CHAPTER ELEVEN

# Fabrics: Foliations and Lineations



## 11.1 INTRODUCTION

In everyday language, we commonly use the word "fabric." When talking about fabrics that are used to make garments, we mean a patterned cloth made by weaving fibers in some geometric arrangement. But the word "fabric" is not used only to refer to material products. In a philosophical moment we might consider the "fabric of life," by which we mean the underlying organization of life. As we found to be the case with many terms, the word fabric has a related yet somewhat different meaning in geology. To a structural geologist, the **fabric** of a rock is the geometric arrangement of component features in the rock, seen on a scale large enough to include many samples of each feature. The features themselves are called **fabric elements.** Examples of fabric elements include mineral grains, clasts, compositional layers, fold hinges, and planes of parting. Fabrics that form as a consequence of tectonic deformation of rock are called **tectonic fabrics,** and fabrics that form during the formation of the rock are called **primary fabrics** (Chapter 2). It may sound picky, but structural geologists also make a distinction between microstructure and texture. Although texture is sometimes used as a synonym for microstructure, for example igneous texture, here we restrict the term **texture** to crystallographic orientation patterns in an aggregate of grains (see Chapter 13) and **microstructure** to their geometric arrangement.

Tectonic fabrics provide clues to the strain state of the rock, the geometry of associated folding, the processes involved in deformation, the kinematics of deformation, the timing of deformation (if the fabric is defined by an arrangement of datable minerals), and ultimately about the tectonic evolution of a region. The purpose of this chapter is to explore two common tectonic fabric elements in rocks, **foliations** and **lineations,** and to introduce you to the characteristics and interpretation of these elements.

## 11.2 FABRIC TERMINOLOGY

Let's start by developing the inevitable vocabulary to discuss tectonic fabrics (see Table 11.1). If there is no preferred orientation (i.e., alignment) of the fabric elements, then we say that the rock has a **random fabric**

| TABLE 11.1 | TECTONIC FABRIC TERMINOLOGY |
|---|---|
| Axial plane cleavage | Cleavage that is parallel or subparallel to the axial plane of a fold; it is generally assumed that the cleavage formed roughly synchronous with folding and is subparallel to the *XY*-plane of the bulk finite strain ellipsoid. |
| Cleavage | A secondary fabric element, formed under low-temperature conditions, that imparts to the rock a tendency to split along planes. |
| Cross-cutting cleavage | Cleavage that is not parallel to the axial plane of a fold (also *nonaxial plane cleavage*); the term *transecting cleavage* is used when cleavage and folding are considered roughly synchronous in a transpressional regime. |
| Fabric | The geometric arrangement of component features in a rock, seen on a scale large enough to include many samples of each feature. |
| Foliation | The general term for any type of "planar" fabric in a rock (e.g., bedding, cleavage, schistosity). |
| Gneissosity | Foliation in feldspar-rich metamorphic rock, formed at intermediate to high temperatures, that is defined by compositional banding; the prefixes "ortho" and "para" are used for igneous and sedimentary protoliths, respectively. |
| Lineation | A fabric element that is best represented by a line, meaning that one of its dimensions is much longer than the other two. |
| Migmatite | Semichaotic mixture of layers formed by partial melting and deformation. |
| Mylonitic foliation | A foliation in ductile shear zones that is defined by the dimensional preferred orientation of flattened grains; the foliation tracks the *XY* flattening plane of the finite strain ellipsoid and is therefore at a low angle to the shear-zone boundary. |
| Phyllitic cleavage | Foliation that is composed of strongly aligned fine-grained white mica and/or chlorite; the mineralogy and fabric of phyllites give the rock a distinctive silky appearance, called *phyllitic luster.* |
| Schistosity | Foliation in metamorphic rock, formed at intermediate temperatures, that is defined by mica (primarily muscovite and biotite), which gives the rock a shiny appearance. |
| Texture | The pattern of crystallographic axes in an aggregate of grains; also *crystallographic fabric.* |

(Figure 11.1a). Undeformed sandstone, granite, or basalt are rocks with random fabrics. Deformed rocks typically have a **nonrandom fabric** or a **preferred fabric,** in which the fabric elements are aligned in some manner and/or are repeated at an approximately regular spacing (Figure 11.1b). There are two main classes of preferred fabrics in rock. A planar fabric, or **foliation** (Figure 11.1c), is one in which the fabric element is a planar or tabular feature (meaning it is shorter in one dimension than in the other two), and a linear fabric, or **lineation** (Figure 11.1d), is one in which the fabric element is effectively a linear feature (i.e., it is longer in one dimension relative to the other dimensions). Structural geologists may use the word "fabric" alone to imply the existence of a preferred fabric (as in, "that rock has a strong fabric"), but you should use appropriate modifiers if your meaning is not clear from context alone.

We're not quite done with terminology yet! Fabrics are complicated features, and there are lots of different adjectives used by structural geologists to describe them. For example, if you can keep splitting the rock into smaller and smaller pieces, right down to the size of the component grains, and can still identify a preferred fabric, then we say that the fabric is **continuous** (Figure 11.2a). In practice, if the fabric elements are closer than 1 mm (that is, below the resolution of the eye), the fabric is continuous. When there is an obvious spacing between fabric elements, we say that the fabric is **spaced** (Figure 11.2b).

Rocks with a **penetrative** tectonic fabric are also called **tectonites.** When linear fabric elements dominate, the rock is called an **L-tectonite,** whereas a rock with dominantly planar fabrics is called an **S-tectonite,** and, not surprisingly, rocks with both types of fabric

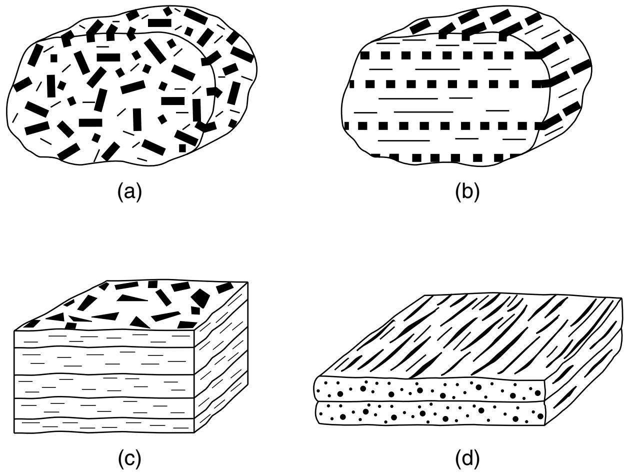



**FIGURE 11.1** The basic categories of fabrics. (a) A random fabric. The fabric elements are dark, elongate crystals. The long axes of these crystals are not parallel to one another. (b) A (1-dimensional) preferred fabric, in which the long axes of elongate crystals are aligned with one another. (c) A foliation. The fabric elements are planar and essentially parallel to one another, creating a 2-dimensional fabric. (d) A lineation. The fabric elements are linear; in this example, we show the alignment of fabric elements in a single plane.

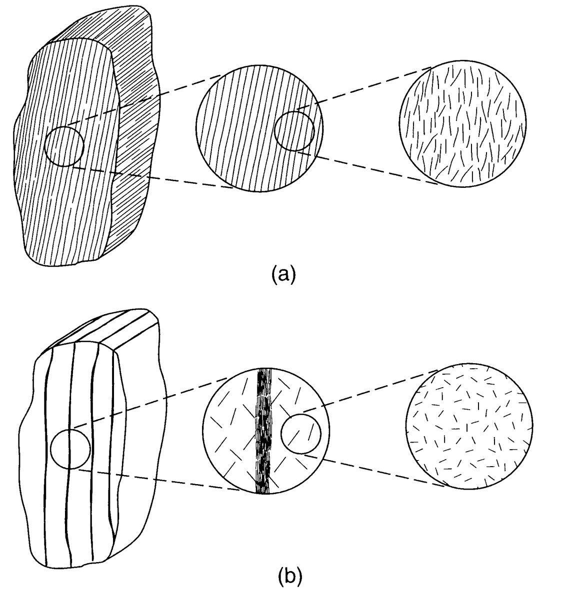



**FIGURE 11.2** The distinction between continuous and spaced fabrics. (a) A continuous fabric. The lines represent a planar fabric element that continues to be visible no matter how small your field of view (at least down to the scale of individual grains). (b) A spaced fabric. The rock between the fabric elements does not contain the fabric. The circled areas represent enlarged views.

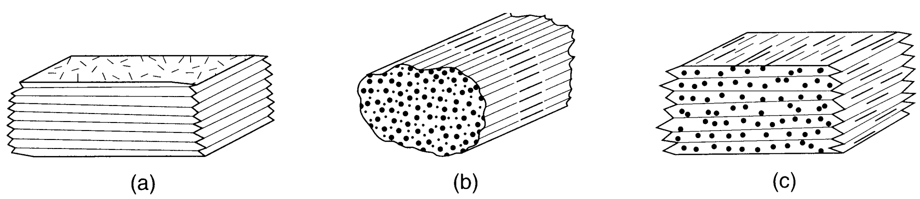



**FIGURE 11.3** The nature of tectonites. (a) An S-tectonite. This fabric is dominantly a foliation, and the rock may tend to split into sheets parallel to the foliation. Within the planes of foliation, linear fabrics are not aligned, or are not present at all. (b) An L-tectonite. The alignment of linear fabric elements creates the dominant fabric, so the rock may split into rod-like shapes. In L-tectonites, there is not a strong foliation. (c) An L/S-tectonite. The rock possesses a strong foliation and a strong lineation.

elements are called **LS-tectonites**[1] (Figure 11.3). Why create such jargon? Simply to highlight rocks whose internal structure has been substantially changed during deformation. Typically, the deformation that leads to the formation of a tectonite is accompanied by metamorphism, so the fabric is defined by grains that have been partially or totally recrystallized, and/or by new minerals that have grown during deformation (called neomineralization). Because most rocks are deformed, tectonites are among the most common rocks you will see (Figure 11.4); some examples are slates, schists, and mylonites, all of which will be discussed in a later section.

## 11.3 FOLIATIONS

A **foliation** is any type of planar fabric in a rock. We are admittedly a bit loose in our use of the term planar. Since, strictly speaking, a plane does not contain any curves or changes in orientation, the terms curviplanar or surface would be more appropriate. Although foliations are generally not perfectly planar, structural geol-

[1]"L" stands for "lineation"; "S" stands for "surface" in English, "schistosité" in French, or "Schieferung" in German.

**FIGURE 11.4** Disjunctive cleavage in mica-rich rock (Rhode Island, USA). Note the variation in cleavage spacing between steeply dipping beds. Width of view is ~ 4 m.

ogists nonetheless are in the habit of talking about planar fabrics. Thus, bedding, cleavage, schistosity, and gneissosity all qualify as foliations. Fractures, however, are not considered to be foliations, because fractures are breaks through a rock and are not a part of the rock itself. A rock may contain several foliations, especially if it has been deformed more than once. To keep track of different foliations, geologists add subscripts to the foliations: $S_0$, $S_1$, $S_2$, and so on. $S_0$ is used to refer to bedding, $S_1$ is the first foliation formed after bedding, and $S_2$ forms after $S_1$. The temporal sequence of foliation development is defined by cross-cutting relationships, but in complexly deformed areas, it may be quite a challenge to determine which foliation is which unless independent constraints on (relative) time are available.

There are many types of tectonic foliations that are distinguished from one another simply on the basis of what they look like. The physical appearance of a foliation reflects the process by which it formed, and the process, in turn, is controlled partly by the composition and grain size of the original lithology, and partly by the metamorphic conditions under which the foliation formed. Different names are used for the different types of foliations. In Table 11.2 and the following discussion, we will introduce different types of foliation roughly in sequence of increasingly higher metamorphic conditions—cleavage first, then schistosity, then gneissic layering.

### 11.3.1 What Is Cleavage?

**Cleavage** in rocks has been defined in different ways by different people, so the use of this term in the literature is confusing. We advocate a nongenetic definition,

**TABLE 11.2 FOLIATION CLASSIFICATION SCHEME**

| | |
|---|---|
| Spaced cleavage | (a) Disjunctive cleavage (e.g., stylolitic cleavage)<br>(b) Crenulation cleavage |
| Continuous cleavage | (a) Coarse cleavage (e.g., pencil cleavage)<br>(b) Fine cleavage (e.g., slaty cleavage) |
| Phyllitic cleavage | Continuous cleavage with a distinctive silky luster in low-grade metamorphic rock (lower greenschist facies) |
| Schistosity | Mica-rich foliation with a distinctive high sheen in low- to medium-grade metamorphic rock (greenschist facies) |
| Gneissic layering or gneissosity | Coarse compositional banding in high-grade metamorphic rock (upper amphibolite and granulite facies) |

in which cleavage is defined as *a secondary fabric element, formed under low-temperature conditions, that imparts on the rock a tendency to split along planes.* The point of this definition is to emphasize

1. That cleavage is a feature of the rock that forms subsequent to the origin of the rock.
2. That the term "cleavage" is used, in practice, for tectonic planar fabrics formed at or below lower greenschist facies conditions (i.e., ≤ 300°C). The term "cleavage" is not used when referring to the fabric in schists or in gneiss.
3. That in a rock with cleavage, there are planes of weakness across which the rock may later break when uplifted and exposed at the surface of the earth, even though cleavage itself forms without loss of cohesion. By this definition, an array of closely spaced fractures is not a cleavage.

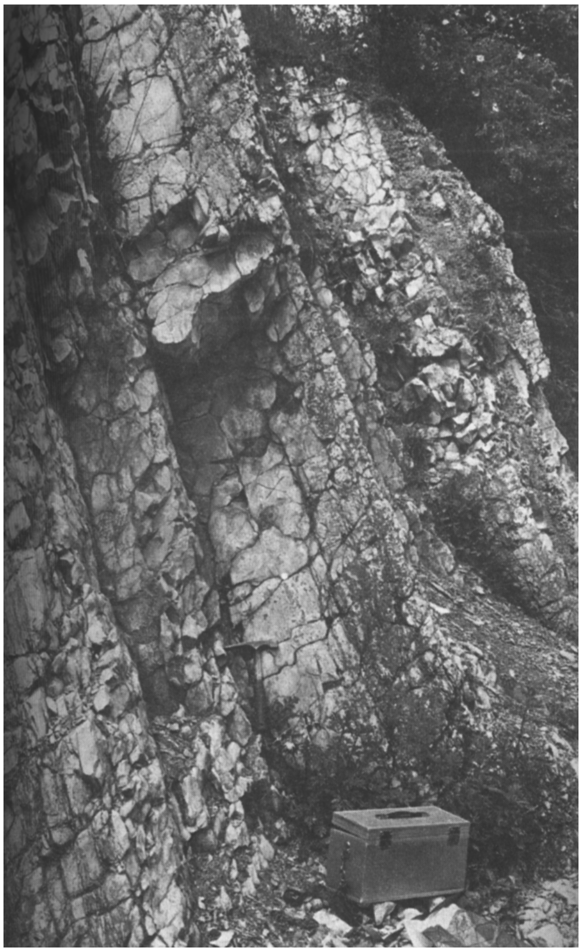
We recognize four main categories of cleavage that are differentiated from one another by their morphological characteristics (or, by how they look in outcrop). These are disjunctive cleavage, pencil cleavage, slaty cleavage, and crenulation cleavage.

## 11.3.2 Disjunctive Cleavage

**Disjunctive cleavage** is a foliation that forms mostly in sedimentary rocks[2] that have been subjected to a tectonic differential stress under sub–greenschist facies metamorphic conditions. It is defined by an array of subparallel fabric elements, called **cleavage domains,** in which the original rock fabric and composition have been markedly changed by the process of pressure solution. Domains are separated from one another by intervals, called **microlithons,** in which the original rock fabric and composition are more or less preserved (Figure 11.5). The adjective "disjunctive" implies that the cleavage domains cut across a preexisting foliation in the rock (usually bedding), without affecting the orientation of the preexisting foliation in the microlithons. Because pressure solution is always involved in the formation of a disjunctive cleavage, other terms such as **(pressure-)solution cleavage** and **stylolitic cleavage** have been used for this structure. If the context is clear, some geologists may refer to the structure simply as spaced cleavage, though, as we see later, crenulation cleavage is also a type of spaced cleavage. Earlier in this century, many geologists incorrectly considered cleavage domains to be brittle fractures formed by loss of cohesion. The old term "fracture cleavage" should therefore be avoided when referring to disjunctive cleavage, because a cleavage cannot be composed of fractures. Such arrays of closely spaced fractures should be called a fracture or joint array.

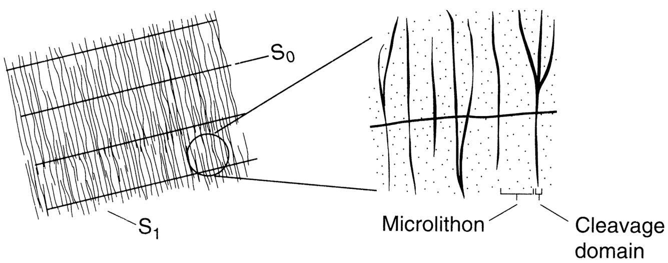
Now that we've gotten through the cleavage terminology, let's see how disjunctive cleavage forms. Consider a horizontal bed of argillaceous (clay-rich) limestone or sandstone that is subjected to a compressive stress ($\sigma_1$ in Figure 11.6). Dissolved ions created by pressure solution are transported away from the site of dissolution through a water film that adheres to the grain surfaces. The ions may then precipitate at crystal faces where compressive stress is less, precipitate nearby in the **pressure shadows** adjacent to rigid grains, or enter the pore fluid system to be carried out of the local rock environment entirely. Note that in order for pressure solution to occur, a thin layer of water molecules must be chemically bonded to grain surfaces. If the water is not bonded, a grain in the water can only sustain isotropic stress, because fluids cannot support shear stresses, and pressure solution won't occur.

[2]There are descriptions of disjunctive cleavage in igneous rocks, but it is most typical of sedimentary rocks.



(a)

(b)




(c)

**FIGURE 11.5** Spaced disjunctive cleavage (or solution cleavage) in limestone from the Harz Mountains of Germany. The cleavage is marked by the narrow dark bands which cut across the original, lighter-colored argillaceous limestone. (a) Outcrop view; (b) close-up view of central portion. In (c) the cleavage domain and microlithon of spaced cleavage are illustrated in gently left-dipping beds.

The distribution of clay in rocks is not uniform. Pressure solution occurs more rapidly where the initial concentration of clay is high. How clay affects pressure solution rates remains enigmatic. Perhaps swelling clay, which contains interlayers of bonded water, increases the number of diffusion pathways available for ions and thus multiplies the diffusion rate. Alternatively, the highly charged surfaces of clay grains may act as a chemical catalyst for the dissolution reaction. Field observations suggest that a rock must contain >10% clay in order for solution or stylolitic cleavage to develop. Where it occurs, the process preferentially removes more soluble grains. Thus, in an argillaceous limestone, calcite is removed, and clay and quartz are progressively concentrated. In argillaceous sandstone, the process is effectively the same as that in argillaceous limestone, except that quartz is the mineral that preferentially dissolves, and clay alone is concentrated. As the framework grains of carbonate and quartz are removed, the platy clay grains collapse together like a house of cards. Concentration of clay in the domain further enhances the solubility of

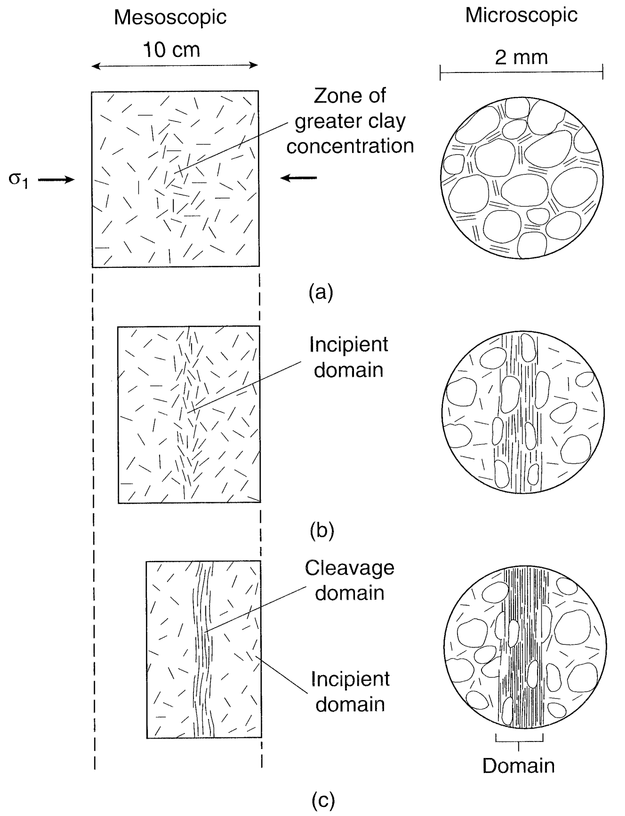



**FIGURE 11.6** Evolution of spaced disjunctive cleavage. (a) Precleavage fabric of the rock. In the area indicated by the arrow in the mesoscopic image, there happens to be a greater initial concentration of clay. The microscopic image indicates that the clay flakes are randomly oriented. (b) As shortening and pressure solution occur, the zone in which there had initially been a greater clay concentration evolves into an incipient cleavage domain. At this stage, grains are being preferentially dissolved on the faces perpendicular to $S_1$, and the clay flakes are collapsing together. (c) Ultimately, a clearly defined cleavage domain is visible. In the domain, the clay flakes are packed tightly together and only small relicts of the soluble mineral grains are visible.

the rock, so there is positive feedback. Eventually, a discrete domain develops in which there is a **selvage,** the material filling the domain, composed of mostly clay (and quartz) with some relict corroded calcite grains. In the selvage, the clay flakes are packed together with a dimensionally preferred orientation. If deformation continues, the domain continues to thicken as pressure solution continues along its edges. As a result, compositional contrast between cleavage domains and microlithons becomes so pronounced that it defines a new stratification in the rock that nearly obscures the original bedding. From this description you see that spaced cleavage formation is identical to the processes by which bedding-parallel stylolites form as a consequence of compaction loading; hence the use of the term stylolitic cleavage.

The **spacing** of cleavage domains in a rock depends on the initial clay content. If the clay content is high, the domains are closely spaced. Spacing also changes with progressive deformation. As strain increases, more cleavage domains nucleate and thus the domain spacing decreases. Thirdly, domain spacing may also be related to the magnitude of differential stress, for experiments suggest that the rate of pressure solution is proportional to the magnitude of differential stress. During the process of domain formation, the fabric of the microlithons remains relatively unaffected, though high-resolution microscopic analysis (such as transmission electron microscopy) may indicate the presence of incipient pressure solution features or newly crystallized grains of the soluble mineral in microlithons.

Before we leave the topic of disjunctive cleavage, we need to present the vocabulary that geologists use to describe disjunctive cleavage in the field. The classification and description of disjunctive cleavage is based on the characteristics of surface morphology of domains, and on domain spacing (Figure 11.7). If the clay content in the host rock is low, domain surfaces are severely pitted. In cross section, such domains resemble the toothlike or jagged sutures on a skull; this type of cleavage domain is called a **sutured domain.** If the clay content is high, cleavage domains tend to have thick selvages that have smooth borders; these domains are called **nonsutured.** When viewed under high magnification with a microscope, you will see that thick domains, in some cases, are composed of a dense braid of threadlike sutured domains. Cleavage domains can be either *wavy,* if the domain undulates, or *planar,* if the domain does not. If wavy domains are closely spaced, such that the domains merge and bifurcate and give the fabric the appearance of braided hair, the cleavage is called **anastomosing.**

The average spacing between domains is also a useful criterion for classification of cleavage. Figure 11.7e shows a simple field classification based on spacing. If the spacing between domains is greater than about 1 m, then the rock really doesn't have an obvious fabric. Such isolated pressure-solution domains, when formed in response to tectonic stress, are called **tectonic stylolites.** The adjective "tectonic" distinguishes these structures from bedding-parallel stylolites formed by compaction. In general practice, if domains are spaced

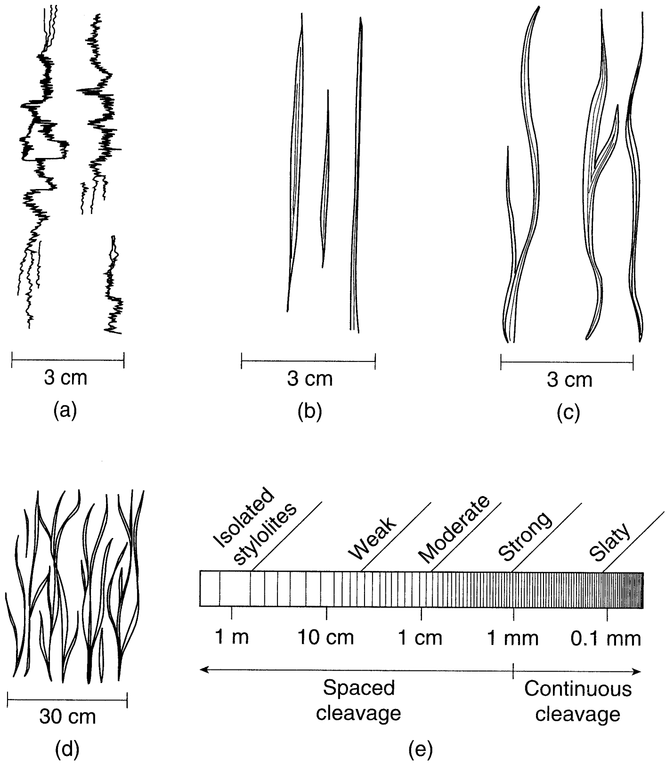



**FIGURE 11.7** Cross-sectional sketches of morphological characteristics used for cleavage description and classification. (a) Sutured domains; (b) planar domains; (c) wavy domains; (d) an anastomosing array of wavy domains. In (e) the description of spaced cleavage based on domain spacing is shown.

between about 10 cm and 1 m apart, we call the feature a **weak cleavage;** a spacing of 1–10 cm defines a **moderate cleavage,** and a spacing of less than 1 cm denotes a **strong cleavage.** When cleavage domains are less than 1 mm apart, the cleavage is **continuous** (see slaty cleavage, below). We caution you again that different authors use these adjectives differently, so our definitions are only generalizations.

### 11.3.3 Pencil Cleavage

If a fine-grained sedimentary rock (shale or mudstone) breaks into elongate pencil-like shards because of its internal fabric, we say that it has a **pencil cleavage** (Figure 11.8). Typically, pencils are 5–10 cm long and 0.5–1 cm in width. In outcrop, pencil cleavage looks as if it results from the interaction of two fracture sets (and in some locations, it is indeed merely a consequence of the intersection between a fracture set and bedding), but actually the parting reflects an internal alignment of clay grains in the rock.

**FIGURE 11.8** Pencil cleavage in shale, with horizontal bedding trace (Knobs Formation, Virginia, USA).

Pencil cleavage forms because of the special characteristics of clay. The strong shape anisotropy of clay flakes creates a preferred orientation parallel to bedding when they settle out of water and are compacted. This preferred orientation imparts the tendency for clay-rich rocks to break on bedding planes that is displayed by *shale.* Shale, mudstone, and slate are all names that are used to describe clay-rich rocks; mudstone is the general term for these rocks, whereas shale and slate are foliated clay-rich rocks. After compaction, a strain ellipsoid representing the state of strain in the shale would look like a pancake parallel to bedding (Figure 11.9). Now, imagine that the shale is subjected to layer-parallel shortening. Cleavage formation processes begin to take place: large detrital phyllosilicates *fold and rotate,* while fine grains undergo *pressure solution* along domains perpendicular to the shortening direction, and new clay *crystallizes. Microfolding* may occur during this stage of the process, but because of the fine grain size these microfolds are only visible under the microscope. In addition, quartz may begin to dissolve, and as these framework grains are removed, clay flakes collapse so that their basal planes are perpendicular to the plane of shortening. The plane defined by new or rotated clay grains is roughly perpendicular to the shortening direction, so it forms a tectonic foliation at high angles to the original bedding (Figure 11.9c). At an early stage during this process, the new tectonic fabric is comparable in degree of development to the initial bedding-parallel fabric. At this stage, the strain ellipsoid representing this state would look something like a big cigar, and the rock displays pencil cleavage (Figure 11.9b). In sum, pencil

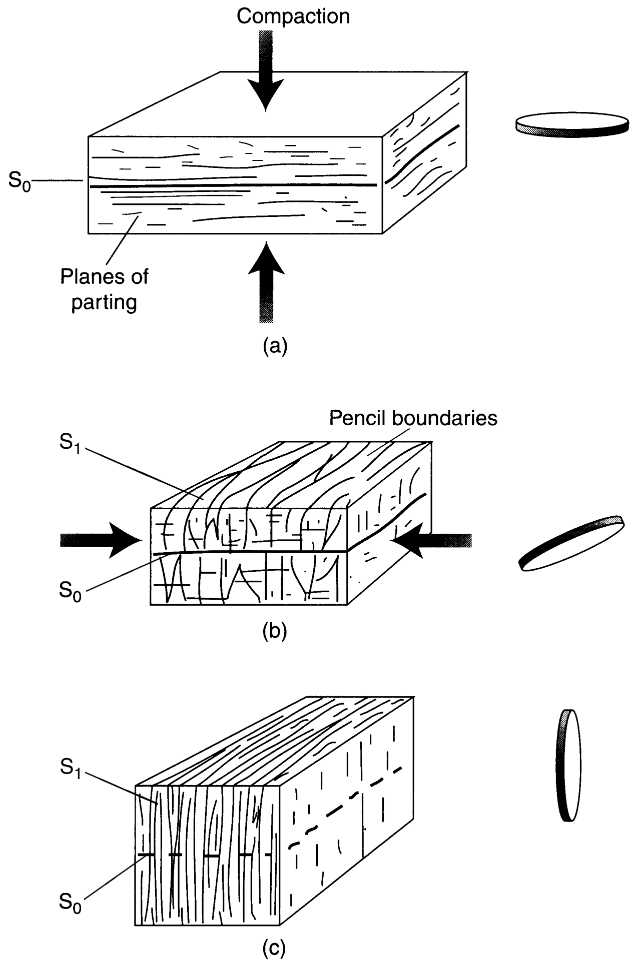



**FIGURE 11.9** Sketches illustrating the progressive development of slaty cleavage via the formation of pencil structure. (a) Compaction during burial of a sedimentary rock produces a weak preferred orientation of clay parallel to bedding. A representative strain ellipsoid is like a pancake in the plane of bedding. (b) Shortening parallel to layering creates an incipient tectonic fabric. Superposition of this fabric on the primary compaction fabric leads to the formation of pencils. The representative strain ellipsoid is elongate, with the long axis parallel to the pencils. (c) Continued tectonic shortening leads to formation of slaty cleavage at a high angle to bedding. The phyllosilicates are now dominantly aligned with the direction of cleavage, and a representative strain ellipsoid is oblate and parallel to cleavage.

cleavage is a fabric found in weakly deformed shale in which the tendency of the shale to part on bedding planes is about the same as the tendency for it to part on an incipient tectonic cleavage that is at a high angle to bedding. Deformation in most areas produces a fabric that is much stronger than the original bedding parting, so pencils are not as common as the more evolved stage, represented by slaty cleavage.

### 11.3.4 Slaty Cleavage

Pencil cleavage may be considered a snapshot of an early stage in the process by which slaty cleavage develops. As shortening perpendicular to cleavage planes accumulates, clay throughout the rock develops a preferred orientation at an angle to the original sedimentary fabric and this orientation dominates over the primary fabric (Figure 11.9c). The finite strain ellipsoid for cleavage development at this stage has the shape of a pancake that parallels the tectonic fabric. Formation of slaty cleavage occurs by much the same process as did the formation of disjunctive cleavage in argillaceous sandstone or limestone, but the resulting domains are so closely spaced that effectively there are no uncleaved microlithons, and the entire rock mass displays the tectonically induced preferred orientation. When a rock has this type of continuous fabric, we say that it has **slaty cleavage** (Figure 11.10). In other words, slaty cleavage is defined by strong dimensionally preferred orientation of phyllosilicates in a very clay-rich rock, and the resulting rock, which is considered to be a low-grade metamorphic rock, is called a *slate.* Slaty cleavage tends to be smooth and planar. This characteristic, coupled with the penetrative nature of slaty cleavage, means that slates can be split into thin sheets, which made them popular roofing materials in the nineteenth century and early parts of the twentieth century.

At high magnifications the detailed character of slaty cleavage becomes apparent. In many cases only a network of anastomosing surfaces around partially dissolved quartz or feldspar grains is preserved (Figure 11.11a), but in some instances remnants of the original bedding orientation can be found (Figure 11.11b). Because slaty cleavage forms under temperature conditions that mark the onset of metamorphism (250°C–350°C), the mineralogy of slate tends to resemble that of shale. However, there is a notable decrease in the amount of interlayered water in clays; that is, smectite, the water-bearing clay, transforms to illite. As metamorphic temperatures increase, the illite becomes more micalike in structure, though it is still fine grained. Thus, rocks with higher-grade slaty cleavage display a distinct sheen on cleavage planes. They are also significantly harder than shale, so they ring when hit with a hammer. These are the slates that provide excellent material for the construction of pool tables.

### 11.3.5 Phyllitic Cleavage and Schistosity

When metamorphic conditions reach the lower greenschist facies, the clay and illite in a pelitic rock react to

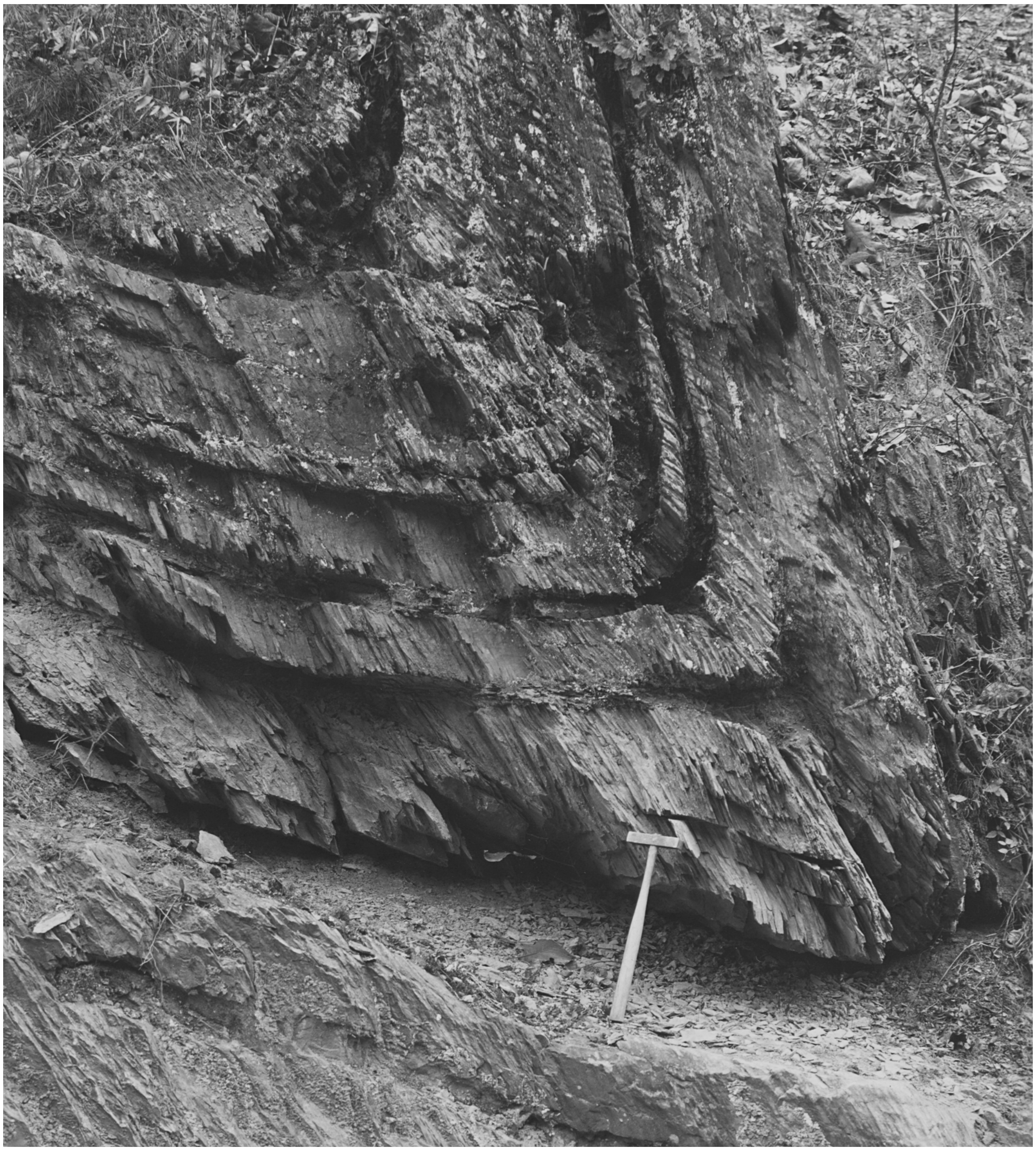


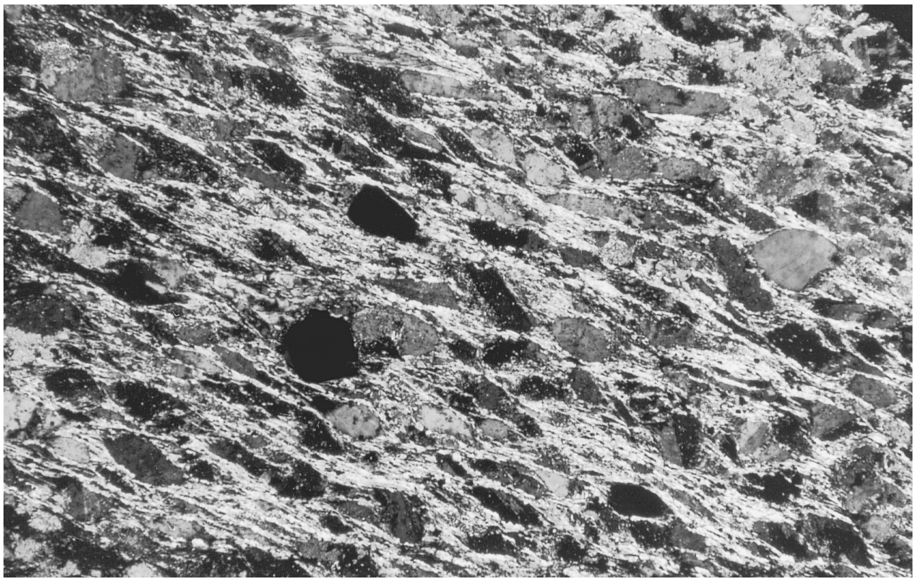
**FIGURE 11.10** An overturned syncline with a well-developed axial plane slaty cleavage (southern Appalachians, USA).

form white mica[3] and chlorite. If reaction occurs in an anisotropic stress field, these phyllosilicates grow with a strong preferred orientation. Rock that is composed of strongly aligned fine-grained white mica and/or chlorite is called **phyllite**[4] and the foliation that it contains is called **phyllitic cleavage.** The mineralogy and fabric of phyllites give the rock a distinctive silky luster. Phyllitic cleavage is intermediate between slaty cleavage and schistosity.

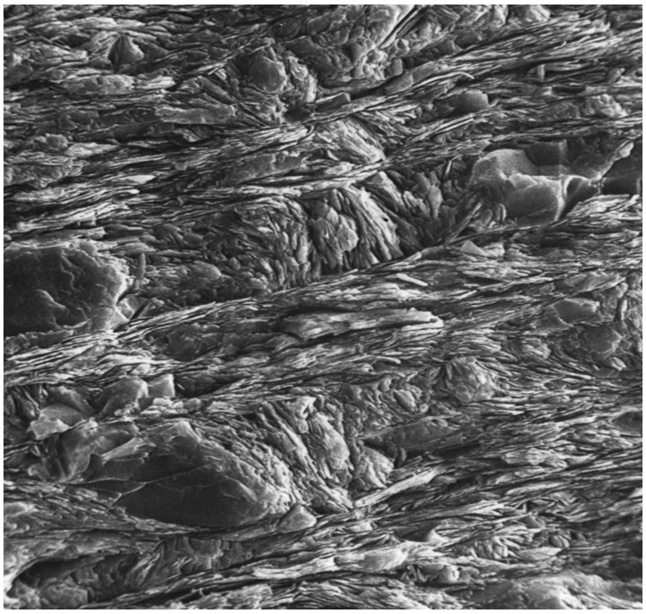
When metamorphic conditions get into the middle greenschist facies, the minerals in a pelitic lithology react to form coarser-grained mica and other minerals. When these reactions again take place in an anisotropic stress field, the mica has a strong preferred orientation. The resulting rock is a **schist,** and the foliation it displays is called **schistosity.** The specific assemblage of minerals that forms depends not only on the pressure and temperature conditions at which metamorphism

[3]"White mica" is fine-grained muscovite. It contains more potassium and aluminum than illite, and is better ordered.

[4]Phyllite is not the same as *phyllonite,* which is a low-grade mylonitic rock.



(a)



(b)

**FIGURE 11.11** (a) Photomicrograph of a continuous cleavage from Newfoundland (Canada); width of view is 2 mm. (b) Scanning electron micrograph from a slate in the Rheinische Schiefergebirge (Germany). Note the microcrenulation of bedding-parallel micas, creating distinct microlithons with mica roughly parallel to bedding and cleavage domains with mica parallel to cleavage. Width of view is 100 μm.

takes place, but also on the chemical composition of the protolith and on the degree to which chemicals are added or removed from the rock by migrating fluids. Conveniently, a schist is named by the assemblage of metamorphic index minerals that it contains (e.g., a garnet-biotite schist). In schist that contains **porphyroclasts** (relict large crystals) or **porphyroblasts** (newly grown large crystals), the schistosity tends to be wavy, as the micas curve around the large crystals.

### 11.3.6 Crenulation Cleavage

A lithology containing a closely and evenly spaced foliation that is shortened in a direction at a low angle to this foliation will crinkle like the baffles in an accordion (Figure 11.12). In fine-grained lithologies like slate or phyllite, these microfolds are closely spaced and the spacing tends to be very uniform. The axial planes of the crenulations define a new foliation called

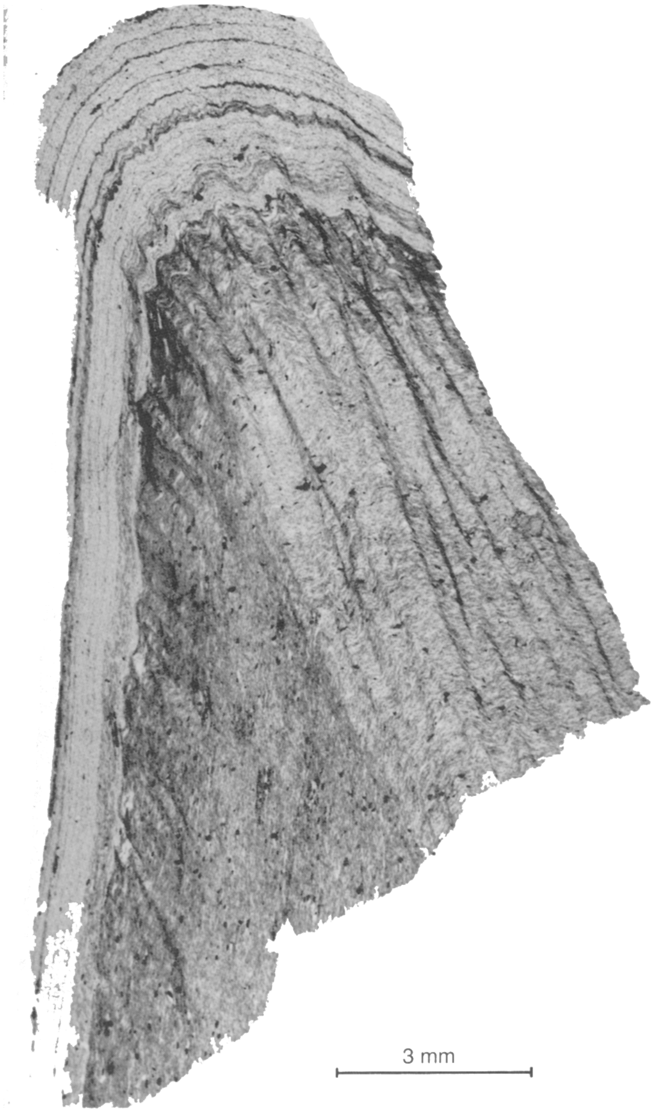



**FIGURE 11.12** Photomicrograph of crenulation cleavage in a phyllite (Vermont, USA). The spaced crenulation cleavage deforms an earlier continuous cleavage, with cleavage spacing and intensity that are less in the sandy layers (white).

**crenulation cleavage** (Figure 11.13). Like mesoscopic folds, crenulations can be symmetric (Figure 11.13b) or asymmetric (Figure 11.13c). In a given outcrop, both symmetric and asymmetric crenulation cleavages may occur. For example, on the limbs of a fold, asymmetric crenulation occurs, whereas in the hinge zone of the fold, the crenulation is symmetric (Figure 11.12).

A prerequisite for the formation of crenulation cleavage is the existence of a preexisting strong lamination or foliation. Crenulation cleavage won't form in a sandstone, but may form in a micaceous shale with a strong bedding-plane foliation, or in a rock that already contains a slaty or phyllitic cleavage (Figure 11.14). Crenulation cleavage in outcrop is typically an $S_2$ foliation that has been superimposed on an earlier ($S_1$) foliation.

Crenulation cleavage forms under conditions that are also amenable to the occurrence of pressure solution. When the starting rock contains a mixture of quartz and clay or fine-grained mica, the quartz is preferentially removed from the limbs of the microfolds and precipitates in the hinges as the crenulations form (Figure 11.15). Gradually, phyllosilicates concentrate on the limbs and quartz is concentrated in the hinges. This mineralogical differentiation can be so complete that the old foliation disappears entirely and is replaced entirely by a new foliation, that is defined not only by preferred orientation of the phyllosilicates, but also by microcompositional layering (Figure 11.15b). This process is a type of **transposition** (see Chapter 10), by which a preexisting foliation is transposed into a new orientation. If quartz is largely removed by progressive pressure solution, a new foliation eventually develops and the crenulated appearance of the rock fades (Figure 11.15c). If this happens, all traces of the original fabric, the one predating the crenulation, may be destroyed. Thus, crenulation cleavage forms in a rock

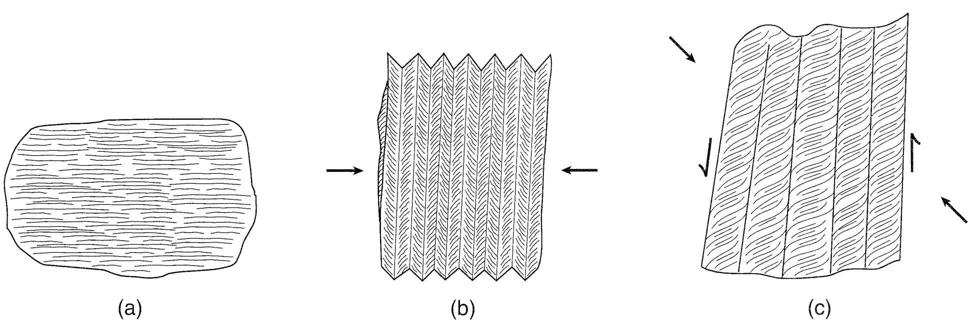



**FIGURE 11.13** The two basic categories of crenulation cleavage. (b) Symmetric crenulation cleavage; (c) asymmetric (sigmoidal) crenulation cleavage. The arrows indicate a possible component of shear associated with this crenulation geometry.

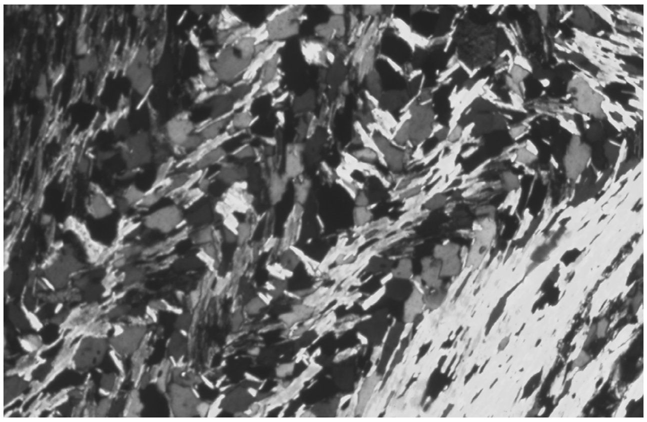


**FIGURE 11.14** Photomicrograph of incipient differentiation in crenulation cleavage (Pyrenees). Width of view is ~0.5 mm

with a preexisting cleavage, but, paradoxically, it can evolve into a rock with seemingly only one foliation.[5]

### 11.3.7 Gneissic Layering and Migmatization

Foliated **gneiss** is a metamorphic rock in which the foliation is defined by compositional banding (Figure 11.16). Commonly, light and dark bands of felsic and mafic mineralogy alternate. The light-colored layers are rich in feldspar and quartz, whereas darker layers contain more of the minerals amphibole/pyroxene (and/or biotite). This color banding in gneiss is called **gneissosity.** Under the metamorphic condition for gneiss formation

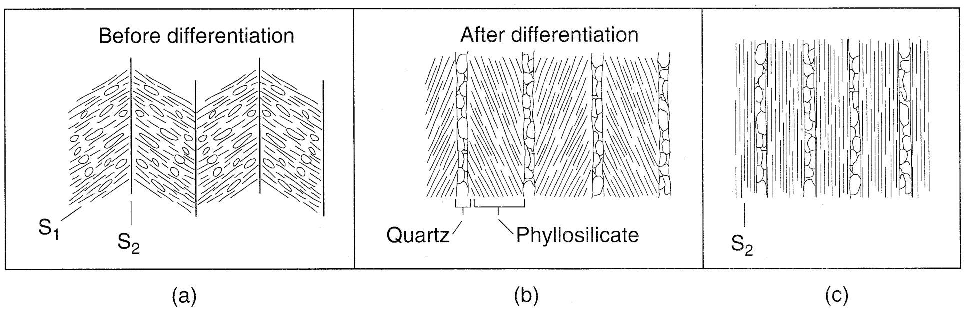



**FIGURE 11.15** Differentiation during the formation of crenulation cleavage. (a) Fairly homogeneous composition, before migration of the quartz. (b) Quartz accumulates in the hinges of the crenulations, and the phyllosilicates are concentrated in the limbs; the result is the formation of compositionally distinct bands in the rock. (c) Complete transposition of the $S_1$ foliation into a new $S_2$ cleavage or schistosity.

**FIGURE 11.16** Fold hinges in transposed gneiss near Parry Sound, Grenville Orogen (Ontario, Canada).

[5]To some extent, crenulation cleavage is a matter of scale of observation; many slates show evidence of microfolds in detrital mica flakes.

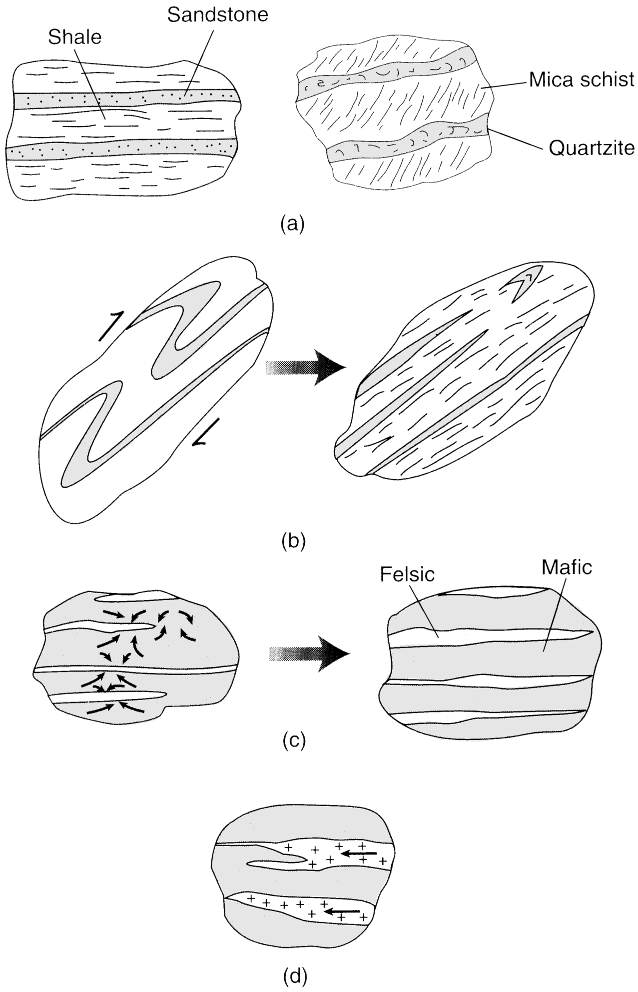



**FIGURE 11.17** Mechanisms of formation of a gneiss. (a) Inheritance from an original lithology; (b) creation of new compositional banding via transposition; (c) metamorphic differentiation; (d) *lit-par-lit* intrusion.

(amphibolite to granulite facies), muscovite reacts to form feldspar, so the rock contains no schistosity. Gneiss can be derived from a sedimentary protolith, in which case it is called a **paragneiss,** or an igneous protolith, in which case it is called an **orthogneiss.** It is often difficult to decide whether a particular rock is an orthogneiss or a paragneiss, however; this requires careful field and petrologic analysis. A special type of gneiss, called **augen gneiss,** contains relatively large feldspar clasts floating in a finer-grained matrix.[6]

How does the compositional banding in gneiss form? There may be several processes involved in the formation of gneissic layering (Figure 11.17). First, it may occur by *inheritance* from original compositional contrasts. If the protolith (the rock from which the gneiss formed) was a stratified sequence with layers of different composition (such as alternating sandstone and shale), then metamorphism will transform this sequence into a compositionally banded metamorphic rock. Secondly, it may result from *transposition* via folding of an earlier layering. Transposition is a common process during deformation under high-temperature conditions, and is discussed in several places in this book (see also Chapters 10 and 12). Rock containing compositional layering that is subjected to intense, high-grade deformation may develop isoclinal folds. If the hinges of the fold are detached and a new sequence of compositional layers has formed, then we say that the new layering is a type of transposed foliation. The layering in a rock with a transposed foliation does not represent the original stratigraphy of the rock, though it may have been derived from it. In other words, the sequence of compositional layers in a transposed rock does not represent the original stratigraphic succession. Thirdly, gneissic layering may be formed by *metamorphic differentiation* when the thermodynamics governing diffusion during metamorphism causes certain ions to be excluded from the formation of new metamorphic minerals in a layer. The excluded ions accumulate to form different minerals in an adjacent layer. Thus, minor differences in the original composition of successive layers may be amplified into major compositional changes after metamorphism. The resulting rock with alternating layers of different composition is a gneiss. Finally, gneissic banding may originate from an igneous process called **lit-par-lit intrusion** (French for "layer-by-layer"). Melts inject as thin sills along many weak planes in the protolith, and this interlayering of sills and host rock defines the gneiss. Usually the process of injection is accompanied by passive folding, so the igneous nature of the contacts is often not evident.

When metamorphic temperatures are sufficiently high, a rock begins to melt; but not all minerals melt at the same temperature. Quartz, some feldspar, and muscovite melt at lower temperatures than mafic minerals like amphibole, pyroxene, and olivine. Therefore, when a rock of intermediate composition begins to melt, certain minerals become liquid while others remain solid. The minerals that stay solid until higher temperatures are achieved are called **refractory minerals.** When only part of a rock melts, we say that it has undergone *partial melting* or **anatexis.** A rock that is undergoing partial melting is a mixture of pockets of melt and lenses of solid, both of which are quite soft. Shortening will cause the mass to flow, and the contrasting zones of melt and solid fold and refold much like chocolate and vanilla batter in marble cake. When this happens in rock, the resulting semichaotic mixture of light and dark layers is called a **migmatite.** Because

[6]Because the altered feldspar grains look like eyes, or "Augen" in German.

of their origin and chaotic nature, the analysis of structures in migmatite may provide little or no information about the kinematics of regional deformation.

### 11.3.8 Mylonitic Foliation

Mylonites are fine-grained, foliated and/or lineated fault rocks that form by crystal-plastic deformation processes. They are discussed in detail in the chapter on shear zones (Chapter 13), but we briefly want to include mylonitic foliation here. The foliation in mylonites is defined by shape-preferred orientation of flattened grains (usually quartz and calcite, but also feldspar and olivine) and mica. Typically, transposition via folding occurs during mylonitization, so the lithologic banding in a mylonite is not the original layering of the rock. **Mylonitic foliation** is associated with these folds, but fold hinges vary in orientation or are strongly curved in the foliation plane as a consequence of differential flow. Like other fine laminations, mylonitic foliation is susceptible to the formation of crenulations, giving rise to structures that aid recognizing the kinematic history of shear zones. Mylonites rival foliations in their importance for structural analysis, so we direct you to Chapter 13 for much more on mylonitization.

## 11.4 CLEAVAGE AND STRAIN

Does the study of a foliation in the field provide any constraints on the nature of strain in the region? Unfortunately, we can only offer a wishy-washy answer: sometimes yes, sometimes no. In order to determine the relationship of cleavage to strain, we need to look at strain markers in a cleaved rock. Red slate is a good lithology for such strain studies, because it may contain reduction spots. **Reduction spots** (Figure 11.18) are small zones where iron in the rock is reduced and therefore is greenish in color. Assuming these spots start out as early diagenetic spheres around an inclusion, they make ideal strain markers because they behave in a totally passive manner, meaning that they have no mechanical contrast with the host rock. In studies of reduction spots in slate, geologists find that the deformed spots are flattened ellipsoids, and that the plane of flattening (i.e., the $XY$ principal plane of the finite strain ellipsoid) is essentially parallel to the slaty cleavage. In these cases, therefore, cleavage appears to approximate the orientation of a principal plane of strain, with the shortening direction being perpendicular to cleavage. But cleavage is probably not strictly parallel to the $XY$ principal plane of strain in all situations. For example, when cleavage occurs in flexural slip folds, or adjacent to fault zones, there may be a component of shear on the cleavage planes themselves, and when this happens the cleavage by definition is not a principal plane. Moreover, a spaced cleavage may initiate as a principal plane of strain, but subsequent folding of the bed in which cleavage occurs may rotate it away from the bulk principal strain directions. Reduction spot studies give estimates of a total shortening strain ($\mathbf{e}_1$) for the formation of slaty cleavage in the range of 50–60%.

**FIGURE 11.18** Reduction spots in slate that are ellipsoidal in shape (elliptical in section); Appalachians, USA. They can be used as strain markers if they were formed as spherical regions around a reducing phase prior to deformation; coin for scale.

The most controversial issue about low-grade cleavage formation is the question of whether cleavage formation is a volume-constant strain or a volume-loss strain process. Imagine a cube of rock that is 10 cm long on each edge. If it is shortened in one direction, but does not stretch in the other directions ($X = Y = 1 > Z$), it must lose volume during deformation (Figure 11.19). Alternatively, if shortening in one direction causes it to expand by an equal proportion in one other direction ($X > Y = 1 > Z$ or plane strain) or in two other directions ($X > Y > Z$), then the strain may be volume-constant. Remember that formation of low-grade cleavage involves significant activity in the way of dissolution and new growth of grains. If the ions of soluble minerals enter the pore water system, they can be carried out of the local rock system by the movement of groundwater. Thus, volume loss during cleavage formation is a likely possibility; in fact, studies of deformed markers (such as the fossil graptolites, whose regular protrusion spacing can be used to measure finite strain) and geochemical studies demonstrate that rock volume may have decreased by up to 50% during cleavage formation

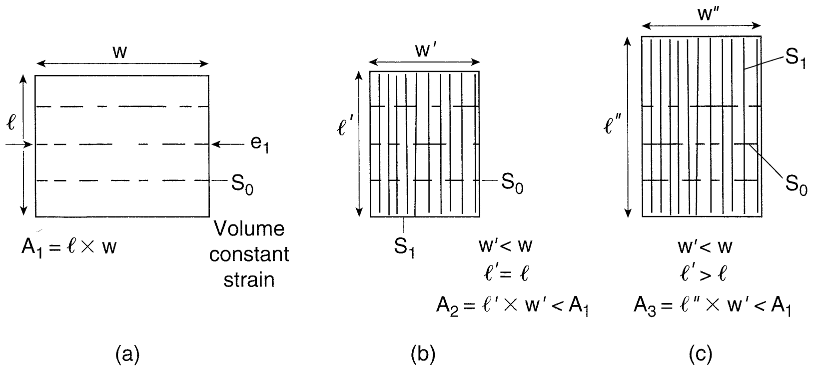
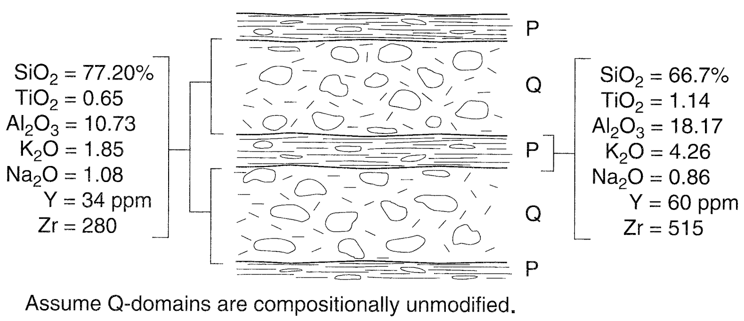



**FIGURE 11.19** To lose or not to lose volume; that's the question. (a) A block of height *l* and width *w* is shortened and forms a cleavage. If there is volume loss strain, then $w' < w$, but $l' = l$ (we are assuming no change in the third dimension; that is, the intermediate strain axis ($Y$) equals 1). If there is volume constant strain, then $l'' > l$, as in (c).




$$\% \text{ volume loss from P-domains} = \frac{(\% \, TiO_2 \text{ in P}) - (\% \, TiO_2 \text{ in Q})}{\% \, TiO_2 \text{ in P}} \times 100$$

**FIGURE 11.20** Volume loss may occur by preferential removal of certain elements. If the composition of the microlithons (Q-domains) is assumed to be constant, then the amount of volume loss can be calculated (see equation); based on relatively immobile elements $TiO_2$, Y, and Zr, the volume loss is calculated as ~45%.

(Figure 11.20). Interestingly, these values of bulk strain and volume loss suggest that plane strain conditions are representative for cleavage formation, and that flattening strains are apparent rather than real (field of apparent flattening; Figure 4.16). But not all strain analyses of cleaved rocks yield indications of large volume loss. Probably the degree of volume loss is affected by the degree to which the rock is an open or closed geochemical system during deformation, and whether the fluids passing through the rock during deformation are saturated or undersaturated with respect to the soluble mineral phase.

Volume-loss scenarios also have an important implication concerning the nature of **rock–water interactions** during deformation. If we assume that most of the volume loss reflects dissolution and removal of certain minerals, then we can calculate the minimum volume of fluid required for a given amount of volume loss if we know the solubility of the mineral and the amount that can be held in solution (saturation). Such calculations suggest that the amounts of fluid needed for the proposed amounts of volume loss are very large. The results of these calculations are usually expressed in terms of the ratio of fluid volume to rock volume (called the **fluid/rock ratio**). In the example in Figure 11.20[7] this ratio is >>100, but whether such large ratios are reasonable remains a matter of debate.

## 11.5 FOLIATIONS IN FOLDS AND FAULT ZONES

We have so far discussed tectonic foliations mainly from a descriptive point of view, without examining how they are related to other structures. Tectonic foliations do not occur in isolation; rather, they are integral components of the suite of structures that represent the manifestation of deformation in a region. In this section we will look at how cleavage develops during progressive deformation, and introduce a number of powerful field characteristics of foliations.

[7]Given the solubility of quartz is 0.019 kg/l and a 20% undersaturated fluid.

First we consider the geometric relationship between low-grade cleavage (slaty, spaced disjunctive, and crenulation cleavage) and folds. Low-grade cleavage forms in regions where metamorphic conditions are low (hence the name), a condition that is prevalent in the foreland region of collisional orogens or in convergent plate margins. Such regions display deformation of the fold-thrust belt style, in which a part of the upper crust several kilometers thick is shortened above a basal detachment. Shortening is partitioned among several mechanisms: thrust faulting, folding (which is generally upright to inclined), and the formation of cleavage and intragranular strains (calcite twinning, quartz deformation bands, kinked micas, and so on). These different strain mechanisms are geometrically related because they all accommodate the same regional shortening.

Imagine a region of flat-lying strata that is subjected to layer-parallel shortening (Figure 11.21a). At stresses that are insufficient to initiate faults, pressure solution is activated. As a consequence of pressure solution, spaced disjunctive cleavage domains are formed in argillaceous sandstone and limestone, and pencil cleavage develops in shale. Because of the orientation of the shortening direction, the cleavage domains are oriented approximately perpendicular to bedding. The initial spacing of the domains, you will recall, depends on the original clay content, so in units that are more clay-rich, the cleavage domains tend to be more closely spaced (Figure 11.21a inset). With continued deformation, individual domains get thicker, and new domains are formed. As a consequence, cleavage gets stronger (i.e., domains are more closely spaced). In shales, an incipient slaty cleavage develops. If a detachment fault occurs at depth, deformation of the strata may be decoupled from the deformation of rock below the detachment. In fact, there may be movement on the detachment simply to accommodate the required shortening and cleavage formation in the overlying strata.

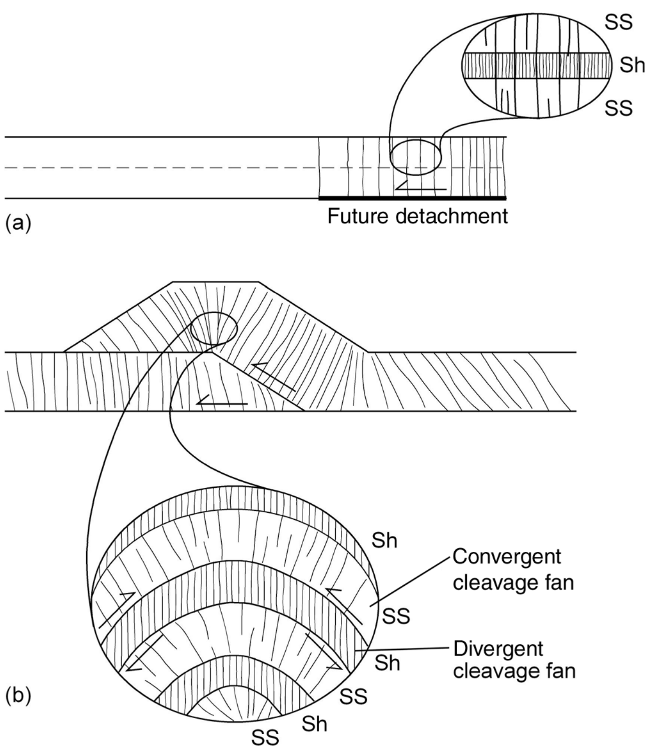



**FIGURE 11.21** Evolution of cleavage in a foreland fold-thrust belt. (a) Initiation of cleavage during layer-parallel shortening. Cleavage starts by being nearly perpendicular to bedding. Inset shows that cleavage is parallel in the sandstone (SS) and shale (Sh) beds, but that the initial spacing is different. After formation of a ramp anticline (b), regional shear has caused cleavage to be inclined with respect to bedding. In the fold itself, cleavage refracts as it passes from lithology to lithology. Cleavage in the shale beds is axial planar with respect to the folds, but fans around the folds in the sandstone layers.

Eventually, if stresses get high enough for faulting to begin, the package of strata containing the cleavage is carried in the hanging wall of a thrust and is folded as it navigates a ramp. During folding, flexural slip occurs between layers, with weaker shale layers caught in-between more competent sandstone or limestone layers. As a consequence, the cleavage rotates and becomes inclined to bedding in shales (Figure 11.21b), while in the more rigid layers it maintains its original orientation at a high angle to bedding. At the ramp, cleavage in the more rigid layers fans around the folds, whereas cleavage in the weaker (shaly) horizons is roughly parallel to the axial plane of the fold (Figure 11.21b inset). During the final phases of deformation, the fold as a whole is flattened and the limbs are squeezed together. In steep to overturned limbs, late-stage tectonic cleavage forms, that is just about parallel to the steeply dipping beds.

Because cleavage fans change from convergent in competent beds (sandstone, limestone) to divergent or axial planar in incompetent beds (shale, marl), cleavage changes orientation from bed to bed, a pattern that is called **cleavage refraction** (Figure 11.21b inset and Figure 11.22). Cleavage refraction is the change in cleavage attitude that occurs where cleavage domains cross from one lithology into another of different competency, and reflects variation in the local strain field between beds. In graded beds, this change in cleavage orientation occurs gradually across the bed, producing curved cleavage surfaces (Figure 11.22). Changes in domain spacing typically accompany cleavage refraction, as this is also controlled by lithology.

If the entire unit being deformed is dominantly shale, then the regional slaty cleavage forms approximately

**FIGURE 11.22** Cleavage refraction in a sandstone—shale sequence (Eifel, Germany). Note that the spacing and nature of cleavage changes between rock types; width of view is ~15cm.

parallel to the axial plane of regional folds. Often, in such tectonic settings, folds are overturned in the direction of regional tectonic transport, so the regional cleavage dips toward the hinterland. In cases where the cleavage is not parallel to the axial plane, but cuts obliquely across the folds, we say that the cleavage is **cross-cutting.** The occurrence of cross-cutting cleavage may indicate that the cleavage was superimposed on preexisting folds, or that there were local complexities in the strain field. For example, rotation of a thrust sheet during folding may cause fold hinges to become oblique to the regional shortening direction. **Transecting cleavage** (Figure 11.23) is a term for cross-cutting cleavage that forms in transpressional environments (meaning there are components of both pure and simple shear). Counterclockwise transecting cleavage, in which the cleavage cuts counterclockwise relative to roughly synchronous fold hinges (Figure 11.23), indicates a component of dextral shear (dextral transpression). Similarly, a component of sinistral shear may produce a clockwise transecting cleavage. However, the use of transecting cleavage as a shear-sense indicator is controversial.

Besides the special situation above, cleavage-bedding relationships provide powerful clues to the relative position of an outcrop with respect to a large regional fold, and cleavage refraction can help you determine the facing ("younging") of folds. Take the example of an initially recumbent fold ($F_1$) with an $S_1$ axial planar cleavage that is folded by a second folding generation

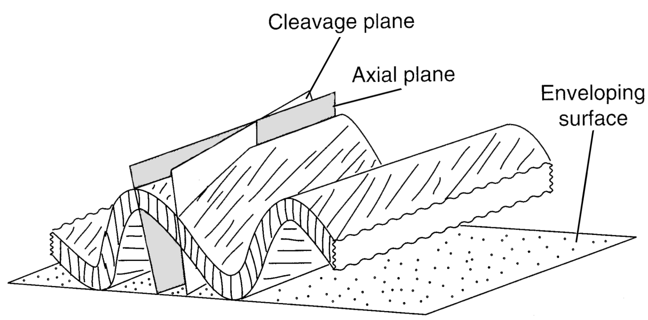



**FIGURE 11.23** The relationship between cleavage, axial plane, and enveloping surface in folds with transecting cleavage. The counterclockwise cleavage transection that is illustrated may be indicative of dextral transpression. Note the obliquity of the bedding-cleavage intersection lineation to the hinge line.

($F_2$), creating an upright, open fold (Figure 11.24). As a consequence, a portion of the $F_2$ fold faces up, while another portion of the fold faces down. If the strata in this fold still contain the $S_1$ cleavage, and if this cleavage is axial planar to the $F_1$ fold in clay-rich horizons, then, on the overturned limb of the upward-facing part of the $F_2$ fold (A), bedding dips more steeply than cleavage, and on the upright limb (B) cleavage dips more steeply than bedding. On the upright limb of the downward-facing part of the large fold (C), bedding dips more

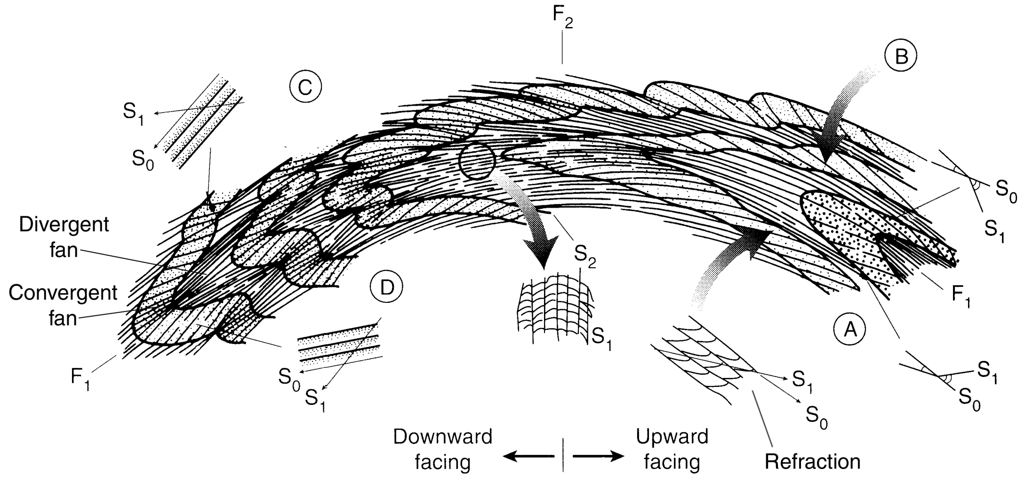



**FIGURE 11.24** Cleavage-bedding relationships and cleavage refraction in upright and overturned limbs of upward-facing and downward-facing folds.

steeply than cleavage, whereas on the overturned limb cleavage is steeper (D). You might also find a crenulation cleavage ($S_2$) that is axial planar to the $F_2$ fold, especially in its hinge region. The geometry of cleavage refraction further indicates the younging direction (i.e., upright and overturned limbs) in the large fold. Figure 11.24 is another one of those rich diagrams that explore a variety of geometries and relationships that will prove to be quite helpful in the field; that is, once you have figured it out. After taking a suitable amount of time to study this final illustration, let's now turn to another element of deformed rocks: lineations.

## 11.6 LINEATIONS

A **lineation** is any fabric element that can be represented by a line, meaning that one of its dimensions is much longer than the other two. There are many types of lineations. Some are associated with other structures (such as folds or boudins), some are visible only on specific surfaces in a rock body, while others reflect the arrangement and shape of mineral grains or clasts within the rock. Some lineations reflect strain axes or kinematic trajectories, while others appear to have no kinematic significance. For their description, we broadly group lineations into three categories: form lineations, surface lineations, and mineral lineations.

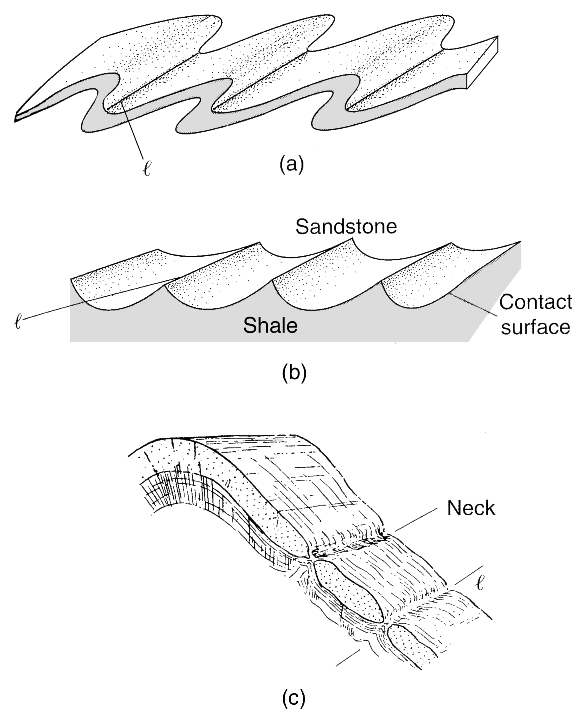



**FIGURE 11.25** Examples of form lineations. (a) Fold and crenulation hinges, (b) mullions, and (c) boudins.

### 11.6.1 Form Lineations

The hinge of any fold is a linear feature. If folds are closely spaced, the fold hinges effectively define a rock fabric that we can measure as a **fold hinge lineation** (Figure 11.25a). Similarly, a **crenulation lineation** is defined by the hinge lines of the microfolds

**FIGURE 11.26** Elongate pebbles (arrows) in a stretched conglomerate (Narragansett, Rhode Island, USA); knife for scale.

in a crenulated rock. Why bother measuring crenulation lineations if a crenulation cleavage is present in the rock? The reason is that sometimes the interval in which crenulations occur is not thick enough for cleavage domains to be measurable. For example, deformation and metamorphism of thick sandstone beds separated by thin interbeds of shale will create quartzite beds separated by thin layers of phyllite. Even if the phyllite layer is so thin that a crenulation cleavage plane cannot be measured, it may be possible to see and measure the crenulation lineation.

When intense deformation detaches the limbs of folds, as occurs during fold transposition, isolated fold hinges may be left in the rock. Such hinges are called **rods** and typically occur in a multilayer composed of phyllite (or schist) and quartzite; the quartz layers are relatively rigid and define visible folds, the limbs of which may be thinned so severely that they pinch out, and the quartz flows into the hinge zone, where it is preserved as a rod. Rodding may also occur in mylonites, because the progressive folding in mylonites may generate rootless isoclinal folds whose limbs are detached and whose hinges (the rods) have rotated into parallelism with the shear direction (see Chapter 13). **Mullions** are cusplike corrugations that form at the contact between units of different competencies in a deformed multilayered sequence (Figure 11.25b); the axes of mullions are a lineation. Typically, the more rigid lithology occurs in convex bulges that protrude into the ductile lithology, and the bulges connect in pointed troughs. Because of their mechanical origin, mullions cannot be used as a facing indicator.

**Boudins** are tablet-shaped lenses of a relatively rigid lithology, embedded in a weaker matrix, that have collectively undergone layer-parallel stretching (Figure 11.25c). In the third dimension, these long tabular bodies are separated by narrow **boudin necks** that are linear objects. In rare cases we find boudins that record extension in two directions, lovingly called chocolate-tablet boudinage. Other elongate objects in rock that are useful lineations include elongate pebbles and elongate pumice fragments. Again, when the long axes of such elongate objects in a rock are aligned, then they define a measurable lineation. An example of stretched-pebble conglomerate is shown in Figure 11.26. The elongation of these embedded objects is generally a manifestation of deformation in some form, which we used in Chapter 4 for strain quantification. When using these objects for structural analysis, however, it is important to make sure that their alignment is a tectonic and not a primary feature (i.e., alignment during deposition in flowing water or air).

### 11.6.2 Surface Lineations

An **intersection lineation** is a linear fabric element formed by, as the name suggests, the intersection of two planar fabric elements (Figure 11.27a). An intersection lineation that structural geologists often use in the field is the **bedding-cleavage intersection,** which

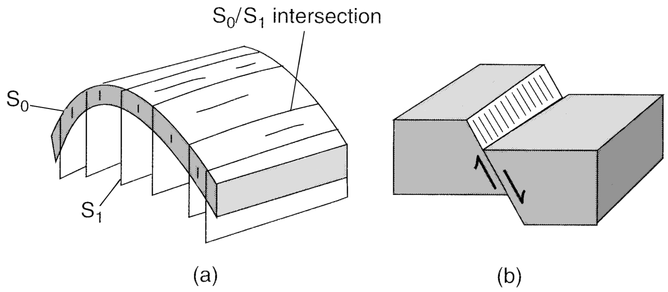



**FIGURE 11.27** Sketches of surface lineations. (a) Intersection lineation of bedding ($S_0$) and (axial plane) cleavage ($S_1$) in a fold, and (b) slip lineation on a (normal) fault surface.

is manifested by the traces of cleavage domains on a bedding plane (or vice versa). When cleavage is parallel to the axial plane of a fold, the bedding-cleavage intersection must parallel the hinge line in mostly cylindrical folds. The field application is powerful, as it allows one to predict regional fold geometry in areas with otherwise sparse outcrop.

**Slip lineations** form on surfaces that move in opposite directions (Figure 11.27b). This occurs, for example, on fault surfaces, but also at the interface between beds in flexural slip folds. There are two basic types of slip lineations: **groove lineations,** formed by plowing of surface irregularities, and **fiber lineations** that are formed when vein mineral fibers precipitate along a sliding surface (see Chapter 8). Slip lineations, of course, are parallel to the slip direction and their roughness may indicate slip sense.

### 11.6.3 Mineral Lineations

When geologists say that a rock contains a **mineral lineation,** they mean that the fabric element defining this lineation is the size of a mineral grain or a cluster of mineral grains (Figure 11.28). Mineral lineations commonly occur in the foliation plane of metamorphic rocks, on shear surfaces, or in the plane of mylonitic foliation. There are several types of mineral lineations. Not all of these have the same tectonic significance, so it is important to determine what type of mineral lineation is present in the rock.

Some minerals, such as kyanite and amphibole, grow such that they are very long in one direction relative to the other two directions. If the long axes of the crystals are aligned in a rock, they create a mineral lineation. The alignment may be due either to growth of the crystal in a preferred direction (controlled by differential stress or by flow-controlled diffusion) or because elongate grains are rotated toward a principal strain direction during deformation. This type of linear fabric is taken to indicate the direction of stretching in the rock and is therefore called a **stretching lineation.** It is quite common in deformed metamorphic terranes to find a consistently oriented mineral lineation that reflects the regional transport direction of the rocks; for example, in areas of regional thrusting.

We earlier mentioned elongate objects, such as stretched pebbles in a conglomerate. Elongate grains or grain clusters produce the same phenomenon, only on a smaller scale. Quartz, which deforms quite readily under metamorphic conditions, may deform into long ribbons that define a distinct lineation in outcrop.

### 11.6.4 Tectonic Interpretation of Lineations

In structural analysis, the **bedding-cleavage intersection lineation** is widely used, because this lineation offers a clue to the orientation of folds in a region where the hinges themselves may not be exposed. Other types of lineations, however, may be more difficult to interpret, because there are at least two alternative interpretations for their origin. First, a lineation can parallel a principal strain; specifically, the direction of stretching or elongation. When we talk about **stretching lineations,** as defined by elongate mineral grains or pebbles, we are implying that the lineation is roughly parallel to the direction of maximum elongation (the *X*-axis of the finite strain ellipsoid). Other lineations, like boudin necks, are roughly perpendicular to the stretching direction. Second, a lineation can be parallel to a shear direction, meaning that it is parallel to a vector defining the motion of one part of a rock with respect to another. Slip lineations (fibers or grooves) are good examples of shear-direction lineations. However, the shear direction is not parallel to the stretching direction, except in special cases. Only in zones of high shear strain does the shear direction approach the finite elongation direction; in other words, grains are stretched and mineral clusters are smeared out in the direction of shear. Clearly, you must be careful when interpreting lineations, and take into account that strain and kinematic indicators can be different.

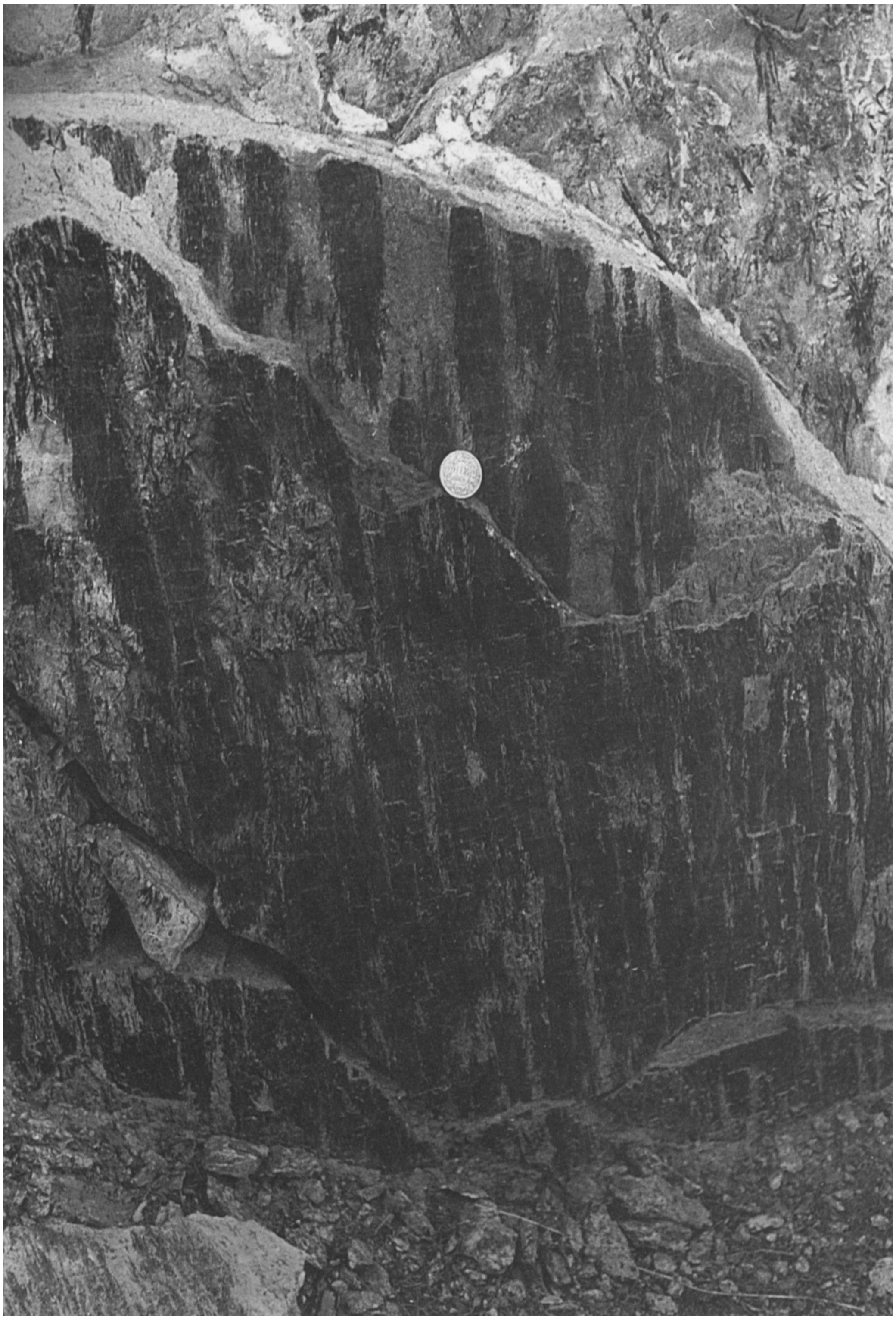


**FIGURE 11.28** A steeply plunging mineral lineation that is defined by hornblende (Gotthard Massif, Switzerland); coin for scale.

## 11.7 OTHER PHYSICAL PROPERTIES OF FABRICS

We focused in this chapter primarily on dimensional and, to a lesser extent, crystallographic aspects of rock fabrics, but these characteristics also impart an anisotropy that may be manifested by other physical properties. It is beyond the scope of this text to explain the underlying principles of these properties or to discuss details of these methods, so we merely mention two of these properties in closing: seismic properties and magnetic anisotropy. **Seismic velocity** varies as a function of material characteristics (such as density and elasticity), so we can use seismic velocity to obtain increasingly more detailed images of the Earth's deep structure (see Chapter 14). Seismic velocity also varies with direction in a single crystal or aggregate of crystals, because the spacing of atoms in the crystal lattice varies in different directions. Thus, when a foliation is defined by the preferred orientation of crystals, the seismic velocity of the sample will vary (by a few percent) in different directions (this phenomenon is called wave splitting). Similarly, the **magnetic properties** of rocks vary as a function of direction, because the ability of a rock to be magnetized depends on the geometric arrangement of atoms in individual crystals and/or the orientation of its mineral constituents. The geometric anisotropy of a foliated rock produces a magnetic fabric that is anisotropic. **Anisotropy of magnetic susceptibility** (AMS) is one measure of a rock's magnetic fabric that is easy to obtain in the laboratory, and is a quick indicator of even the weakest fabric element. In Figure 11.29, we show the distribution of the maximum and minimum susceptibility axes in a progressively deformed sequence of rocks. Notably the distribution of the minimum axes is a sensitive indicator of cleavage intensity. While the orientation of the strain ellipsoid and the ellipsoid representing directional variability of the magnetic fabric are often parallel, the relative magnitudes of these ellipsoids depend strongly on the magnetic phases and processes involved. The relationships between finite strain and magnetic fabric are therefore not straightforward. Nevertheless, the orientation and intensity of magnetic fabrics offer a sensitive measure of fabric elements in natural rocks.

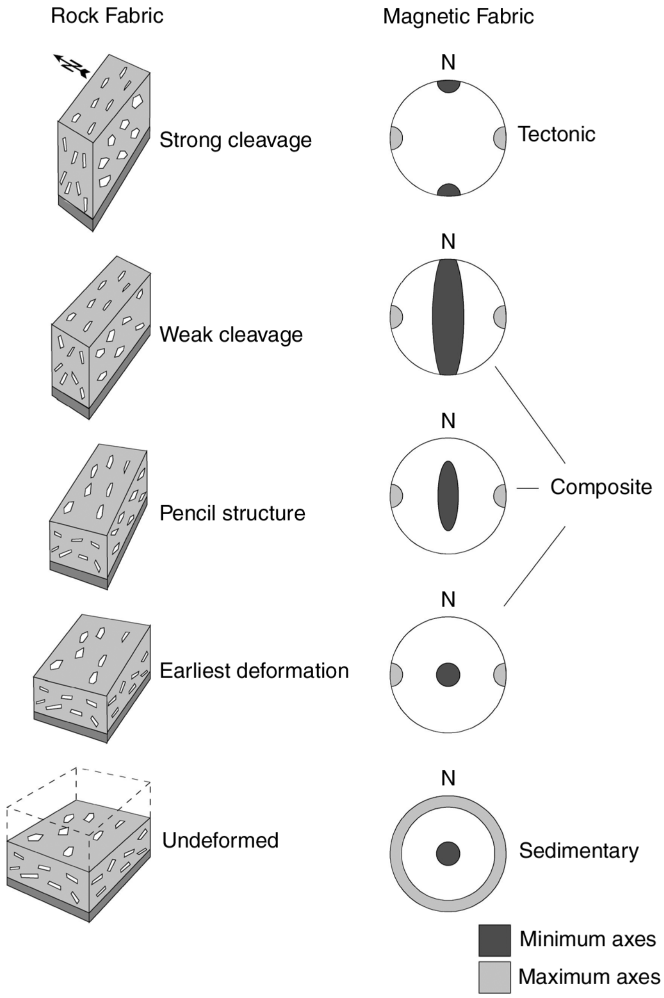



**FIGURE 11.29** Magnetic fabrics in progressively cleaved mudrocks. The degree of foliation development, from compaction fabric to strong cleavage, is characterized by the distributions of the maximum and minimum susceptibility axes.

## 11.8 CLOSING REMARKS

You may find it surprising that after nearly 150 years of research on foliations and lineations (dating from the early studies of Sorby, Darwin, and their contemporaries in the nineteenth century), many questions about these fabric elements remain unresolved. Perhaps one reason for this uncertainty is our inability to create these fabric elements in the laboratory under conditions and processes similar to those in nature; fluid migration and material transfer, especially, pose experimental limitations. So, many of our ideas about tectonic fabrics are based on field evidence, which by its nature is often circumstantial. In the past few decades, however, new progress has been made in our understanding of fabric formation using electron microscopy (scanning and transmission electron microscopy). These high-resolution studies, accompanied by microchemistry, reveal the detailed structure of mineral grains (Figure 11.30). Complementary data obtained from X-ray techniques (texture goniometry) further support a general view of fabric formation based on strain magnitude and metamorphic grade. And all scientists agree? Well, it remains common for a group of geologists to stand on an outcrop and argue about the origin and the meaning, or even the very existence of a fabric. We have attempted to curtail our own biases, but they have undoubtedly crept into the text and may be criticized by your instructor. No matter, debate is what keeps this a vibrant and exciting field of study!

**FIGURE 11.30** Transmission electron micrograph of microfolds in mica from some of the earliest studied slates in southern Wales (UK). Width of view is ~ 0.4 μm.

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