Elements of Map-Scale Structure

1.1 Introduction

The primary objective of structural map making and map interpretation is to develop an internally consistent three-dimensional picture of the structure that agrees with all the data. This can be difficult or ambiguous because the complete structure is usually undersampled. Thus an interpretation of the complete geometry will probably require a significant number of inferences, as, for example, in the interpolation of a folded surface between the observation points. Constraints on the interpretation are both topological and mechanical. The basic elements of map-scale structure are the geometries of folds and faults, the shapes and thicknesses of units, and the contact types. This chapter provides a short review of the basic elements of the structural and stratigraphic geometries that will be interpreted in later chapters, reviews some of the primary mechanical factors that control the geometry of map-scale folds and faults, and examines the typical sources of data for structural interpretation and their inherent errors.

1.2 Representation of a Structure in Three Dimensions

A structure is part of a three-dimensional solid volume that probably contains numerous beds and perhaps faults and intrusions (Fig. 1.1). An interpreter strives to develop

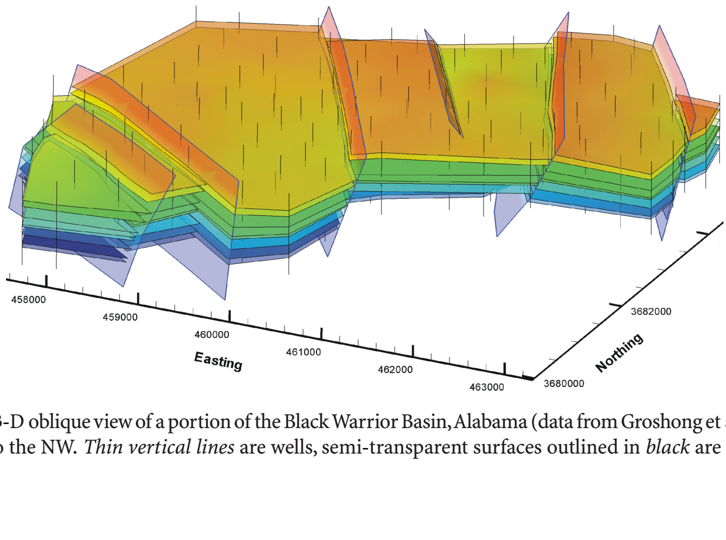


Fig. 1.1. 3-D oblique view of a portion of the Black Warrior Basin, Alabama (data from Groshong et al. 2003b), viewed to the NW. *Thin vertical lines* are wells, semi-transparent surfaces outlined in *black* are faults

a mental and physical picture of the structure in three dimensions. The best interpretations utilize the constraints provided by all the data in three dimensions. The most complete interpretation would be as a three-dimensional solid, an approach possible with 3-D computer graphics programs. Two-dimensional representations of structures by means of maps and cross sections remain major interpretation and presentation tools. When the geometry of the structure is represented in two dimensions on a map or cross section, it must be remembered that the structure of an individual horizon or a single cross section must be compatible with those around it. This book presents methods for extracting the most three-dimensional interpretive information out of local observations and for using this information to build a three-dimensional interpretation of the whole structure.

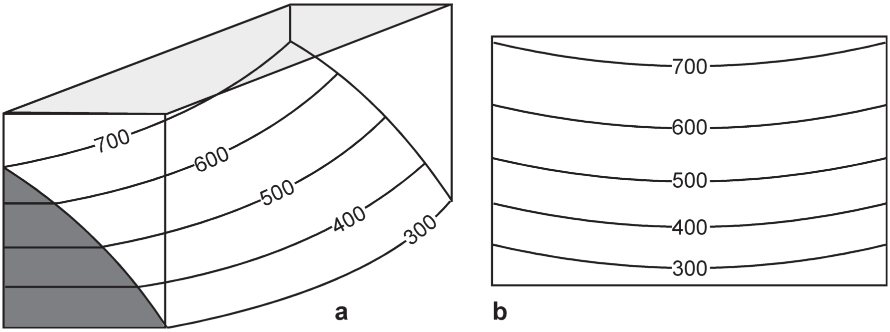
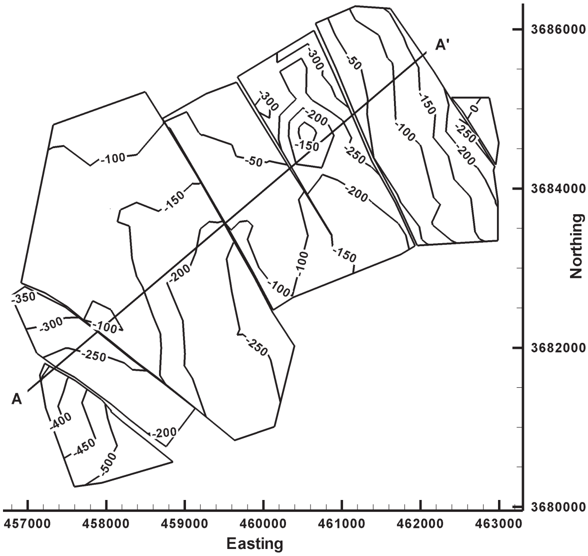


Fig. 1.2. Structure contours. **a** Lines of equal elevation on the surface of a map unit. **b** Lines of equal elevation projected onto a horizontal surface to make a structure contour map

Fig. 1.3.
Structure contour map of the faulted upper horizon from Fig. 1.1. Contours are at 50 ft intervals, with negative elevations being below sea level. Faults are indicated by gaps where the horizon is missing



1.2.1 Structure Contour Map

A structure contour is the trace of a horizontal line on a surface (e.g., on a formation top or a fault). A structure contour map represents a topographic map of the surface of a geological horizon (Figs. 1.2, 1.3). The dip direction of the surface is perpendicular to the contour lines and the dip amount is proportional to the spacing between the contours. Structure contours provide an effective method for representing the three-dimensional form of a surface in two dimensions. Structure contours on a faulted horizon (Fig. 1.3) are truncated at the fault.

1.2.2 Triangulated Irregular Network

A triangulated irregular network (TIN; Fig. 1.4) is an array of points joined by straight lines that define a surface. In a TIN network, the nearest-neighbor points are connected to form triangles that form the surface (Banks 1991; Jones and Nelson 1992). If the triangles in the network are shaded, the three-dimensional character of the surface can be illustrated. This is an effective method for the rendering of surfaces by computer. The TIN can be contoured to make a structure contour map.

1.2.3 Cross Section

Even though a structure contour map or TIN represents the geometry of a surface in three dimensions, it is only two-dimensional because it has no thickness. To completely represent a structure in three dimensions, the relationship between different horizons must be illustrated. A cross section of the geometry that would be seen on the face of a slice through the volume is the simplest representation of the relationship between

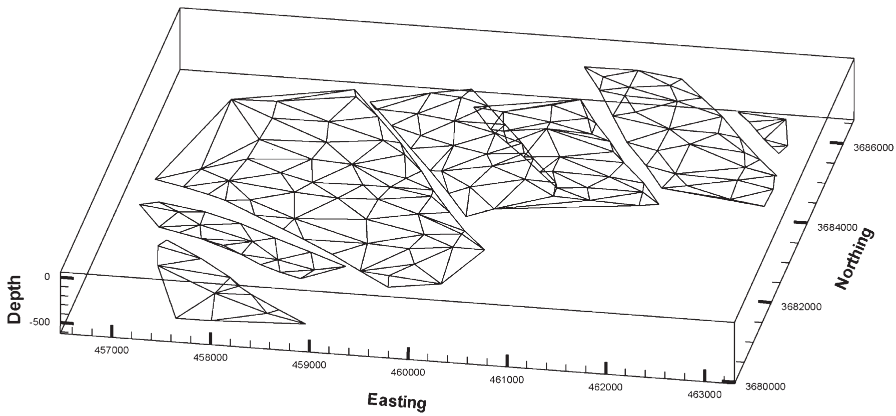


Fig. 1.4. Triangulated irregular network (TIN) of points used to form the upper map horizon in Figs. 1.1 and 1.3. 3-D perspective view to the NW, 3× vertical exaggeration. Vertical scale in ft, horizontal scale in meters

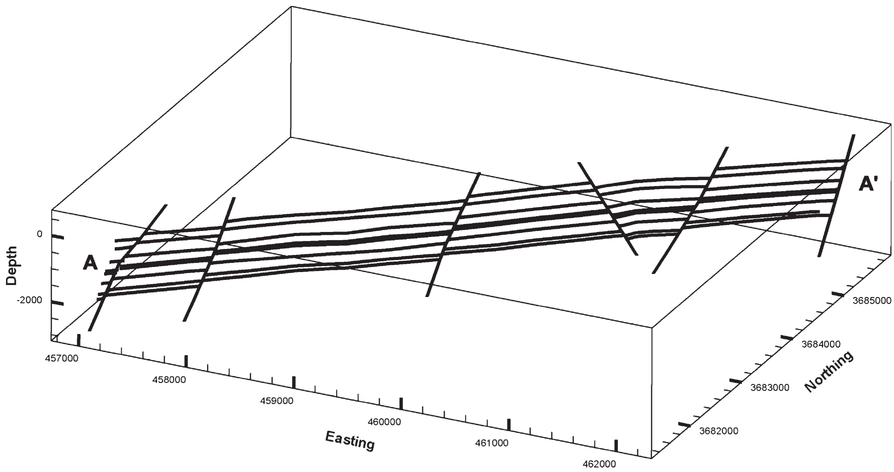


Fig. 1.5. East-west cross section across the structure in Fig. 1.1 produced by taking a vertical slice through the 3-D model. Line of section shown on Fig. 1.3

horizons. In this book cross sections will be assumed to be vertical unless it is stated otherwise. A cross section through multiple surfaces illustrates their individual geometries and defines the relationships between the surfaces (Fig. 1.5). The geometry of each surface provides constraints on the geometry of the adjacent surfaces. The relationships between surfaces forms the foundation for many of the techniques of structural interpretation that will be discussed.

1.3 Map Units and Contact Types

The primary concern of this book is the mapping and map interpretation of geologic contacts and geologic units. A contact is the surface where two different kinds of rocks come together. A unit is a closed volume between two or more contacts. The geometry of a structure is represented by the shape of the contacts between adjacent units. Dips or layering within a unit, such as in a crossbedded sandstone, are not necessarily parallel to the contacts between map units. Geological maps are made for a variety of purposes and the purpose typically dictates the nature of the map units. It is important to consider the nature of the units and the contact types in order to distinguish between geometries produced by deposition and those produced by deformation. Units may be either right side up or overturned. A stratigraphic horizon is said to face in the direction toward which the beds get younger. If possible, the contacts to be used for structural interpretation should be parallel and have a known paleogeographic shape. This will allow the use of a number of powerful rules in the construction and validation of map surfaces, in the construction of cross sections, and should result in geometries that can be restored to their original shapes as part of the structural validation process. Contacts that were originally horizontal are preferred. Even if a restoration is not actually done, the concept that the map units were originally horizontal is implicit in many structural interpretations.

1.3.1
Depositional Contacts

A depositional contact is produced by the accumulation of material adjacent to the contact (after Bates and Jackson 1987). Sediments, igneous or sedimentary extrusions, and air-fall igneous rocks have a depositional lower contact which is parallel to the pre-existing surface. The upper surface of such units is usually, but not always, close to horizontal. A conformable contact is one in which the strata are in unbroken sequence and in which the layers are formed one above the other in parallel order, representing the uninterrupted deposition of the same general type of material, e.g., sedimentary or volcanic (after Bates and Jackson 1987).

Lithologic boundaries that represent lateral facies transitions (Fig. 1.6a), were probably not horizontal to begin with. Certain sedimentary deposits drape over pre-existing topography (Fig. 1.6b) while others are deposited with primary depositional slopes (Fig. 1.6c). The importance of the lack of original horizontality depends on the scale of the map relative to the magnitude of the primary dip of the contact. Contacts that dip only a few degrees might be treated as originally horizontal in the interpretation of a local map area, but the depositional contact between a reef and the adjacent basin sediments may be close to vertical (Fig. 1.7), for example, and could not be considered as originally horizontal at any scale. Depositional contacts that had significant original topographic relief (Fig. 1.7) should be restored to their original depositional geometry, not to the horizontal.

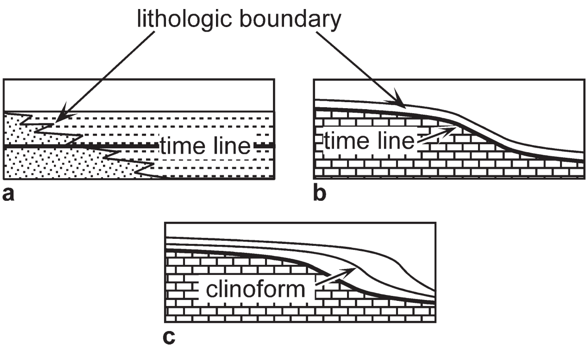


Fig. 1.6. Cross sections showing primary depositional lithologic contacts that are not horizontal. **a** Laterally equivalent deposits of sandstone and shale. The depositional surface is a time line, not the lithologic boundary. **b** Draped deposition parallel to a topographic slope. **c** Primary topography associated with clinoform deposition

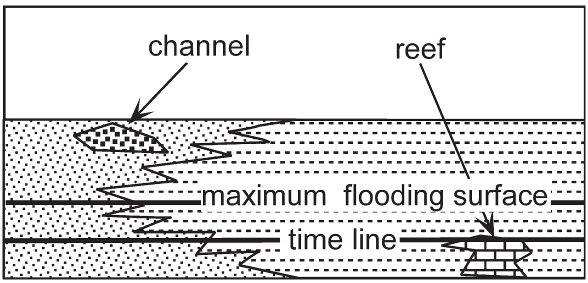


Fig. 1.7. Cross sections showing primary sedimentary facies relationships and maximum flooding surface. All time lines are horizontal in this example

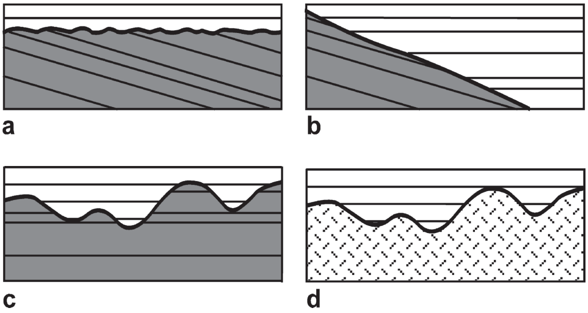


Fig. 1.8.
Unconformity types. The unconformity (*heavy line*) is the contact between the older, underlying *shaded* units and the younger, overlying *unshaded* units. **a** Angular unconformity. **b** Buttress or onlap unconformity. **c** Disconformity. **d** Nonconformity. The patterned unit may be plutonic or metamorphic rock

1.3.2 Unconformities

An unconformity is a surface of erosion or nondeposition that separates younger strata from older strata. An angular unconformity (Fig. 1.8a) is an unconformity between two groups of rocks whose bedding planes are not parallel. An angular unconformity with a low angle of discordance is likely to appear conformable at a local scale. Distinguishing between conformable contacts and low-angle unconformities is difficult but can be extremely important to the correct interpretation of a map. A progressive or buttress unconformity (Fig. 1.8b; Bates and Jackson 1987) is a surface on which onlapping strata abut against a steep topographic scarp of regional extent. A disconformity (Fig. 1.8c) is an unconformity in which the bedding planes above and below the break are essentially parallel, indicating a significant interruption in the orderly sequence of sedimentary rocks, generally by an interval of erosion (or sometimes of nondeposition), and usually marked by a visible and irregular or uneven erosion surface of appreciable relief. A nonconformity (Fig. 1.8d) is an unconformity developed between sedimentary rocks and older plutonic or massive metamorphic rocks that had been exposed to erosion before being covered by the overlying sediment.

1.3.3 Time-Equivalent Boundaries

The best map-unit boundaries for regional structural and stratigraphic interpretation are time-equivalent across the map area. Time-equivalent boundaries are normally established using fossils or radiometric age dates and may cross lithologic boundaries. Volcanic ash fall deposits, which become bentonites after diagenesis, are excellent time markers. Because an ash fall drapes the topography and is relatively independent of the depositional environment, it can be used for regional correlation and to determine the depositional topography (Asquith 1970). It can be difficult to establish time-equivalent map horizons because of the absence or inadequate resolution of the paleontologic or radiometric data, lithologic and paleontologic heterogeneity in the depositional environment, and because of the occurrence of time-equivalent nondeposition or erosion in adjacent areas. Time-equivalent map-unit boundaries may be based on certain aspects of the physical stratigraphy. A *sequence* is a conformable succession of genetically related strata bounded by unconformities and their correlative conformities (Mitchum 1977; Van Wagoner et al. 1988). A parasequence is a subunit within a sequence that is bounded by marine flooding surfaces (Van Wagoner et al. 1988) and the approximate time equivalence along flooding surfaces makes them suitable for structural mapping. A maximum flooding surface (Fig. 1.7; Galloway 1989) can be the best for regional correlation because the deepest-water deposits can be correlated across lithologic boundaries. At the time of maximum flooding, the sediment input is at a minimum and the associated sedimentary deposits are typically condensed sections, seen as radioactive shales or thin, very fossiliferous carbonates.

1.3.4 Welds

A weld joins strata originally separated by a depleted or withdrawn unit (after Jackson 1995). Welds are best known where a salt bed has been depleted by substratal dissolution or by flow (Fig. 1.9). If the depleted unit was deposited as part of a stratigraphically conformable sequence, the welded contact will resemble a disconformity. If the depleted unit was originally an intrusion, like a salt sill, the welded contact will return to its original stratigraphic configuration. A welded contact may be recognized from remnants of the missing unit along the contact. Lateral displacement may occur across the weld before or during the depletion of the missing unit (Fig. 1.9b). Welded contacts may crosscut bedding in the country rock if the depleted unit was originally crosscutting, as, for example, salt diapir that is later depleted.

1.3.5 Intrusive Contacts and Veins

An intrusion is a rock, magma, or sediment mass that has been emplaced into another distinct unit. Intrusions (Fig. 1.10) may form concordant contacts that are parallel to the layering in the country rock, or discordant contacts that crosscut the layering in the country rock. A single intrusion may have contacts that are locally concordant and discordant.

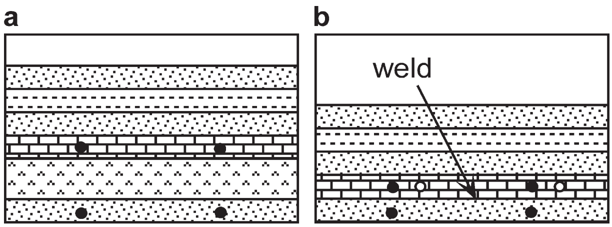


Fig. 1.9. Cross sections illustrating the formation of a welded contact. *Solid dots* are fixed material points above and below the unit which will be depleted. **a** Sequence prior to depletion. **b** Sequence after depletion: *solid dots* represent the final positions of original points in a hangingwall without lateral displacement; *open circles* represent final positions of original points in a hangingwall with lateral displacement

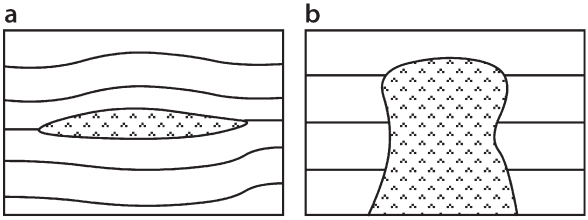


Fig. 1.10.
Cross sections of intrusions. Intrusive material is *patterned* and *lines* represent layering in the country rock. **a** Concordant. **b** Discordant

A vein is a relatively thin, normally tabular, rock mass of distinctive lithologic character, usually crosscutting the structure of the host rock. Many veins are depositional and represent the filling of a fracture, whereas others are the result of replacement of the country rock. Veins are mentioned here with intrusions because the contact relationships and unit geometries may be similar to those of some intrusions.

1.3.6 Other Boundaries

Many other attributes of the rock units and their contained fluids can be mapped, for example, the porosity, the oil-water contact, or the grades of mineral deposits. Most of the mapping techniques to be discussed will apply to any type of unit or contact. Some interpretation techniques, particularly those for fold interpretation, depend on the contacts being originally planar boundaries and so those methods may not apply to nonstratigraphic boundaries.

1.4 Thickness

The thickness of a unit is the perpendicular distance between its bounding surfaces (Fig. 1.11a). The true thickness does not depend on the orientation of the bounding surfaces. If a unit has variable thickness, various alternative measurements might be used, such as the shortest distance between upper and lower surfaces or the distance measured perpendicular to either the upper or lower surface. The definition used here is based on the premise that if the unit was deposited with a horizontal surface but a variable thickness, then the logical measurement direction would be the thickness measured perpendicular to the upper surface, regardless of the structural dip of the surface (Fig. 1.11b).

Thickness variations can be due to a variety of stratigraphic and structural causes. Growth of a structure during the deposition of sediment typically results in thinner stratigraphy on the structural highs and thicker stratigraphy in the lows. Both growth folds and growth faults occur. A sedimentary package with its thickness influenced by an active structure is known as a growth unit or growth sequence. The high part of a growth structure may be erosional at the same time that the lower parts are depositional. Thickness variations may be the result of differential compaction during and after deposition. If the composition of a unit undergoes a facies change from relatively uncompactable (i.e., sand) to relatively compactable (i.e., shale) then after burial and compaction the unit thickness will vary as a function of lithology. Deformation-related thickness changes are usually accompanied by folding, faulting, or both within the unit being mapped. Deformation-related thickness changes are likely to correlate to position within a structure or to structural dip.

1.5 Folds

A fold is a bend due to deformation of the original shape of a surface. An antiform is convex upward; an anticline is convex upward with older beds in the center. A synform is concave upward; a syncline is concave upward with younger beds in the center. Original curves in a surface, for example grooves or primary thickness changes, are not considered here to be anticlines or synclines.

1.5.1 Styles

Folds may be characterized by domains of uniform dip, by a uniform variation of the dip around a single center of curvature, or may combine regions of both styles. Regions of uniform dip (Fig. 1.12a) are called dip domains (Groshong and Usdansky 1988). Dip domains are separated from one another by axial surfaces or faults. Dip domains have also been referred to as kink bands, but the term kink band has mechanical

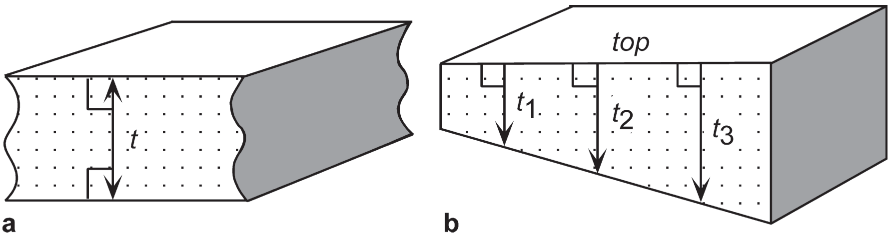


Fig. 1.11. Thickness (*t*). **a** Unit of constant thickness. **b** Unit of variable thickness

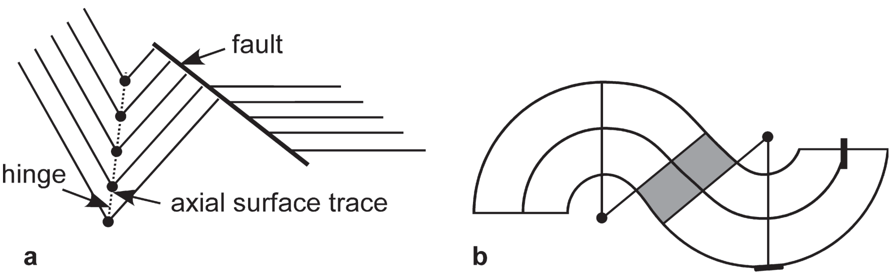


Fig. 1.12. Regions of uniform dip properties. **a** Dip domains. **b** Concentric domains separated by a planar dip domain (*shaded*)

implications that are not necessarily appropriate for every structure. An axial surface (Fig. 1.12a) is a surface that connects fold hinge lines, where a hinge line is a line of maximum curvature on the surface of a bed (Dennis 1967). A hinge (Fig. 1.12a) is the intersection of a hinge line with the cross section. A circular domain (Fig. 1.12b) is defined here as a region in which beds approximate a portion of a circular arc. If multiple surfaces in a circular domain have the same center of curvature, the fold is concentric. Dips in a concentric domain are everywhere perpendicular to a radius through the center of curvature. The center of curvature is determined as the intersection point of lines drawn perpendicular to the dips (Busk 1929; Reches et al. 1981). A circular curvature domain does not possess a line of maximum curvature and thus does not strictly have an axial surface. Dips within a domain may vary from the average values. If the dips are measured by a hand-held clinometer, variations of a few degrees are to be expected due to the natural variation of bedding surfaces and the imprecision of the measurements.

As a generality, the structural style is controlled by the mechanical stratigraphy and the directions of the applied forces. Mechanical stratigraphy is the stratigraphy described in terms of its physical properties. The mechanical properties that control the fold geometry are the stiffness contrasts between layers, the presence or absence of layer-parallel slip, and the relative layer thicknesses. Stiff (also known as competent) lithologies (for example, limestone, dolomite, cemented sandstone) tend to maintain constant bed thickness, and soft (incompetent) lithologies (for example, evaporites, overpressured shale, shale) tend to change bed thickness as a result of deformation (Fig. 1.13). Very stiff and brittle units like dolomite may fail by pervasive fracturing, however, and then change thickness as a unit by cataclastic flow. Deformed sedimen-

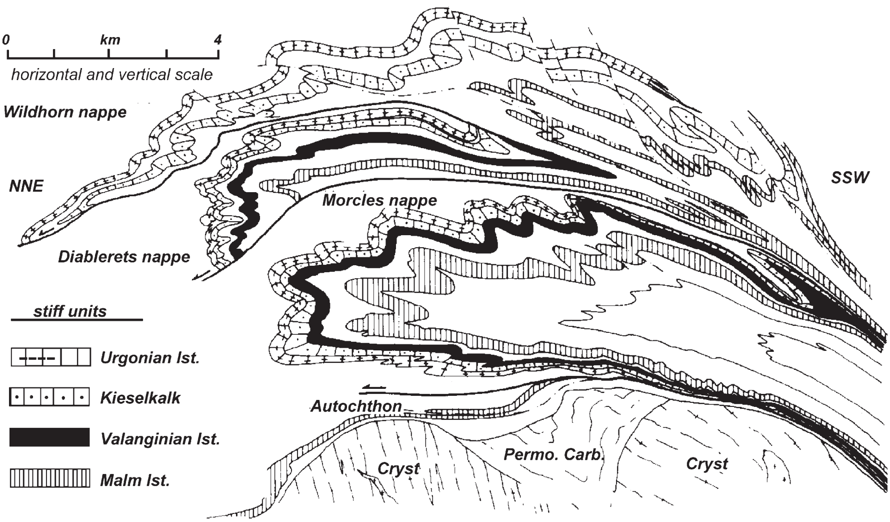


Fig. 1.13. Cross section of the Helvetic Alps, central Switzerland. Mechanical stratigraphy consists of thick carbonate units (stiff) separated by very thick shale units (soft). The folds, especially at the hinges, are circular in style. (After Ramsay 1981)

tary rocks tend to maintain relatively constant bed thickness, although the thickness changes that do occur can be very important. If large thickness changes are observed in deformed sedimentary rocks other than evaporites or overpressured shale, primary stratigraphic variations should be considered as a strong possibility.

In cross section, folds may be harmonic, with all the layers nearly parallel to one another (Fig. 1.14), or disharmonic, with significant changes in the geometry between different units in the plane of the section (Figs. 1.13, 1.15). The fold geometry is controlled by the thickest and stiffest layers (or multilayers) called the dominant members (Currie et al. 1962). A stratigraphic interval characterized by a dominant (geometry-controlling) member between two boundary zones is a structural-lithic unit (Fig. 1.15;

Fig. 1.14. Dip-domain style folds in an experimental model having a closely spaced multilayer stratigraphy. The *black* and *white layers* are plasticene and are separated by grease to facilitate layer-parallel slip. The *black layers* are slightly stiffer than the *white layers*. (After Ghosh 1968)

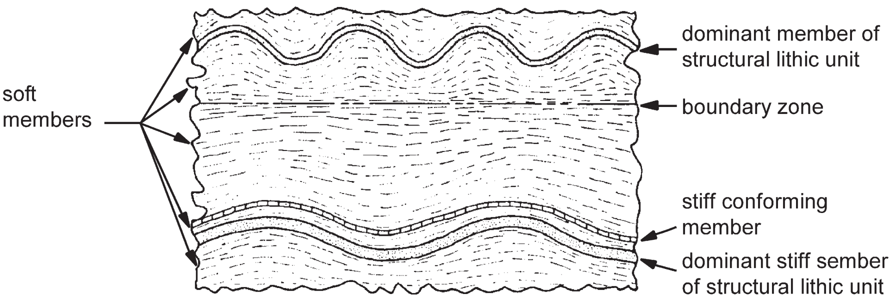


Fig. 1.15. Structural-lithic units. (After Currie et al. 1962)

Currie et al. 1962). A short-wavelength structural lithic unit may form inside the boundary of a larger-wavelength unit and be folded by it, in which case the longer-wavelength unit is termed the dominant structural-lithic unit and the shorter-wavelength unit is a conforming structural-lithic unit (Currie et al. 1962).

The dip changes within structural-lithic units may obscure the map-scale geometry. For example, the regional or map-scale dip in Figure 1.15 is horizontal, although few dips of this attitude could be measured. Where small-scale folds exist, the map-scale geometry may be better described by the orientation of the median surface or the enveloping surface (Fig. 1.16; Turner and Weiss 1963; Ramsay 1967). The median surface (median line in two dimensions) is the surface connecting the inflection points of a folded layer. The inflection points are located in the central region of the fold limbs where the fold curvature changes from anticlinal to synclinal. An enveloping surface is the surface that bounds the crests (high points) of the upper surface or the troughs (low points) of the lower surface on a single unit. Figure 1.16 shows that the dip observed at a single location need not correspond to the dip of either the median surface, the enveloping surface, or to the trace of a formation boundary. The dip of the median surface may be more representative of the dip required for map-scale interpretation than the locally observed bedding dips.

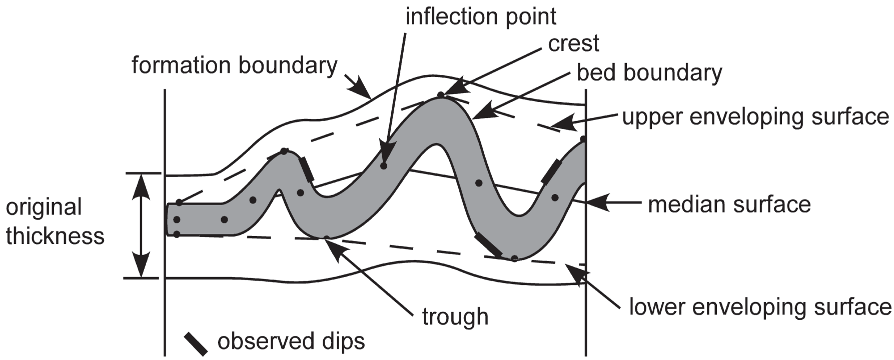


Fig. 1.16. Terminology for folded surfaces

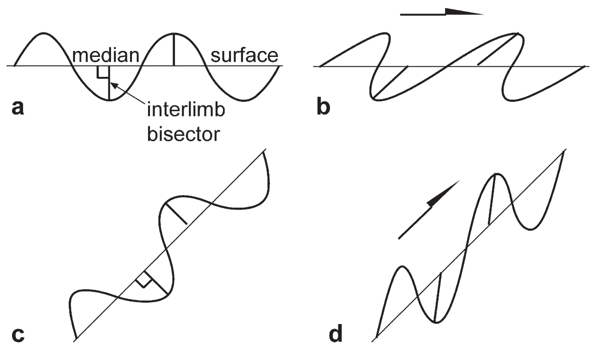


Fig. 1.17. Fold symmetry. **a** Symmetrical, upright. **b** Asymmetrical, overturned. **c** Symmetrical, overturned. **d** Asymmetrical, upright. *Arrow* gives direction of vergence

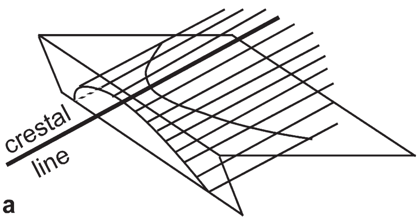
The symmetry of a fold is determined by the angle between the plane bisecting the interlimb angle and the median surface (Fig. 1.17a; Ramsay 1967). The angle is close to 90° in a symmetrical fold (Fig. 1.17a,c) and noticeably different from 90° in an asymmetrical fold (Fig. 1.17b,d). An essential property of an asymmetrical fold is that the limbs are unequal in length. Fold asymmetry is not related to the relative dips of the limbs. The folds in Fig. 1.17b,c have overturned steep limbs and right-way-up gentle limbs, but only the folds in Fig. 1.17b are asymmetric. This is a point of possible confusion, because in casual usage a fold with unequal limb dips (Fig. 1.17b,c) may be referred to as being asymmetrical. Folds may occur as regular periodic waveforms as shown (Fig. 1.17) or may be non-periodic with wavelengths that change along the median surface.

The vergence of an asymmetrical fold is the rotation direction that would rotate the axial surface of an antiform from an original position perpendicular to the median surface to its observed position at a lower angle to the median surface. The vergence of the folds in Fig. 1.17b,d is to the right.

1.5.2 Three-Dimensional Geometry

A cylindrical fold is defined by the locus of points generated by a straight line, called the fold axis, that is moved parallel to itself in space (Fig. 1.18a). In other words, a cylindrical fold has the shape of a portion of a cylinder. In a cylindrical fold every straight line on the folded surface is parallel to the axis. The geometry of a cylindrical fold persists unchanged along the axis as long as the axis remains straight. A conical fold is generated by a straight line rotated through a fixed point called the vertex (Fig. 1.18b). The cone axis is not parallel to any line on the cone itself. A conical fold changes geometry and terminates along the trend of the cone axis.

The crest line is the trace of the line which joins the highest points on successive cross sections through a folded surface (Figs. 1.18, 1.19a; Dennis 1967). A trough line is the trace of the lowest elevation on cross sections through a horizon. The plunge of a cylindrical fold is parallel to the orientation of its axis or a hinge line (Fig. 1.19b). The most useful measure of the plunge of a conical fold is the orientation of its crest line or trough line (Bengtson 1980).



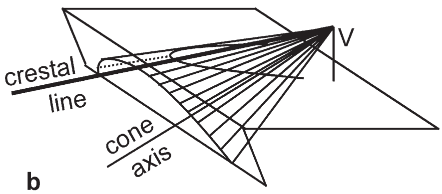


Fig. 1.18. Three-dimensional fold types. **a** Cylindrical. All *straight lines* on the cylinder surface are parallel to the fold axis and to the crestal line. **b** Conical. *V* vertex of the cone. *Straight lines* on the cone surface are not parallel to the cone axis

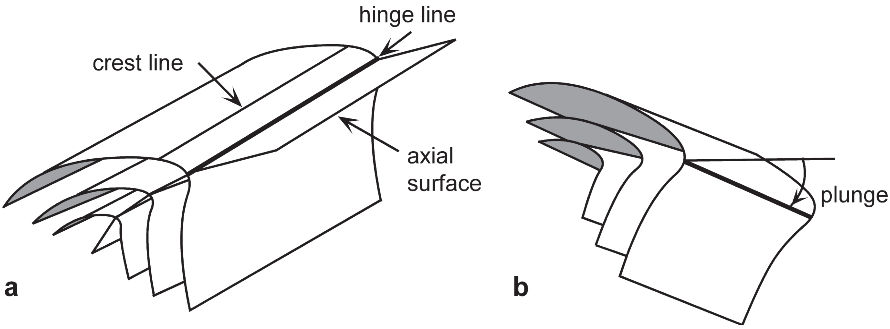


Fig. 1.19. Cylindrical folds. **a** Non-plunging. **b** Plunging

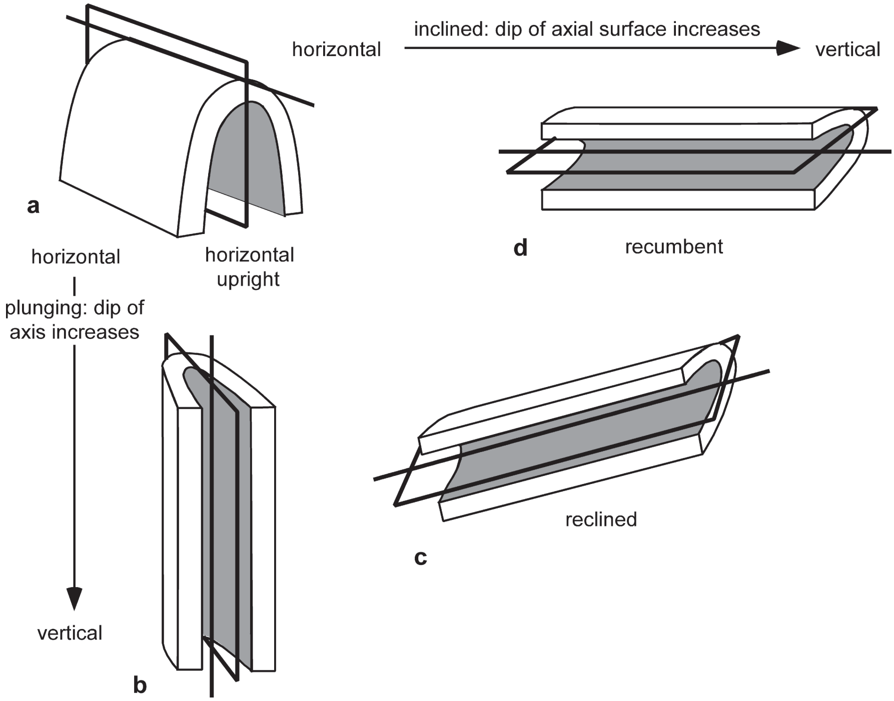


Fig. 1.20. Fold classification based on orientation of the axis and axial surface. **a** *Horizontal upright:* horizontal axis and vertical axial surface. **b** *Vertical:* vertical axis and vertical axial surface. **c** *Reclined:* inclined axis and axial surface. **d** *Recumbent:* horizontal axis and axial surface. (After Fleuty 1964)

The complete orientation of a fold requires the specification of the orientation of both the fold axis and the axial surface (Fig. 1.20). In the case of a conical fold, the orientation can be specified by the orientation of the axial surface and the orientation of the crestal line on a particular horizon. Common map symbols for folds are given in Fig. 1.21.

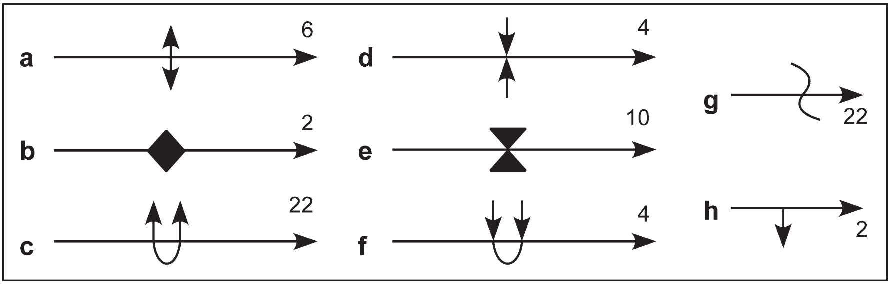


Fig. 1.21. Common map symbols for folds. Fold trend is indicated by the *long line*, plunge by the *arrowhead*, with amount of plunge given. The fold trend may be the axial trace, crest or trough line or a hinge line. **a**, **b** Anticline. **c** Overturned anticline; both limbs dip away from the core. **d**, **e** Syncline. **f** Overturned syncline; both limbs dip toward the core. **g** Minor folds showing trend and plunge of axis. **h** Plunging monocline with only one dipping limb

1.5.3 Mechanical Origins

The fundamental mechanical types of folding are based on the direction of the causative forces relative to layering (Ramberg 1963; Gzovsky et al. 1973; Groshong 1975), namely longitudinal contraction, transverse contraction, and longitudinal extension (Fig. 1.22). If the stratigraphy is without mechanical contrasts, forces parallel to layering produce either uniform shortening and thickening or uniform extension and thinning. If some shape irregularity is pre-existing, then it is amplified by layer-parallel shortening to give a passive fold. If the stratigraphy has significant mechanical contrasts, then a mechanical instability can occur that leads to buckle folding in contraction and pinch-and-swell structure (boudinage) in extension. If the forces are not equal vertically, then a forced fold is produced, regardless of the mechanical stratigraphy. Longitudinal contraction, transverse contraction, and longitudinal extension are end-member boundary conditions; they may be combined to produce folds with combined properties.

Buckle folds normally form with the fold axes perpendicular to the maximum principal compressive stress, σ_1. The folds are long and relatively unchanging in geometry parallel to the fold axis but highly variable in cross section. Buckle folds are characterized by the presence of a regular wavelength that is proportional to the thickness of the stiff unit(s). A single-layer buckle fold consists of a stiff layer in a surrounding confining medium. The dip variations associated with a given stiff layer die out into the regional dip within the softer units at a distance of about one-half arc length away from the layer (Fig. 1.15). In the author's experience, buckle folds in sedimentary rocks typically have arc-length to thickness ratios of 5 to 30, with common values in the range of 6 to 10.

As buckled stiff layers become more closely spaced, the wavelengths begin to interfere (Fig. 1.23) resulting in disharmonic folds. Once the layers are sufficiently closely spaced, they fold together as a multilayer unit. A multilayer unit has a much lower buckling stress and an appreciably shorter wavelength than a single layer of same thickness (Currie et al. 1962). Stiff units, either single layers or multilayers, tend to

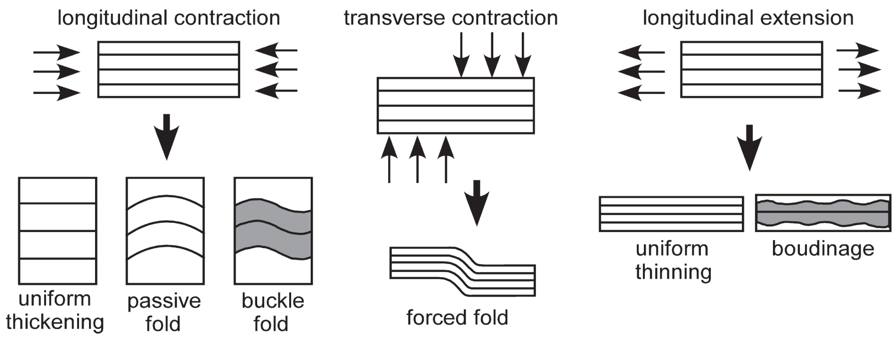


Fig. 1.22. End-member displacement boundary conditions showing responses related to the mechanical stratigraphy. *Shaded* beds are stiff lithologies, *unshaded* beds are soft lithologies

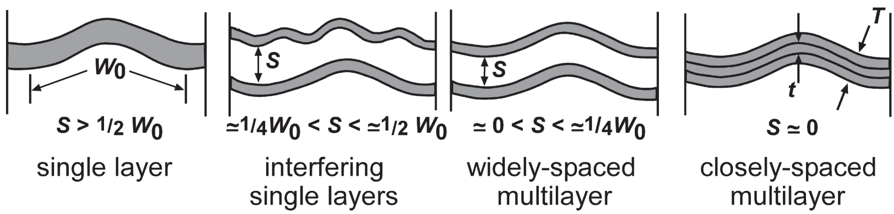


Fig. 1.23. Transition from single layer folding to multilayer folding as space between stiff layers decreases. The stiff layers are *shaded*

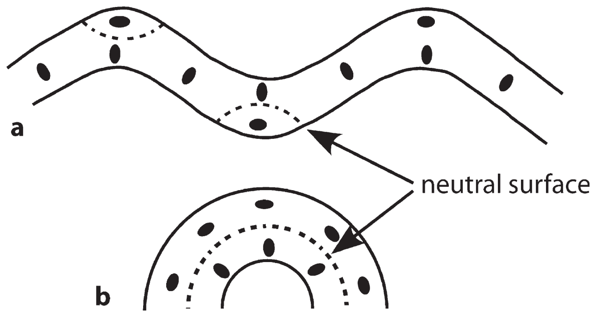


Fig. 1.24. Strain distribution shown by *strain ellipses* in folded stiff layers. The trace of the neutral surface of no strain is *dashed*. **a** Buckle fold. **b** Pure bend (Drucker 1967)

produce circular to sinusoidal folds if enclosed in a sufficient thickness of softer material (Fig. 1.13). Stratigraphic sections made up of relatively thin-bedded multilayer units result in folds that tend to have planar dip-domain styles (Fig. 1.14).

The strain distribution in the stiff layer of a buckle fold (Fig. 1.24a) is approximately layer-parallel shortening throughout most of the fold. The neutral surface separates regions of layer-parallel extension from regions of layer-parallel contraction. Only in a pure bend (Fig. 1.24b) are the areas of extension and contraction about equal and the neutral surface in the middle of the layer. The strain in thick soft layers between stiff layers is shortening approximately perpendicular to the axial surface. The strain in thin soft layers between stiff layers may be close to layer-parallel shear strain. The pure bending strain distribution is usually more closely approached in transverse contraction folds, for example above a salt dome, than in buckle folds.

Cleavage planes and tectonic stylolites in a fold can indicate the mechanical origin of the fold because they form approximately perpendicular to the maximum shortening direction by processes that range from grain rotation to pressure solution (Groshong 1988). Cleavage in a buckle fold is typically at a high angle to bedding (Fig. 1.25), being more nearly perpendicular to bedding in stiff units and more nearly parallel to the axial surface in soft units. Cleavage that is approximately perpendicular to bedding produces a cleavage fan across the fold. The line of the cleavage-bedding intersection is approximately or exactly parallel to the fold axis and can be used to help determine the axis.

Folds produced by an unequal distribution of forces in transverse contraction (Fig. 1.22) are termed forced folds (Stearns 1978). Forced folds tend to be round to blocky or irregular in map view. The major control on the form of the fold is the rheology of the forcing member (Fig. 1.26). A stiff and brittle forcing member (i.e., crystalline basement) leads to narrow fault boundaries at the base of the structure and strain that is highly localized in the zone above the basement fault. A soft unit between a stiff forcing member and the cover sequence will cause the deformation to be disharmonic. A soft forcing member (like salt) typically produces round to elliptical structures with deformation widely distributed across the uplift.

Little strain need occur in the uplifted or downdropped blocks associated with a stiff forcing member. Nearly all the strain is localized in the fault zone between the

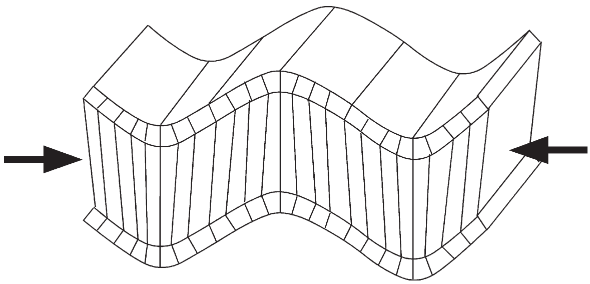


Fig. 1.25.
Cleavage pattern in a buckle fold. The gently dipping surfaces are bedding and the steeply dipping surfaces represent cleavage or stylolites. *Arrows* show directions of the boundary displacements

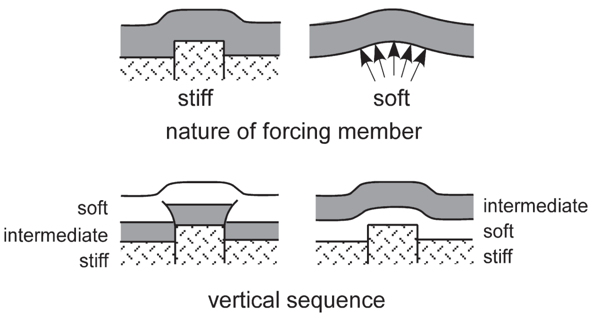


Fig. 1.26.
Effect of mechanical stratigraphy on drape folds. The lowest unit is the forcing member

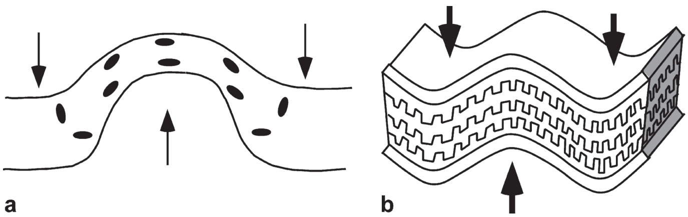
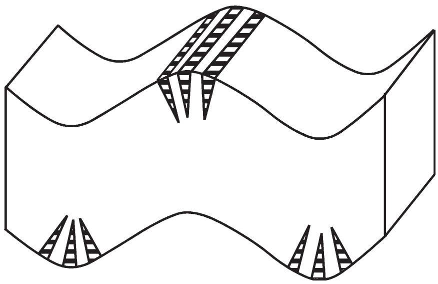


Fig. 1.27. Strain and cleavage patterns in transverse contraction folds produced by differential vertical displacement. **a** Strain distribution above a model salt dome (after Dixon 1975). **b** Cleavage or stylolites parallel to bedding. *Arrows* show directions of the boundary displacements

Fig. 1.28.
Veins due to outer-arc bending stresses



blocks or in the steep limb of the drape fold over the fault zone. A soft forcing member (i.e., salt) will distribute the curvature and strain widely over the uplifted region (Fig. 1.27a). Because cleavage and stylolites form perpendicular to the shortening direction, in folds produced by displacements at a high angle to bedding, the expected cleavage and stylolite direction is parallel to bedding (Fig. 1.27b). In highly deformed rocks, cleavage parallel to bedding might be the result of deformation caused by a large amount of layer-parallel slip or by isoclinal refolding of an earlier axial-plane cleavage.

Extension fractures and veins may form due to the bending stresses in the outer arc of a fold (Fig. 1.28). Such features should become narrower and die out toward the neutral surface. The fracture plane is expected to be approximately parallel to the axis of the fold and the fracture-bedding line of intersection should be parallel to the fold axis. Bending fractures might occur in any type of fold.

1.6 Faults

A fault (Fig. 1.29) is a surface or narrow zone across which there has been relative displacement of the two sides parallel to the zone (after Bates and Jackson 1987). The term displacement is the general term for the relative movement of the two sides of the fault, measured in any chosen direction. A shear zone is a general term for a relatively narrow zone with subparallel boundaries in which shear strain is concentrated (Mitra and Marshak 1988). As the terms are usually applied, a bed, foliation trend, or other marker horizon maintains continuity across a shear zone but is broken and displaced

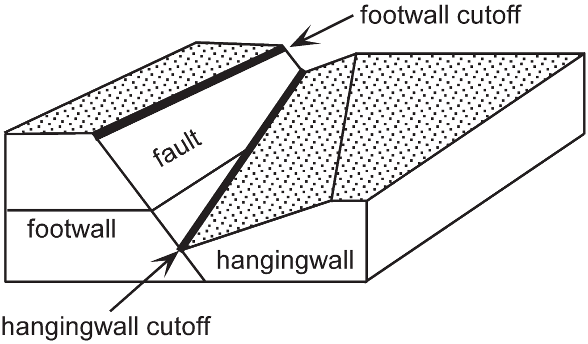


Fig. 1.29.
General terminology for a surface (*patterned*) offset by a fault. *Heavy lines* are hangingwall and footwall cutoff lines

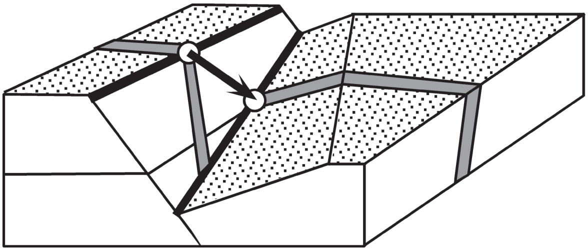


Fig. 1.30.
Fault slip is the displacement of points (*open circles*) that were originally in contact across the fault. Here the correlated points represent the intersection line of a dike and a bed surface at the fault plane

across a fault. It may be difficult to distinguish between a shear zone and a fault zone on the basis of observations at the map scale, and so here the term fault will be understood to include both faults and shear zones.

The term hangingwall refers to the strata originally above the fault and the term footwall to strata originally below the fault (Fig. 1.29). Because of the frequent repetition of the terms, hangingwall and footwall will often be abbreviated as HW and FW, respectively. A cutoff line is the line of intersection of a fault and a displaced horizon (Fig. 1.29). The HW and FW cutoff lines of a single horizon were in contact across a fault plane prior to displacement. Across a fault zone of finite thickness or across a shear zone, the HW and FW cutoffs were originally separated by some width of the offset horizon that is now in the zone.

1.6.1
Slip

Fault slip is the relative displacement of formerly adjacent points on opposite sides of the fault, measured along the fault surface (Fig. 1.30; Dennis 1967). Slip can be subdivided into horizontal and vertical components, the strike slip and dip slip components, respectively. A fault in which the slip direction is parallel to the trace of the cutoff line of bedding can be called a trace-slip fault. In horizontal beds a trace-slip fault is a strike-slip fault. Measurement of the slip requires the identification of the piercing points of displaced linear features on opposite sides of the fault. Suitable linear features at the map scale might be a channel sand, a facies boundary line, a fold hinge line,

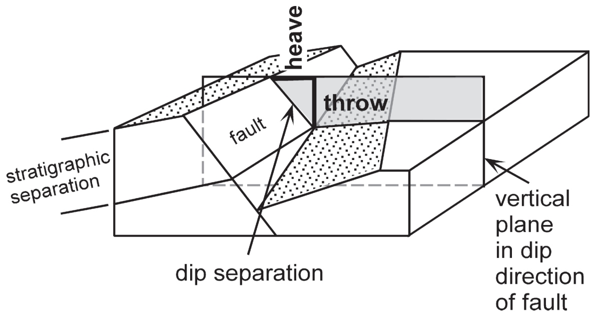


Fig. 1.31.
Fault separation terminology

or the intersection line between bedding and a dike (Fig. 1.30). The displacement of dipping beds on faults oblique to the strike of bedding leads to complex relationships between the displacement and the slip in a specific cross-section direction, such as parallel to the dip of bedding. A strike-slip component of displacement is never visible on a vertical cross section.

1.6.2 Separation

Fault separation is the distance between any two parts of an index plane (e.g., bed or vein) disrupted by a fault, measured in any specified direction (Dennis 1967). The separation directions commonly important in mapping are parallel to fault strike, parallel to fault dip, horizontal, vertical and perpendicular to bedding. It should be noted that the definitions of the terms for fault separation and the components of separation are not always used consistently in the literature. Stratigraphic separation (Fig. 1.31) is the thickness of strata that originally separated two beds brought into contact at a fault (Bates and Jackson 1987) and is the stratigraphic thickness missing or repeated at the point, called the fault cut (Tearpock and Bischke 2003), where the fault is intersected. The amount of the fault cut is always a stratigraphic thickness.

Throw and heave (Fig. 1.31) are the components of fault separation most obvious on a structure contour map. Both are measured in a vertical plane in the dip direction of the fault. Throw is the vertical component of the dip separation measured in a vertical plane (Dennis 1967). Stratigraphic separation is not equal to the fault throw unless the marker horizons are horizontal (see Sect. 5.5.3). Heave is the horizontal component of the dip separation measured in a vertical plane normal to the strike of the fault (Dennis 1967).

1.6.3 Geometrical Classifications

A fault is termed normal or reverse on the basis of the relative displacement of the hangingwall with respect to the footwall (Fig. 1.32). For a normal fault, the hangingwall is displaced down with respect to the footwall, and for a reverse fault the hangingwall is displaced up with respect to the footwall. The relative displacement may be either a slip or a separation and the use of the term should so indicate, for example, a *normal-*

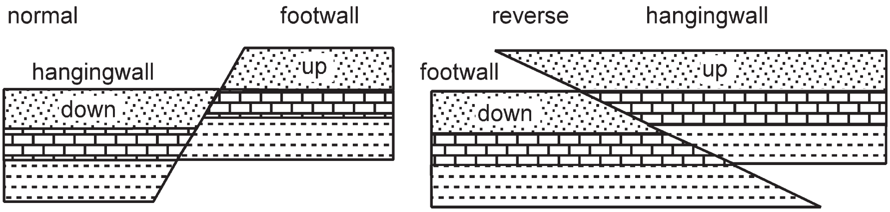
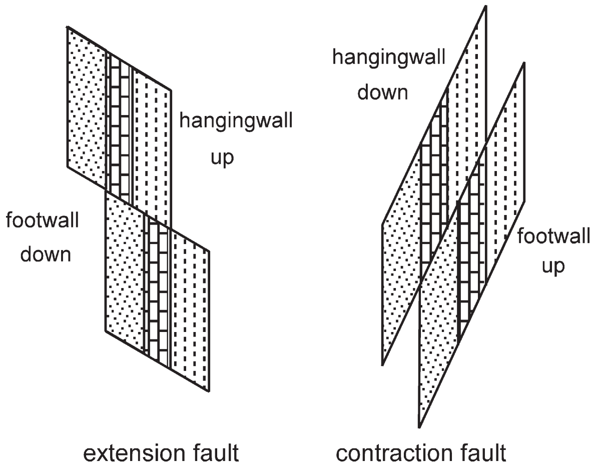


Fig. 1.32. Vertical cross section showing the relative fault displacement terminology with horizontal as the reference plane

Fig. 1.33.
Vertical cross section showing the relative fault displacement terminology with bedding as the reference plane



separation fault. Using the horizontal as the plane of reference (i.e., originally horizontal bedding), a normal-separation fault extends a line parallel to bedding and a reverse-separation fault shortens the line.

Using bedding as the frame of reference is not the same as using a horizontal plane, as illustrated by Fig. 1.33 which shows the faults from Fig. 1.32 after a 90° rotation. With bedding vertical, a reverse displacement (Fig. 1.33) extends the bedding while shortening a horizontal line. The fault might have been caused by reverse slip on a fault formed after the beds were rotated to vertical or by the rotation of a normal fault. Using bedding as the frame of reference (Norris 1958), an extension fault extends the bedding, regardless of the dip of bedding, and a contraction fault shortens the bedding.

A fault cut is the point at which a well crosses a fault. A fault with a component of dip separation has the effect of omitting or repeating stratigraphy across the fault at the fault cut (Fig. 1.34). With respect to a vertical line or a vertical well, a normal fault causes the omission of stratigraphic units (Fig. 1.34a) and a reverse fault causes the repetition of units (Fig. 1.34b). Opposite-sense omissions or repetitions may occur in a well that is not vertical (Mulvany 1992). For example, a well drilled from the footwall to the hangingwall of a normal fault will show repeated section down the well (Fig. 1.34a) and will show missing section down the well if the fault is reverse (Fig. 1.34b).

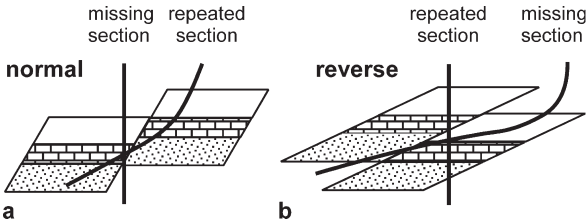


Fig. 1.34.
Effect of well orientation on occurrence of missing or repeated section. All units are right side up and cross sections are vertical. **a** Wells penetrating a normal fault. **b** Wells penetrating a reverse fault

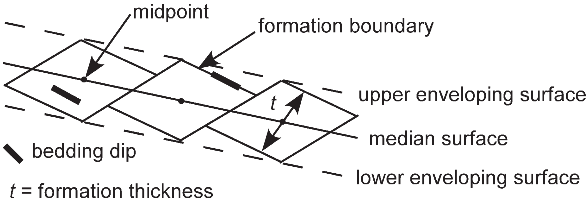


Fig. 1.35.
Regional dip of faulted bedding surface, indicated by enveloping surfaces is different than bedding dip

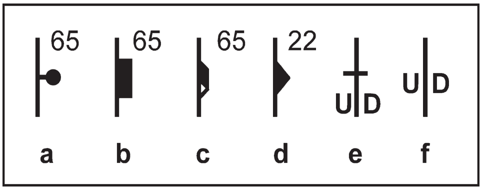


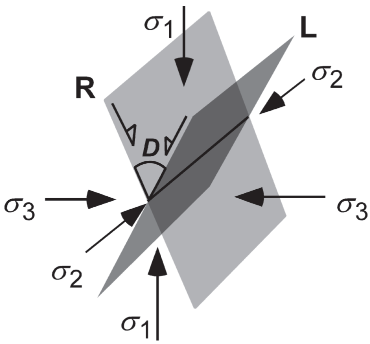
Fig. 1.36.
Map symbols for faults indicating separation. **a–c** Normal separation, *symbol* on hangingwall. **d** Reverse separation, *triangle* on hangingwall. **e** Vertical fault, vertical separation indicated (*U* up; *D* down). **f** Fault of unknown dip, vertical separation indicated

The median surface of a faulted unit (median line in two dimensions) is the surface connecting the midpoints of the blocks in the middle of the reference unit (Fig. 1.35). An enveloping surface is the surface that bounds the high corners or the low corners on a single unit. Dips within the fault blocks may all be different from the dip of either the median surface or the enveloping surface. Within a fault block the original thickness may remain unaltered by the deformation (Fig. 1.35), although the entire unit has been thickened or thinned as indicated by the changed thickness between the enveloping surfaces. Common map symbols for faults are given in Fig. 1.36.

1.6.4
Mechanical Origins

Faults commonly initiate in conjugate pairs (Fig. 1.37). Conjugate faults form at essentially the same time under the same stress state. This geometry has been produced in countless experiments (Griggs and Handin 1960). The acute angle between the two conjugate faults is the dihedral angle which is usually in the range of 30 to 60° but may be significantly smaller if the least principal stress is tensile (Ramsey and Chester 2004). In experiments the maximum principal compressive stress, σ_1, bisects the dihedral angle. The least principal compressive stress, σ_3, bisects the obtuse angle,

Fig. 1.37. Conjugate pair of faults related to orientation of principal stresses. *D:* dihedral angle. The principal stresses are σ_1, σ_2, and σ_3, in order of greatest to least compressive stress. *R* right-lateral fault plane; *L* left-lateral fault plane; *half-arrows* show the sense of shear on each plane



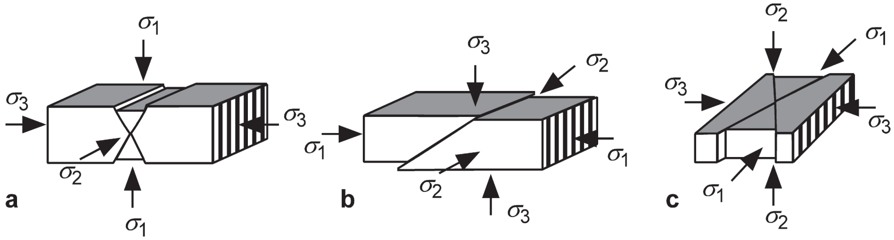


Fig. 1.38. Fault orientations at the surface of the earth predicted from Andersonian stress theory. **a** Normal. **b** Reverse. **c** Strike slip

and the intermediate principal stress, σ_2, is parallel to the line of intersection of the two faults. The slip directions are directly related to the orientation of the principal stresses (Fig. 1.37), with one set being right lateral (dextral) and the other set left lateral (sinestral).

The surface of the ground is a plane of zero shear stress and therefore one of the principal stresses is perpendicular to the surface and the other two principal stresses lie in the plane of the surface (Anderson 1905, 1951). From the experimental relationship between fault geometry and stress (Fig. 1.37), this leads to a prediction of the three most common fault types and their dips (Fig. 1.38). Relative to the horizontal, normal faults typically dip 60°, reverse faults average 30°, and strike-slip faults are vertical.

The predicted dips in Fig. 1.38 are good for a first approximation, but there are many exceptions. Fault orientations may be controlled by lithologic differences, changes in the orientations of the stress field below the surface of the ground, and by the presence of pre-existing zones of weakness. True triaxial stress states can result in the formation of two pairs of conjugate faults having the same dips but slightly different strikes, forming a rhombohedral pattern of fault blocks (Oertel 1965). Oertel faults are likely to be arranged in low-angle conjugate pairs that are 10–30° oblique to each other. Faults will rotate to different dips as the enclosing beds rotate. Even with all the exceptions, it is still common for faults to have the approximate orientations given in Fig. 1.38.

1.6.5
Fault-Fold Relationships

A planar fault with constant displacement (Fig. 1.39a) is the only fault geometry that does not require an associated fold as a result of its displacement. Of course, all faults eventually lose displacement and end. A fault that dies out without reaching the surface of the earth is called blind, and a fault that reaches the present erosion surface is emergent, although whether it was emergent at the time it moved may not be known. Where the displacement ends at the tip of a blind fault, a fold must develop (Fig. 1.39b). Displacement on a curved fault will cause the rotation of beds in the hangingwall and perhaps in the footwall and will produce a fold (Fig. 1.39c). A generic term for the fold is a ramp anticline. The fold above a normal fault is commonly called a rollover anticline if the hangingwall beds near the fault dip toward the fault.

Fault dips may be controlled by the mechanical stratigraphy to form ramps and flats, although at the scale of the entire fault, the average dip may be maintained (i.e., 30° for a reverse fault). A flat is approximately parallel to bedding, at an angle of say,

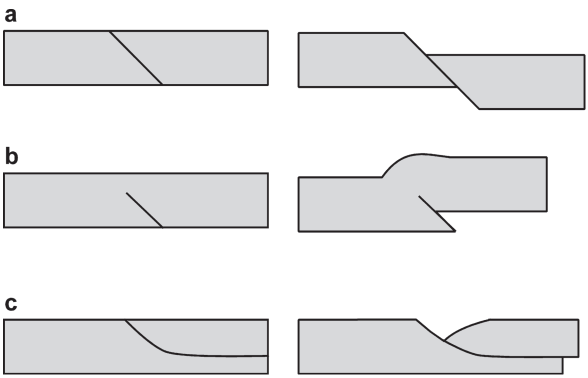


Fig. 1.39. Relationships between folds and faults. **a** Constant slip on a planar fault does not cause folding. **b** Slip on either a plane or curved fault that dies out produces a fold in the region of the fault tip. **c** Slip on a curved fault causes folding in the hangingwall

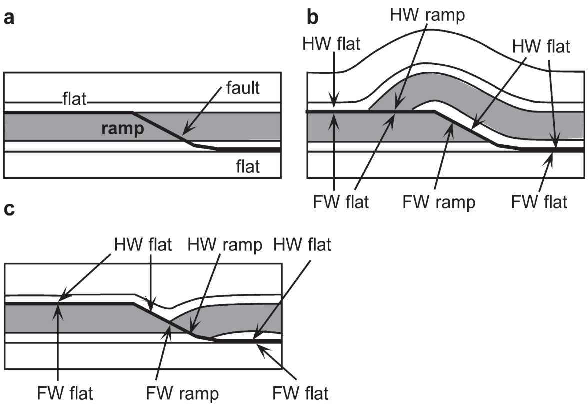


Fig. 1.40. Ramp-flat fault terminology. *HW* is hangingwall; *FW* is footwall. **a** Before displacement. **b** After reverse displacement. **c** After normal displacement. (After Woodward 1987)

10° or less. A ramp crosses bedding at an angle great enough to cause missing or repeated section at the scale of observation, say 10° or more. Characteristicly, but not exclusively, ramps occur in stiff units such as limestone, dolomite, or cemented sandstone, whereas flats occur in soft units, such as shale or salt. Both normal and reverse faults may have segments parallel to bedding and segments oblique to bedding (Fig. 1.40). After displacement, hangingwall ramps and flats no longer necessarily match across the fault (Fig. 1.40b,c).

Another common fault geometry is listric (Fig. 1.41a) for which the dip of the fault changes continuously from steep near the surface of the earth to shallow or horizontal at depth. Both normal and reverse faults may be listric. The lower detachment of a listric fault is typically in a weak unit such as shale, overpressured shale or salt. Faults that flatten upward are comparatively rare, except in strike-slip regimes, and may be termed antilistric. Antilistric reverse faults have been produced by the stresses above a rigid block uplift (Sanford 1959).

In three dimensions, fault ramps are named according to their orientation with respect to the transport direction (Dahlstrom 1970; McClay 1992). Frontal ramps are approximately perpendicular to the transport direction, lateral ramps are approximately parallel to the transport direction and oblique ramps are at an intermediate angle (Fig. 1.42a). Displacement of the hangingwall produces ramp anticlines having orientations that correspond to those of the ramps (Fig. 1.42b). The ramp terminology was developed for thrust-related structures but is equally applicable to normal-fault structures. Lateral and oblique ramps necessarily have a component of strike slip and lateral ramps may be pure strike-slip faults.

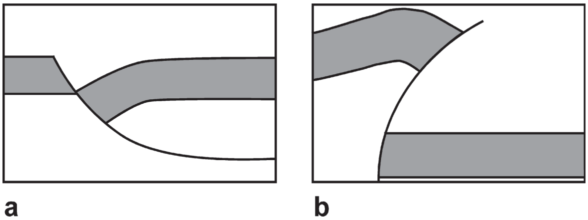


Fig. 1.41. Typical curved fault shapes in cross section. **a** Listric. **b** Antilistric

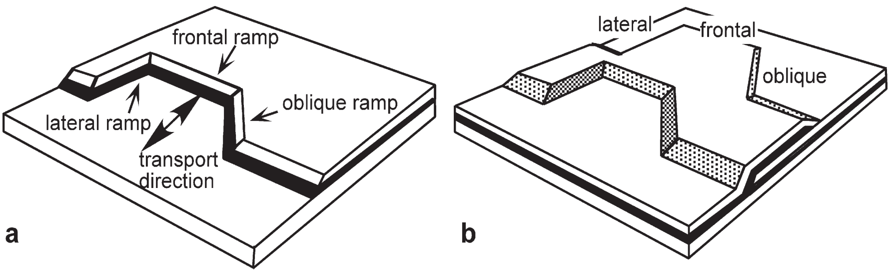


Fig. 1.42. Ramps and ramp anticlines in three dimensions. **a** Ramps in the footwall. *Arrows* indicate transport direction. **b** Thrust ramp anticlines. (After McClay 1992)

1.7 Sources of Structural Data and Related Uncertainties

The fundamental information generally available for the interpretation of the structure in an area is the attitude of planes and locations of the contacts between units. The primary sources of this information are direct observations of exposures, well logs, and seismic reflection profiles. These data are never complete and may not be correct in terms of the exact locations or attitudes. Before constructing or interpreting a map, it is worth considering the uncertainties inherent in the original data.

1.7.1 Direct Observations

Outcrop- and mine-based maps are constructed from observations of the locations of contacts and the attitudes of planes and lines. Good practice is to show on the working map the areas of exposure at which the observations have been made (Fig. 1.43a). Exposure is rarely complete and so uncertainties typically exist as to the exact locations of contacts. Surface topography is usually directly related to the underlying geology and should be used as a guide to contact locations. Contacts that control topography may be traced with a reasonable degree of confidence, even in the absence of exposure, at least where the structure is simple. The assignment of a particular exposure to a specific stratigraphic unit may be in doubt if diagnostic features are absent. Bedding attitudes measured in small outcrops might come from minor folds, minor fault blocks, or cross beds and not represent the attitude of the formation boundaries. The connectivity of the contacts in the final map is usually an interpretation, not an observation (Fig. 1.43b).

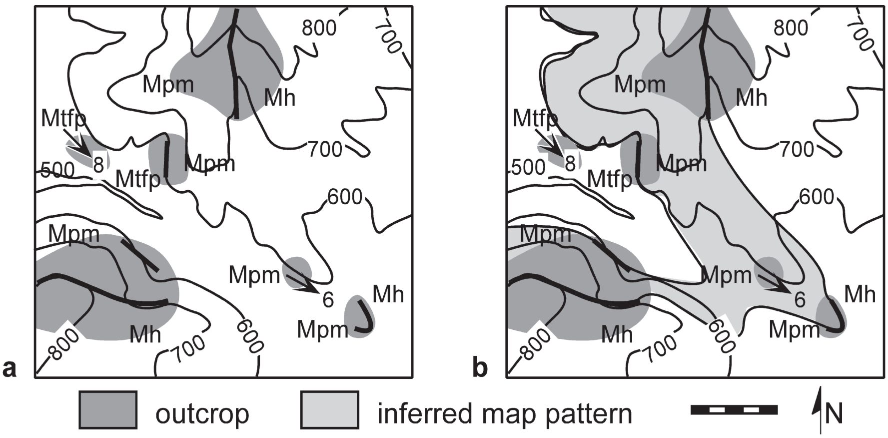


Fig. 1.43. Geologic map on a topographic base. *Contours* are in feet and the scale bar is 1 000 ft. Three formations are present, from oldest to youngest: *Mtfp*, *Mpm*, *Mh*. Attitude of bedding is shown by an *arrow* pointing in the dip direction, with the dip amount indicated. **a** Outcrop map showing locations of direct observations (*shaded*). **b** Completed geologic map, contacts *wide* where observed, *thin* where inferred. *Lighter shading* is the interpreted outcrop area of the *Mpm*

1.7.2 Wells

Wells provide subsurface information on the location of formation boundaries and the attitude of planes. Measurements of this information are made by a variety of techniques and recorded on well logs. Sample logs are made from cores or cuttings taken from the well as it is drilled. In wells drilled with a cable tool, cuttings are collected from the bottom of the hole every 5 or 10 feet and provide a sample of the rock penetrated in that depth interval. In wells that are rotary drilled, drilling fluid is circulated down the well and back to remove the cuttings from the bottom of the hole. The drilling fluid is sampled at intervals as it reaches the surface to determine the rock type and fossil content of the cuttings. Depths are calculated from the time required for the fluid to traverse the length of the hole and are not necessarily precise.

A wire-line log is a continuous record of the geophysical properties of the rock and its contained fluids that is generated by instruments lowered down a well. Lithologic units and their contained fluids are defined by their log responses (Asquith and Krygowski 2004; Jorden and Campbell 1986). Two logs widely used to identify different units are the spontaneous potential (SP) and resistivity logs (Fig. 1.44). In general, more permeable units show a larger negative SP value. The resistivity value depends on presence of a pore fluid and its salinity. Rocks with no porosity or porous rocks filled with oil generally have high resistivities and porous rocks containing saltwater have low resistivities. A variety of other log types is also valuable for lithologic interpretation, including gamma-ray, neutron density, sonic and nuclear magnetic resonance logs. The gamma-ray log responds to the natural radioactivity in the rock. Very radioactive (hot) black shales are often widespread and make good markers for correlations between wells. A caliper log measures the hole diameter in two perpendicular directions. Weak lithologies like coal or fractured rock can be recognized on a caliper log by intervals of hole enlargement. In wells drilled with mud, fluid loss into very porous lithologies or open fractures may cause mud cakes that will be recognized on a caliper log by a reduced well-bore size.

Logs from different wells are correlated to establish the positions of equivalent units (Levorsen 1967; Tearpock and Bischke 2003). Geologic contacts may be correlated from well to well to within about 30 ft in a lithologically heterogeneous sequence or to within inches or less on high-resolution logs in laterally homogeneous lithologies. The cable that lowers the logging tool into the well stretches significantly in deep wells. The recorded depth is corrected for the stretch, but the correction may not be exact. Different log runs, or a log and a core, may differ in depth to the same horizon by 20 ft at 10 000 ft. Normally, different log runs will duplicate one of the logs, for example the SP, so that the runs can be accurately correlated with each other.

The orientation of the well bore is measured by a directional survey. Some wells, especially older ones, may be unintentionally deviated from the vertical and lack a directional survey, resulting in spatial mislocation of the boundaries recorded by the well logs (if interpreted as being from a vertical well) which will lead to errors in dip and thickness determinations. The most common effect is for a well to wander down dip with increasing depth.

A dipmeter log is a microresistivity log that simultaneously measures the electrical responses of units along three or more tracks down a well (Schlumberger 1986). The

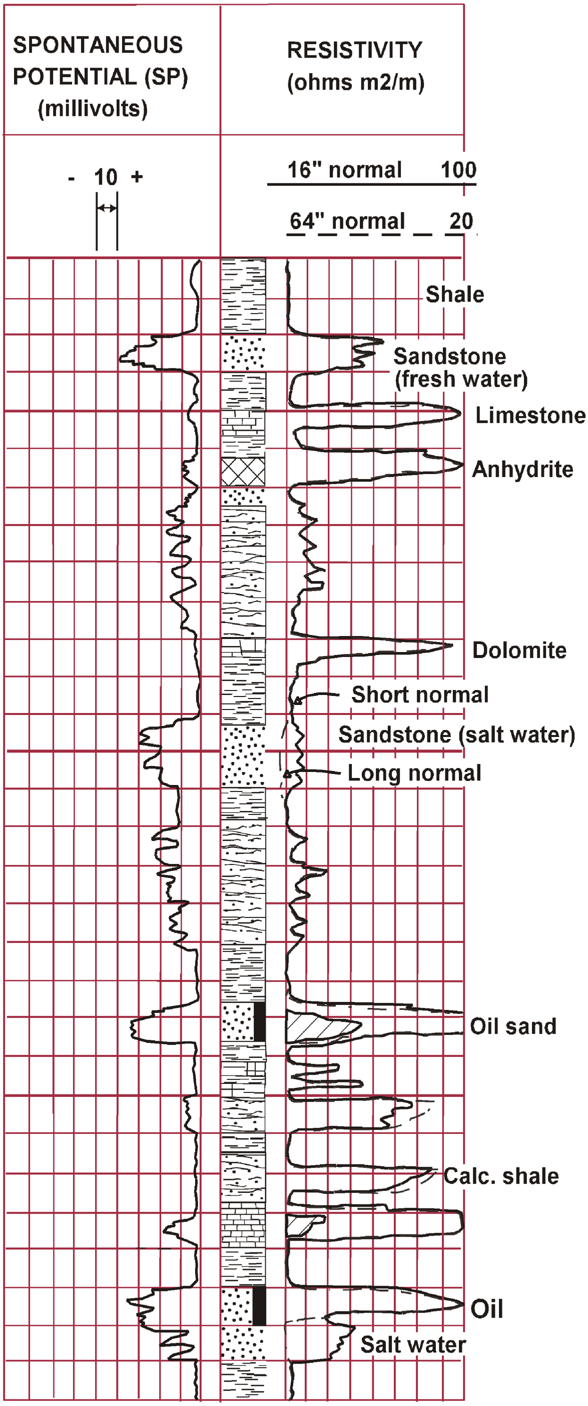


Fig. 1.44.
Typical example of electric logs used for lithologic and fluid identification. The interpreted lithologic column is in the *center*. Short normal (16 in) and long normal (64 in) refer to spacing between electrodes on the resistivity tool. (After Levorsen 1967)

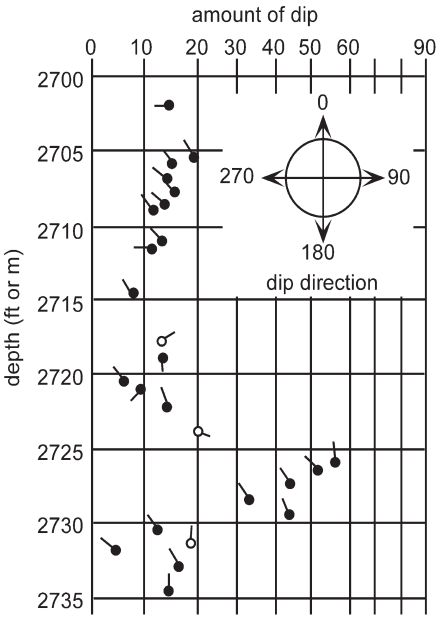


Fig. 1.45. Representative segment of a dipmeter log. The depth scale could be in feet or meters. *Solid points* indicate the higher quality correlations, *open points* lower quality correlations

responses are correlated around the borehole, and the dip of the unit is determined by a version of the 3-point method (Sect. 2.4.2). Dips may be calculated for depth intervals as small as 8–16 cm. A typical record (Fig. 1.45) shows the dips as "tadpoles", the heads of which mark the amount of dip and the tails of which point in the direction of dip. The dips presented on the log are corrected for well deviation.

The correlations required to determine the dip for a dipmeter log are not always possible and may not always be correct. On the printed log, solid points (Fig. 1.45) indicate the highest quality correlations and open points indicate lower quality correlations. Sparse data or gaps on the dipmeter record indicate that no correlations were possible, a likely occurrence in a very homogeneous lithology (including fault gouge). Closely spaced dips that are scattered in amount and direction, such as between the depths of 2715 and 2725 in Fig. 1.45, suggest miscorrelations or perhaps small-scale bedding features, and are probably not reliable dips for structural purposes. A log may use a special symbol to show dips that are consistent over vertical intervals five or more times that of the minimum correlation interval. These large-interval dips are more likely to represent the structural dip. The correlations in a dipmeter log are made by scanning some distance (the scan angle) up and down the individual tracks to look for correlations. If the angle between the well bore and bedding is small (equivalent to a steep dip in a vertical well), the correlative units may lie outside the search interval. Thus dipmeter logs rarely show dips that are at angles of less than 30–40° from the well bore (50–60° dip in a vertical well) unless they were specifically programmed to look for them. If the dipmeter interpretation program is unable to make good correlations across the well it will probably show either no dips in the interval or may have made false correlations and so show low-quality dips at a high angle to the well bore.

1.7.3 Seismic Reflection Profiles

Many interpretations of subsurface structure are based on seismic reflection profiles. Sound energy generated at or near the earth's surface is reflected by various layer boundaries in the subsurface. The time at which the reflection returns to a recorder at the surface is directly related to the depth of the reflecting horizon and the velocity of sound between the surface and the reflector. Seismic data are commonly displayed as maps or cross sections in which the vertical scale is the two-way travel time (Fig. 1.46).

The geometry of a structure that is even moderately complex displayed in travel time is likely to be significantly different from the true geometry of the reflecting boundaries because of the distortions introduced by steep dips and laterally and vertically varying velocities (Fig. 1.47). Reflections from steeply dipping units may return

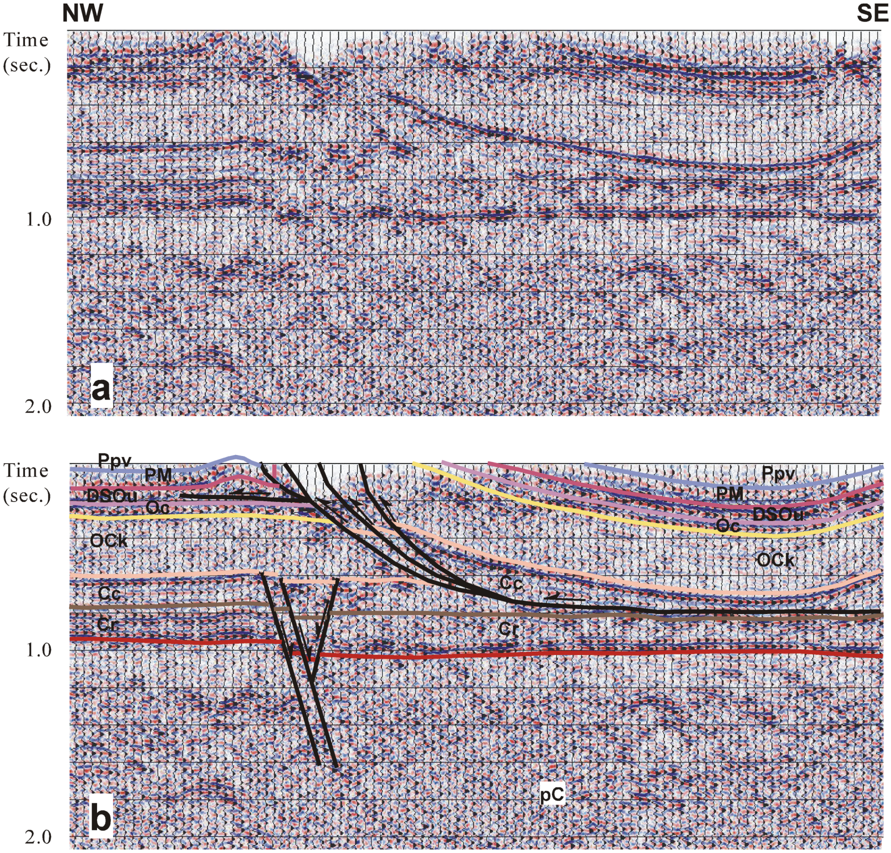


Fig. 1.46. Time-migrated seismic profile from southern Appalachian fold-thrust belt (Maher 2002), displayed with approximately no vertical exaggeration. The vertical scale is two-way travel time in seconds. **a** Uninterpreted. **b** Interpreted

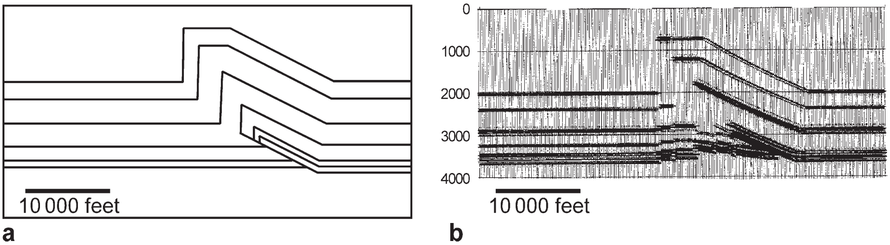


Fig. 1.47. Seismic model of a faulted fold. **a** Geometry of the model, no vertical exaggeration. **b** Model time section based on normal velocity variations with lithology and depth. Vertical scale is two-way travel time in milliseconds. (After Morse et al. 1991)

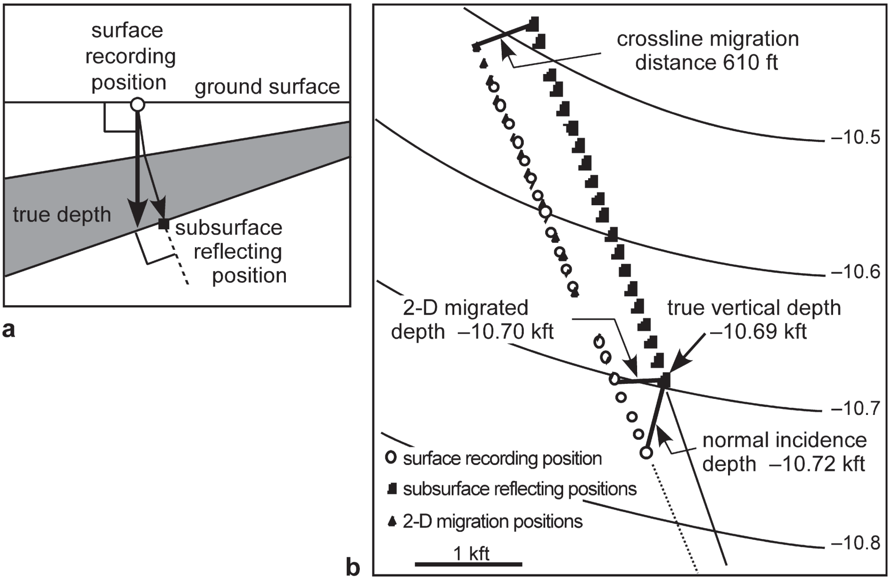


Fig. 1.48. Mislocation of seismic reflection points caused by dip of the reflector. **a** Ray path end points in vertical cross section. **b** Structure contours on a seismic reflector, depths subsea in kilofeet, showing actual and interpreted locations of the reflecting points on a seismic line. (After Oliveros 1989)

to the surface beyond the outer limit of the recording array and so are not represented on the seismic profile. The structural interpretation of seismic reflection data requires the conversion of the travel times to depth. This requires an accurate model for the velocity distribution, something not necessarily well known for a complex structure. The most accurate depth conversion is controlled by velocities measured in nearby wells (Harmon 1991).

If the trend of a seismic line is oblique to the dip of the reflector surface, two-dimensional reflection data have location problems similar to those of unknowingly deviated wells. This is in addition to the location problems associated with the conversion of travel time to depth. Reflections are interpreted to originate along ray paths that are normal to the reflector boundaries (Fig. 1.48a). A normal-incidence seismic ray is deflected up the dip (Fig. 1.48a). The true location of the reflecting point is up the dip and at a shallower depth than a point directly below the surface recording position. When the locations of reflecting points are plotted as if they were vertically below the surface recording stations, there will be a decrease in the calculated depth (or two-way travel time) relative to the true depth. Two-dimensional time migration is a standard processing procedure that corrects for the apparent dip of reflectors in the plane of the seismic line, but does not correct for the shift of the reflector positions in the true dip direction. Two-dimensional depth migration may give the correct depth to the reflecting point, but still does not correct for the out-of-plane position shift. For example (Fig. 1.48b), four degrees of oblique dip leads to a 400- to 600-ft shift in the true position of the reflection points on a seismic section at depths of 10 500–10 800 ft.