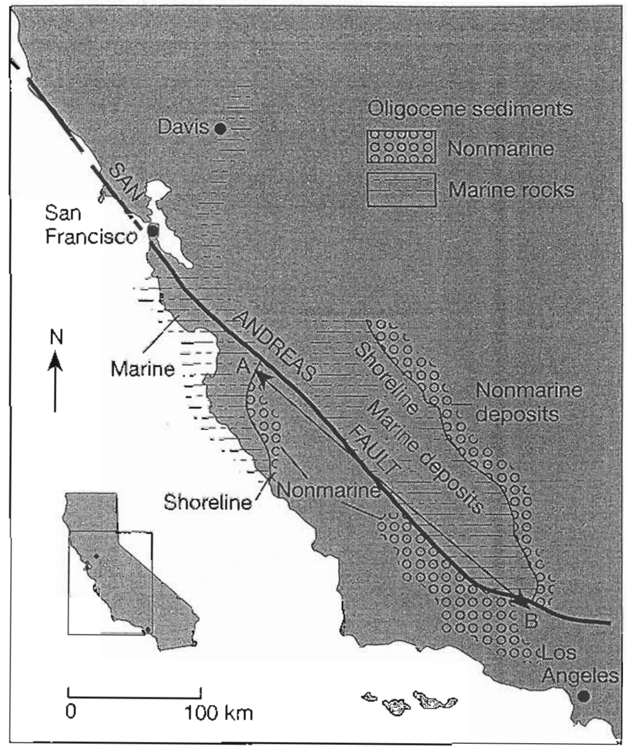


Figure 4.18 Shorelines of an Oligocene sedimentary basin offset about 300 km along the San Andreas fault, California. Points A and B were originally adjacent to each other.

Isopach maps are maps showing the contours of thickness of a geologic unit. (The term *isopach* is derived from the Greek words *isos*, which means equal, and *pachos*, which means thickness.) If there is a regular variation in the thickness of a layer, fault offset of the layer (Figure 4.19*A*) may show up as a discontinuity on an isopach map (Figure 4.19*B*). Each isopach is a unique line of constant layer thickness, and matching isopachs across the discontinuity should in principle make it possible to determine the horizontal component (*H*) or the strike-slip component (*S*) of the displacement. If the data are from well logs, however, they are not very closely spaced. Thus the locations of isopachs and their intersection with the fault may be only approximate. Moreover, unless the fault strike is known, the strike-slip component of displacement cannot be determined accurately. Because isopach maps do not include elevation information, we cannot use them to determine the dip-slip and the vertical components of displacement.

Structure contour maps are maps of elevation contours on a particular geologic surface at depth, generally a stratigraphic horizon. Faults in the subsurface can be identified from discontinuities in structure contours (Figure 4.20). If the structure contours display a linear

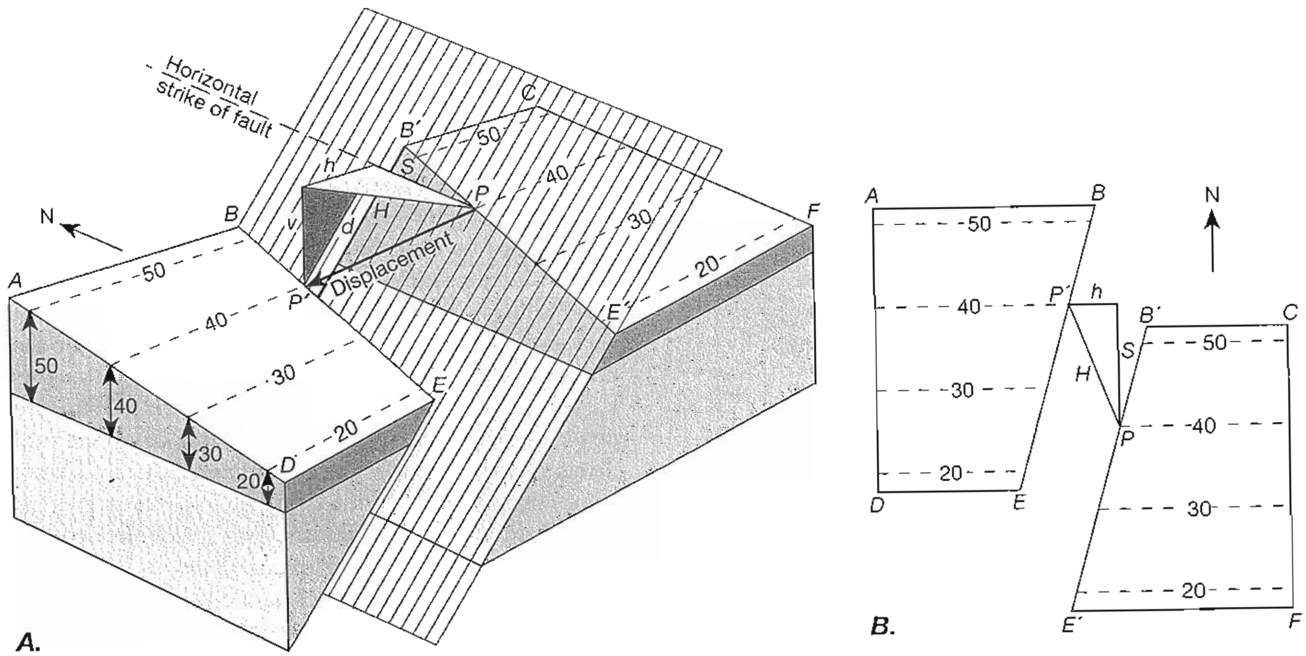


Figure 4.19 Interpretation of a fault from a discontinuity on an isopach map. *A*. Block diagram showing a bed of varying thickness cut by a fault. Contours of equal thickness (isopach lines) are drawn on the top surface of the bed. The true displacement and the strike-slip (*S*), dip-slip (*d*), horizontal (*H*), and vertical (*v*) components are shown. For ease of interpreting the three-dimensional drawing, we show a special case for which the isopachs are parallel to the strike of the layer surface and perpendicular to the strike of the fault. *B*. Isopach map of the structure shown in part *A*. The horizontal component of displacement (*H*) is determined by connecting the map projections of rwo points *P* and *P′* that mark where the same isopach on opposite sides of the fault intersects the fault surface. The strike-slip component of displacement (*S*) is determined by connecting rhe extensions of equal isopach lines with a line parallel to the srrike of the fault.

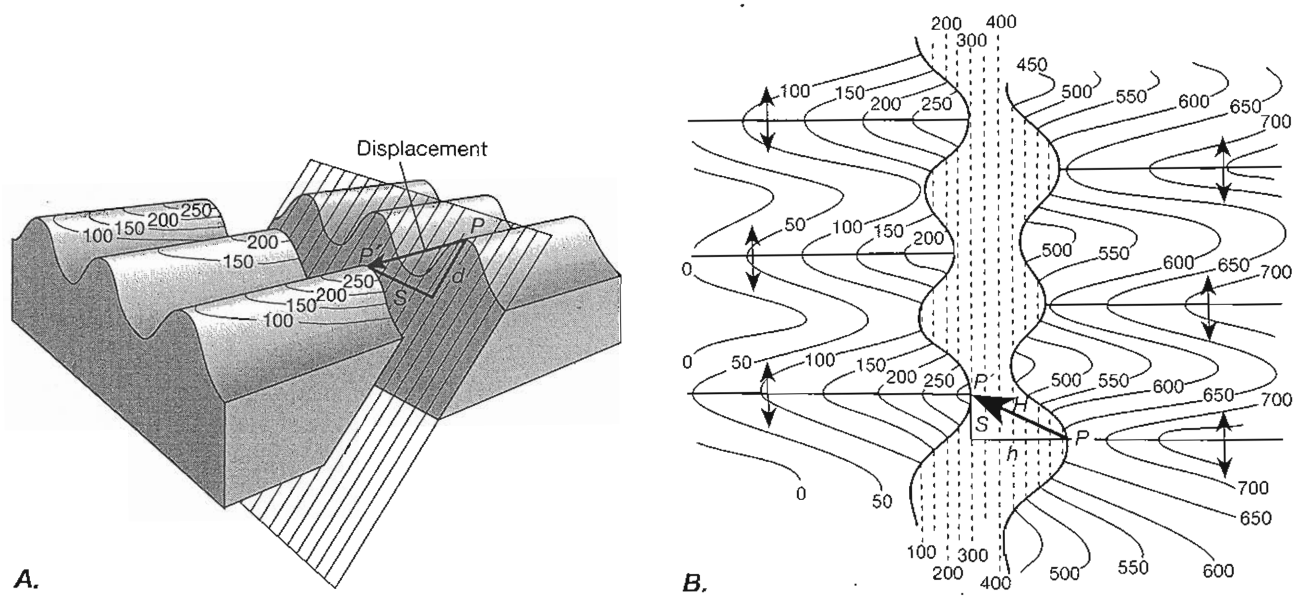


Figure 4.20 Interpretation of a fault from a structurè contour map. *A*. Three-dimensional diagram showing the folded surface of a stratigraphic contact that has been cut and displaced by a fault. The contacr has contours of equal elevation (strueture conrours) drawn on ir. *B*. A structure contour map of the same structure shown in part *A*. The horizontal component of displacement (*H*) is determined by joining the points on the map that are the vertically projected piercing points *P* and *P′* of the fold hinge on opposite sides of the fault. The strike-slip component (*S*) is parallel to the faulr strike, and the horizontal dip-slip component (*h*) is normal to the fault strike.

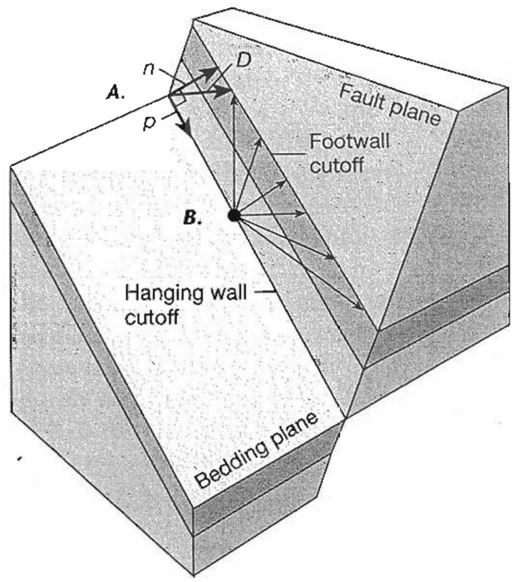


Figure 4.21 Faulted planar features are nonunique indicators of fault displacement. *A*. The true displacement on the fault (*D*) can be specified as the sum of the component normal to the cutoff line of the planar feature (*n*) and the component parallel to the cutoff line (*p*). *B*. Because the parallel component of displacement (*p*) does not produce any offset in the plane, any value of *p* results in the same geometry for the faulted plane, and the displacemeut *D* is therefore not uniquely defined.

structure such as a fold hinge that can be matched uniquely across the fault (Figure 4.20*A*), the fault displacement can be determined. The horizontal component of fault displacement (*H*) (Figure 4.20*B*) is determined by matching the map pattern of the structure contours across the fault; the vertical component is determined from the different elevations of two initially adjacent structural features. If the strike of the fault can be determined, we can use trigonometry to find 1) the strike-slip component of displacement (*S*) from the horizontal component (*H*) and 2) the actual displacement (*D*) from the strike-slip (*S*) and dip-slip (*d*) components (Figure 4.20*A*). As with isopachs, however, the determination is subject to the accuracy of the contours themselves.

Nonunique Constraints on Displacement

Frequently, the principal evidence for a fault consists of the offset of a planar structure, typically sedimentary bedding. It is very important to realize that this offset alone can never define the displacement on the fault, regardless of the appearance of the outcrop pattern. The reason for this is not difficult to understand. We express the true displacement (*D*) as the sum of its components in the fault plane normal (*n*) and parallel (*p*) to the cutoff line, which is the intersection of the planar feature with the fault (Figure 4.21*A*). The normal component (*n*) is the normal distance between the matching cutoff lines on the hanging wall and the footwall, and this component can be measured easily. The component parallel to the cutoff line (*p*), however, produces no change in the orientations or locations of the cutoff lines and thus cannot be observed. The complete description of the displacement (*D*) is therefore indeterminant. Figure 4.21*B* shows six of the infinite number of displacement vectors that could produce the same geometry for the faulted bedding plane. Included among these six vectors are components of normal, reverse, dextral, and sinistral displacement. Each vector has the same component of displacement normal to the cutoff line (*n*) but a different component parallel to the cutoff.

Thus, if the only information available about a fault is the offset of parallel planar features, we cannot talk about the slip or displacement on the fault because we cannot determine it. We speak instead of the separation, which is the distance measured in a specified direction between the same planar feature on opposite sides of the fault (Figure 4.22). The separation enables us to determine only the component of displacement normal to the cutoff line. Any component parallel to that line is iudeterminate. Illustrated in the figure are two of the common separations and the directions in which each is measured—the strike separation, measured parallel to the strike of the fault, and the **dip** separation, measured parallel to the dip of the fault. Other separations used in some cases include the stratigraphic separation, measured normal to a bedding plane; the vertical separation, measured in a vertical direction (it is the vertical component of both the dip and the stratigraphic separations); and the **horizontal** separation, measured in a plane normal to the strike of

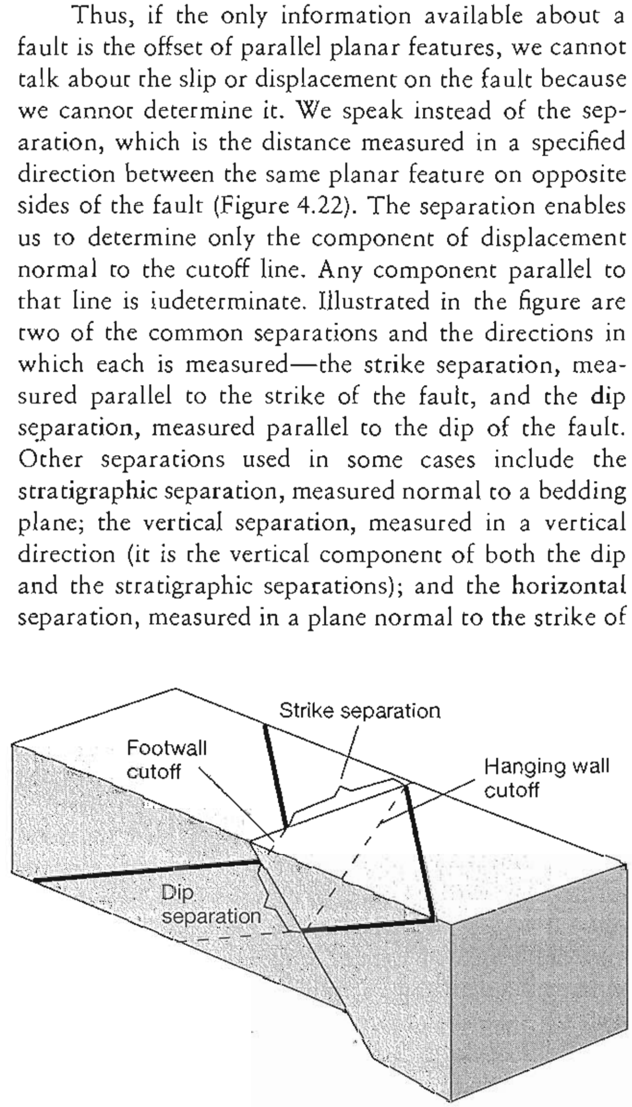


Figure 4.22 Block diagram indicating the strike separation and the dip separation of a faulted layer on a dip-slip normal fault where the footwall block has been eroded down to the same elevation as the hanging wall block.

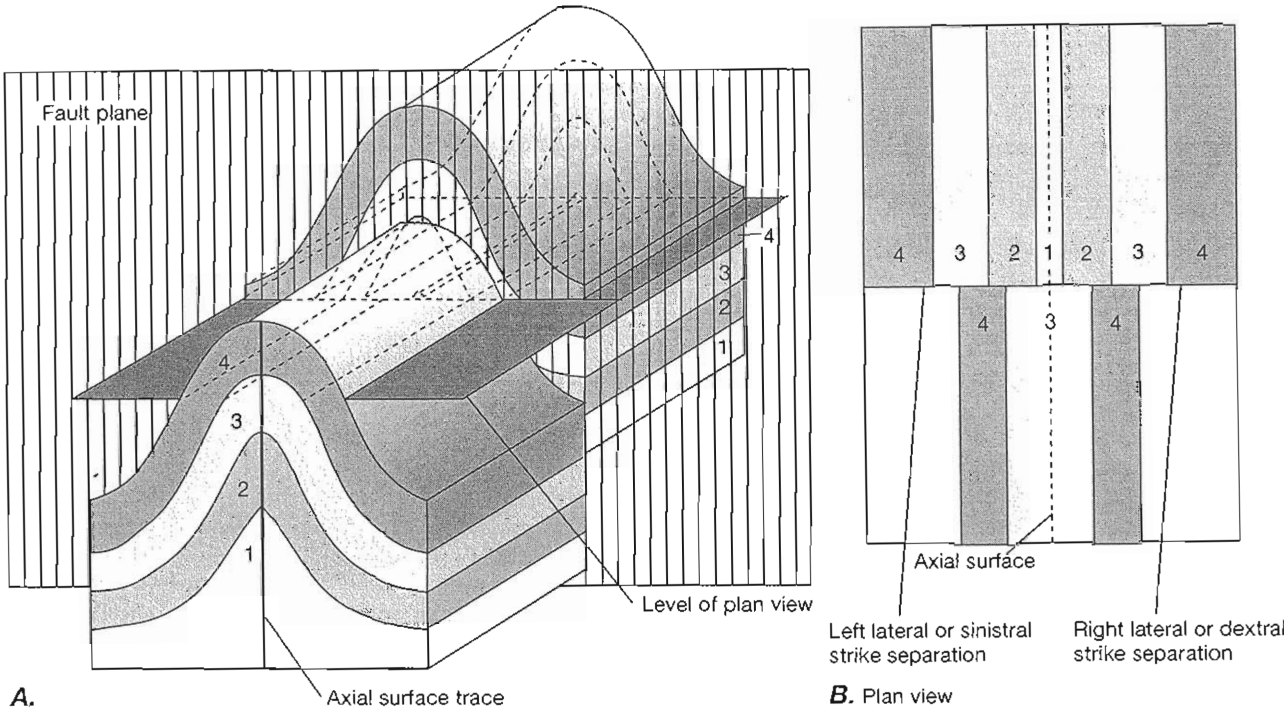


Figure 4.23 Example of different separations produced by the faulting of folded layers. *A*. Fold offset by dip-slip movement along a vertical fault. *B*. Map view at the level of the horizontal plane in part *A*, showing the opposite separations obtained on opposite sides of the fold.

the fault (it is the horizontal component of the dip separation).

On the same faulr, a separation measured for one plane is different from the separation measured in the same direction for a plane of a different orientation. In fact, if the two planar features are appropriately oriented, one strike separation can be right-lateral and the other left-lateral. For example, opposite limbs of a faulted fold can be very different in separation (Figure 4.23). Similarly, opposite senses of dip separation can develop for appropriately oriented planes and displacements.

These examples emphasize the nonuniqueness of the separation and the difficulty, or impossibility, of using it to characterize a fault. If, however, the separation in a particular direction can be determined for two planes of different orientation offset by a fault, then it is possible to determine a unique magnitude and direction for the displacement vector. In effect, the intersection of the two planes defines a line that intersects the fault at unique piercing points on the hanging wall and footwall (sec Figure 4.14*A*).

4.4 Faults in Three Dimensions

All faults are three-dimensional, somewhat irregular surfaces of finite extent. It is all too easy to lose track of this fact, because generally faults appear in outcrop as lines on a relatively two-dimensional surface, and they are represented on the printed page as lines on a map (Figure 4.24*A*) or on a cross section (Figure 4.24*B*). Although such representations are certainly useful, they encourage us to ignore the three-dimensional consequences of fault geometry (Figure 4.24*C*).

An individual fault surface generally is not a flat or smoothly curved surface but instead may have **fault ramps** connecting segments of the fault (Figure 4.25*A*). A **frontal ramp** is oriented such that its intersection with the main fault surface is approximately perpendicular to the displacement direction on the fault. On strike-slip faults, frontal ramps are called **jogs** or **bends**. A **lateral ramp** is oriented such that its intersection with the main fault surface is oblique or parallel to the displacement direction on the fault. If parallel, it is also a **step** or **sidewall ramp**; otherwise, it is an **oblique ramp**.

In general, displacement of a fault block over a ramp induces in the block a deformation whose characteristics depend on the orientation of the fault and the displacement. A frontal ramp or jog can be extensional (Figure 4.25*B*) or contractional (Figure 4.25*C*), depending on whether the material is pulled apart across the ramp or pushed together by the dominant shear on the fault zone.

During faulting at a ramp, the location of the ramp can migrate as the fault surface cuts in discrete jumps into one fault block or the other. The result is a **fault duplex**, characterized by a stack of horses bounded by

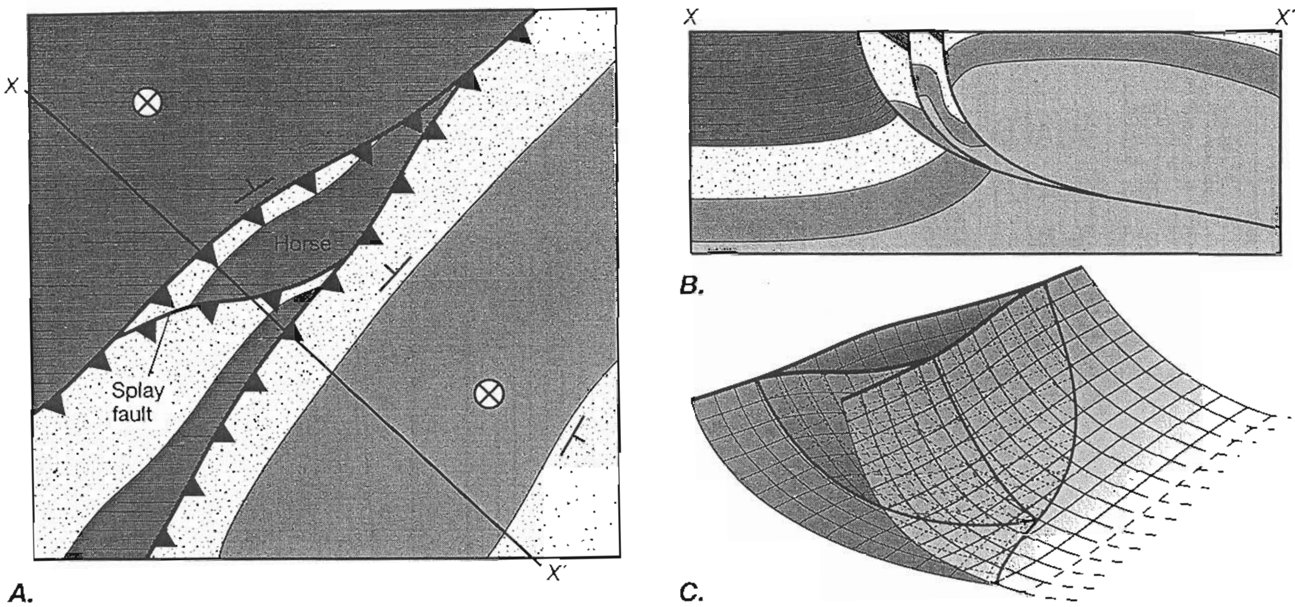


Figure 4.24 Three-dimensional representation of faults. *A*. Geologic map of branching imbricate thrust faults connected by a subsidiary splay fault which isolates a horse. *B*. Cross section along XX′ through the fault system in part *A*. *C*. Portrayal of the three-dimensional geometry of the faults in parts *A* and *B*.

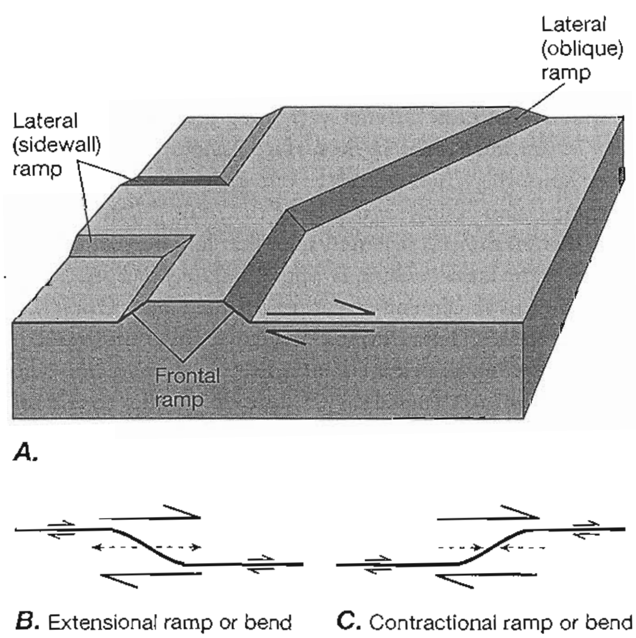


Figure 4.25 Geometry of fault ramps. Any of the diagrams can be oriented arbitrarily and therefore can apply to any fault type. *A*. Schematic shape of a fault surface. *B*. An extensional fronral ramp or jog. The dashed arrows indieate that material tends ro be pulled apart at the ramp. *C*. A conrractional frontal ramp or jog. The dashed arrows indicate rhat material tends to be pushed together at the ramp.

traces of the main fault. Duplexes can be normal (Figure 4.26*B*), thrust (Figure 4.26*A*), or strike-slip (Figure 4.26*C*). We discuss the characteristics of duplex formation that are specific to the different fault types in Chapters 5 through 7.

Every fault surface, no matter what its type, must come to an end in every direction, and the end is marked by a termination line. A termination line must be continuous and must form a closed line about the fault surface; it cannot simply end. It has different featutes, depending on the geometry of the termination.

The termination of a fault at the sutface of the Earth is the fault trace on the topographic surface (Figures 4.27 through 4.29). It may be the original boundary of the fault or the intersection of an originally deeper part of the fault with a surface of erosion. It is in essence the cutoff line of the Earth's surface on the fault.

At a brittle–fluid or brittle–ductile interface, the displacement diseontinuity on a fault is easily accommodated by the flow of the fluid or ductile material. Thus the discontinuity cannot extend beyond the brittle material, and the cutoff line of the interface defines the termination of the fault.

Termination of one fault against another tesults in two relationships between the termination line and the displacement on the younger fault. If a younger fault terminates against an older one, its displacement vector

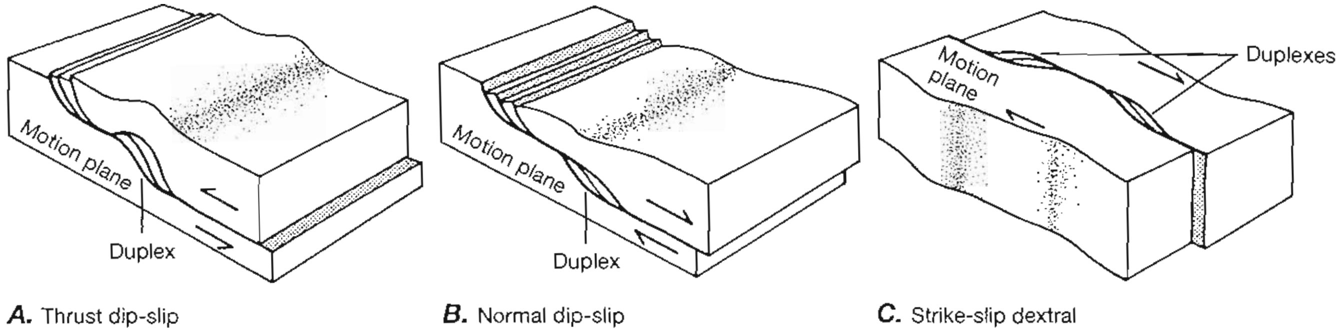


Figure 4.26 Block diagram illustrating duplexes. The motion plane contains the slip direction and the line perpendicular to the fault.

must be parallel to the termination line. In Figure 4.27*A*, a vertical strike-slip fault terminates on a coeval horizontal fault. The termination line must be parallel to the displacement vectors on the strike-slip fault. Similarly, in Figure 4.27*B*, a normal fault terminates against an older vertical fault. The vertical fault initially was a strike-slip fault between block C and the unfractured pair of blocks A and B. Subsequently, however, part of that fault was reactivated as an oblique-slip fault between blocks B and C. In this case the dip-slip displacement on the younger fault must be parallel to the termination line of that fault.

If, on the other hand, a younger fault has cut and displaced an older fault, the termination line of the older fault is its cutoff line on the younger fault, and the displacement vector is not related to the termination line. For example, in Figure 4.27*C*, a right-lateral strike-slip fault cuts an older normal fault. The termination line of the normal fault against the strike-slip fault is not parallel to the displacement vectors on either fault.

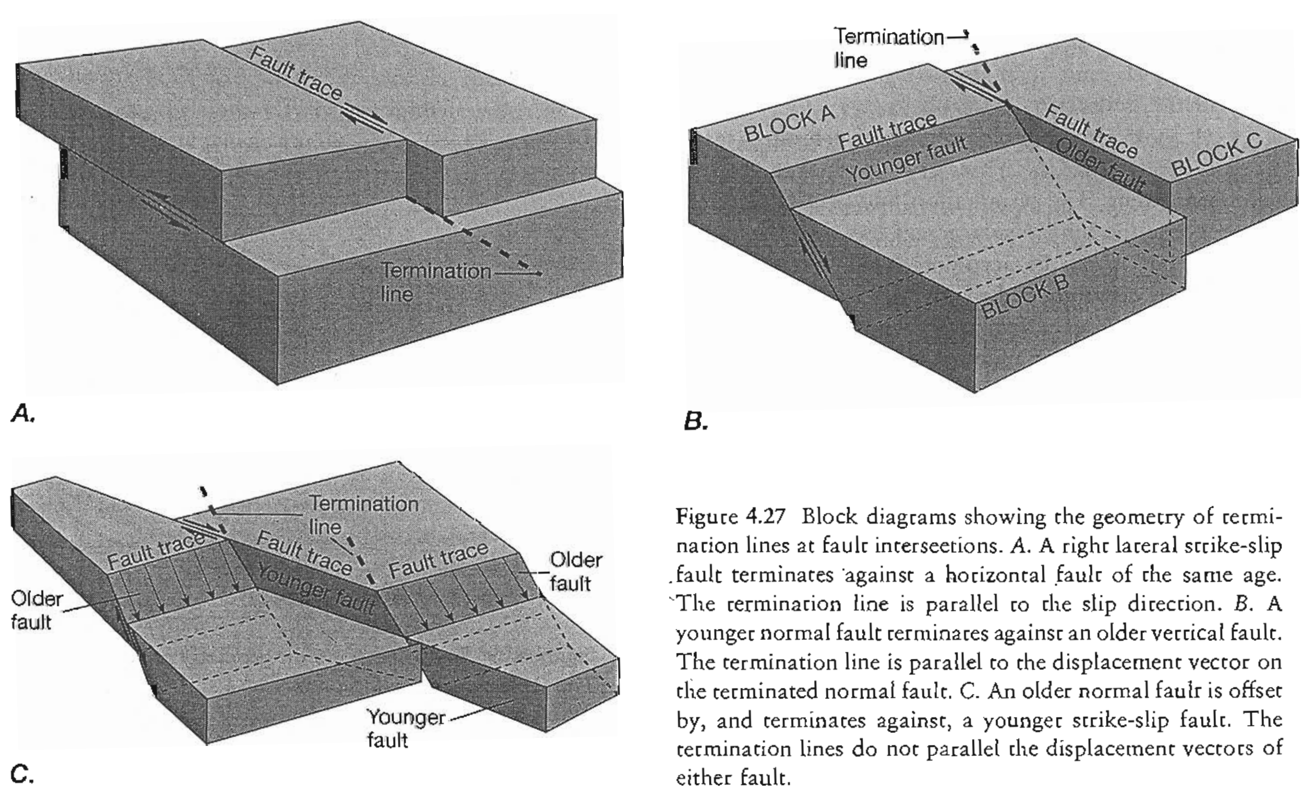


Figure 4.27 Block diagrams showing the geometry of termination lines at fault intersections. *A*. A right lateral strike-slip fault terminates against a horizontal fault of the same age. The termination line is parallel to the slip direction. *B*. A younger normal fault terminates against an older vertical fault. The termination line is parallel to the displacement vector on the terminated normal fault. C. An older normal fault is offset by, and terminates against, a younger strike-slip fault. The termination lines do not parallel the displacement vectors of either fault.

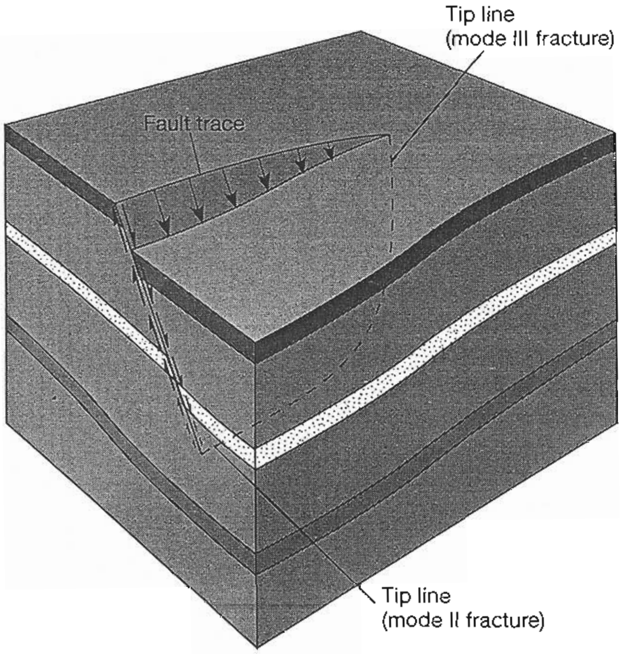


Figure 4.28 Geometry of tip lines. Displacement on a normal fault dies out along the strike and down the dip. The tip line is a continuous quasi-elliptical line. The angle it makes with the displacement vector changes around the perimeter of the fault, changing the fault from a mode III fracture (tip line parallel to displacement) to a mode II fracture (tip line perpendicular to displacement).

For very large strike-slip displacements, the observed fault intersections might resemble those in Figure 4.27*B*, which could result in an erroneous interpretation of the age of the faults.

A look at any geologic map of a faulted terrane demonstrates that the traces of individual faults are of limited extent. The termination line of a fault is a tip line where the fault displacement has decreased to the extent that it can be accommodated by coherent deformation distributed through the solid rock (Figures 4.28 and 4.29). If a fault trace ends without running into another fault, it must end at a tip line. In Figure 4.28 the tip line is parallel to the displacement vector on the fault where it intersects the top surface, and it is perpendicular to the displacement where it intersects the vertical side. Because tip lines of faults are often roughly elliptical in shape at depth, all relative orientations of displacement and tip line between these two end-member types occur on a single fault. Thus, a single fault has the characteristics of both mode II and mode III fractures (compare Figure 3.1) depending on the relative orientation of the displacement and the tip line.

Below the surface, then, a fault can be bounded on all sides by a continuous tip line that connects at both ends with the surface trace of the fault. A blind fault, however, does not break the surface anywhere and thus is completely surrounded by a termination line that is either a tip line or a branch line (see below). Later, possibly long after active faulting ceases, the tip line at the top of the fault surface may be eroded away to create a trace of the fault on the Earth's surface.

A branch line is a line of intersection where a fault surface splits into two fault surfaces of the same type, or two fault surfaces of the same type merge into one. All segments of the fault shown in Figure 4.29*A* are completely surrounded by a fault trace at the surface, a tip line, or a branch line, except, of course, for the trace of the fault on the vertical left side of the block, which is an artificial line on the cross section. In Figure

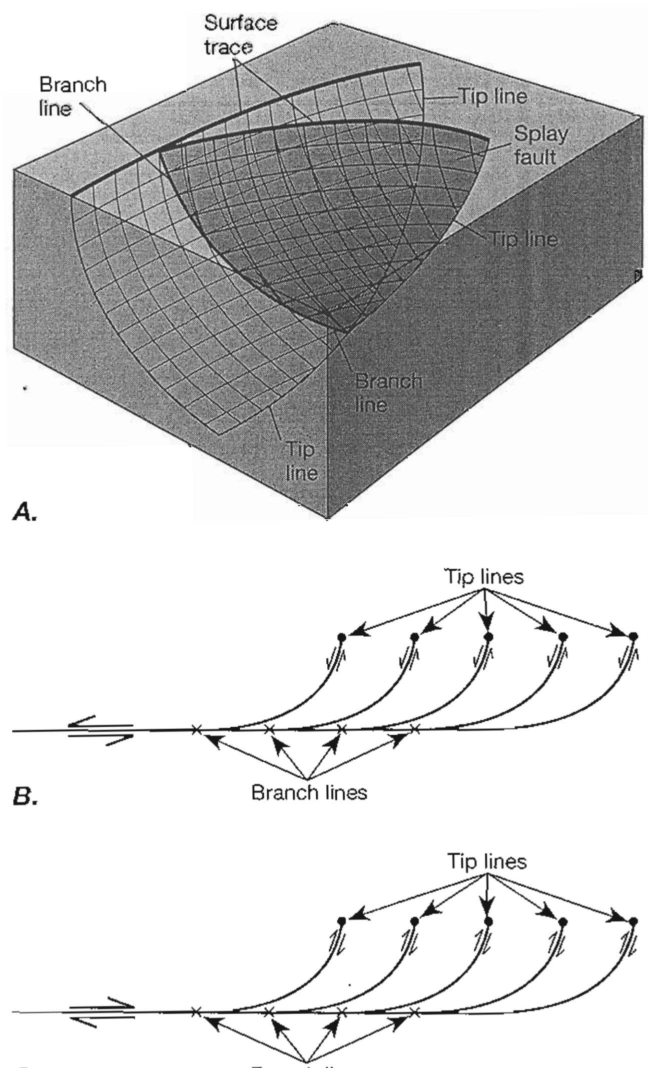


Figure 4.29 Splay faults and the geometry of branch lines. *A.* The three-dimensional geometry of a splay fault shows how the fault surface is completely bounded by a surface trace, a branch line, or a tip line. *B.* Extensional imbricate fans. *C.* Contractional imbricate fans.

4.24C, the horse is bounded on all edges below the surface by branch lines.

Faults of all types commonly die out in a set of splay faults, which are smaller, subsidiary faults that branch off from the main fault (Figure 4.29*A*). Where splay faults branch off from the main fault at fairly regular intervals and have comparable geometries, they form an **imbricate fan**, which can be either extensional (Figure 4.29*B*) or contractional (Figure 4.29*C*). Each splay in turn dies out at its own tip line or intersects a free surface. Where a single fault ends in a series of splay faults, the effect is to distribute the deformation at the end of the main fault over a larger volume of rock, thereby decreasing its intensity.

Two circumstances occur in which a fault can end, in a sense, without being bounded by a termination line. If the fault surface curves, the nature of the fault may change completely, but no termination line exists. Figure 4.30 illustrates a normal fault that changes orientation to become a vertical oblique-slip fault.

The other circumstance involves faults that extend deep into the crust or even into the upper mantle. With increasing depth in the Earth, the temperature and pressure rise, and if they are sufficiently high, rocks become ductile. This ability of rock to flow ultimately limits the depth to which a fault can maintain its identity as a shear zone. As we trace a fault deeper into the crust, we expect it first to change from a zone of brittle deformation into a ductile shear zone. At some depth, which we do not know well, the zone of deformation must spread out, until the nature of the fault as a discrete shear zone is lost, and the displacement is accommodated by the slow, widespread flow of the rocks. In this circumstance, the boundary of the fault is indistinct.

The depth at which a fault loses its identity is not well known, but it probably depends in part on the

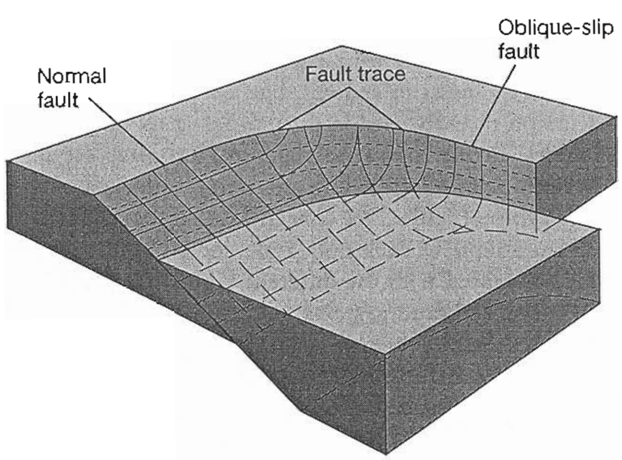


Figure 4.30 A change in orientation of a fault surface modifies a normal fault into a vertical oblique-slip fault. No termination line can be identified.

magnitude and rate of the displacement on the fault. Brittle fracturing is replaced by ductile flow at depths of about 10 to 20 km. Fault zones have been traced by seismic reflection techniques into the lower crust to depths of about 25 km, and in some places, slices of the upper mantle are exposed at the surface along faults, suggesting that faults have extended at least to the Moho. Subduction zones, in fact, are major thrust faults some of which can be traced hundreds of kilometers into the mantle, although we do not expect most crustal faults to extend to such depths. We have limited opportunity to observe rocks that have deformed near and below the base of the crust, however, so our knowledge is very poor.

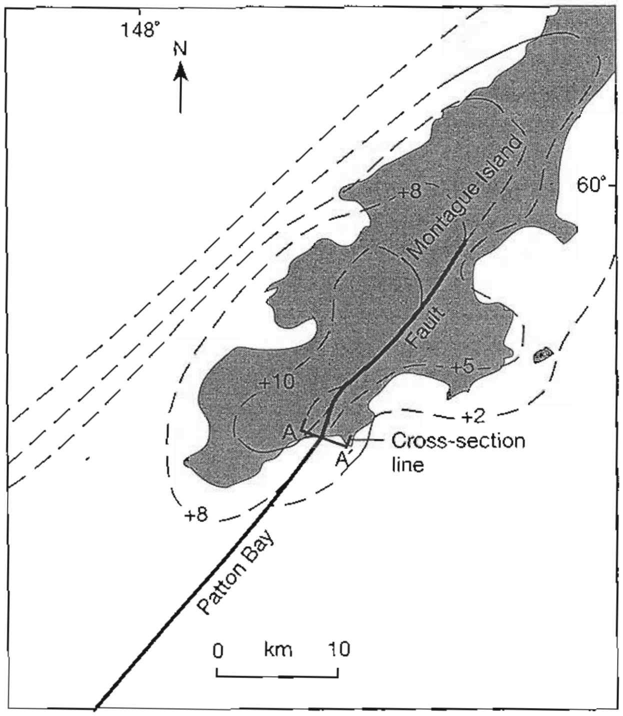
4.5 Refinements

All our discussion about fault movement has concerned movement of one side *relative* to the other. An absolute sense of movement can rarely be obtained, because we generally do not have a reference point independent of both of the fault blocks. Thus, for example, it generally cannot be determined whether the hanging wall of a normal fault has decreased in elevation as a result of faulting, or the footwall has increased in elevation.

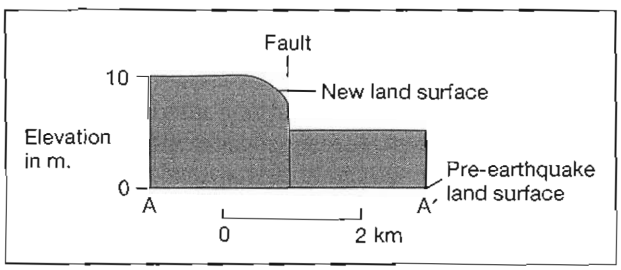
Faulting associated with the Alaska earthquake of 1964 near Anchorage provided one interesting exception to this statement. Measurement of motion of fault blocks relative to sea level showed that in one place where a normal fault had formed, *both* sides had moved *upward*. Figure 4.31*A* shows part of the area affected by the fault and gives contours of the amount of uplift that occurred during faulting. Note that the uplift contours show both sides of the Patton Bay Fault to have moved up but that the northwest side moved up farther than the southeast side. This effect is more clearly shown in Figure 4.31*B*, which is a plot of the amount of uplift along the line AA′ that crosses the fault. This interesting result on a historical earthquake makes one cautious about assuming that the relative motion of the hanging wall is also the absolute motion.

Another exception to our general inability to determine absolute movements on faults occurs with strike-slip faults of very large displacement. In this case, we can take the geomagnetic field as an absolute reference and use paleomagnetic studies of the rocks to determine which side of a fault has changed latitude.

One of the major features used for classifying faults and interpreting their tectonic significance is the original orientation of both the fault plane and the displacement vector at the time the fault moved. The original orientation is important, because many faults are rotated into different attitudes by later tectonic activity. Thus,



A.



B.

Figure 4.31 Faulting of Montague Island, southern Alaska, associated with the 1964 Alaska earthquake. The map shows that the uplift contours are positive on both sides of the Patton Bay Fault. Contours are marked in meters of uplift. The cross section shows a plot of the uplift magnitudes along the line A–A' that crosses the Patton Bay Fault. The fault is a normal fault resulting from uplift by different amounts on opposite sides of the fault.

for example, a thrust fault with the hanging wall up could be rotated about a horizontal axis to such an extent that it resembled a normal fault with the hanging wall down. The present attitude may or may not correspond to the original one.

Cross sections in sedimentary sequences that have not undergone significant deformation parallel to the regional strike must preserve the original cross-sectional area of the undeformed section. Such cross sections are balanced. The requirement that cross-sectional area be preserved places significant constraints on possible interpretations of structure at depth, because any proposed interpretation must make it possible to restore the fault to an acceptable undeformed state characterized by continuous layers with no gaps or overlaps in the stratigraphic section. The cross section must be balanced between two pinning points, which are vertical reference lines chosen to pass through undeformed sections of a stratigraphic sequence. Thus the shape of these reference lines is assumed to have been unchanged by the deformation. In highly deformed regions, the pinning points can be chosen through the bottoms of flat synclines or the tops of flat anticlines, where presumably there has been no slip between beds.

If the deformation has not changed the thicknesses of units in the stratigraphic section, then balancing the section can be achieved by line balancing, which requires that the length of each contact between the pinning points be the same before and after deformation. If deformation has shortened and thickened units within the plane of the section, however, the lengths of contact lines can change, and balancing must be done by area balancing which requires that the area of each unit between pinning points be conserved during deformation.

Although the balancing requirement imposes an important constraint on the construction of cross sections of unmetamorphosed sedimentary rocks, in other geologic situations there may be significant deformation normal to any possible cross section. Under those conditions, material moves in or out of the cross-section plane, and the fundamental assumption of constant cross-sectional area, which is the basis for constructing balanced cross sections, is no longer valid. Some techniques have been developed to account roughly for this deformation, but the process requires large amounts of detailed information, and balancing cross sections under these conditions becomes an increasingly complicated and approximate procedure.

In this chapter, we have described many features common to faults in general, but we have limited our discussion to individual fault surfaces or to simple sets of faults. In the following chapters, we complete our description of faults, dealing separately with each of the major types of faults. We concentrate on those features that are unique to each fault type, including the structures associated with the different faults, the geometry of complicated fault systems, and the tectonic settings in which the different types of faults are found. Despite the differences in tectonic setting of the three major types of faults, the geometric characteristics discussed in this chapter are remarkably consistent for all fault systems, and the differences that do exist are largely attributable to the difference in orientation of the faults with respect to the Earth's surface.

Additional Readings

Fleuty, M. J. 1975. Slickensides and slickenlines. *Geol. Mag.* 112: 319–322.

Petit, J. P. 1987. Criteria for the sense of movement on fault surfaces in brittle rocks. *J. Struct. Geol.* 9: 597–608.

Sibson, R. H. 1977. Fault rocks and fault mechanisms. *J. Geol. Soc. Lond.* 133: 190–213.

Sibson, R. H. 1983. Continental fault structure and the shallow earthquake source. *J. Geol. Soc. Lond.* 140: 741–767.

Simpson, C. 1986. Determination of movement sense in mylonites. *J. Geol. Educ.* 34: 246–261.

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

Woodcock, N. J., and M. Fischer. 1986. Strike-slip duplexes. *J. Struct. Geol.* 8: 725–735.

Woodward, N. B., S. E. Boyer, and J. Suppe. 1989. Balanced geological cross-sections: An essential technique in geological research and exploration. Short Course in Geology: Volume 6. Washington, D.C.: American Geophysical Union.