

CHAPTER

14

Formation of Foliations and Lineations

The diverse foliations and lineations described in the previous chapter form in many different ways, and their significance is not always obvious. The principal causes of foliations and lineations are ductile flattening and elongation of the rock itself; mechanical rotation; solution and precipitation; and recrystallization. The latter two mechanisms commonly involve chemical reaction. The composition of the rock also influences the type of foliation developed. Stylolitic foliations, for example, are largely restricted to limestones and marbles, although they also form in other rocks, principally calcareous or argillaceous sandstones. A quartz-rich sandstone characteristically develops a rough to smooth disjunctive foliation, but never a fine continuous foliation. Similarly, a crenulation foliation generally forms in rocks that contain a high proportion of platy minerals, although it may also develop in finely laminated rocks.

In this chapter we discuss the principal mechanisms of formation of foliations and lineations. To do so, we need first to introduce a few concepts of the geometry of deformation, which is treated in more detail in Chapter 15.

14.1 Shortening and Lengthening

At the beginning of Chapter 12, we introduce the concepts of flattening and of shear in two dimensions by considering the deformation of a square of material.

When the square is homogeneously flattened, it changes into an elongate rectangle (see Figure 12.2A, B). When the square is homogeneously sheared, it changes into a parallelogram (see Figure 12.1A, B).

Generalized to three dimensions, a flattening deformation changes a cube of material into a rectangular prism (Figure 14.1A), of which one dimension has been shortened and the other two have both lengthened. If two dimensions of the cube are shortened and only one is lengthened the deformation is a constriction (Figure 14.1A). Shear of a cube parallel to one face and one edge produces a rhombohedron (Figure 14.1A). In each case the original volume of the cube can be, but is not necessarily, conserved. The shaded faces of the blocks show the two-dimensional deformation in the plane that contains the directions of maximum shortening and lengthening.

Instead of using a cube, it is more convenient to represent the effect of deformation by looking at a circle or a sphere, because all lines in all directions start out the same length. The left sides of the blocks in Figure 14.1A illustrate the relationship between the deformation of a square and of a circle inscribed in the square. Figure 14.1B shows the deformation of a sphere equivalent to the deformation shown in Figure 14.1A. The circle is deformed into an ellipse called the *strain ellipse*, and the sphere is deformed into an ellipsoid called the *strain ellipsoid*. We can define the geometry of an ellipse (or an ellipsoid) by the lengths of the two (or three) principal axes. These axes are the *principal axes of strain*. In three

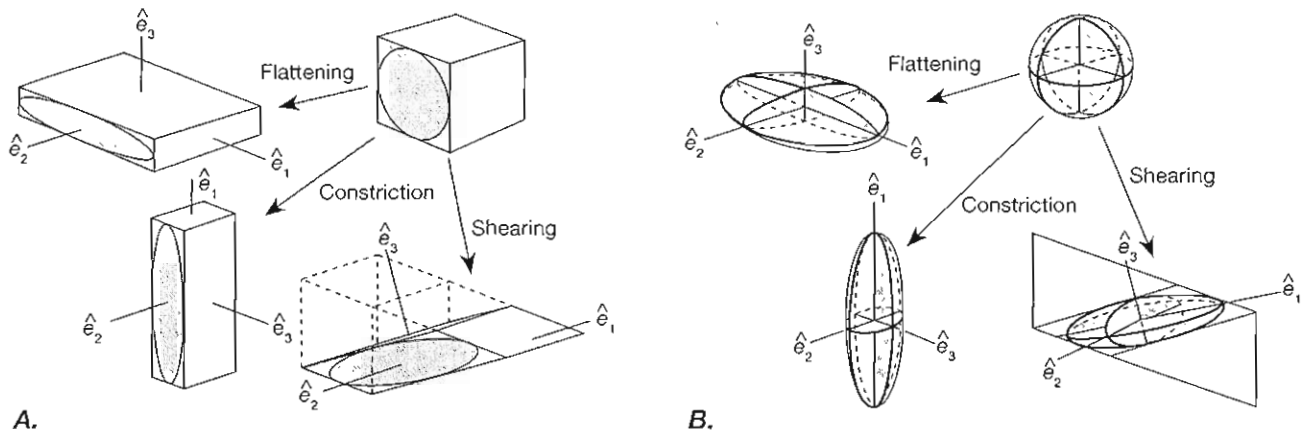


Figure 14.1 Types of homogeneous deformation. Here \hat{e}_1 , \hat{e}_2 , and \hat{e}_3 are the directions of maximum, intermediate, and minimum extension. A. A cube undergoes flattening if two of its dimensions are lengthened and one is shortened; it undergoes constriction if two of its dimensions are shortened and one is lengthened; and it undergoes shearing if the cube is deformed so that its cross section is a rhomboid. The shaded faces of the blocks show the deformation in the \hat{e}_1 - \hat{e}_3 plane. A circle is deformed into the strain ellipse whose maximum and minimum axes are parallel to \hat{e}_1 and \hat{e}_3 . B. The same geometries of deformation as in part A imposed on a sphere produce ellipsoids called strain ellipsoids whose principal axes are the principal directions of extension $\hat{e}_1 \geq \hat{e}_2 \geq \hat{e}_3$. If subjected to flattening, a sphere is deformed into a pancake-shaped (oblate) ellipsoid; if subjected to constriction, it becomes a cigar-shaped (prolate) ellipsoid; and if subjected to shearing, it becomes an ellipsoid with axes inclined relative to the shear plane and with no deformation parallel to the \hat{e}_2 direction. The strain ellipses in part A are sections through the strain ellipsoids.

dimensions, each of the three principal extensions¹ $\hat{e}_1 \geq \hat{e}_2 \geq \hat{e}_3$ (see Equation 9.1) is the change in length of the principal ellipsoid radius divided by the length of the original radius of the sphere. In two dimensions, we generally use only \hat{e}_1 and \hat{e}_3 . The plane of flattening is the \hat{e}_1 - \hat{e}_2 plane, and the direction of maximum principal extension \hat{e}_1 is also called the direction of maximum stretch. We discuss these concepts in more detail in Chapter 15.

A rock is made up of mineral grains² and commonly contains fossils or other deformable objects. Objects that initially are spherical including ooids, radiolarian tests, and alteration spots, are changed into ellipsoids by a deformation such as a flattening. These deformed objects are aligned parallel to the plane of flattening (\hat{e}_1 - \hat{e}_2) and to the axis of maximum stretching (\hat{e}_1) (Figure 14.2A), providing the most straightforward mechanism for formation of a foliation or lineation, respectively. A lineation that is parallel to \hat{e}_1 is called a stretching lineation.

Other features, such as equant mineral grains, clusters of mineral grains, and clasts in a conglomerate, may

not have an initially spherical shape or random distribution of orientations. A deformation such as a flattening, however, also changes these features to produce foliations and lineations (Figure 14.2B, C), although the effect of an initial preferred orientation can never be completely eliminated (Figure 14.2C).

Boudins form during deformation if there is a component of lengthening parallel to a competent layer in an incompetent matrix (Figure 14.3A). Ductile extension of the competent layer cannot keep pace with that of the incompetent matrix, and the layer tends to pull apart into boudins. As the difference in competence increases from small to large, the form of the boudins changes from a pinch-and-swell structure, through separated boudins with pronounced necks, to boudins with sharp ends that may actually be fractures (as illustrated in the progression from top to bottom in Figure 14.3A). If the matrix is too competent to flow around the separating boudins, the region of low stress between the boudins becomes a favorable site for precipitation of a mineral such as quartz or calcite.

In simplified terms (Figure 14.3B), the incompetent material exerts a shear stress σ_s on each side of the layer, which provides a total force $2F_s$ parallel to the layer and proportional to the length L on which it acts ($F_s = \sigma_s L$). A deviatoric tensile stress $\Delta\sigma_n$ acting across the thickness T of the competent layer provides a force

¹ As in the notation for stress, we use a circumflex to indicate a principal value.

² Excepting, of course, coal and volcanic rocks composed of glass.

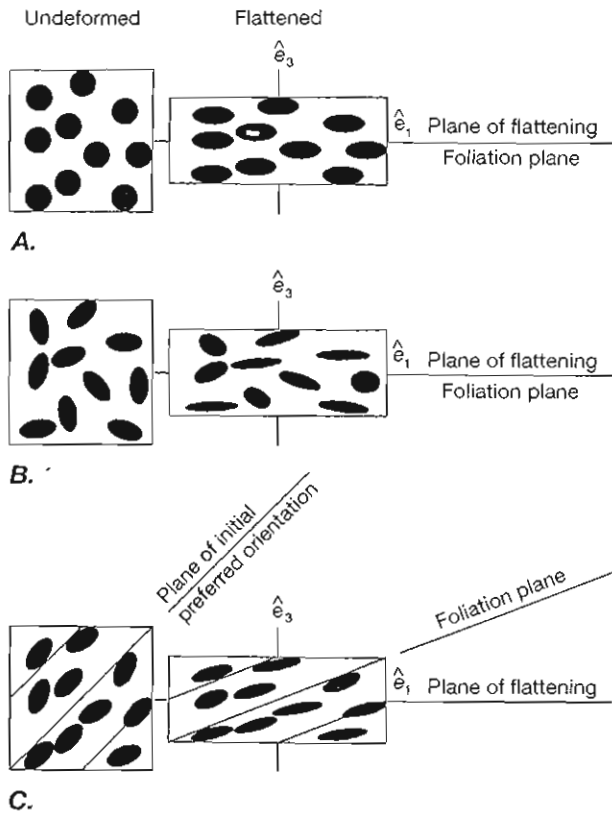
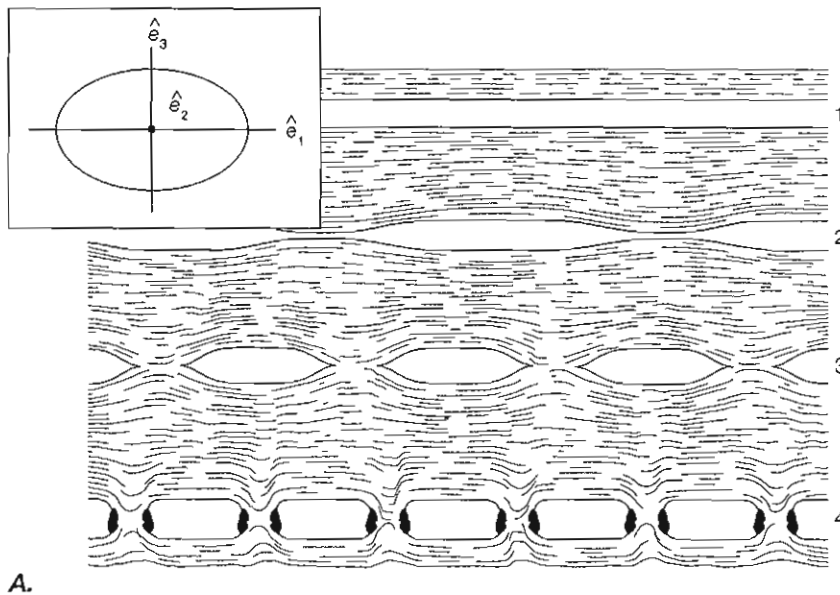
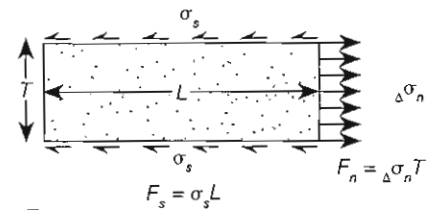


Figure 14.2 Foliation formed by the deformation of discrete objects in the rock. A. Flattening of initially spherical objects produces a foliation parallel to the plane of flattening (the \hat{e}_1 - \hat{e}_2 plane). B. Flattening of randomly oriented elliptical objects produces a foliation statistically parallel to the plane of flattening. C. Flattening of initially elliptical objects with an initial preferred orientation produces a foliation in an orientation different from the plane of flattening.



A.



B.

Figure 14.3 Formation of boudins. A. Competent layers imbedded in an incompetent matrix and oriented parallel to the axis of maximum stretch accommodate the lengthening by segmenting into boudins. The contrast in competence between the layer and the matrix increases from zero (top) to high (bottom), resulting in a progression from (1) uniform stretching and thinning through (2) pinch and swell and (3) necked boudins to (4) fractured boudins. B. The length of each boudin is determined by the yield or tensile strength of the layer and the balance of forces created by the shear stresses σ_s on the layer surface and the deviatoric tensile stress within and parallel to the layer $\Delta\sigma_n$.

$F_n = \Delta\sigma_n T$ that balances the force $2F_s$. The length L^* at which $\Delta\sigma_n$ reaches the yield stress $\Delta\sigma_n^*$ of the competent material determines the location of failure of the layer and the characteristic length of the boudin in cross section.

$$F_n = 2F_s$$

$$\Delta\sigma_n^* T = 2\sigma_s L^*$$

$$L^* = \frac{\Delta\sigma_n^* T}{2\sigma_s}$$

14.2 Mechanical Rotation

The occurrence of mechanical rotation of mineral grains during deformation is demonstrated by "snowball" garnets (Figure 4.17E) and curvilinear fibrous overgrowths. Several lines of evidence indicate that rotation of mineral grains into a preferred orientation is an important mechanism of formation of foliations and lineations. For example, originally detrital mica grains, which are commonly parallel to bedding in undeformed sediments, are parallel to a tectonite foliation in the deformed equivalents. Mica grains in some crenulation foliations are rotated or bent into parallelism with the new foliation. The experimental deformation of rocks that contain randomly oriented micas has produced a preferred orientation under conditions for which rotation is the only possible mechanism of reorientation.



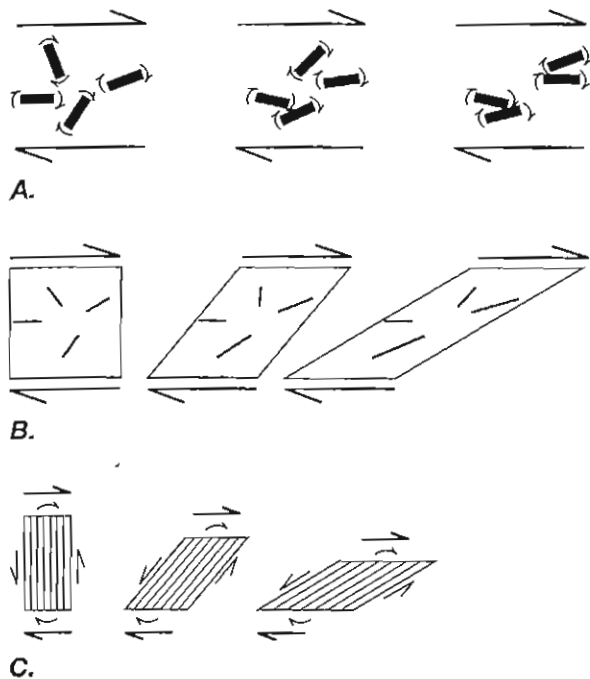


Figure 14.4 Mechanisms of rotation of crystal grains in a deforming rock. A. The Jeffrey model: rotation of rigid elongate mineral grains in a ductile matrix. B. The March model: rotation of mineral grains that act as passive markers during the deformation. C. The Taylor-Bishop-Hill model: rotation of a mineral grain accompanying ductile shear on a single set of slip planes within the grain.

Such a rotation can occur in three ways: A mineral grain can rotate as a rigid particle surrounded by a ductile matrix (the Jeffrey model,³ Figure 14.4A); it can act as a strictly passive marker in the deforming rock (the March model,⁴ Figure 14.4B); or it can shear on crystallographic slip planes and rotate such that its deformation is compatible with that of the surrounding matrix (the Taylor-Bishop-Hill model,⁵ Figure 14.4C).

The consequences of the Jeffrey model for a simple shearing deformation are apparent when a rigid rectangular plate is suspended in a shearing fluid. The plate rotates continually about an axis parallel to the shear plane and perpendicular to the shear direction. Although it never comes to rest, the plate rotates most slowly when it is parallel to the shear plane because in this orientation, the torque applied to the plate by the shearing fluid is a minimum. The plate rotates most rapidly

³ After the British physicist G. B. Jeffrey, who in 1923 investigated theoretically the motion of rigid grains suspended in a deforming viscous fluid.

⁴ After the German physicist A. March, who in 1932 analyzed theoretically the development of preferred orientation in deformable rods and plates.

⁵ Named for the metallurgists G. I. Taylor, J. F. W. Bishop, and R. Hill who developed the theory to explain the formation of crystallographic preferred orientations in ductile deformed metals.

when it is perpendicular to the shear plane because in this orientation the applied torque is a maximum. The shearing of a matrix could thus produce a concentration of the orientations of suspended rigid particles parallel to the shear plane. Such a preferred orientation defines a foliation.

For the March model, the rotation of passive particles in a continuum is different from that of rigid particles in a viscous matrix (Figure 14.4B). For example, during simple shearing of the medium, the passive markers do not rotate continually but instead approach a limiting orientation parallel to the shear plane (Figure 14.4B). They cannot rotate past the shear plane, and the rotation rate approaches zero as the markers approach parallelism with the shear plane. The resulting concentration of passive markers subparallel to the shear plane produces a preferred orientation and thus a foliation.

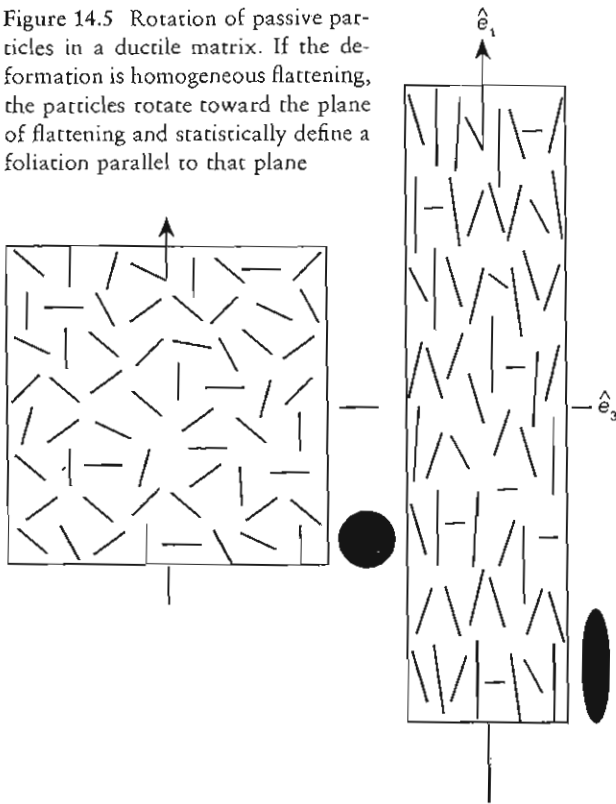
The Taylor-Bishop-Hill model shows how ductile shear on internal crystallographic slip planes of a mineral grain can result in rotation of the grain. For example, Figure 14.4C shows a hypothetical crystal grain that can deform by shear parallel to only one slip plane. Slip on this slip system, however, does not conform to the externally imposed geometry of shear, so the crystal must combine shear on its slip system with a rotation to make the internal and external deformations compatible. When the internal slip system coincides in orientation with the externally imposed shear, the crystal does not rotate further. This mechanism therefore tends to rotate crystals into a preferred orientation that can define a foliation.

During homogeneous flattening (Figure 14.1), any of the three rotation mechanisms results in the rotation of platy or elongate particles, such as mica plates or amphibole needles, toward parallelism with the plane of flattening (the \hat{e}_1 - \hat{e}_2 plane) to produce a foliation. Figure 14.5 shows this effect for randomly oriented grains that behave passively during the deformation. During homogeneous constriction (Figure 14.1), platy or elongate grains rotate toward parallelism with the extension direction \hat{e}_1 to produce a lineation. Grains parallel to any of the principal planes do not rotate, and those initially subparallel to \hat{e}_3 remain at high angles to the foliation.

14.3 Solution, Diffusion, and Precipitation

Rock deformation that produces foliations and lineations depends in part on the mobility of mineral species through the rock. The mechanisms involved include the breakdown of minerals by solution and chemical reaction, the migration of the chemical components

Figure 14.5 Rotation of passive particles in a ductile matrix. If the deformation is homogeneous flattening, the particles rotate toward the plane of flattening and statistically define a foliation parallel to that plane



through the rock, and the formation of new mineral grains by precipitation and recrystallization. Many spaced foliations result in part from such processes.

Stylolitic foliations, commonly found in deformed limestones and marbles, are perhaps the most familiar example. The irregular stylolite cleavage domains may truncate fossils, which indicates that part of the fossil has been dissolved (Figure 13.5B), and that solution has accommodated shortening across the stylolite. The material that fills these stylolites—largely clay minerals, iron oxides, and carbonaceous matter—is the insoluble residue from limestone solution and may include some secondary minerals as well.

In some deformed sandy argillites that have a rough disjunctive foliation, truncation of detrital sand grains against cleavage domains (C in Figure 13.7) results from solution rather than from shear displacement along the cleavage. Originally equant detrital grains, such as quartz, may ultimately be almost completely dissolved away into thin, platelike grains parallel to and partly defining the foliation (T in Figure 13.7). An insoluble residue of platy minerals and oxides forms the cleavage domains.

The dissolved material migrates through the rock, probably by grain boundary diffusion over short distances or by transport in a fluid flowing through pores or fractures over large distances. The dissolved material

often reprecipitates locally, possibly in microlithons where it accommodates a local dilation (a volume increase); as overgrowths on preexisting minerals or particles in the rock (B in Figures 13.7 and 13.26B); as slickenfibers on shear surfaces (Figure 13.26A); or as fibrous or massive deposits in veins. In some cases, however, the bulk composition of the rock may be permanently changed by the removal or introduction of one or more chemical components.

Two major factors affect the dissolution of minerals. First, any deformed mineral is more soluble than an undeformed one because of its higher locked-in strain energy (see Chapter 19). Second, a crystal subjected to a differential stress tends to dissolve more readily at surfaces on which the normal-stress component is a maximum. This mechanism is commonly called Riecke's principle,⁶ and the process is called pressure solution or solution transfer. A corollary of Riecke's principle is that minerals tend to precipitate at surfaces on which the normal-stress component is a minimum.

Volume-loss folding is accomplished by solution and removal of material from the folding layers (Section 12.3 and Figure 12.10). In the resulting folds (see, for example, Figure 12.11), the foliation develops as an axial foliation composed of solution seams.

Crenulation foliations commonly exhibit a compositional banding associated with the cleavage domains and microlithons. Cleavage domains tend to be enriched in platy minerals and depleted in quartz, compared both to the microlithons and to the uncrenulated rock (Figure 13.9A, C). In some cases the microlithons are enriched in quartz, suggesting solution of quartz in the cleavage domains and precipitation in the microlithons.

The banding shown in Figure 13.9A, C, for example, could result either from preferential solution of more highly deformed minerals in the cleavage domains or from pressure solution, particularly if quartz dissolved preferentially at a quartz–mica interface. If the maximum compressive stress $\hat{\sigma}_1$ were at a high angle to the limbs of the crenulations, and therefore to the quartz–mica interfaces, quartz would readily dissolve in the crenulation limbs. In the microlithons, however, the quartz–mica interfaces would be subparallel to $\hat{\sigma}_1$ and at a high angle to the minimum compressive stress $\hat{\sigma}_3$, and on those surfaces the quartz would tend to precipitate. This mechanism could therefore account for the migration of quartz from cleavage domain to microlithon.

Oriented overgrowths, or pressure shadows, on mineral grains and particles may originate in the same way. If the particles in question behaved relatively rigidly during ductile deformation of the surrounding ma-

⁶ After the nineteenth-century German physicist E. Riecke.

terial, a zone of abnormally low stress would develop where the particle boundary is at a high angle to the minimum compressive stress $\hat{\sigma}_3$. Minerals in solution diffuse to the low-stress area and precipitate as either nonfibrous or fibrous overgrowths (B in Figures 13.7 and 13.26B).

14.4 Recrystallization

Recrystallization is the creation of new crystal grains out of old ones. During deformation, recrystallization can result in a preferred orientation of mineral grains. Two kinds of recrystallization are important in structural geology. In coherent recrystallization either old deformed grains are progressively transformed into new undeformed grains as a grain boundary migrates through the old crystal lattice, or old grains are subdivided into many new grains by the rotation of small internal domains called subgrains. The crystal structure and the composition of old and new grains are the same, although new grains have different lattice orientations from the old. In reconstructive recrystallization, the old crystal structure breaks down—for example, during a chemical reaction—and a new structure forms that generally has a different composition. The distinction between the solution/precipitation process and reconstructive recrystallization is not always well defined.

Both types of recrystallization can change the shape and arrangement of grains. A foliation or lineation can develop by solution or chemical reaction, for example, either by a selective destruction of old grains that leaves only grains with a particular orientation, or by the production of new grains that grow in a preferred orientation.

The evolution of slaty cleavage shown in Figure 13.11 is an example of the effect of reconstructive recrystallization. The initial foliation is a zonal crenulation foliation (Figure 13.11A). With increasing amounts of recrystallization of the platy clay minerals, the structure of the foliation changes to a discrete crenulation foliation, and then gradually to a disjunctive foliation with a very strong fabric in the microlithons but no remnant of the initial crenulations (Figure 13.11B). These slates are still low-grade metamorphic rocks.

Some phyllites contain relict sedimentary mica grains as well as newly crystallized micas. The relict grains are characterized by large, irregular grain shapes; the new grains are smaller with very regular grain boundaries and a strong preferred orientation that defines the new foliation. If recrystallization is extensive, it can obliterate all clues to the nature and preferred orientation of the original grains.

Mimetic growth (the Greek word *mimetikos* means “imitation”) is the growth of new crystals that nucleate on older crystals of similar structure in an orientation governed by the orientation of the older crystal. In this way, the growth of new crystals can enhance any preexisting preferred orientation of the old crystals.

An existing foliation in a rock can also control the orientation of new mineral grains, because growth may occur parallel to the foliation more easily than across it. Micas, for example, grow most rapidly parallel to their cleavage plane, and new mica grains that nucleate with cleavage planes parallel to a foliation grow more rapidly than micas in other orientations, thereby enhancing the preexisting foliation.

Without some external controlling factor, reconstructive recrystallization generally does not produce a preferred orientation, and it can even destroy a preexisting one. Recrystallization in association with deformation, however, can produce a preferred orientation or enhance an existing one, and because these processes are very commonly associated, the interaction between them strongly influences the fabrics that result.

14.5 Steady-State Foliations

Different types of foliation are not independent of one another and in fact may develop from one another, as is evident from Figures 13.9, 13.11, and 13.17. Thus crenulation foliations develop from the deformation of an earlier foliation, and with progressive deformation and recrystallization, a crenulation foliation can evolve into a continuous foliation. Continuous foliations in turn can become crenulated (Figure 13.9) and evolve into a crenulation foliation.

Such sequences of development raise the question of whether a foliation ever reaches a “final” state of evolution, and if so, under what circumstances that could occur. In some regions, two or more cycles of foliation evolution have been deciphered, each of which includes the crenulation of an initial foliation, whether of sedimentary or tectonic origin, followed by the formation of a crenulation foliation, commonly with compositional differentiation accentuating the difference between cleavage domains and microlithons, and followed in turn by increasing recrystallization and the development of a new continuous foliation.

Such circumstances are probably common, and make the realistic identification of different generations of foliation, and the correlation of particular generations from one area to another, highly suspect. The designation of different generations of foliation by numerical subscripts such as S_1 , S_2 , and so on thus may only have very local significance. If applied over a large area, such

a system of notation indicates a possibly erroneous conclusion concerning the correlation of different foliations and the simplicity of the deformational history.

14.6 Slickenside and Mineral Fiber Lineations

Slickensides on faults generally contain lineations that parallel the direction of slip on the fault surface (Section 4.2). The various types include structural slickenlines, mineral slickenlines, and slickenfibers. Mineral fiber lineations occur not only as slickenfibers but also as fibrous vein fillings and fibrous overgrowths. These lineations originate by a variety of different mechanisms, which we discuss below.

Structural Slickenlines

Fault surfaces are never perfectly flat but contain minor irregularities or protrusions called asperities (the Latin word *asper* means “rough”) (see Figure 19.2). If asperities are particularly strong and resistant to abrasion and fracture, they can scratch and gouge the opposite surface of the fault, giving rise to one type of structural slickenline (Figure 14.6A). Scratches and gouge marks end where an asperity breaks off the opposite surface of the fault. The length of these lineations is a lower bound for the displacement on the fault surface.

Small ridges can develop where the fracture plane is deflected behind a hard asperity. A corresponding groove, of course, must form on the opposite surface. Similarly, ridge-in-groove lineations, or fault mullions, form if the fault surface is an irregular surface rather than planar, and if the irregularities are linear and par-

allel to the slip direction (Figure 14.6B; see also Figure 4.8B). In both cases, the length of the lineation is not necessarily related to the amount of displacement on the fault, because the ridge and the matching groove form as part of the fracture propagation process, not as a result of the displacement on the fault. Thus they cannot be used to constrain the magnitude of the displacement.

The displacement on some fault surfaces is accommodated by solution where there is a component of shortening across the fault. The mechanism is similar to the production of stylolites, except that the solution surface, called a slickolite surface, is subparallel to the displacement direction rather than approximately normal to it, as for stylolites. The counterpart of the tooth structure on stylolites (Figure 13.5) is a spike on a slickolite surface (Figure 14.6C).

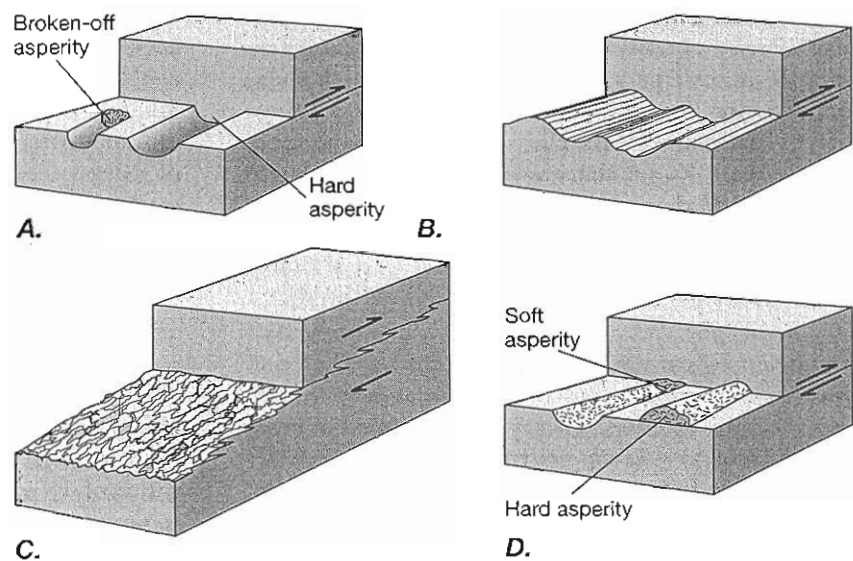
Mineral Slickenlines

Slickenlines defined by streaks on a slickenside result from the smearing out of mineral grains and soft asperities (Figure 14.6D). They may also accumulate behind hard asperities and may form in combination with scratches and gouge lineations.

Mineral Fiber and Slickenfiber Lineations

Mineral fiber lineations occur as slickenfibers on fault surfaces (Figure 14.7A, B), as fibrous vein fillings (Figures 14.7C, D and 14.9), and as fibrous overgrowths on grains or particles (Figure 14.8). The continuity and morphology of the mineral fibers across the vein or shear surface imply that fiber growth accompanied and kept pace with gradual displacement and crack opening.

Figure 14.6 Structural and mineral slickenlines. A. Structural slickenlines formed by scratching and gouging of one side of the fault by hard asperities in the other side. B. Ridge-in-groove structural slickenlines formed by linear irregularities in the fault surface that parallel the slip direction. C. Spikes on a slickolite. The slickolite is a solution surface subparallel to the direction of displacement. The spikes are irregularities in the solution surface that parallel the slip direction and are comparable in origin to the teeth on stylolites. D. Mineral streak lineations form from the wearing down and smearing out of mineral grains and soft asperities or from the collection of gouge behind a hard asperity.



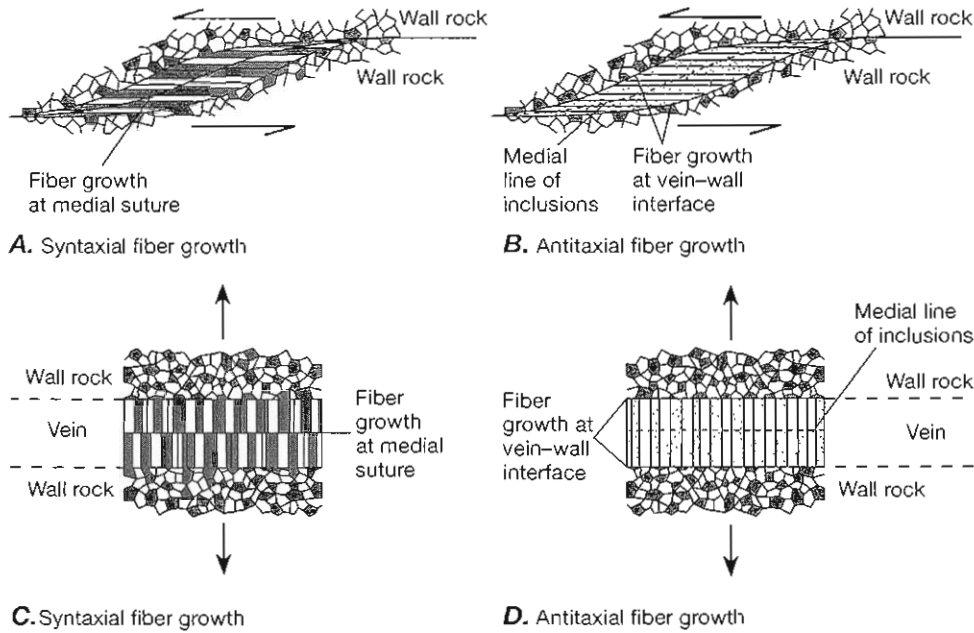


Figure 14.7 Comparison of syntaxial and antitaxial mineral fiber lineations in faults and veins. Syntaxial growth occurs if the mineral making up the fibers is also a common mineral in the host rock. If the mineral fibers are different from minerals in the host rock, antitaxial growth occurs. The arrows indicate the direction of displacement. *A.* Syntaxial growth of slickenfiber lineations on a fault surface. Growth occurs along the medial suture. *B.* Antitaxial growth of slickenfiber lineations on a fault surface. Growth occurs along the interface between the fibers and the wall of the fault. *C.* Syntaxial fiber growth in a vein occurs at the medial suture in the vein. *D.* Antitaxial fiber growth occurs at the interface between fibers and wall rock.

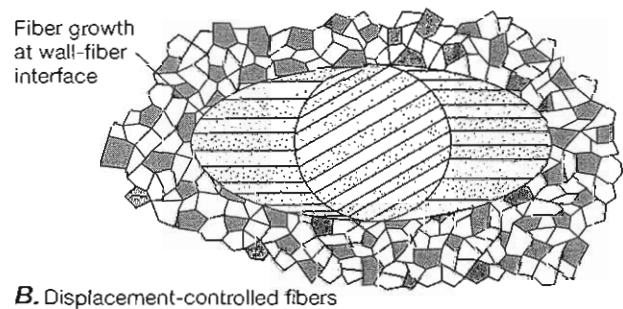
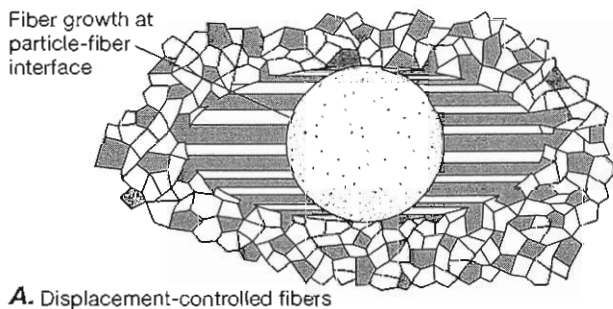
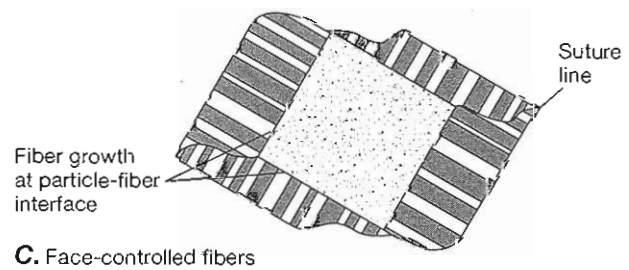


Figure 14.8 Fibrous overgrowths or pressure shadows. *A.* If the fiber mineral is similar to host rock minerals and different from the particle mineral, fibers grow in optical continuity with similar mineral grains in the wall rock. Growth occurs at the fiber-particle interface. *B.* If the fiber mineral is the same as the particle but different from the minerals in the host rock, fibers grow in optical continuity with the particle, here illustrated by a twinned calcite grain. Growth occurs at the fiber-wall rock interface. *C.* Face-controlled fiber orientation in a fibrous overgrowth on a pyrite cube. The mineral fibers grow perpendicular to the crystallographic faces of the pyrite. Growth occurs at the fiber-pyrite interface, and the suture line between differently oriented groups of fibers indicates the displacement of the corner of the grain.

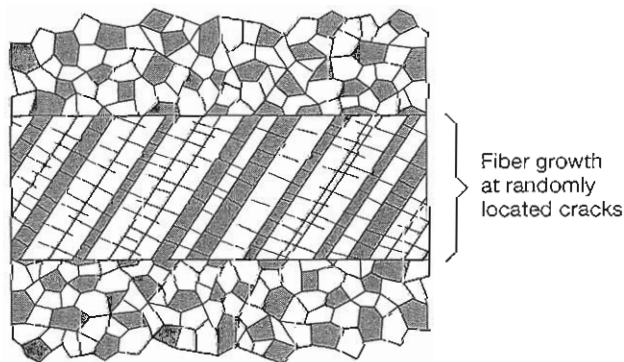


The transport of the mineral constituents to the ends of the fibers occurs either by diffusion along grain boundaries or, more probably, through a fluid phase in which the constituents are dissolved. Slickenfibers on fault surfaces therefore imply slow aseismic creep rather than large, rapid displacements that would produce earthquakes.

Whether the fibers are at low angles to fault surfaces (Figure 14.7A, B), at high angles to vein surfaces (Figure 14.7C, D; see also Figure 14.9), or attached to mineral grains or particles as overgrowths (Figure 14.8), they generally grow in an orientation parallel to the direction of displacement at the time of fiber growth (see Sections 16.3 and 17.5). In some cases, however, the fibers grow normal to the crystal faces of a mineral grain, such as pyrite (Figure 14.8C). In this case, the suture line between the differently oriented fibers records the displacement history of the corner of the grain. The mechanism is not well understood, but the different fiber orientations at different crystal faces make this case easy to identify. Four types of displacement-controlled fiber growth are recognized that reflect the process by which growth has taken place.

In syntaxial growth, fibers tend to grow in optical continuity with mineral grains of the same composition. (In Greek, *syn* means "with, together" and *tassein* means "to arrange.") Thus for this structure to develop, the fiber mineral must be a mineral that is present in the host rock. For slickenfibers (Figure 14.7A) and vein fillings (Figure 14.7C), fibers extending from mineral grains in opposite walls meet at a medial suture where there is both a structural and an optical discontinuity. The suture is the site of latest growth of the fibers.

In fibrous overgrowths around a particle, the fibers can grow syntaxially on mineral grains in the host rock, with fiber growth occurring at the fiber-particle interface (Figure 14.8A), or they can grow syntaxially on the particle itself, as illustrated for a particle of twinned calcite in Figure 14.8B. In this case, fiber growth occurs at the interface between fiber and wall rock.



Antitaxial growth occurs when the fiber mineral is absent or uncommon in the host rock. In slickenfibers (Figure 14.7B) and vein fillings (Figure 14.7D), a medial suture may contain inclusions of host rock composition, but the fibers are optically and structurally continuous across the suture. Fiber growth occurs along the margins of the vein or fault, where there is a discontinuity in mineral composition.

Composite growth occurs when fibers of two different minerals grow, one of which is common, and one rare or absent, in the host rock. The fiber structure in veins may then show a central antitaxial band of the mineral that is rare in the host, flanked on both sides by syntaxial hands of the mineral that is common in the host. The fibers in both types of bands grow at the interface between the different bands.

Although we have discussed only straight fibers here, curved fibers on fault planes (Figure 13.26A) and in veins and overgrowths (Figure 13.26B) are relatively common. The curvature of the fibers in most cases records a component of rotation in the deformation during fiber growth.

Fibers may also grow by the crack-seal mechanism, which involves repeated microfracturing of the fiber along its length, followed by the deposition of optically continuous overgrowths that heal the fracture. Evidence for this mechanism includes abundant subplanar arrays of microscopic fluid inclusions crossing the fibers at the healed cracks (Figure 14.9). The result is sometimes called stretched crystals. They can occur if the fiber mineral is the same as the dominant mineral in the host rock. Crystal fibers are structurally and optically continuous across the whole vein, and they often connect, and are optically continuous with, two fragments of crystal grain on opposite sides of the vein that were originally a single grain. Because an increment of growth may occur at any place along the fiber, the orientation and shape of the fiber do not necessarily record the history of the displacement—only the net result.

Figure 14.9 Crack-seal growth forming a mineral fiber lineation of "stretched" crystals in a vein. Growth does not occur at any particular surface but by repeated cracking followed by deposition of fiber mineral to close the crack.

Additional Readings

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