CHAPTER

2 Techniques of Structural Geology and Tectonics

Investigations in structural geology require familiarity with basic techniques of observation and of reporting and displaying three-dimensional formation. Thus we use standardized methods for measuring the orientations of planes and lines in space. Techniques for distinguishing the relative ages of strata in a sequence of layered sediments are crucial to interpreting their significance. Various graphical displays have proved most useful for plotting and interpreting orientational or three-dimensional data. Such displays include geologic maps and cross sections that present data in a geographic framework, as well as histograms and spherical projections that portray only orientational data without regard to geographic location. Geophysical data such as seismic reflection profiles and gravity measurements are essential to the interpretation of many large-scale structures, and a structural geologist must understand how these data are obtained and interpreted in order to take their limitations into account. Most discussions of structural geology and tectonics assume the reader is familiar with all these techniques.

Proper interpretation of geologic structures depends critically on the ability to visualize spatial relationships among various features. This ability to think in three dimensions does not necessarily come easily, but one can learn it with practice. Learning to "think in 3-D" is a major goal of laboratory courses in structural geology.

2.1 The Orientation of Structures

Many of the structures observed in outcrops are approximately planar or linear features. Planar features include bedding, fractures, fault planes, dikes, unconformities, and planar preferred orientations of micas. Linear features include grooves and streaks on a surface, intersections of two planar features, and linear preferred orientations of mineral grains. We can represent features that are not planar, such as folded surfaces and folded linear features, by measuring a series of tangent planes or lines around the structure.

Thus the attitude of a plane or a line—that is, its orientation in space—is fundamental to the description of structures. We specify the attitudes both of planes and of lines with two angles measured, respectively, from geographic north and from a horizontal plane. The attitude of a plane is specified by its strike and its dip, or by the trend of the dip line and the dip angle. The attitude of a linear feature is given by its trend and its plunge.

The strike is the horizontal angle, measured relative to geographic north, of the horizontal line in a given planar structure (Figure 2.1*A*). This horizontal line is the strike line, and it is defined by the intersection of a horizontal plane with the planar structure. It has a unique orientation for any given orientation of plane

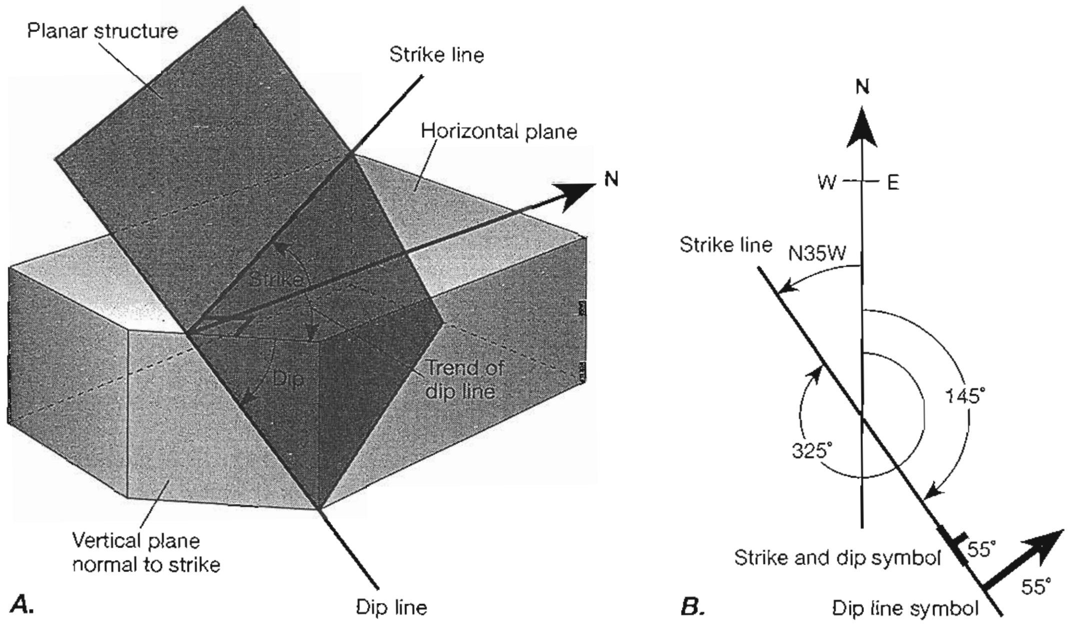


Figure 2.1 The strike and dip of a planar structure. *A.* The strike of a planar structure is the azimuth of a horizontal line iu the plane and is defined by the intersection of a horizontal plane with the planar structure. The dip angle is measured between horizontal and the planar structure in a vertical plane normal ro the strike line. The dip direction in this diagram is ro the NE. *B.* The strike shown here is measured on a quadrant compass as N50W or on a 360° compass as either 325° or 145°. The attitude is indicated on a map by a T-shaped symbol with the stem parallel to the dip direction. The dip angle is written beside this symbol.

except a horizontal one. The **dip** is the slope of a plane defined by the dip angle and the dip direction. The **dip angle**, also referred to simply as the dip, is the angle between a horizontal plane and the planar structure, measured in a vertical plane that is perpendicular to the strike line (Figure 2.1*A*). It is the largest possible angle between the horizontal plane and the inclined plane. For a given strike, a particular value of the dip angle identifies two planes which slope in opposite directions. To distinguish between them, we specify the approximate dip direction by giving the quadrant (NE, SE, SW, NW) or the principal compass direction (N, E, S, W) of the down-dip direction.

The dip line is perpendicular to the strike line and is the direction of steepest descent on the planar structure. In some cases, particularly in British usage, the attitude of a plane is specified by the trend and plunge (defined below) of the dip line. The plunge of the dip line is the same as the dip of the plane.

For the attitude of a linear structure, the trend is the strike of the vertical plane in which the linear structure lies; it is unique except for an exactly vertical linear structure (Figure 2.2*A*). The **plunge** is the angle between the horizontal plane and the linear srructure, measured in the vertical plane (Figure 2.2*A*). The direcrion of plunge, like the dip direction, can be specified by the quadrant or by the principal compass direction of the down-plunge direction. Alternatively, it can be specified implicitly by using the convention that the trend always be measured in the down-plunge direction.

We generally use one of two conventions to record a strike or a trend. We can specify the angle as a **bearing** (the angle measured between 0° and 90° east or west of north or, in some cases, south) or as an **azimuth** (the angle berween 0° and 360°, increasing clockwise from north). Measurements differing by 180° have the same orientation. Thus a northeast strike or trend could be reported as bearings N45E or S45W and as azimuths 045° or 225°, each pair differing by 180°. Using the same conventions, we could give a northwest strike or trend as bearings N45W or S45E and as azimuths 135° or 315°. It is good practice always to write the azimuth as a three-digit number, using preceding zeros where necessary, to distinguish it from dip or plunge angles which are always berween 0° and 90°.

A variety of conventions are in common use for writing the attitudes of planes and lines. We generally write the strike and dip in the order strike, dip, dip direction, regardless of whether bearing or azimuth is used. Strike bearings are always reported relative to north; no distinction is made between azimuths differing by 180°. Thus N35W;55NE, 325;55NE, and 145;55NE

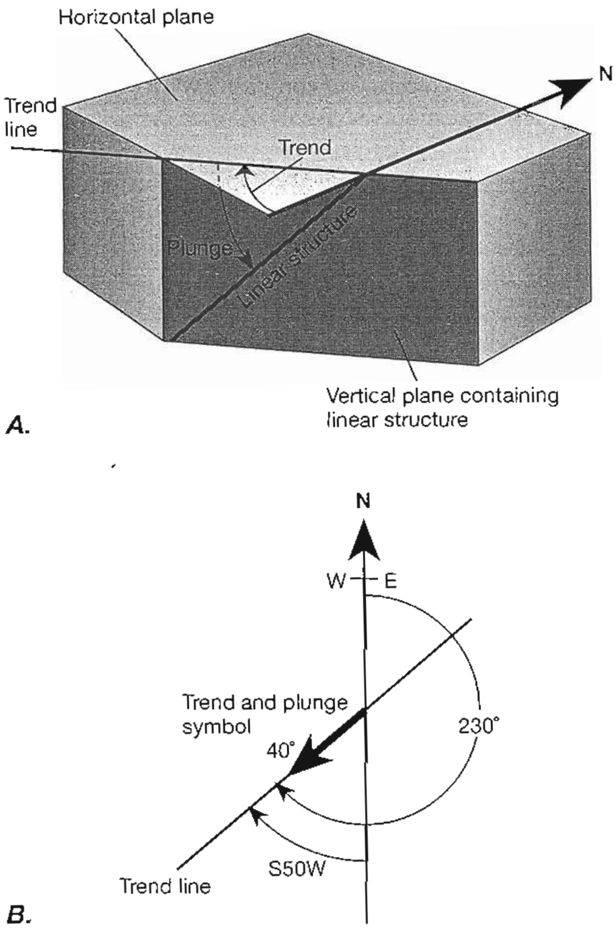


Figure 2.2 The trend and plunge of a linear structure. *A*. The trend is the angle between north and the strike of the vertical plane that contains the linear structure. The plunge is the angle between horizontal and the linear structure, measured in the vertical plane that contains the structure. *B*. The trend shown here is given as N20W on a quadrant compass or as either 340° or 160° on a 360° compass. If by convention the trend is recorded in the down-plunge direction, then it can be given only by S20E or 160°. On a map, linear features are plotted as an arrow parallel to the trend and pointing in the down-plunge direction. The plunge angle is written beside this arrow.

all specify the same attitude of plane (Figure 2.1*B*). The dip direction is always in a quadrant adjacent to that of the strike.

The most convenient convention for the attitudes of lines is to measure the trend in the down-plunge direction, in which case that direction need not be stated explicitly. We first write the plunge and then the trend (the reverse order is also used) so that the measurements 40;S50W and 40;230 both refer to the same attitude (Figure 2.2*B*).

We plot the attitude of a plane on a map by drafting a line parallel to strike with a short bar indicating the down-dip direction. If the trend and plunge of the dip line is measured, we plot the attitude with an arrow pointing in the down-dip direction (Figure 2.1*B*). The attitude of a line is indicated by an arrow pointing in the down-plunge direction (Figure 2.2*B*; the arrow symbols used to indicate the attitude of a line and of the dip line of a plane should be different). In all cases, the value of the dip or plunge, as appropriate, is written beside the symbol so that the attitude can be seen at a glance.

2.2 Geologic Maps

Geologic maps are the basis of all studies of structure and tectonics. They are two-dimensional representations of an area of the Earth's surface on which are plotted a variety of data of geologic interest. These data are based on observations from many outcrops and on the judicious inference of relationships that are not directly observable. The data plotted may include the distribution of the different rock types, the location and nature of the contacts between the rock types, and the location and attitude of structural features.

Contacts are lines on the topographic surface of the Earth where the boundary surface between two different rock types intersects the topography. Contacts can be **stratigraphic**, where one unit lies depositionally upon another; **tectonic**, where the units are faulted against one another; or **intrusive**, where one unit invades another. On a geologic map, different types of contacts and the reliability of the contact location are indicated by different styles of lines. The shape of the contact on the map depends both on the geometry of the contact surface (for example, whether it is planar or folded and what its attitude is) and on the topography that the contact surface intersects. An experienced observer can determine the geometry of structures in an area, as well as the quality of information available, simply by careful inspection of a good geologic map.

All geologic maps are smaller than the area they represent. Exactly how much smaller is represented by the **scale** of the map, which is the ratio of the distance on the map to the equivalent distance on the ground. A scale of 1:25,000 (one to twenty-five thousand), for example, indicates that 1 unit of distance on the map (such as a centimeter or an inch) represents a horizontal distance of 25,000 of the same unit on the ground. Because the scale is a ratio, it applies to any desired unit of measurement. The scales of most maps used in structural and tectonic work range between 1:1000 and 1:100,000, though other scales are also used. In particular, maps of large regions such as states, provinces, countries, and continents are published at scales between 1:500,000 and 1:20,000,000.

As the scale of a map changes, the size of the features and the amount of detail that can be represented on the map also change. If a map is of a very small region (an area a few meters in dimension for example), then correspondingly small features and great detail can be portrayed. If the map represents a large region (an area hundreds of kilometers in dimension for example), then only large features and little detail can be shown.

Unfortunately, the word *scale* is used in two different and opposite ways, which can lead to confusion. With regard to geologic features, the term refers to the dimensions of the feature. Thus small-scale features have a characteristic dimension roughly in the range of centimeters to perhaps hundreds of meters. Large-scale features have a characteristic dimension of roughly hundreds of meters to thousands of kilometers. With regard to maps, however, *scale* refers to the distance on a map divided by the equivalent distance on the ground. Thus small-scale maps (such as 1:100,000) cover larger areas than large-scale maps (such as 1:1000). Confusion arises because large-scale features are portrayed on small-scale maps, and vice versa.

Geologic mapping is usually done on an accurate topographic base map. Most countries publish topographic maps at a scale between 1:25,000 and 1:50,000. If an area is very small, however, more detailed base maps must generally be prepared by using surveying or photographic techniques.

Because the Earth's surface is very nearly spherical, any planar map of the surface is a distortion of true shape. For small areas, even up to standard 1:24,000 or 1:25,000 quadrangles (approximately 17 km in a north–south direction and in midlatitudes 11 km to 15 km east to west), this distortion is minor and usually is ignored. For larger regions, however, distortions become significant. Many different types of projections of the spherical surface onto a plane are used. Each represents a compromise between minimizing the distortion of the shape of the region and minimizing the distortion of its area.

Small-scale regional maps are useful in structure and tectonics in order to portray large-scale structural features. Such maps are constructed by combining the large-scale maps on which the geology was originally mapped into a single map, usually at a smaller scale. Thus it is common to see features originally mapped at a scale of 1:24,000, for example, compiled on a map at 1:250,000, 1:1,000,000, or even 1:10,000,000 covering an entire continent. As the map scale decreases, the amount of detail must also decrease. Such compilations therefore require that many choices be made about what information to represent and what to omit, and the resulting map is highly interpretive. Often the information available from adjacent map sheets is not entirely consistent, and the compiler must contend with problems such as different levels of detail on different maps, disagreements on the nature of map units, and the inconsistent location of contacts. In cases where such discrepancies cannot be resolved, discontinuities may appear on the smaller-scale regional map.

Other differences among geologic or tectonic maps of a particular area may be due to the particular purpose for which each map is made. A map of soil and surficial deposits would look very different from a map of the bedrock in the same area. Maps may also be compiled to emphasize various aspects of geology. A geologic map emphasizes the distribution of lithologies and their ages, whereas a tectonic map of the same area combines units of similar tectonic significance.

2.3 Cross Sections: Portrayal of Structures in Three Dimensions

A geologic map provides the basis for detailed understanding of the structural geometry of an area. The map, however, is only a two-dimensional representation of three-dimensional structures. Cross sections, on the other hand, show the variation of structure with depth, usually as it would appear on a vertical plane that cuts across the area of a geologic map. Fundamentally, cross sections are interpretations of structure at depth extrapolated from data available at the surface, but they may also be constrained by data on the regional stratigraphy or lithology, by direct observation from drill holes or mines, and by geophysical data (see Section 2.6). Without such independent constraints, cross sections are *highly interpretive,* because they are based on the assumption that the structure at depth is a simple projection of the structure observed at the surface and the attitudes and geometry do not change along the line of projection. The validity of that assumption varies a great deal, depending on the characteristics of the local geology.

Cross sections ideally are oriented normal to the dominant strike of planar structures in an area. In this orientation, the dip of those structures is accurately represented on the section. In some cases, however, if the structure is complicated with a variety of attitudes, no one orientation of cross section adequately represents all the structures, and the apparent dip of any plane whose strike is not perpendicular to the cross-section line is less than the true dip of that structure.

In order to show an undistorted view of structures at depth, we must take the vertical scale ratio equal to the horizontal scale ratio. In some cases, however, the features of interest are best shown by using vertical exaggeration, for which the vertical scale ratio is larger than the horizontal scale ratio. The relatively small changes in topography and in stratigraphic thickness that commonly occur over large distances can be shown clearly only with vertical exaggeration. As a result, vertically exaggerated cross sections are standard in marine geology and stratigraphy.

The habitual use of vertical exaggeration to portray certain features, however, gives a false impression of the nature of those features. Figure 2.3, for example, shows cross sections of a subduction zone and an adjacent volcanic island arc. The true scale section is shown in Figure 2.3*A*. Note that the topographic variation is almost impossible to see! A properly vertically exaggerated section is shown in Figure 2.3*B*. The vertical exaggeration necessary to emphasize the topography also exaggerates the surface slopes so that their appearance is not at all representative of actual slopes. The effect of vertical exaggeration is just as dramatic on the dip of planar features (compare the angle of the subducting lithospheric slab in Figure 2.3*A* with that in Figure 2.3*B*). Published informal cross sections often give an even more confusing view by combining vertically exaggerated topography with no vertical exaggeration below the surface, as shown in Figure 2.3*C*.

Vertical exaggeration makes dipping structures appear much steeper than they are. The effect is much stronger on shallowly dipping planes than on steeply dipping ones, and at high values of the vertical exaggeration, the distinction between shallow and steep true dips effectively disappears.

Another effect of vertical exaggeration is to cause beds of the same thickness but different dip to appear to differ in thickness. In Figure 2.3*B* the subducted lithospheric slab appears to be thinner where it is dipping than where it is horizontal, an effect caused simply by exaggeration of the vertical dimension.

Thus vertical exaggeration causes distortions of the geometry of features that seriously alter the way they look. Because much of structural geology involves visualizing the true shape of features in three dimensions, the use of vertical exaggeration with structural cross sections should be avoided whenever possible.

In areas of complex structure where a single cross section cannot be constructed to reveal the true angular

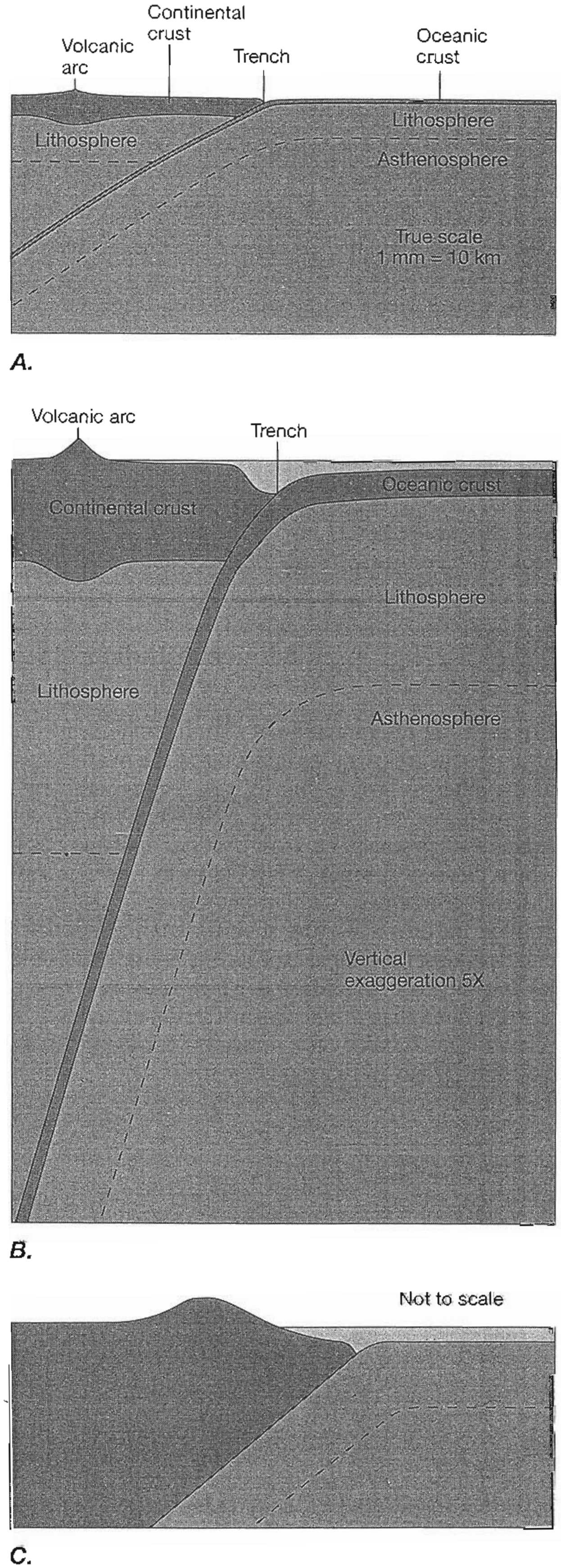


Figure 2.3 (*Right*) The effect of vertical exaggeration on cross sections. *A*. True-scale cross section of a subduction zone in which the horizontal and vertical scales are equal (1 mm = 10 km). Note that topographic variations are almost imperceptible. *B*. A cross section of the same subduction zone as in part *A*, here shown at a vertical exaggeration of 5×. That is, the vertical scale is 5 times the horizontal scale. Note the exaggeration of the dip of the subducted slab. *C*. Schematic cross section that combines vertical exaggeration for the topography with no vertical exaggeration for structures below the surface. This mixing of vertical scales precludes the construction of an accurate cross section.

relationships of all the structures, multiple cross sections at different orientations may be used. Two constructions are regularly employed—block diagrams and fence diagrams. Block diagrams show a perspective diagram of a block of crust with appropriate geometric representation of the structure on the top, which is the map view, and a view of the sides, which includes two cross sections generally at high angles to each other. We have already used a number of block diagrams to illustrate structural features (see Figures 2.1*A* and 2.2*A*). Although these views of the structure on three different surfaces enable us to visualize the three-dimensional geometry, sections themselves are not accurate representations of the angular relationships because of the necessary distortion of perspective. Fence diagrams represent three-dimensional structures by showing a perspective view of intersecting cross sections. They are difficult to draw and require a great deal of information; thus they are relatively rare in the published structural geologic literature.

2.4 Stratigraphic Sequence Indicators

Crucial to the interpretation of a stratigraphic sequence of deformed rocks is determination of the relative ages of the different layers—that is, the "stratigraphic up" or "younging" direction in the sequence. Consider, for example, the schematic cross sections in Figure 2.4, which illustrate a series of layered rocks folded in two different geometries. Without knowledge of the stratigraphic up directions, the simplest interpretation we could make of the limited exposure in Figure 2.4*B* would be that shown in Figure 2.4*A*, which is incorrect.

Features that are useful in the determination of relative age include some primary structures in sedimentary or igneous rocks, unconformities, and an independent knowledge of the stratigraphic sequence.

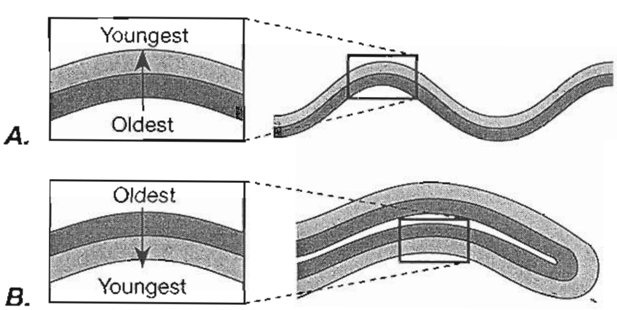


Figure 2.4 Cross sections showing "stratigraphic up" directions. *A*. A fold with beds right side up, as indicated by the arrow. The structure is simple, as shown schematically to the right. *B*. A fold with upside-down beds, as indicated by the arrow. The structure must be more complex, as shown to right.

Primary Structures in Sedimentary Rocks

Many structures formed during or shortly after sedimentation are useful in determining relative stratigraphic age. The most common of these structures are bottom markings, graded bedding, cross bedding, and scour-and-fill or channel structures.

Bottom markings are features formed mainly in association with turbidity currents and preserved on the underside of many sandstone beds in interlayered sandstone–shale sequences. They represent casts of the small-scale surface topography imposed on mud by an overlying sand layer, or upon which the sand layer was deposited. Bottom markings include flute casts and load casts. Flute casts form from the deposition of sand in spoon-shaped depressions scoured out of the underlying mud by high-velocity currents. Subsequent lithification of the sand preserves a cast of the depression on the bottom of the sandstone layer (Figure 2.5*A*, *B*). Flute casts tend to be markedly asymmetric in longitudinal cross section; they indicate the direction in which the

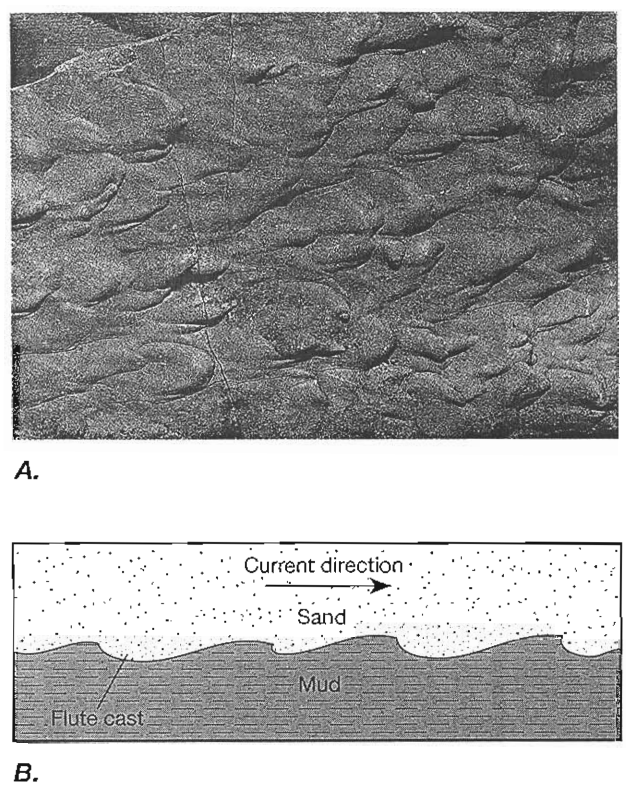


Figure 2.5 Flute casts. *A*. Flute casts on the bottom of a sandstone bed in a turbidite. *B*. Cross section of flute casts, shown right side up.

current was moving at the time the scours were cut (Figure 2.5*B*).

Load casts are sandstone casts of depressions in underlying mud that form as the denser sand sinks into the soft, water-saturated mud. They appear as bulges on the bottoms of sandstone layers and, in some cases, can actually become isolated balls of sand in a mud matrix. If the mud forms sharp peaks pointing upward into the sand, and the sand forms rounded masses protruding down into the mud, the structures are described as flame structures which result from the down-slope shearing of the sediments. The directionality of these features serves as an indicator of the relative age of the sediment layers.

Graded bedding in sediments or sedimentary rocks is characterized by the gradual change in grain size and mineralogy within a single bed, usually from coarse and clay-poor at the bottom to fine and clay-rich at the top (see Figure 2.6). Grading occurs in a suspension of unsorted sediment because the coarsest fraction settles out faster than the finest fraction. Thus the oldest part of the layer is the coarsest, and the youngest is the finest. In some cases, however, metamorphism can actually reverse the direction of grading. Because the finest fraction has a higher surface area and is therefore more chemically reactive, that part of the layer may grow coarser crysrals during the metamorphism than the originally coarser part of the layer.

Cross bedding consists of thin beds that occur at angles of as much as 20° to 30° to the principal bedding planes. These beds result from deposition of material in ripples or dunes. Characteristically, the cross beds are concave upward, and they are tangent to the lower major bedding surface. Where subsequent erosion has removed the top of the bed, the crossbeds are truncated against the upper major bedding surface. In this case, they are useful as indicators of the stratigraphic up direction.

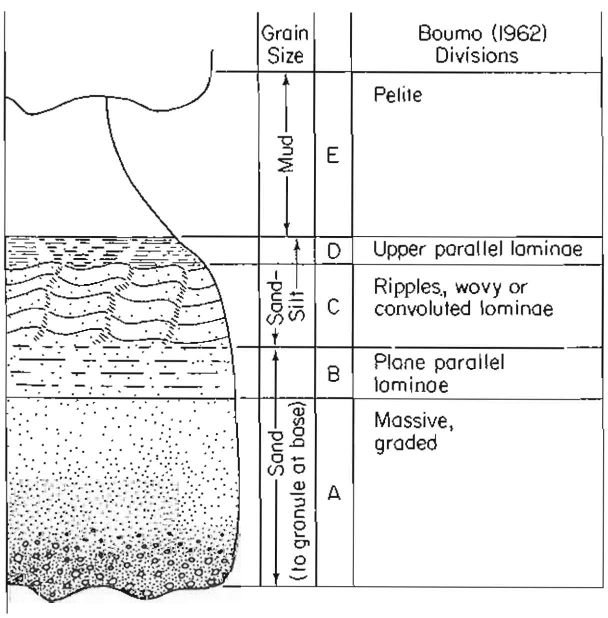


Figure 2.6 Ideal graded structure of a turbidite bed, showing grading, ripples, and upper mud-rich layer (pelite).

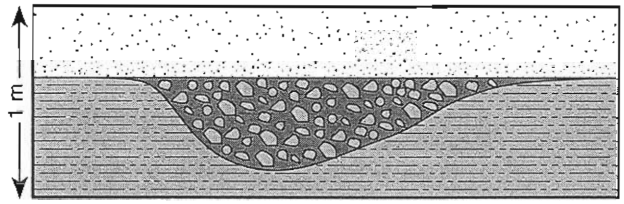


Figure 2.7 Sketch illustrating a cross section of channel or scour-and-fill structure. The depression was eroded in the lower unit (siltstone) and first conglomerate, and then sand deposited on top.

Channel structures, or scour-and-**fill structures**, are deposits that fill in stream or current channels cut by erosion into underlying sediment. Most channel deposits are conglomerates, but finer-grained sediments are also found. These structures can be identified by the characteristic truncation of layers of the underlying sediment against the side of the channel (Figure 2.7). Identifying which layer has been eroded, and which subsequently deposited, makes it possible to determine the stratigraphic up direction.

Primary Structures in Igneous Rocks

Structures that indicate the stratigraphic up direction in igneous rocks are more abundant in extrusive than in intrusive rocks, but they are not so common as in sedimentary rocks. Telltale features in extrusive rocks include flow-top breccia, pillow lava structures, and filled cavities.

Flow-top breccia develops at the upper surface of basalt flows, thereby making it possible to identify the original top of the layer (Figure 2.8). It forms by the breakup of a solidified layer during renewed movement of the underlying lava. In some cases, the breccia formed on top can roll under the flow as it advances. Thus caution is advisable in the use of this indicator.

Many volcanic rocks contain vesicles—cavities formed by solidification around gas bubbles (Figure 2.8). Vesicles may be filled by minerals precipared from solution, in which case they are called amygdules (the Greek word *amygdalon* means "almond"). In rare instances, sediment or late magma may be deposited in the vesicles. Because the cavities fill from the bottom up, partially filled amygdules provide an indicator of

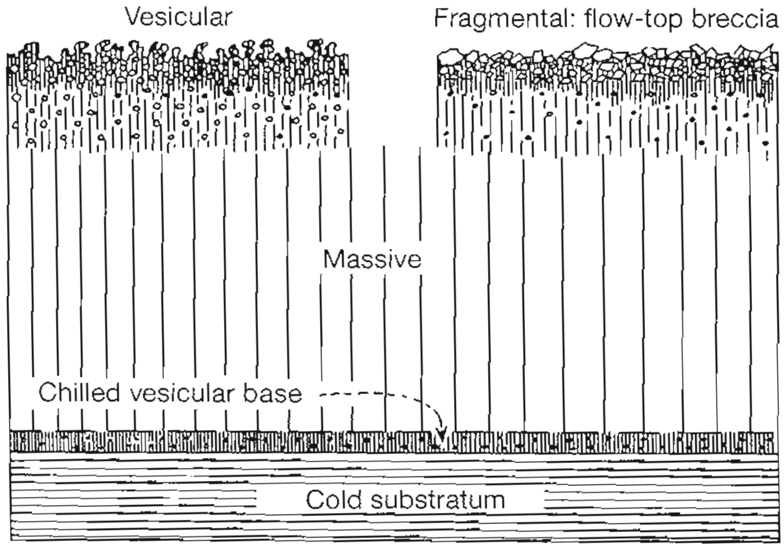


Figure 2.8 Diagram of top and bottom of subaerial lava flow. Vesicular lava may be found at the top or at the bottom of a flow. Flow-top breccia forms on top by solidification and subsequent breakup of the rop during flow.

the stratigraphic up direction, as do composite amygdules if the filling sequence is known.

Pillow lavas form during extrusion of lavas under water. Lava often issues from a central vent at the top of a mound and flows downward onto the horizontal floor. Because the hot lava is quenched by the water, it flows by forming tubelike fingers that have a chilled exterior which insulates the lava within. The bottom of the top is convex upward. In cross section many tubes have a distinctive pillow-shaped upper surface and a downward pointing cusp on the lower surface (Figure 2.9), thereby indicating the original up direction.

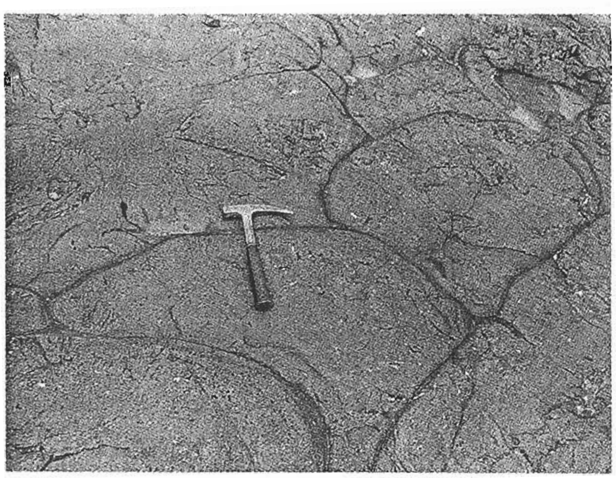


Figure 2.9 Pillow lavas in cross section. Smartville complex, Sierra Nevada, California.

Structures that reveal the original up direction are much less common in plutonic rocks than in volcanic rocks or sediments. In rare cases, however, plutonic rocks form by crystal settling at the bottom of a magma chamber, and sedimentary structures may develop that can be used to infer the stratigraphic up direction. Size grading of grains of a single mineral in a layer can be used as a stratigraphic up indicator (Figure 2.10*A*). The existence of gradations between two layers of different mineralogy, or phase layering, is more common than size grading (Figure 2.10*B*). It probably does not result from crystal settling, however, so cannot be used as an indicator of the stratigraphic up direction. Scour-and-fill structures are present in some layered igneous rocks. As in sedimentary rocks, they indicate the presence of currents during deposition of the rocks and are a reliable indicator of the stratigraphic up direction.

In some areas, magmas with a significant proportion of suspended crystals have intruded or extruded as sills or flows. As the molten material came to rest, the crystals settled, forming crystal accumularions toward the bottom of the sill or flow.

Unconformities

Unconformities provide a valuable means of determining relative age in a stratigraphic section. Unconformities may be disconformities, which are time gaps within a sequence of parallel layers (Figure 2.11*A*); angular unconformities, which are erosional surfaces that cut across older beds at an angle and are overlain by parallel beds (Figure 2.11*B*); or nonconformities, which are con-

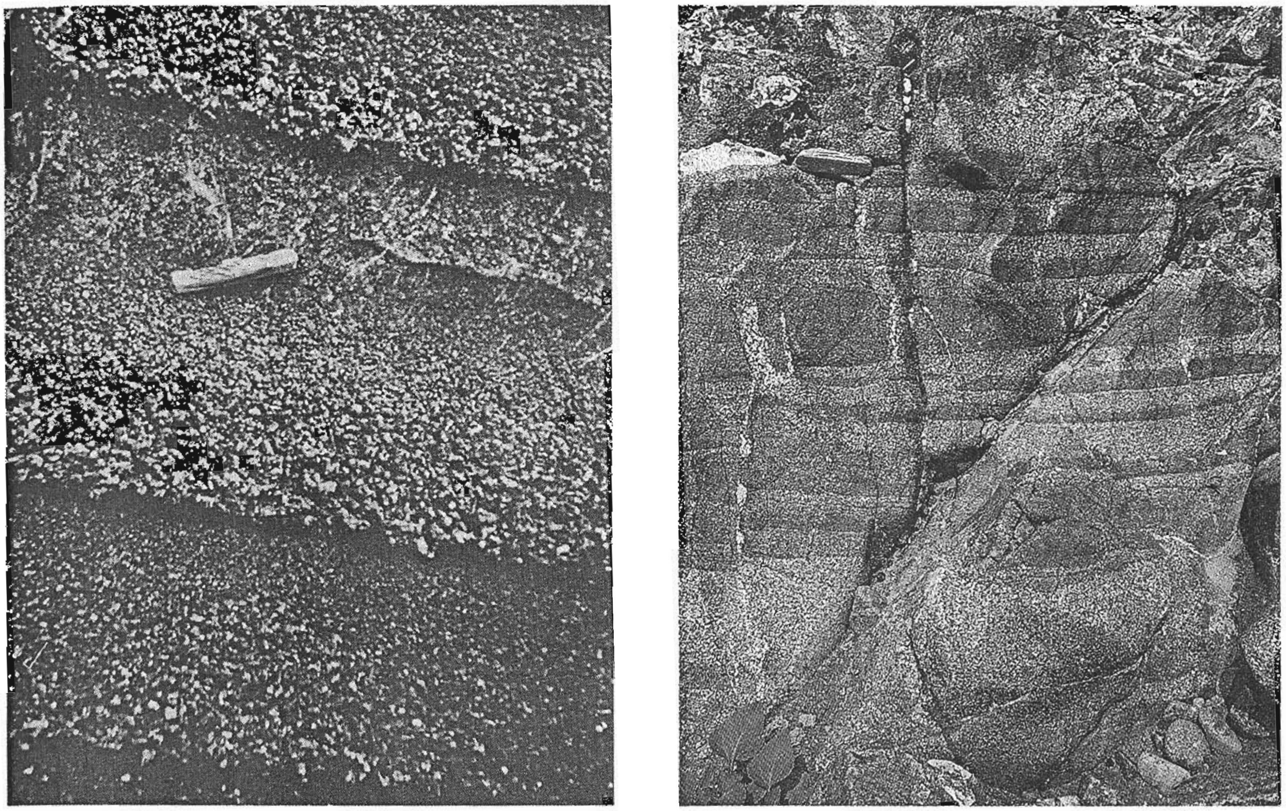

A. B.

Figure 2.10 Size and phase grading in plutonic igneous rocks. *A*. Size grading in olivine–clinopyroxene cumulate, Duke Island ultramafic complex. Pocket knife for scale. *B*. Photograph showing repeated graded phase layering in gabbro, Vourinos ophiolite complex, northern Greece. This gradational alteration of pyroxene-rich (dark) and plagioclase-rich (light) layers may not be the result of gravity settling.

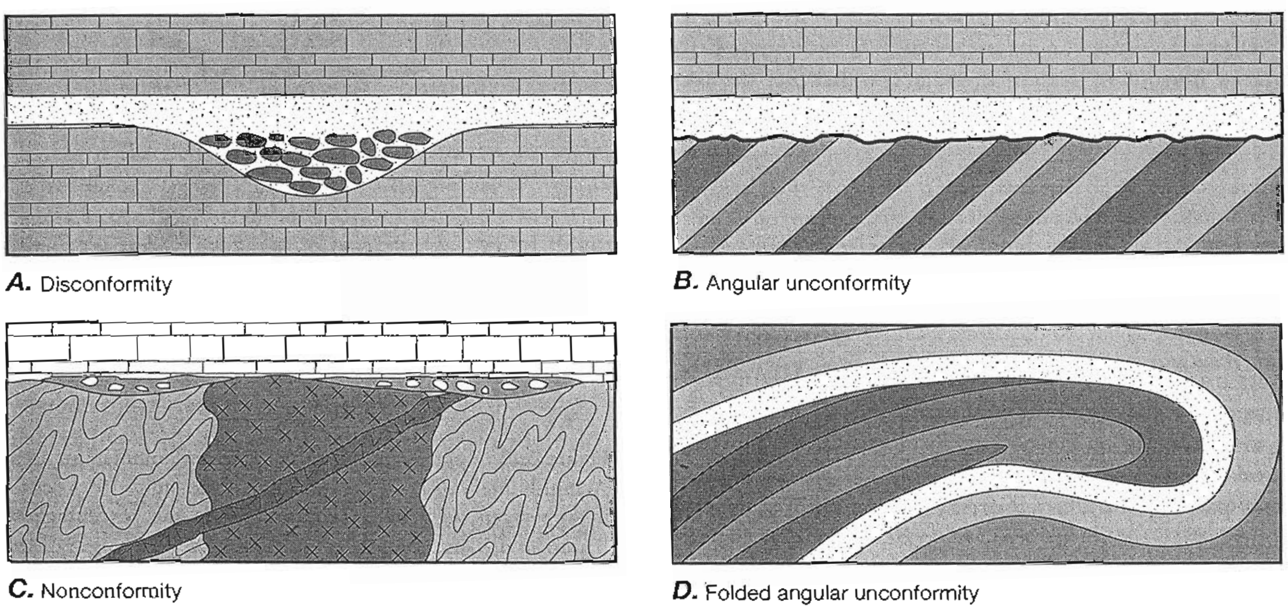


Figure 2.11 Types of unconformities. Parts *A–C* are undeformed.

tacts between sedimentary rocks and underlying igneous or metamorphic rocks (Figure 2.11*C*). Figure 2.11*D* shows how an angular unconformity might appear when folded into a large fold. Note the lack of parallelism of the beds below and above the unconformity.

Some unconformities display fossil soil horizons. If one is present, it provides a means of establishing the relative stratigraphic age. Such a horizon also makes it easy to distinguish between a sedimentary unconformity and a tectonic contact such as a fault; rbese can be difficult to tell apart, especially if the tocks have been deformed. Even in some metamorphosed rocks, fossil soils are sometimes preserved as a thin band of aluminum-rich metamorphic rocks sandwiched between a metasedimentary sequence and an older metamorphic or igneous terrane.

Other Indicators

Many orogenic zones include belts of rocks characterized by deformed, but little metamorphosed, fossiliferous sediments. In these rocks, it is possible to determine the stratigraphic sequence by means of biostratigraphic analysis of the sediments themselves. In highly deformed and/or metamorphosed areas, it may be necessary to infer stratigraphic relationships from the stratigraphy in less deformed or metamorphosed areas.

In some cases the rocks are unfossiliferous, the stratigraphy is unknown, and the rocks do not possess any structures that indicate the relative age. In such situations, radiometric age determinations may yield the only age information available. Sedimentary processes do not reset radiometric clocks, so sediments cannot be dated directly this way. Metamorphic or igneous events can be dated, and the age of formation of sttuctural or tectonic features can be established if radiometrically dated igneous or metamorphic events bracket the tectonic event in time.

2.5 Graphical Presentation of Orientation Data

Often it is desirable to present orientation data in such a way that the distribution of orientations is emphasized independently of the geographic location of the data. For example, it may be useful to know whether there is a pattern of preferred orientation of beds, joints, or linear features in an area, regardless of how the orientations vary across a map. The types of diagrams most frequently used to present such information are histograms, rose diagrams, and spherical projections.

Orientation histograms (in Greek, *histos* means "mast" or "web," and *gram* means "line") ate plots of one part of the orientation data, such as strike azimuth, against the frequency of orientations that are found within particular orientation intetvals. The frequency may be plotted as a percent of all observarions or as the number of observations witbin each interval. The plots characteristically consist of a series of rectangles, where the width of the rectangle represents the orientation interval and its height represents the frequency.

Rose diagrams are essentially histograms for which the orientation axis is transformed into a circle to give a true angular plot. The intervals of angle are plotted as pie-shaped segments of a circle in their true orientation, and the length of the radius is proportional to the frequency of that orientation. The use of the true angle conveys an intuitive sense of the orientation distribution. Rose diagrams are used for displaying such features as the direction of sediment transport and the strike of vertical joints (see, for example, Figure 2.12).

Both histograms and rose diagrams can present only one aspect of the attitude of planar or linear fea-

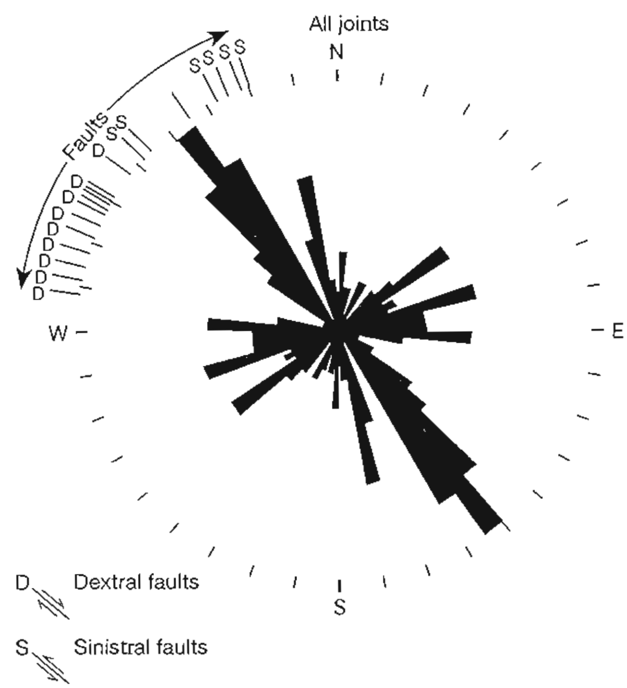


Figure 2.12 Rose diagram of the faults and joints in an area. Azimuth intervals are 5°. The length of the radius from center in any given segmenr is proportional to rhe percentage of features with an orientarion in thar sector. The length of the longest radius represents 40 percent of the measurements. The diagram has a twofold axis of symmetry; rhat is, a 180° rotation of the diagram is identical to rhe original diagram.

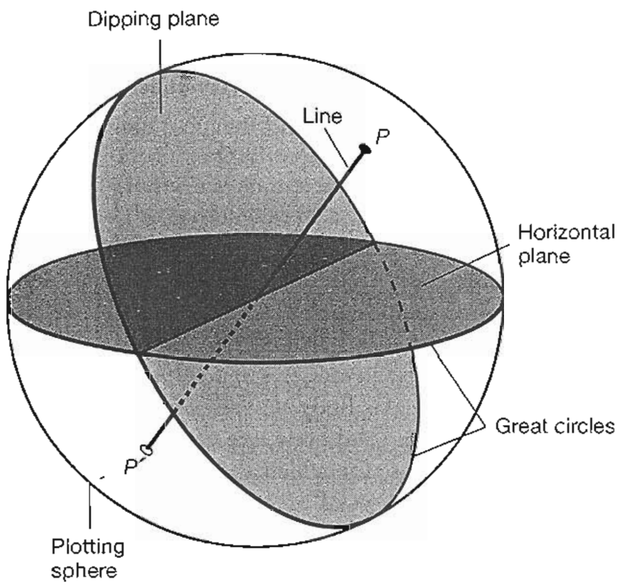


Figure 2.13 Plotting orientation data on a plotting sphere. Planar and linear features are considered to pass through the center of the sphere and to intersect its surface. Thus the attitude of a plane is defined by the great circle that is the intersection of the plane with the plotting sphere. A horizontal plane and a dipping plane are shown. The attitude of a line is defined by two poinrs (*P* and *P'*) that are the intersections of the line with the plotting sphere.

tures, such as the strike or bearing, respectively. In cases where the dip of the planar features is always essentially vertical or the plunge of linear features is always essentially horizontal, or some other constant angle, this limitation is of no consequence. If the dip or plunge of the feature is an important variable in defining its attitude, however, the best method of plotting orientation data is the spherical projection.

When orientation data are plotted on a spherical projection, all planes and lines are considered to pass through the center of the plotting sphere, and their attitudes are then defined by their intersections with the surface of the sphere. For planes, the intersection is a great circle; for lines, the intersection is a point (Figure 2.13). Using the full sphere is actually redundant; a hemisphere is all that is needed. Applications for structural geology and tectonics generally use the hemisphere below the horizontal plane (the lower hemisphere), whereas applications in mineralogy usually employ the upper hemisphere. In both cases, the hemisphere is projected onto a flat plane to permit the convenient graphical presentation of data.

There are a variety of methods of projecting a sphere onto a plane, although two projections are most often encountered in structural geology and tectonics. The first may be referred to as a stereographic projection or as an equal-angle projection; the other is a Lambert projection or an equal-area projection. They differ only in the way the hemisphere is projected onto a plane called the image plane.

For both types of projections, the image plane for the projection is tangent to the hemisphere at *T* (Figure 2.14) and parallel to the plane containing the edge of the hemisphere. The equal-angle projection (Figure 2.14*A*) is constructed by using the highest point on the sphere opposite the image plane, the zenith point *Z*, as the projection point. The projection of a point on the hemisphere is defined by constructing a line from the zenith point *Z* through the point on the hemisphere (*P* or *Q*) to the image plane (*P'* or *Q'*). Thus any orientation of line has a unique projection on the image plane. A great circle on the hemisphere is projected by drawing

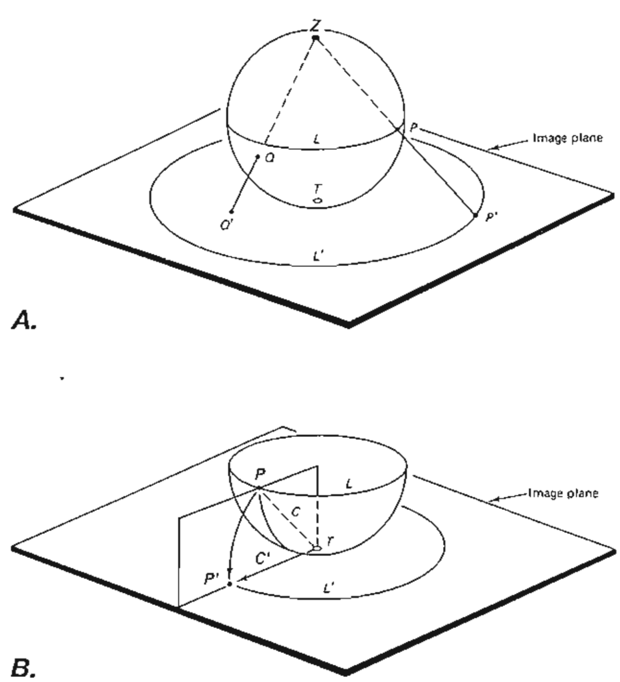


Figure 2.14 Projection of the lower half of the plotting sphere onto the image plane. *A*. The principle of stereographic, or equal-angle, projection. The zenith point *Z* is the projection point. Points *P* and *Q* on the plotting hemisphere are projected to points *P'* and *Q'* on the image plane by a line passing from *Z* through *P* and *Q*, respectively, to the image plane. A great circle is described on the image plane by projecting each point of the great circle on the plotting hemisphere to the image plane. *B*. The principle of Lambert, or equal-area, projection. The point *P* is projected to the image plane by constructing a cord of the sphere from T (the tangent point of the image plane) to *P* and rotating that cord in a vertical plane about *T* down to C' in the image plane. The end of the rotated cord *P'* is the projected point. The great circle is projected by rotating each point of the great circle on the plotting hemisphere down to the image plane in a similar manner.

lines from the zenith point through all points along that great circle. The locus of intersections of those lines with the image plane defines the projection, which again is unique for any given orientation of plane.

The advantage of this type of projection is that angles between lines on the hemisphere are not distorted by the projection. Moreover, circles on the hemisphere remain circles on the projection, although the center of the circle on the hemisphere does not project to the center of the circle on the image plane. All great circles are also arcs of circles on the image plane. Areas that are equal on the hemisphere, however, are in general not equal on the image plane.

An equal-area projection is constructed by using the point of tangency T between the hemisphere and the image plane as a center of rotation (Figure 2.14*B*). The projection of a point on the hemisphere is determined by constructing a chord C from T to the point P on the hemisphere and rotating this chord in a vertieal plane about T down to C' in the image plane. The end of the chord P' in the image plane defines the projection of the point. Projections of great circles are constructed, in principle, by rotating all points on the great circle from the hemisphere down to the image plane in a similar manner. This projection has the advantage that any two different but equal areas on the hemisphere are also equal on the image plane. It is therefore used to present data when the statistical concentration of points is important to the interpretation, because those concentrations are not distorted by the projeetion. The shapes of areas on the hemisphere are not preserved by the projection, however, so angular relationships are distorted, although they can still be determined if the angles are measured along a great circle.

2.6 Geophysical Techniques

Although mapping rocks that are exposed at the surface provides good information about the three-dimensional structure near the surface, it cannot reveal the structure of areas covered by alluvium, deep soils, vegetation, or water such as lakes, seas, and oceans. Nor can surface mapping provide information about structure at great depth. Information about the shapes of major faults at depth, the presence of magma chambers at depth, the location of the crust–mantle boundary, or the thickness and nature of the lithosphere and the lower mantle can come only from the interpretation of geophysical measurements, and especially from seismic, gravity, and magnetic measurements. We review briefly the application of these aspeets of geophysics to large-scale structure and tectonics because they have become essential, and because a structural geologist must at least be aware of the techniques and their limitations. Adequate coverage of these topics, however, would require at least a separate book, and we encourage students to take appropriate courses in geophysics.

Seismic Studies

Seismic waves are oscillations of elastic deformation that propagate away from a source. Waves from large sources such as major earthquakes and nuclear explosions can be detected all around the world. Small explosions are often used as sources to investigate structure at a more local scale. Body waves, which can travel anywhere through a solid body, are of two kinds: compressional (P) waves, for which the particle motion is parallel to the direction of propagation, and shear (S) waves, for which the particle motion is normal to the direction of propagation. (The designations "P" and "S" refer to primary and secondary waves, so named beeause of the normal sequence of arrival of the waves as revealed on a seismogram.) P waves travel faster than S waves. They are therefore the first waves to arrive at a detector from a source and thus are the easier to recognize and measure. For this reason, and because explosions generate mostly P waves, they are the waves predominantly used to investigate the structure of the Earth. The propagation of seismic waves may be described by the seismic rays, which are lines everywhere perpendicular to the seismic wave fronts. Three types of seismic studies are particularly important in structure and tectonics: seismic refraction, seismic reflection, and first-motion studies.

Seismic refraction studies investigate the structure of the Earth by means of those seismic rays that are transmitted through boundaries at which seismie velocity changes. The change in seismic velocity refracts, or bends, the rays (see Box 2.1) so that in the Earth, where seismic veloeity generally increases with depth, the ray paths tend to be concave upward. The travel time of rays from the source to different receivers is plotted against the distances of the reeeivers from the source. Such plots make it possible to determine the ray velocity in the deepest layer through which the ray travels. By measuring travel times for rays that penetrate to greater and greater depths, the investigator can determine the veloeity structure of the Earth.

The velocities of P and S waves depend on the density and the elastic eonstants of the rock. Thus knowing how seismic P and S velocities vary with depth provides information about the distribution of the density and elastic properties in the Earth, and locating where changes in seismic velocity occur reveals where the rock type changes.

Although seismic refraction studies give a good "reconnaissance" view of the structure of a large area,

Box 2.1 Seismic Refraction

The time required for seismic rays to travel directly from a source to different detection stations distributed around the source is affected by the particular paths the seismic rays take, and these in turn are determined by the structure and the seismic velocity of the material along each path. If a seismic ray travels obliquely across a boundary from a low- to high-seismic-velocity material, it is refracted away from the normal to the boundary (Figure 2.1.1). If the ray travels from high- to low-velocity material, it is refracted toward the normal to the boundary.

Travel-time measurements can be interpreted to reveal the variation of seismic wave velocity with depth. The principle is illustrated in Figure 2.1.2, which shows the location of a seismic source and an array of detectors. Some of the ray paths shown stay within the crust; others travel in part through the upper mantle. A time–distance plot indicates the arrival times of those different rays at the detectors. Because the seismic velocity in the mantle is higher than that in the crust, mantle rays reach distant detectors before crustal rays. For the layered structure shown, the difference in arrival times at the different detectors reflects the speed of the rays through the deepest layer along the ray path. Thus the slopes of the two lines on the time–distance plot are the inverse of the velocities in the crust and mantle, respectively.

If a layer that has a lower seismic velocity occurs at depth between rocks that have higher seismic velocities, rays are bent toward the normal to the boundary upon entering the layer and away from the normal upon leaving (Figure 2.1.1). Seismic rays, therefore, can never reach their maximum depth in that layer, and the seismic velocity of that layer—and therefore its very existence—cannot be detected on a time–distance plot.

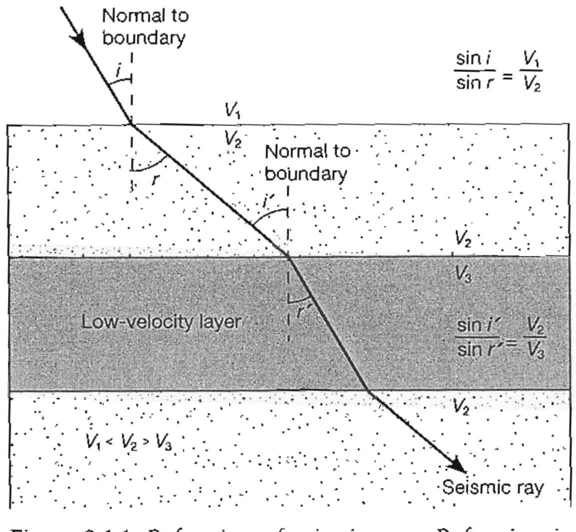


Figure 2.1.1 Refraction of seismic rays. Refraction is away from the normal to the boundary if the ray travels from a low to a high-seismic-velocity material (here V_1 to V_2 or V_3 to V_2). Refraction is toward the normal to the boundary if the ray travels from a high- to a low-seismic-velocity material (V_2 to V_3).

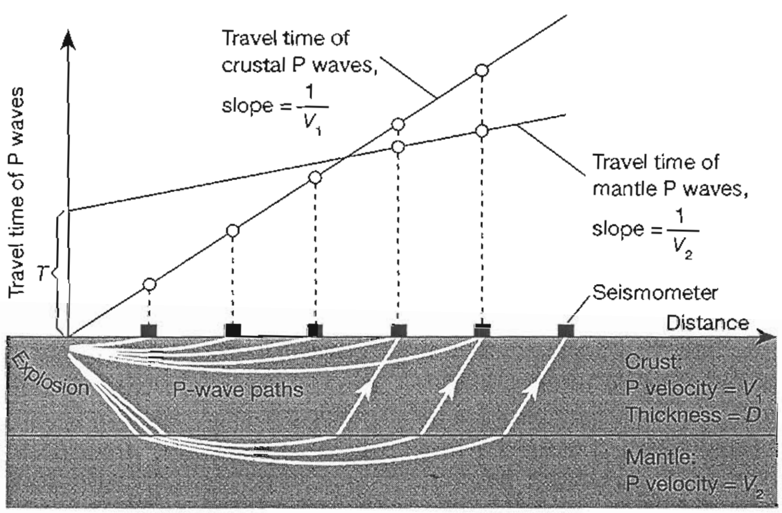
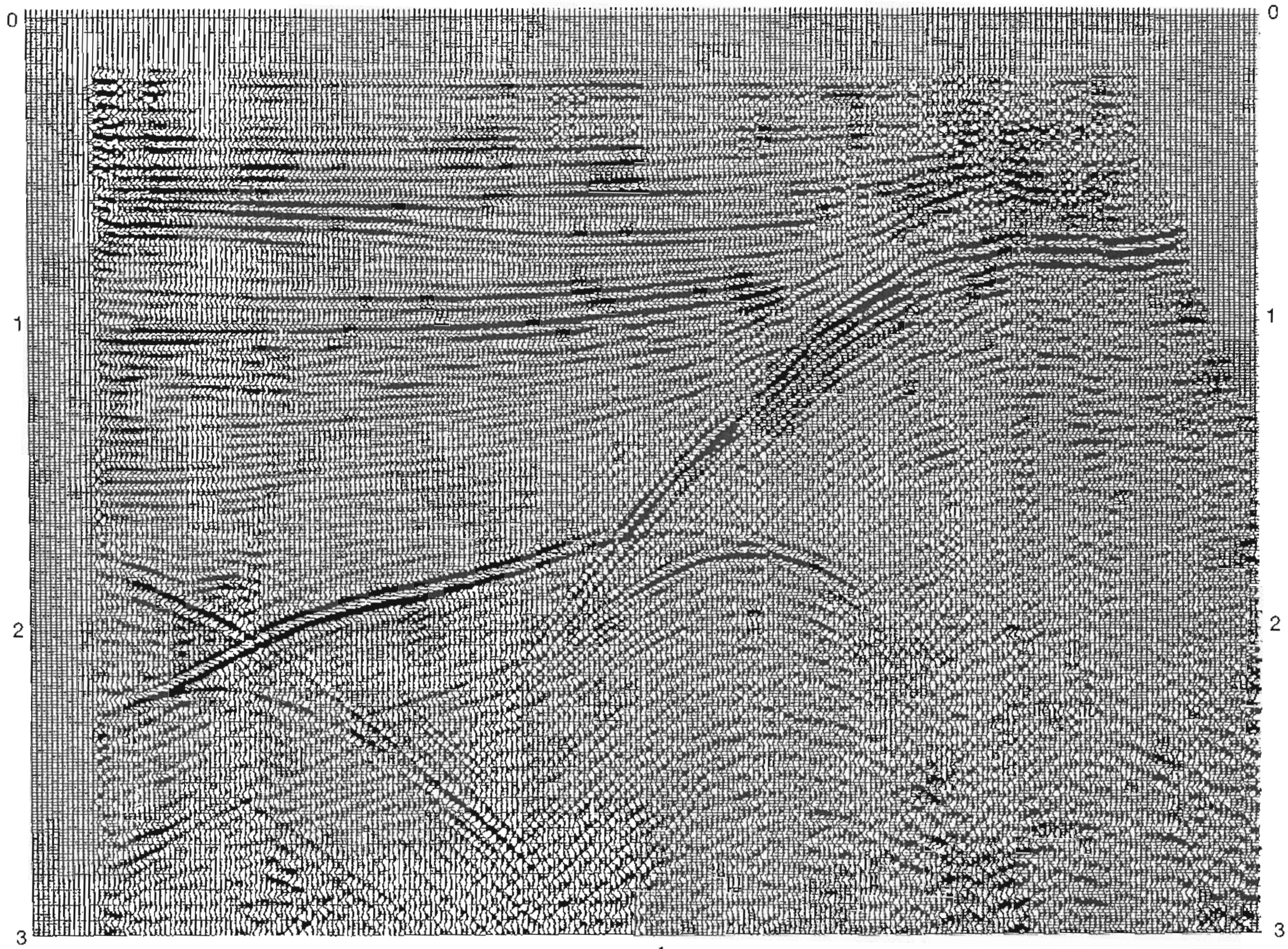


Figure 2.1.2 Illustration of the principle of seismic refraction in a two-layer structure. The diagram shows ray paths for P waves through the structure. The travel-time plot indicates the arrival times of the rays at the different detectors.



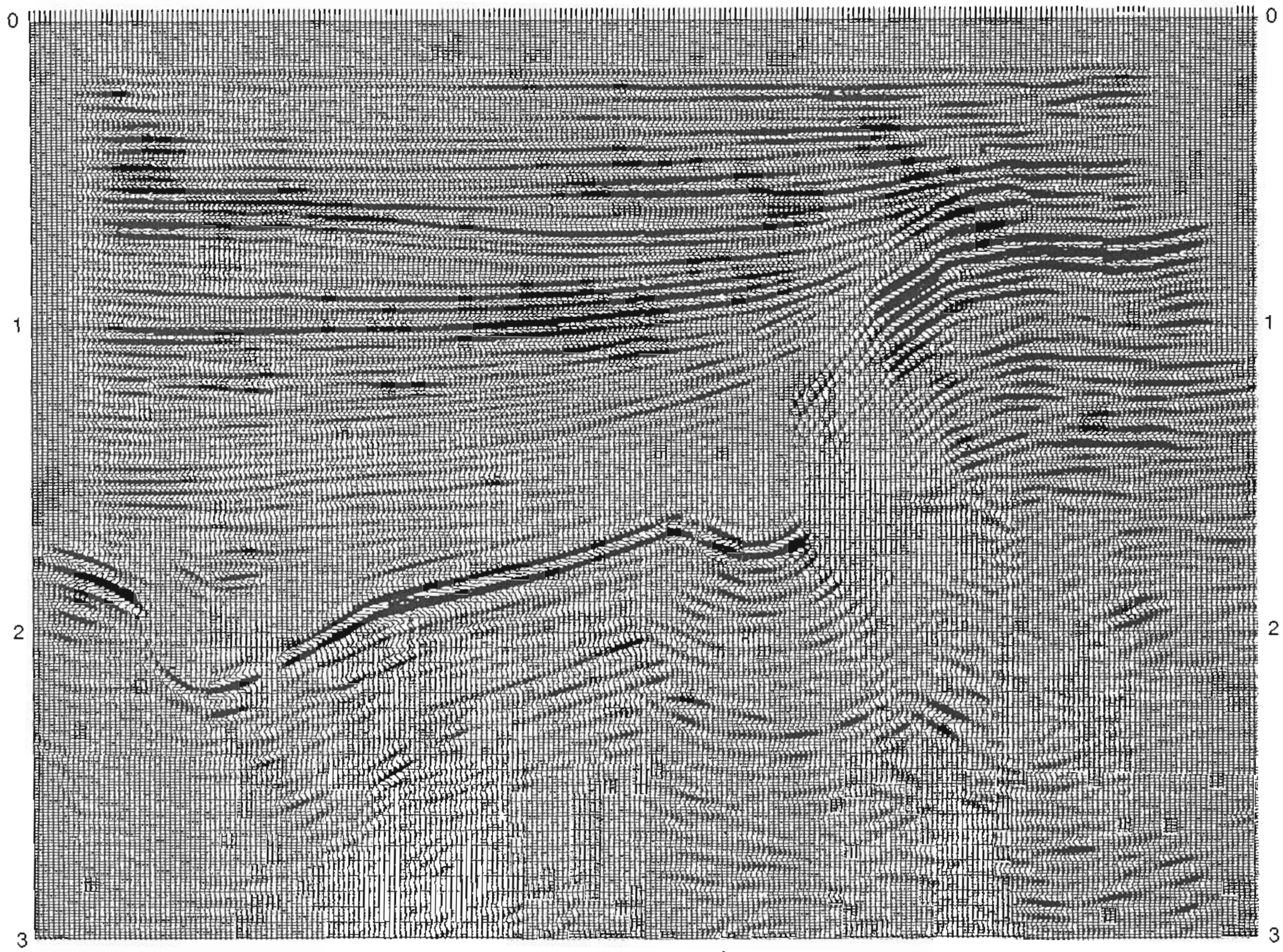
A.

Figure 2.15 Seismic reflection profiles. Individual seismic records are the wavy vertical lines plotted side by side along the distance axis. The vertical axis is the two-way travel time. Peaks in each record are shaded black to show up reflecrors rhat can be traced from one record to the next. Good horizontal reflectors are particularly evidenr below about 1.7 s in the left half of the profile. *A.* Unmigrared seismic profile. *B.* (*Facing page*) Migrared seismic profile.

the techniquc has sevcral disadvantages. The presence of low-velocity layers cannot be detected (Box 2.1). Deep structures ordinarily can be detected only at distances from the source that are greater than the depth. The properties of the Earth are averaged over large distances, so details of structure are lost. And nonhorizontal or discontinuous layers and complex structure are difficult or impossible to resolve.

In seismic reflection studies, reflections of P waves off internal boundaries are used to investigate the structure of the Earth. Seismic signals are recorded by as many as several hundred to several thousand geophones at a time. The resulting data are analyzed by computer (Boxes 2.2 and 2.3), and the seismograms are plotted side by side on the distance axis (Figure 2.15*A*). The individual peaks on the seismogram, or events, record the two-way travel time for each reflection, which is the time required for a wave to travel from a surface point to a reflector and back to the same surface point. Peaks in each record are shaded black so that strong signals in adjacent seismograms at the same two-way travel time show up as lines that indicate a continuous reflector.

Sophisticated computerized digital processing of the seismic records, including the very important processes of stacking and migration, allow complex structures to be resolved (Figure 2.15*B*) and therefore yield an incomparable image of the subsurface structure. The stacking of seismic records is a method of enhancing the signal-to-noise ratio by adding together reflections



B.

that occur at different angles from the same subsurface point (Box 2.2). Migration is a technique that allows the true locations of reflectors to be determined. All reflection signals on a seismogram are plotted as though they were vertical reflections (Figure 2.15*A*). Any given reflection, however, could have come from any point on an arc around the receiver, and the process of migration corrects the seismograms to determine the actual location of the reflector (Figure 2.15*B*; Box 2.3).

We can illustrate the benefits of migration by comparing an unmigrated seismic record (Figure 2.16*A*), a migrated profile (Figure 2.16*B*), and the true geologic cross section (Figure 2.16C). Migration eliminates the artifacts and errors in the unmigrated record. The actual geologic section, however, can be determined only if the two-way travel time can be converted into depth, which requires knowledge of the way velocity varies with depth. Note that if the seismic velocity changes with depth, the seismic (time) section distorts the actual vertical scale.

Despite their obvious value, the utility of reflection seismic studies is limited by their expense. Producing one seismic reflection profile can involve making hundreds of shots, each of which is recorded by hundreds to thousands of geophones.

The first arrival of a P wave can be either a compression of the material (a decrease in volume) or its opposite, a rarefaction (an increase in volume). Whether a compression or a rarefaction arrives first is revealed by the direction of first motion on the seismogram. The

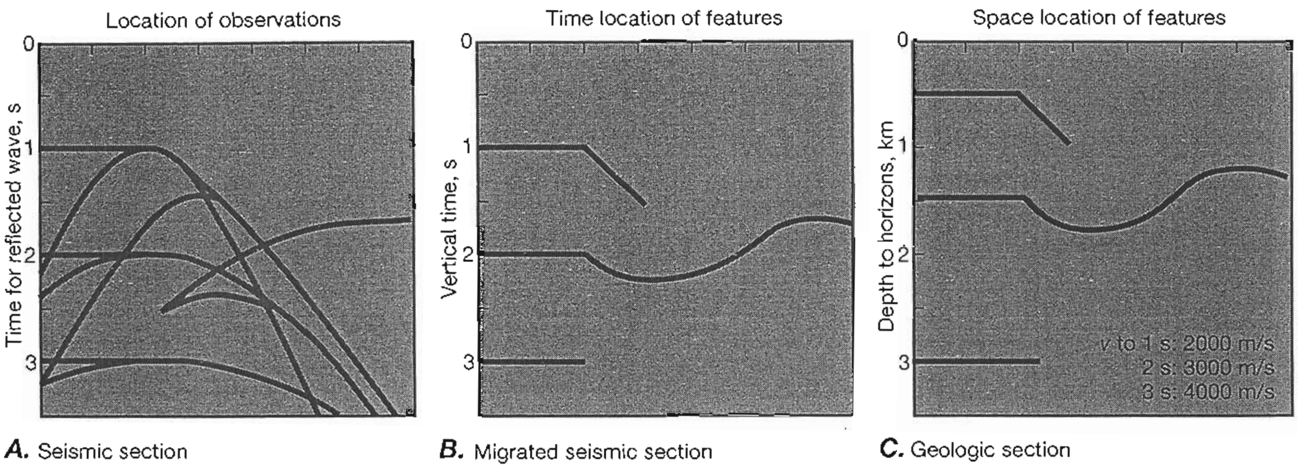


Figure 2.16 Diagrams illustrating the effects of migration. *A.* An unmigrated seismic section with multiple intersecting curved reflections. *B.* The same section as in part *A* after migration. The ambiguities and artifacts of the unmigrated section are all removed. C. The corresponding geologic section. The depth scale is different from the two-way travel-time scale because seismic velocity varies with depth.

pattern of compression or rarefaction first motions that radiate out from a sudden slip event on a fault is characteristic of the orientation of the fault and the sense of slip (Box 2.4). First-motion studies, therefore, are used to determine the orientation of, and sense of slip on, faults at depth. Regional patterns of first motions reveal large-scale tectonic motions of the plates. Figure 2.17 shows the radiation patterns that are characteristic of the three basic types of faulting: normal, thrust, and strike-slip.

Analysis of Gravity Anomalies

Gravity measurements are perhaps the second most important geophysical technique (after seismic techniques) used in structural geology and tectonics. A gravity anomaly (from the Greek word *anomalia,* which means "irregularity or unevenness") is the difference between a measured value of the acceleration of gravity, to which certain corrections are applied, and the reference value for the particular location. The reference value is determined from an internationally accepted formula that gives the gravitational field for an elliptically symmetric Earth. Because gravity anomalies arise from differences in the density of rocks, the goal in structural geology is to relate these differences in density to structural features. If no density contrasts exist, then the structure can have no effect on the gravitational field, and gravity anomalies cannot aid in the interpretation of that structure.

The structure at depth is interpreted by matching the gravity anomaly profile observed along a linear traverse with the anomaly profile calculated from an assumed model of the structure. The model is adjusted until the model anomaly profile shows a satisfactory fit to the observed anomaly profile. Although the model can never be unique, it is usually constrained by surface mapping and possibly by seismic data.

In order to calculate an anomaly, we must correct the measured value to the same reference used for the standard field. All measurements are therefore corrected to sea level as a common reference level. This altitude correction, the free-air correction, results in an increase in most land-based values but leaves surface observa-

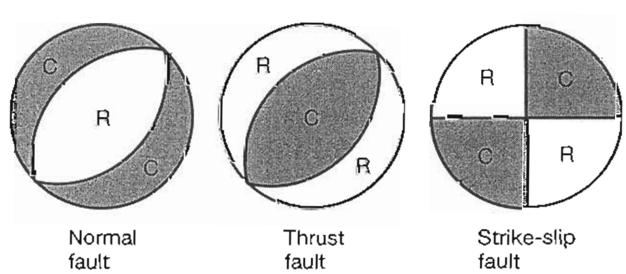


Figure 2.17 Equal-area projections showing the radiation pattern of compression first motions (C) and rarefaction first motions (R) for the three main types of faults. The fault on which the earthquake occurs is assumed to be at the center of the plotting sphere. All orientations that plot within a given sector are the orientations of rays when they leave the source that have the indicated first motion. Planes separating the sectors are nodal planes, one of which must be the fault plane. Material on each side of the fault moves toward the compression sectors, defining the sense of shear on the fault.

Box 2.2 Stacking of Seismic Records

Figure 2.2.1 illustrates the principle involved in the common depth-point stacking of seismic records. If explosions are detonated at shot points S_1, S_2, S_3, and S_4, reflections from the same point P on a horizontal subsurface boundary will be received at geophones G_1, G_2, G_3, and G_4, respectively. The same is true for all other horizontal reflectors below the point p. The corresponding shot points and geophones (S_i and G_i) are equidistant from the point p above P. If the travel times are corrected for the difference in length of the ray paths, the records can be added together, or stacked. The time-corrected signals from the reflections at P reinforce one another, and the signals from random noise tend to cancel out, thereby increasing the signal-to-noise ratio. The result is an enhanced seismogram showing the reflections as they would appear if the shot point and receiver were both at p.

In practice, data are gathered from a large linear array of shot points and geophones. Each geophone records many reflections from different depths, and the stacking is done by computer to produce enhanced seismograms at each point in the profile. Figure 2.15*A* is an example of a stacked seismic profile, which shows abundant horizontal reflectors at shallow depths.

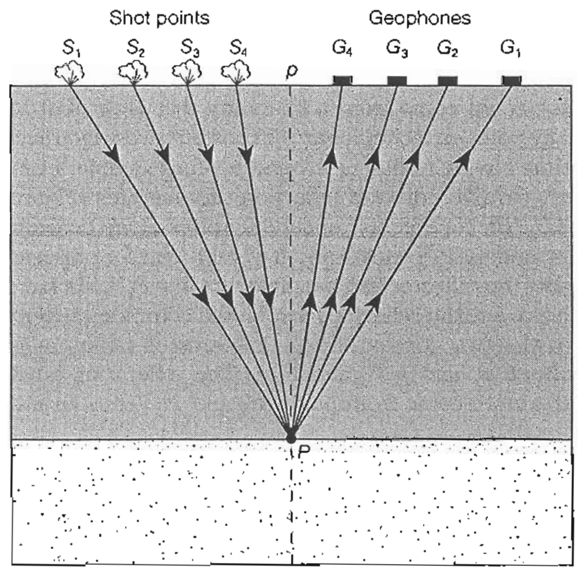


Figure 2.2.1 The principle of stacking of seismic records. Explosions are set off at shot points S_1, S_2, S_3, and S_4. The rays reflected from the same point P on a reflector at depth are received, for each of the explosions, at geophones G_1, G_2, G_3, and G_4, respectively. Adjusting the arrival times of these reflections for the different lengths of ray path allows the four records to be added together, which enhances the signal from the reflection at P and cancels out random noise. The effect is an increase in the signal-to-noise ratio.

tions at sea unchanged. If this is the only correction applied, the calculated anomaly is called a free-air anomaly.

The Bouguer correction is also frequently applied. It is assumed that between sea level and the altitude of the measurement is a uniform layer of continental crustal rock that represents an excess of mass piled on the surface. This assumed excess gravitational attraction is therefore subtracted from land-based measurements. At sea, it is assumed that all depths of water represent a deficiency of mass, because water is less dense than rock. The assumed deficiency in gravitational attraction is therefore added to sea-based measurements. The Bouguer anomaly results from application of both the free-air and Bouguer corrections.

The gravitational effects of local topography, however, differ measurably from those of a uniform layer. Thus a refinement of the simple Bouguer anomaly called the complete Bouguer anomaly requires a terrain correction to account for the local effects.

Thus, the Bouguer anomaly compares the mass of existing rocks at depth to the mass of standard continental crust whose elevation is at sea level. Bouguer anomalies are generally strongly negative over areas of high topography, indicating that there is a deficiency of mass below sea level compared to standard continental crust. They are strongly positive over ocean basins, indicating that there is an excess of mass below the ocean bottom compared to standard continental crust.

The area under a gravity anomaly profile provides a unique measure of the total excess or deficiency of mass at depth, and the shape of the profile constrains the possible distribution of the anomalous mass. The interpretation of mass distribution is not unique, however, because a given anomaly profile can be produced by a wide range of density differences and distributions. Figure 2.18*A*, for example, shows three symmetric bodies of the same density, each of which produces the same symmetric gravity anomaly. Figure 2.18*B*, *C* illustrates how the faulting of different density distributions affects

Box 2.3 Migration of Seismic Records

Although horizontal or very shallowly dipping reflectors are common in undeformed sedimentary basins (shallow parts of Figure 2.15), much of the structure of interest in structural and tectonic investigations is a great deal more complex (deeper parts of Figure 2.15). For example, beds with significant and variable dips and discontinuous beds (possibly truncated by a fault) are common. Such structures give rise to distortions and artifacts in seismic profiles (Figure 2.15), which must be corrected by **migration**.

We shall describe the principle of migration by using an example for which the seismic velocity of the material is constant, and the source and detector are at the same point p (Figure 2.3.1*A*). A particular reflection that apparently plots at P below the detector could come from a boundary that is tangent to any point on a circular arc of constant two-way travel time having radius pP around p—for example, from P'. On two adjacent reflection seismograms (Figure 2.3.1*B*), a reflection apparently plots at P_1 beneath p_1 and at P_2 beneath p_2. It would therefore appear that the reflector had the dip of the line P_1P_2. In fact, however, the reflector must be the common tangent to the two constant-travel-time arcs of radius p_1P_1 about p_1 and p_2P_2 about p_2. The reflections must therefore come from points P'_1 and P'_2. Thus an uncorrected profile shows erroneous locations and dips for dipping reflectors, and the reflection points P_1 and P_2 must be migrated along their respective constant-travel-time arcs to the correct locations at P'_1 and P'_2. The higher the true dip, the greater the distortion. Vertical reflectors plot on unmigrated seismic profiles as an alignment of reflections having a 45° dip.

If the seismic source and the detector are not at the same point, the arc of constant two-way travel time becomes an ellipse, and it is further distorted if the velocity is not constant. These are complications that must be accounted for in any analysis of a real seismic record, although the principle remains the same.

Another problem occurs if a reflector is discontinuous. The end of the reflector acts as a defraction point which takes energy from any angle of incidence and radiates it in all directions as though the point were a new source (point D in Figure 2.3.2). The signals recorded by the nearby detectors—for example, at p_1, p_2, and p_3—then plot along a parabolic arc on an uncorrected seismic profile (the dotted line in Figure 2.3.2; see also Figures 2.15*A*, 2.16*A*), because the two-way travel times for the signal increase as the distance of the detector from the end of the reflector increases. The location for all possible dif-

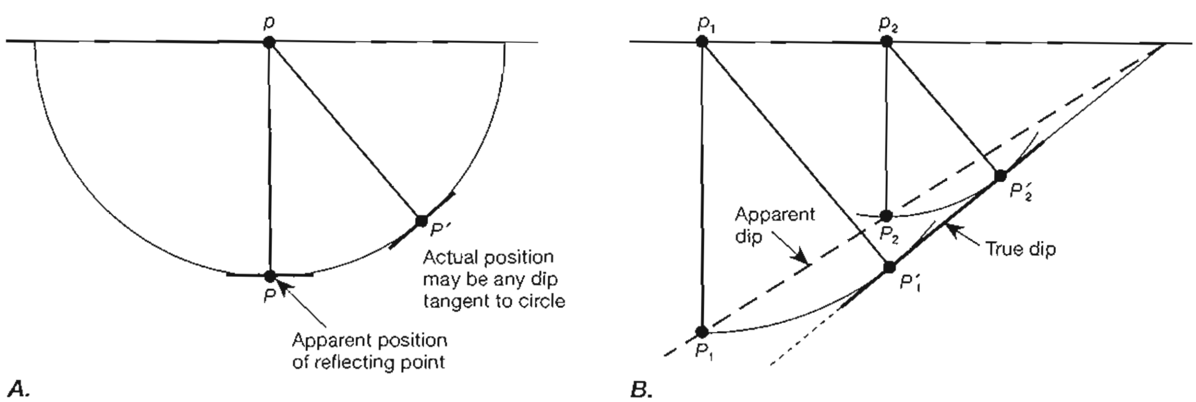


Figure 2.3.1 The migration of seismic signals corrects the seismic records to give the true location and dip of reflectors. In this example, seismic velocity is considered constant, and the shot point and receiver are both located at the same point. *A*. A reflection received at p appears on the seismic record at a two-way travel time that plots at P. In fact, the signal could come from any reflector, such as P', that is tangent to the semicircular arc of radius pP around p. That arc is the locus of constant two-way travel time. *B*. Reflections detected at p_1 and p_2 plot vertically below each point at P_1 and P_2, respectively, giving the reflector the apparent dip and location of the line P_1P_2. The true location and dip of the reflector, however, must be given by the line $P_1'P_2'$, which is the common tangent to the constant two-way travel-time arcs about p_1 and p_2, respectively. Note that P_1' is the actual location of the reflector below p_2.

fraction points that could generate the signal recorded at a given receiver, however, must lie along an arc of constant two-way travel time about that receiver (the dashed arcs in Figure 2.3.2). The true location of the diffraction point is the common intersection of the arcs constructed for several detectors. Thus migrating each signal along its arc to the common point identifies the true location of the diffraction point D.

In practice, the process of migration consists of taking each individual event on a reflection seismogram, migrating it along its arc of constant two-way travel time, and adding that event to any other seismogram intersected by that arc at the point of intersection. The resulting seismic profile is then a series of seismograms, each of which consists of the original record altered by the addition of all the events that migrate to that record. With this procedure, reflecting boundaries appear as coherent traces of events across the section in their correct location, and diffracted signals sum together at the location of the diffraction point. The other additions to the different seismograms tend to cancel each other out and do not produce coherent patterns on the seismic profile.

Determining the constant two-way travel-time arc, of course, requires determining the velocity structure. The amount of computation required to migrate every event in a profile to every seismogram intersected by its constant-travel-time arc is prodigious; in practice, it can be handled only by a computer.

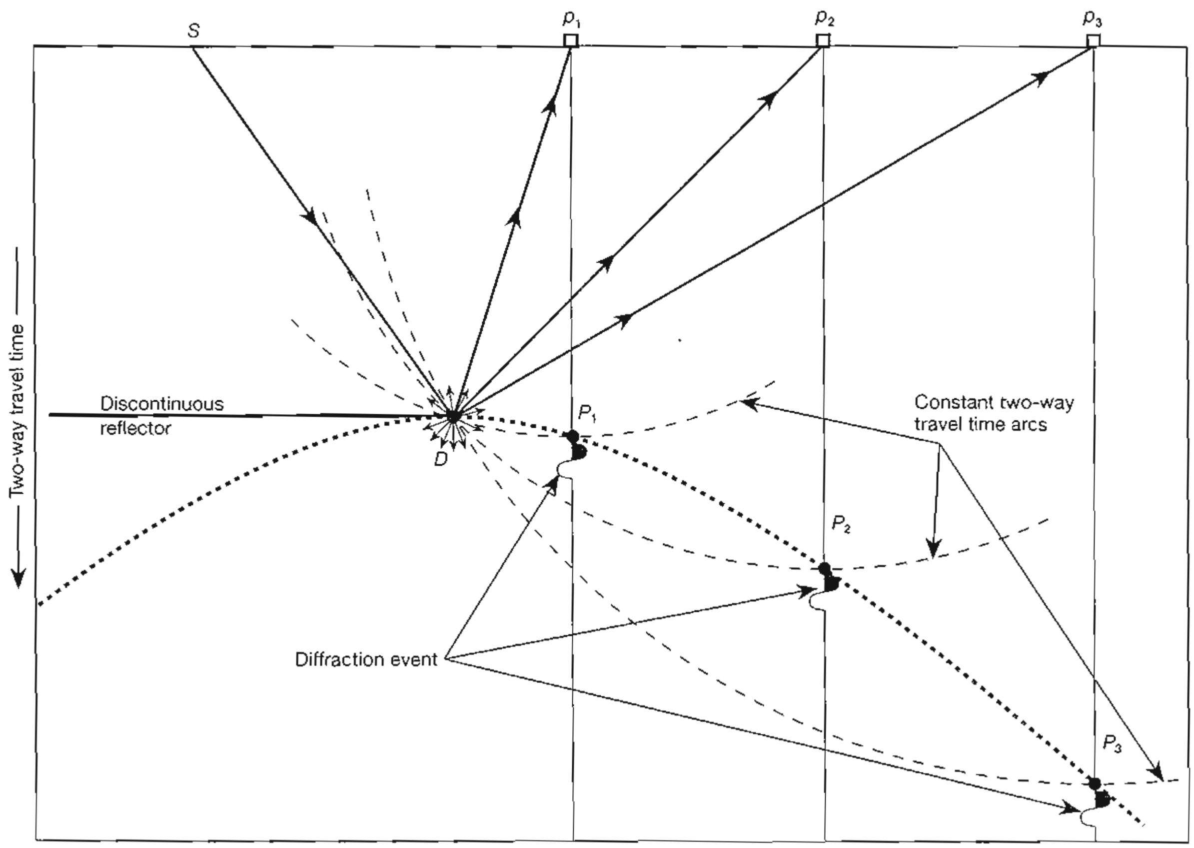
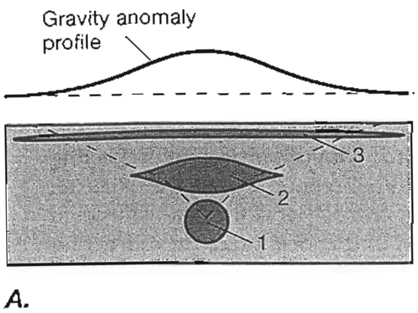
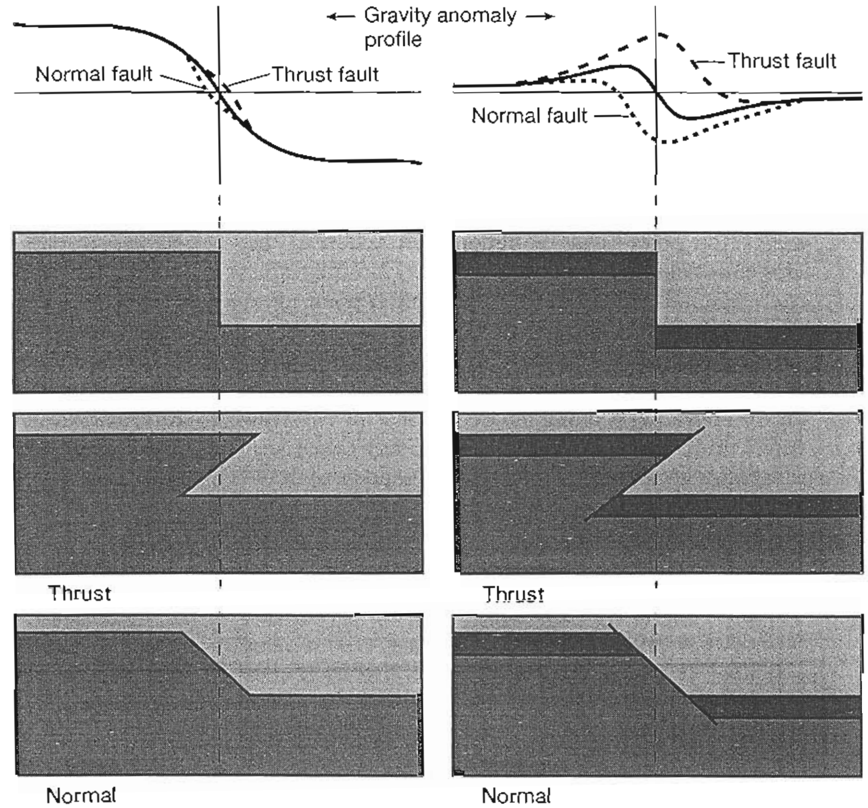


Figure 2.3.2 The end of a discontinuous reflector (D) acts as a diffraction point and radiates seismic energy in all directions for any angle of incidence. The further the receiver is from D, the later the diffracted ray arrives. Thus the diffracted energy arrives at receivers p_1, p_2, and p_3, for example, at times that fall along a parabolic arc at P_1, P_2, and P_3, respectively. The three constant-travel-time arcs constructed about the three receivers with radii p_1P_1, p_2P_2, and p_3P_3, respectively, must intersect at the location of the diffraction point. Thus we must migrate each event along its constant-travel-time arc to the common point at D in order to reconstruct its true location.





B. Gravity effect of a faulted dense basement

C. Gravity effect of a faulted dense layer

Figure 2.18 Illustration of the ambiguity and the nonuniqueness inherent in the interpretation of gravity anomalies. *A*. Three symmetrical bodies of the same density, each of which can produce the observed symmetric anomaly. *B*. The effect on gravity anomalies of vertical, thrust, aud normal displacement of a dense basement overlain by less dense strata. The asymmetry of the anomaly reflects that of the underlying structure, but the distinction among the three different srructures is almost negligible. *C*. The effect on gravity anomalies of vertical, thrust, and normal displacement of a dense layer wirhin less dense layers. The three strucrures produce markedly different gravity anomalies.

the gravity anomaly profile. If a low-density layer overlies a thick higher-density layer and the structure is faulted (Figure 2.18*B*), the gravity anomaly profile is asymmetric, but the different geometries of faulting have only a minor effect on the anomaly shape. If the denser material is in a relatively thin layer (Figure 2.18*C*), the gravity anomaly profile is again asymmetric, though the shape is different from that in Figure 2.18*B*, and the effect of different fault geometry is significant. Thus although the anomaly shape imposes constraints on the possible structure, to be reliable, gravity models should be based on additional structural and geophysical information.

Geomagnetic Studies

A magnetic field is a vector quantity that has both magnitude and direction. For the Earth's magnetic field, the magnitude can be specified by the magnitudes of the horizontal and vertical components of the field. The orientation is specified by the declination and inclination, which are essentially the trend and plunge of the field line, though the inclination also includes the polarity, which defines whether the magnetic vector points up or down. Studies of the Earth's magnetic field include the study of magnetic anomalies and of paleomagnetism.

Magnetic anomalies are measurements of the variation of the Earth's magnetic field relative to some locally defined reference. There is no international standard reference field from which anomalies are measured, because the Earth's magnetic field is not constant and changes significantly even on a human time scale.

Regional maps of magnetic anomalies are made by using both aerial and surface measurements. The principal use of continental magnetic anomaly maps is to infer the presence of rock types and structures that are covered by other rocks, sediments, or water. In some cases, the presence of particular rock types at depth can be inferred on the basis of characteristic patterns on a magnetic anomaly map. For example, the extension of rocks of the Canadian shield beneath thrust faulrs of the Canadian Rocky Mountains can be inferred from the extension of the shield magnetic pattern beneath the thrust front.

Marine magnetic surveys have resulted in the well-known maps of the symmetric patterns of magnetic anomalies which have been so fundamental to the development of plate tectonic theory. When correlated with the magneric reversal time scale, these maps can be interpreted to give a map of the age of ocean basins.

Magnetic anomalies also can be used in a manner similar to graviry anomalies ro infer structure at depth, except that the magnetic anomalies are due to differences in magnetic properties of the rocks rather than in their densities. Modeling of magnetic information is more complex, because a given anomaly in total field intensity can result either from differences in intensity of magnetization of the rocks or from different orientations of the magnetic vector. Models of structure based on magnetic anomalies suffer from the same lack of uniqueness as models based on gravity anomalies, and for similar reasons.

By a variety of processes that include cystallization, cooling, sedimentation, and chemical reaction in the Earth's magnetic field, rocks can become magnetized in a direction parallel to the ambient field and can preserve that magnetism even if the rocks are rotated to new orientations. Studies of paleomagnetism involve measuring the orientation of the magnetic field preserved in rocks and comparing it to the orientation of the present-day field. If the original horizontal plane in the sample is known, these measurements can be interpreted to indicate the declination and inclination of the Earth's field at the time of magnetization. Gently dipping, unaltered sediments and volcanic rocks provide the most reliable paleomagnetic measurements, but more deformed or metamorphosed rocks and plutonic rocks are sometimes useful. Rocks that have been tilted since magnetization are generally assumed to have tilted about a horizontal axis, so they are restored to the original horizontal by rotation about an axis parallel to the strike of the bedding.

The earth's magnetic field is approximately symmetric about the axis of rotation, and the inclination of the field lines varies systematically with latitude from vertically down at the north pole through horizontal at the equator to vertically up at the south pole. Because this relationship is assumed to have been constant throughout geologic time, the paleo-declination determined for the sample in its original horizontal attitude indicates the amount of rotation a rock has undergone about a vertical axis, and the paleo-inclination relative to the original horizontal indicates the latitude at which the sample was magnetized. Such measurements can therefore define the changes in latitude and the rotations about a vertical axis resulting from the large-scale tectonic motions that rocks have experienced since magnetization. They can provide no information, however, on the changes in longitude associated with these motions.

Plotting the apparent paleomagnetic pole position for different time periods from a particular region provides an approximate indication of the movement of that area with respect to the Earth's geographic pole. These results are usually presented in the form of apparent polar wander maps, such as the map of paleopole positions for North America and Europe during the Phanerozoic (Figure 2.19*A*). If the continents are restored to their relative positions before the opening of the Atlantic Ocean, the apparent polar wander paths coincide approximately from Silurian through Triassic, indicating the period of time the continents were joined (Figure 2.19*B*).

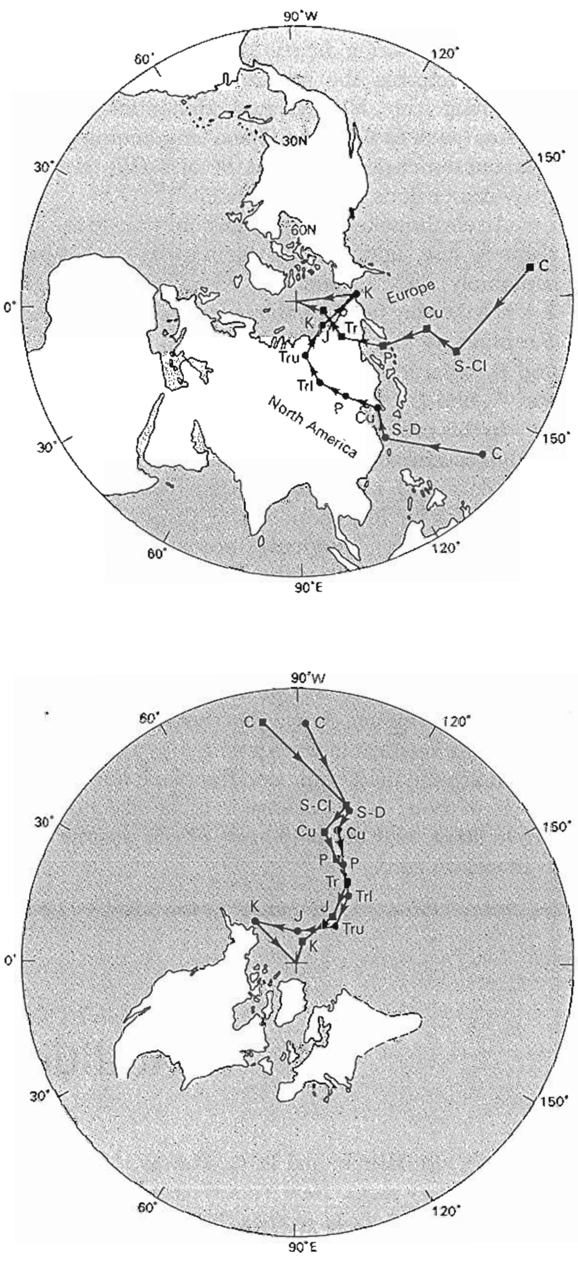


Figure 2.19 Apparent polar wander path (APW) for North America (circles) and for Europe (squares). C = Cambrian, S = Silurian, D = Devonian, Cl and Cu = lower and upper Carboniferous, P = Permian, Tr, Trl, and Tru = Triassic, lower Triassic, and upper Triassic, K = Cretaceous. *A*. Polar wander paths for the continents in their present positions. *B*. Polar wander paths for the continents before the opening of the Atlantic Ocean.

Box 2.4 First-Motion Radiation Pattern from a Faulting Event

The first-motion radiation pattern from a fault slip event can be accounted for by the two-dimensional model shown in Figure 2.4.1. The undeformed state is shown in Figure 2.4.1*A*; it is represented by two squares drawn on opposite sides of an east–west (E–W) line that represents the future location of a fault. Gradual prefaulting deformation of the rock (Figure 2.4.1*B*) deforms the squares into parallelograms, shortening the NW-oriented dimensions of the squares (such as *AD* and *CF*) and lengthening the NE-oriented dimensions (such as *BC* and *DE*). N–S and E–W dimensions remain unchanged.

An earthquake occurs when cohesion on the fault plane is lost, and sudden slip returns each square separately to its undeformed condition (Figure 2.4.1*C*). During faulting, the outer points *A*, *B*, *E*, and *F* remain stationary, while the points on the fault, *C* and *D*, separate into the respective pairs C_N and D_N, and C_S and D_S.

In this process, the NW-oriented dimensions suddenly become longer (for example, D_N moves away from *A*, and C_S moves away from *F*), creating a rarefaction for the first motion. The NE-oriented dimensions, however, suddenly become shorter (for example, C_N moves closer to *B* and D_S moves closer to *E*), creating a compression for the first motion. Again the N–S and E–W dimensions remain unchanged. Thus compressive first motions radiate outward in the NE and SW quadrants, and rarefaction first motions radiate outward in the NW and SE quadrants. The quadrants are separated by nodal planes, which are the fault plane and the plane normal to it, for dimensions do not change in these directions during faulting, and the amplitude of the seismic wave is therefore zero.

The first-motion radiation pattern of compressions and rarefactions thus enables us to identify the fault plane and the nodal plane, and it indicates the sense of slip on either plane that would generate that pattern, because slip would have to be toward the compression quadrant on either side of either nodal plane. The actual fault plane can often be identified by taking geologic considerations into account or by studying the location of aftershocks that occur along the fault plane.

The same principle works in three dimensions, and the nodal planes can be identified from first-motions by using a worldwide array of seismometers that in effect form a three-dimensional array surrounding

Figure 2.4.1 (*Facing page*) A two-dimensional model for the mechanism of first-motion radiation patterns. *A*. Undeformed state represented by squares on either side of a future fault. *B*. Deformed state before faulting. N–S and E–W dimensions of the squares are unchanged, but NE–SW dimensions (such as *BC* and *DE*) are lengthened and NW–SE dimensions (such as *AD* and *CF*) are shortened. *C*. Faulted state: sudden slip on the fault generates an earthquake. N–S and E–W dimensions of the squares are still unchanged, but NE–SW dimensions (such as BC_N and D_SE) are suddenly shortened, and NW–SE dimensions (such as AD_N and C_SF) are suddenly lengthened. Thus first motions are compressions for rays leaving the source in the quadrants marked C and are rarefactions for rays leaving the source in the quadrants marked R. The fault plane and the plane normal to it are nodal planes along which no change in dimension occurs, and the amplitude of the first motion is therefore zero.

Additional Readings

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the earthquake. In Chapter 9 we show that the maximum and the minimum compressive stresses lie in the rarefaction and the compression quadrants, respectively. The stress orientations are sometimes approximated by P and T axes, respectively, where the P axis bisects the nodal planes in the rarefaction quadrants, and the T axis bisects the nodal planes in the compressional quadrants (Figure 2.4.1*C*). It is a potential source of confusion that the minimum compressive stress, which is the deviatoric tensile stress (see Chapter 9), is located in the quadrant of compressional first arrivals.