SECONDARY PLANAR STRUCTURAL ELEMENTS

The **fabric** of any object is the geometric relationship between regularly repetitive, homogeneously distributed constituent parts (grain shape and size, clasts, etc. in rocks) at any scale. Rock fabrics reflect composition, mineralogy and a number of processes that rocks have undergone. **Primary fabric** is a grain configuration developed during the deposition of sediment or the emplacement of a magmatic rock. Mineral grain packing and preferred shape orientation of pebbles of feldspar phenocrysts are examples of primary fabric. **Secondary fabrics** reflect strain and deformation conditions. The basic categories of fabrics are:

- 0-dimensional (no fabric): random.
- 1-dimensional: linear
- 2-dimensional: planar

Accordingly, it is conventional to distinguish

- L fabrics, characterized by elongated elements whose long axes cluster close to a unique line. The rocks, L-tectonites, have a dominant linear fabric.
- S fabrics, where the long axes of essentially flat fabric elements cluster strongly near a plane. S-tectonites have a dominant planar fabric.
- L-S fabrics, where there is moderate clustering near both a plane and a line that lies in that plane. L-S tectonites display equally developed linear and planar fabrics.

In strain terms, these types of fabrics correspond to constriction, flattening, and intermediate strains, the latter yielding k values (definition in finite strain) near 1.
A secondary fabric expressed through the **preferred orientation** of the individual minerals constituting the rock is the most distinctive feature that separates metamorphic rocks from both igneous and sedimentary rocks. This fabric commonly consists of sets of uniformly developed, closely spaced and parallel planes along which the rock splits easily, without regard to the orientation of bedding. These secondary planes, along which the brittle strength of the rock is minimal, are thus **mechanical anisotropies**. Since they have no counter-part in undeformed rocks they must be produced by ductile deformation.

Definition

Foliation is the general term describing the arrangement of any kind of sub-parallel, closely spaced and low-cohesion surfaces that are no strata in deformed rocks (and glaciers). These generally regularly spaced surfaces impart to foliated rocks the facility to split into leaf-like (*folia* = leaf in Latin) planar elements other than bedding.
Foliation planes are reported for a wide range of temperature and pressure conditions, from shallow crustal to deep mantle conditions. Any plane is referred to as S-surface. Where S-surfaces of different generations can be distinguished by type and age (crosscutting relationships, overprinting, absolute age of mineral components), they are given numerical subscripts according to relative timing: S_0 is the primary surface, generally bedding, and S_1, S_2 … S_n are secondary foliation planes in order of determined superposition. Such a foliation-related reference frame helps to unravel the tectonic and metamorphic evolution of the area where S-surfaces are present.

Morphological classification

Morphological features used to characterize and classify foliations are those used for planar features. They refer to:

- Spacing between the planes or planar domains.
- The shape of the planes (rough, smooth, wriggly, etc.).
- The spatial relationship between planes (parallel, anastomosing, conjugate, crosscutting…).
- Characteristics of the boundaries of planar domains (gradational, sharp, discrete, etc.).
- The fabric of the rock between foliation planes (planar, folded, etc.).

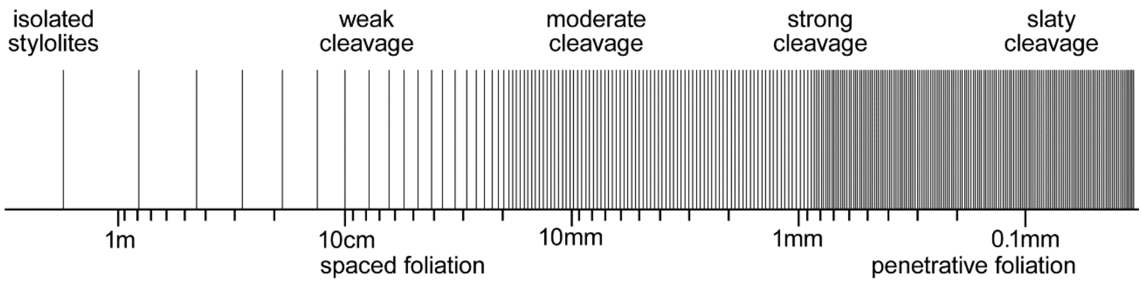
The morphology of foliations reflects formation processes. For example, bedding-parallel foliation may reflect **compaction**, i.e. volume reduction due to pore-water expulsion from fresh sediments under the increasing weight of overburden load. In deeper crustal levels, the foliation morphology varies with metamorphic grade (the temperature and pressure conditions during foliation development), with the position in folds and with rock type and composition. Foliation is classified by mechanisms of formation into two broad types and several endmembers. Morphological gradations between the descriptive terms classified below exist and provide a wide spectrum of intermediate forms.

Spacing scale

Foliation fabrics vary between two broad descriptive categories referring to the overall spacing of foliation planes: **spaced** foliation and **penetrative** foliation.

Spaced foliation

Spaced foliation planes are discrete, tabular **domains** separated by thin slabs of rock without fabric or with a differently oriented, older, primary (original) or secondary fabric. These rock slabs which are thick enough (> ca. 1mm) to be distinguished in hand specimens or outcrop are called **microlithons**. Foliation domains (the *foliae*) are heterogeneously distributed lamellae where the fabric and mineralogy of the host rock have been altered (usually concentration of phyllosilicates and opaques) so that minerals show a preferred shape and/or crystallographic orientation. Foliation domains are those thin planar regions along which the rock splits.



Qualitative classification of foliation based on domain spacing

Penetrative (continuous, pervasive) foliation

In penetrative foliation, all platy grains have a statistically preferred planar orientation. Penetrative means that these fabric elements (equally spaced surfaces that are approximately planar and parallel) appear everywhere throughout the entire rock mass and tend to persistently obliterate earlier structures. Imagine sheets of a book to envision a penetrative foliation. Planes are defined by discontinuities, preferred dimensional orientation of platy minerals, laminar mineral aggregates, or some combination of these structures. The penetrative fabric is visible down to the scale of individual grains and intervals between the foliation planes (or **cleavage)**.

> Note
> The spaced versus the penetrative character of foliation planes is a scale-dependent concept: A penetrative fabric at map and outcrop scales may prove to be spaced (disjunctive) in a thin section.

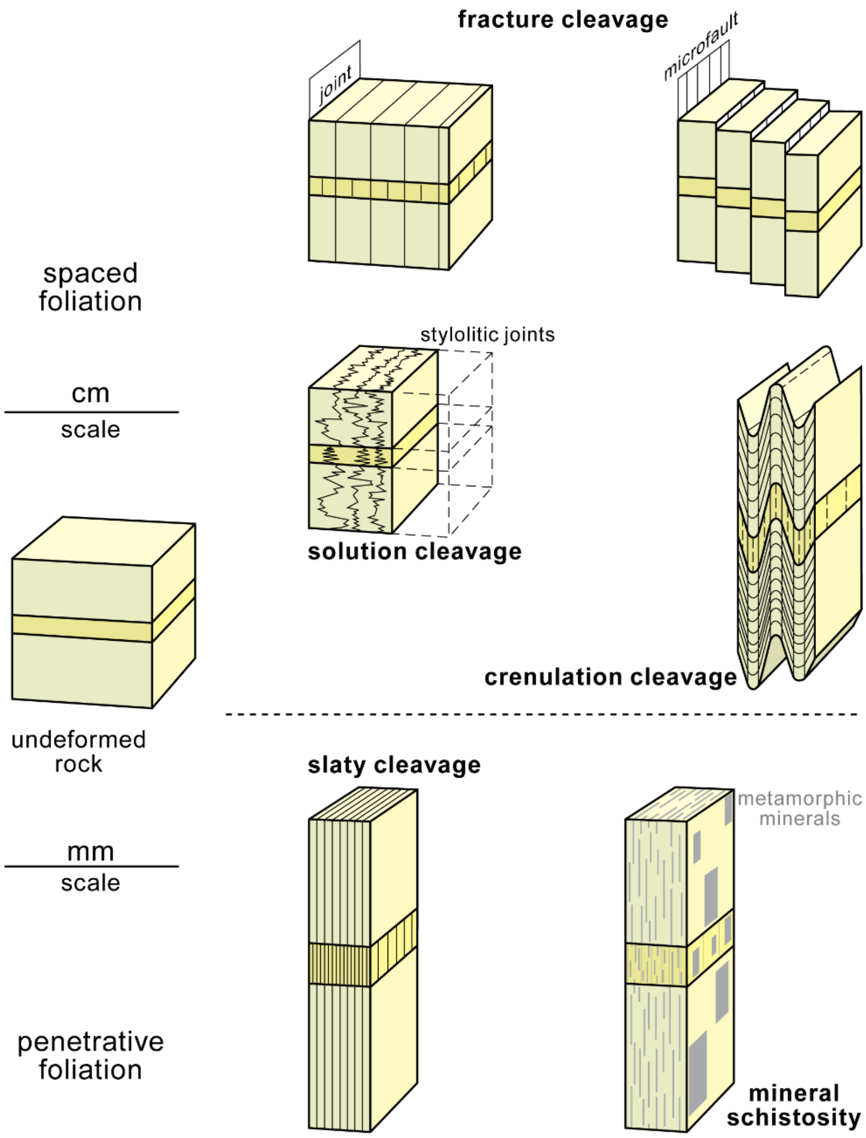
Spaced foliations

Fracture cleavage

Fracture cleavage consists of evenly spaced, planar discontinuities that sharply divide the rock into a series of plate-shaped microlithons that display essentially no internal deformation. Fracture cleavage can be envisioned as a dense population of joints or microfaults generally formed in low metamorphic grade, competent rocks such as sandstone and limestone, where fracture cleavage may coexist with and grade into slaty cleavage in interlayered pelites. Microscopic to meter-scale sets of foliation-like, closely spaced yet non-penetrative fractures may be confused with dense sets of joints. Disjunctive fracture cleavage, occasionally with shear movement, is not a "true" foliation in terms of finite strain: it is a **false cleavage**.

Solution cleavage

Solution cleavage consists of regularly spaced dissolution surfaces (e.g. **stylolitic joints**) that divide the rock into a series of microlithons. Dissolution surfaces, along which some rock mass has been removed often contain dark seams of insoluble residues that may impart a prominent striping to the rock. Stripes denote the spatial variation in mineral composition and/or grain size. Mineral overgrowths, pressure shadows and veins record local mass transfer. Solution cleavage is generally formed in fluid-rich, low metamorphic grade rocks and is common within limestone as regularly spaced stylolitic planes.



Morphological classification of foliations

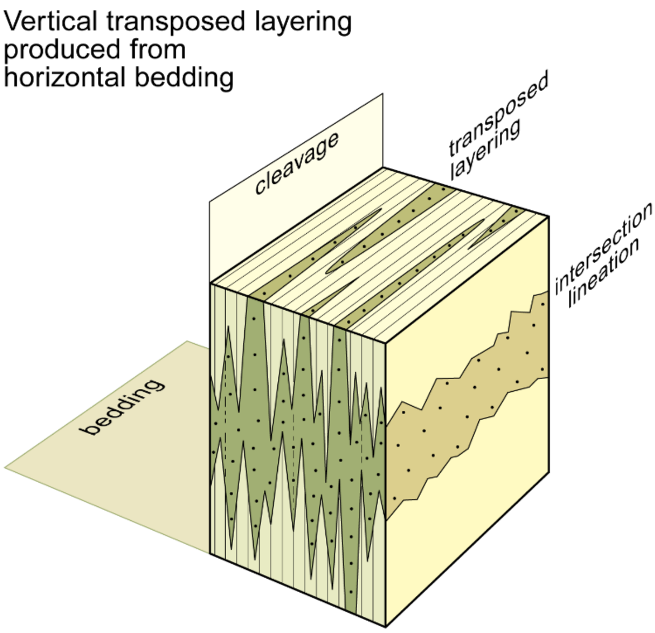
Crenulation Cleavage

Crenulation cleavage is created when an earlier foliation is folded (**crenulated**) on a meso- to micro-scale. The small, regular crinkle folds (10^{-1}-10^{1} mm) may be symmetric but are most commonly asymmetric. The crenulation cleavage is defined by the parallel alignment of grains in the limbs of

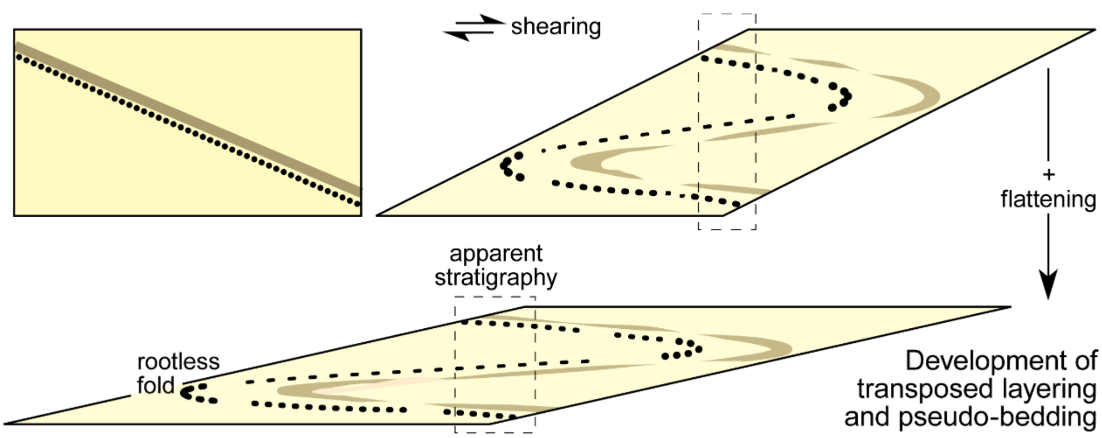
the tight to isoclinal microfolds whose wavelength controls the spacing of successive foliation planes. Where microfolds are asymmetric, the short limb usually becomes the cleavage plane.
The dissolution of soluble material often takes place along the microfold hinges, leaving a concentration of insoluble residue along these limbs while soluble material (generally quartz) is transported and precipitated in hinges.
Crenulation cleavage is sometimes defined by microfaults parallel to the microfold limbs and then referred to as **strain slip cleavage**, but the non-genetic name "crenulation cleavage" is preferable. A **discrete crenulation cleavage** truncates the pre-existing cleavage. Conversely, the pre-existing fabric can be followed from microlithons through a **zonal crenulation cleavage**.
Crenulation cleavages are found in all metamorphic grades.

Transposed layering

Transposition layering is defined by parts of a pre-deformation surface (bedding or an older foliation) which are rotated independently into a new orientation; after intense deformation all of these parts are subparallel.

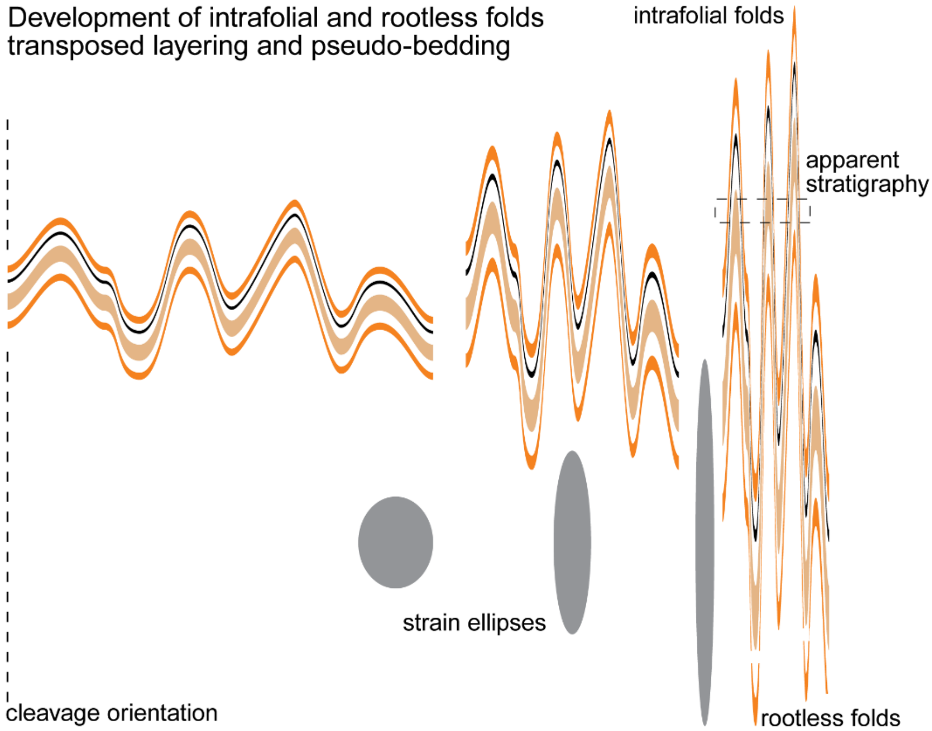


In a laminated sequence of rocks, successive layers generally have different competences. Intense deformation produces appressed to isoclinal folds through rotation and stretching/thinning of limbs until these coincide with the foliation plane.



Fold hinges are sharp and folds are **intrafolial**. Hinges may be torn apart along the stretched limbs that ultimately disappear. The intrafolial folds are **rootless** where hinges are completely detached

from the limbs and relic bedding appears to be restricted to local occurrences in obscured fold hinges. There is then practically no variation in the orientation of the transposed bedding. A transposed sequence may be mistaken for a normal sedimentary succession. Nevertheless, **pseudo-bedding** has no stratigraphic significance.
Recognition criteria for transposition are (1) foliation parallel to bedding, (2) isolated, intrafolial fold hinges, (3) isolated boudins of competent layers, (4) extreme flattening of strain markers and (5) reversals of younging criteria and asymmetry of parasitic folds in close proximity.



Transposition is found on all sizes ranging from hand specimens to several kilometers big structures.

Differentiated layering

Crystallization of metamorphic minerals, metamorphic recrystallization of older minerals and pressure solution may create independently or together a new, generally rough "layering" defined by alternating layers of different composition and/or grain size. In effect, metamorphism reorganizes the chemical components of rock and produces new minerals in new orientations governed by evolving strain. The resulting compositional, **differentiated layering** is visible as distinct light and dark-colored bands in hand specimen. Differentiated layering is found in medium to coarse-grained, granular metamorphic rocks of all grades. Slaty cleavage, crenulation cleavage, and schistosity can be differentiated.
In high-grade rocks, differentiated layering is customarily described as **gneissic layering**, also commonly defined by alternating mafic (dark-colored) and felsic (light-colored) layers. The resulting lithological banding may be more or less modified bedding, and thus reflects either initial sedimentary compositional differences or a foliation entirely due to differentiation during deformation. Gneissic layering can also result from oriented melt segregation during partial melting and/or intimate injection of subparallel igneous veins. **Gneiss** is a generally coarse-grain, metamorphic rock with gneissic layering.
Strongly sheared rocks develop a **mylonitic foliation**, which is both a mineral foliation due to the preferred orientation of platy mineral grains and aggregates and a planar shape fabric defined by the flattened crystals (called ribbons).

Penetrative foliations

The rock and foliation classification (diagenetic, slate, phyllite and schist) is fundamentally based on the increasing size of phyllosilicate grains, hence different degrees of recrystallization (metamorphic grade). The distinction between foliation types is quite arbitrary, with weakly defined boundaries based on personal perception.

Diagenetic Foliation

A foliation may form parallel to bedding during diagenesis and compaction of sediments with clay minerals and/or detrital micas. Such a foliation is observed in very-low to low-grade pelitic rocks that have undergone no or little deformation, as the absence of regional folding demonstrates. Micas lying on the foliation plane commonly have frayed edges. The diagenetic foliation, a primary mechanical anisotropy, may play an important role in the development of secondary foliations if any.

Slaty Cleavage

The word **slate** originated as a quarryman's term for fine-grained rocks that were sufficiently fissile to split into thin, planar slabs suitable for roofing tiles and blackboards. **Slaty cleavage** describes the fabric responsible for the planar, quite a flat parting of rocks whose individual grains are not obvious without a microscope. Slaty cleavage is often remarkably constant in orientation, has a matt aspect and is typical of fine-grained shales deformed under low metamorphic grade. The parallel alignment of phyllosilicate grains (clay minerals, illite, chlorite, micas) too small to be visible to the naked eye produces such a fabric.

Under the microscope, slates have a **domainal structure**, which refers to rock regions with different compositions and fabrics. The cleavage consists of spaced bands of strongly aligned, recrystallized platy minerals **anastomosing** around inter-cleavage domains (microlithons). The microlithons are rich in the major mineral constituents (usually quartz) of the rock other than layer silicates. Grains in these microlithons generally show little or no preferred orientation with phyllosilicate grains subparallel to bedding and showing kinks and undulatory extinction. The microlithons are generally lenticular with their long dimensions parallel to the cleavage. They vary in size and the number of grains. The spacing of slaty-cleavage planes ranges from less than a millimeter to a few millimeters. Seams and accumulation of insoluble residues (often oxides) commonly accentuate the cleavage domains rich in minute, flaky or tabular minerals (mica, chlorite, and talc) whose parallel, planar arrangement defines the cleavage.

Phyllitic fabric

Phyllites are rocks with a shiny, silky to satiny sheen due to flaky grains larger than in slate, yet still too small to be identified with the unaided eye. Consequently, phyllites cleave into thicker slabs than slates. Their mica-dominated planar fabric is flat or crinkled.

Schistosity

Schistosity refers to foliation planes with glittering micas (mainly muscovite and biotite) big enough to be discerned with the unaided eye. Schistosity is common in metamorphic rocks because metamorphic recrystallization tends to enlarge the grain size and produce new, platy and elongated minerals. Therefore, planar fabrics become coarser than slaty cleavage.

Owing to coarser grains, schistosity tends to be wavy and discontinuous. It also occurs in low-grade rocks, particularly in retrograde greenschist facies rocks. **Schists** are metamorphic rocks with schistosity.

Composite foliation

A **composite foliation** refers to one foliation plane and/or gneissic layering resulting from at least two foliation-forming events with the two foliations being mostly parallel. These two foliations can be identified and separated in isoclinal hinge zones that fold the earlier foliation plane. This is frequent in high-grade rocks.

Conjugate crenulations

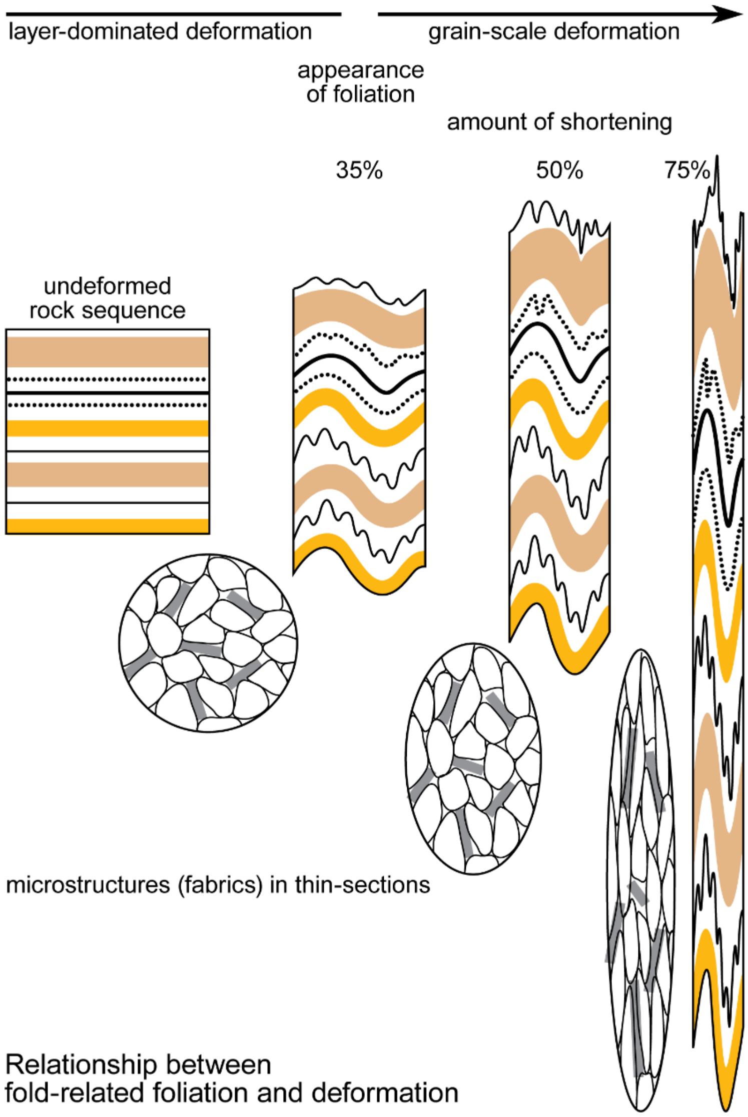
Two simultaneous, identical crenulation cleavages may be mutually oblique at an angle of 60-90°. The cleavage surfaces are generally less regularly spaced and more localized than non-conjugate crenulation cleavage. Usually, one direction is more developed than the other, but either set may be prominent from place to place, and there is no consistency about which accommodated the last deformation. These **conjugate crenulation cleavages** are commonly associated with conjugate folds or kinks.

Strain significance of foliation

Strain analysis data indicate that foliations form over a wide range of strain intensity and deformation regimes.

Orientation with respect to the strain ellipsoid

Finite strain analyses on reduction spots, ooids, flattened fossils, and pillows indicate that the foliation is perpendicular to the short axis λ_3 of the local finite strain ellipsoid (with $\lambda_1 \geq \lambda_2 \geq \lambda_3$). Therefore, the field orientation of foliation planes in rocks is not the actual strain orientation but the geographical orientation of the plane of maximum finite flattening (λ_1, λ_2).



Relationship between fold-related foliation and deformation

Numerical and analog considerations support evidence from measurements of finite strain in natural structures. In complex transpressive or transtensive flows, foliation planes lay somewhere between the (λ_1, λ_2) planes of the instantaneous and finite strain ellipsoids. The orientation depends on the relative rates of development and recovery of the fabric so that cleavage may show only a part of the total finite strain history.

Foliation and strain magnitude

The strain involved in the development of foliations is difficult to measure because foliations are associated with a wide range of strains. Measurements of finite strain in natural structures indicate that slaty cleavage appears when a fine-grain pelite has been compressed (compacted) by about 10%. Foliation becomes conspicuous after 20-30% flattening and intense foliations involve >35% flattening.

Foliation and volume loss

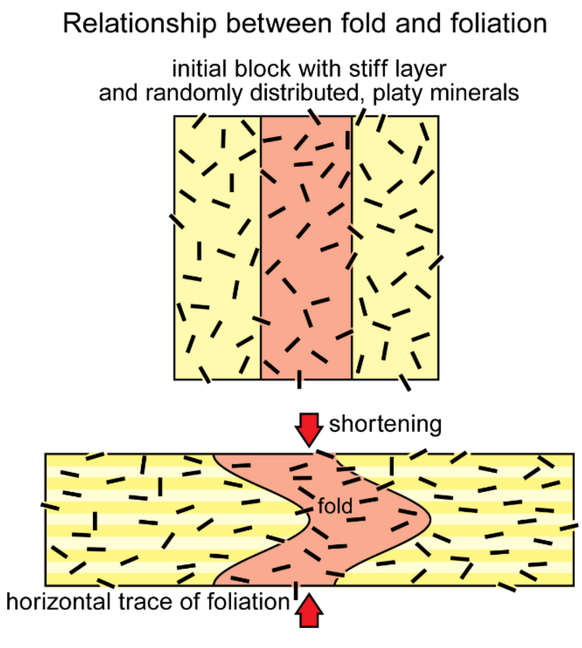
Geometrical measurements and local, chemical balance indicate that the development of slaty cleavage approximately produces >10% (up to 50%) volume loss in slates.

Foliation related to folds

Experiments and detailed case studies of natural folds show that after ca 35%, bulk shortening accommodated at the layer-scale (faulting and buckling) changes to grain-scale, which triggers the appearance of foliation.

Axial plane foliation: Definition

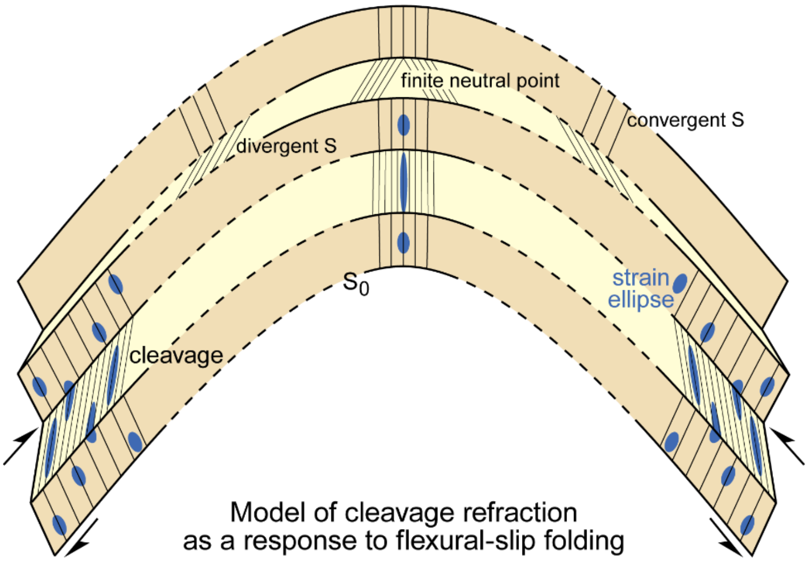
Foliation in folded rocks is subparallel (ideally, parallel) to the axial plane in the fold-hinges produced by the same deformation event. It is accordingly referred to as the **axial plane foliation**, although foliation planes commonly change orientation according to lithology and grossly divergent from the axial surface in the fold limbs. Numerical and analog models of strain patterns within buckle fold profiles provide additional support for parallelism of the XY -plane of finite strain ellipsoid with the axial plane foliation observed in natural folds.



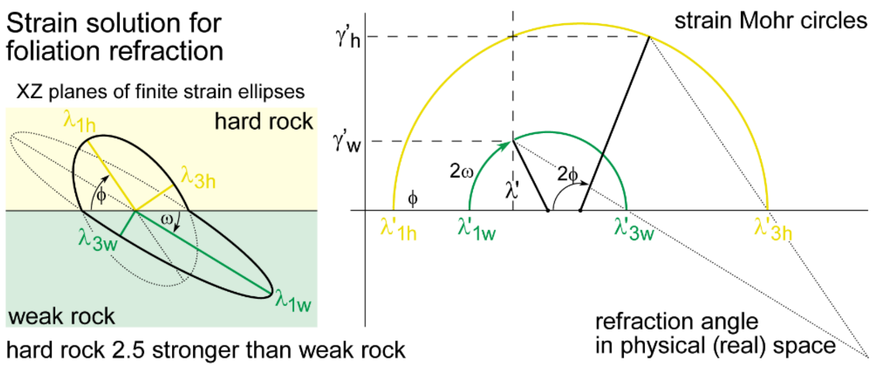
Refraction

Foliation planes typically change orientation at boundaries of layers with different grain-size or composition, i.e. competence. This change of angular relationship between foliation and bedding across lithological boundaries (and occasionally within layers as in graded beds) is termed **refraction**, as for the light that departs from its original direction of travel at the interface between two materials of different refraction index. In layered rocks, refraction occurs because the viscosity contrast along alternating layers produces local shear components during folding (remember flexural slip). These local shear components add local strains so that the flattening plane in the rock is not parallel to that of the bulk strain.

Models in multilayered materials reproduce foliation fans and refraction fitting changes in orientation of the XY plane of finite strain ellipsoid in different parts of a fold and layers of different competence. This similarity suggests again that foliation forms in the XY plane of finite strain. As such, foliation refraction is evidence for a heterogeneous strain with changes in orientation of principal axes and strain intensity across boundaries between layers with different competences.



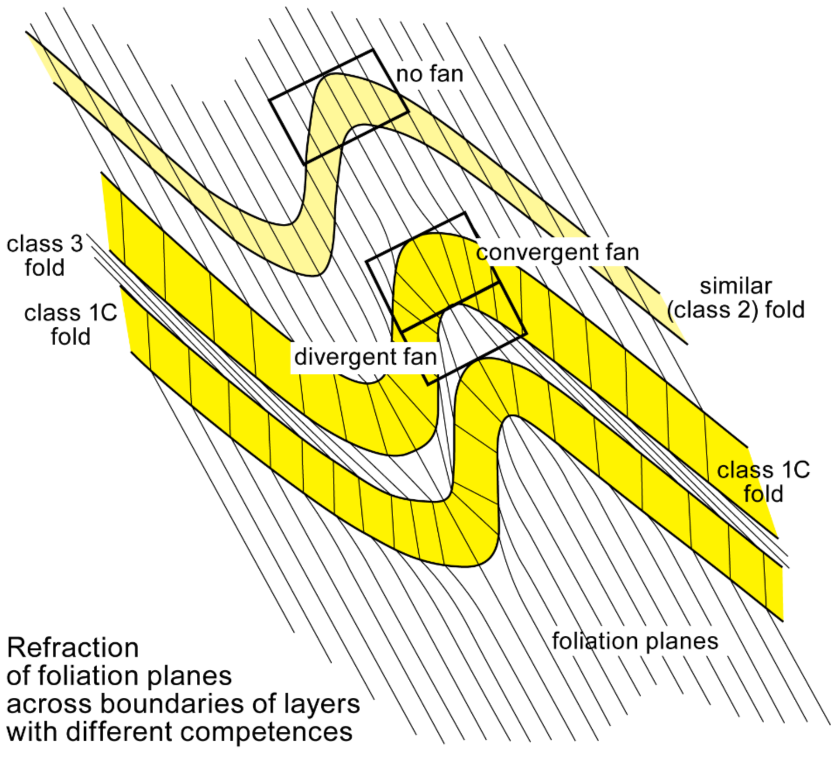
Straight foliation traces in the two-dimensional XZ plane of finite strain denote homogeneous strain. Consider angular refraction across a coherent bedding plane parallel to the Y-axis. Two adjacent homogeneously deformed rocks with different viscosity yield different strain Mohr circles. These circles personify two strain states that must be compatible along the layer boundary, provided there is constant volume (area) deformation and there is no discontinuous slip on the layer boundary. Continuity across the boundary implies that the elliptical sections of the strain ellipsoids on either side are equal on that boundary. In two dimensions, the strain ellipses have different shapes in the two layers. Normalized to be equal area, the longitudinal strains are equal along the interface, i.e. the two strain Mohr circles of the hard and weak rocks must intersect exactly on the layer boundary. However, different angles between this interface direction and the longest axes X indicate differences in shear strain, which the strain Mohr construction can illustrate if one knows strain magnitudes in one layer.



Take the weak layer. The strain circle is plotted for that rock, with the strain point obtained from the intersection with the line at twice the angle between foliation trace and bedding. Since the strain ellipses of the two rocks are coincident on the bedding plane, the stretch is the same for both rocks on that plane. Equal stretch is read along a vertical line since quadratic elongations (actually reciprocals) are plotted along abscissa: in that case, the line passes through the defined strain point. The angle between foliation and bedding can be used to find out the stretch point of the other rock. This point should be the intersection between the line at a double angle to the abscissa and the vertical line of equal stretch. The oriented line can slide along the abscissa, thus requiring more information to be fixed. This can be either the finite strain (then providing the abscissa for the strain Mohr circle at $\left(\lambda'_1+\lambda'_3\right)/2$) or some knowledge of viscosity contrast, which is the ratio of the two viscosities. Multiplying the ordinate of the first point by this ratio provides the ordinate of the sought intersection point, and this ordinate is half the reciprocal shear strain in the second rock.
This graphical exercise shows that foliation refraction angles are a proxy for the relative strengths of adjacent lithologies. The bedding/foliation angle is larger in the competent layers than in the incompetent ones. In other words, foliation takes a shorter way across competent layers such as sandstone than through less-competent layers as shale.

Foliation fans

Owing to refraction, axial plane foliations typically **fan,** i.e. display a radiating pattern within the fold. The fan is **convergent** or **divergent**, depending on whether the foliation converges towards the core or the convex side of a fold, respectively. Both convergent and divergent fans may coexist in a folded multi-layer. Competent layers tend to develop parallel folds in which strain axes are at high angles to the layer boundaries, thereby generating convergent fans. Incompetent layers, tend to develop congruent folds with large amounts of shear imposed by adjacent competent layers and consequently fostering divergent fans. In numerical models, convergent fans are similar to the pattern of the XY plane produced during folding by tangential longitudinal strain, while divergent fans are comparable to the pattern due to flexural slip.

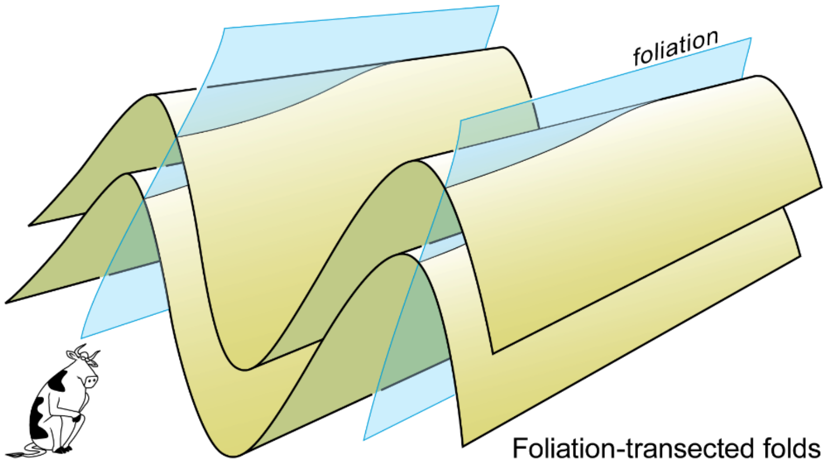


Refraction of foliation planes across boundaries of layers with different competences

Due to fanning foliations, convergent and divergent foliation orientations form triangular regions with no cleavage in hinge regions of the less competent layers of folds: **the finite neutral points**.

Transection

In a few cases, the foliation seems axial planar in profile, but it is oblique to the hinge line and goes across from limb to limb in the third dimension. **Transection** is the oblique intersection between foliation and the coeval fold axes.



Foliation-transected folds

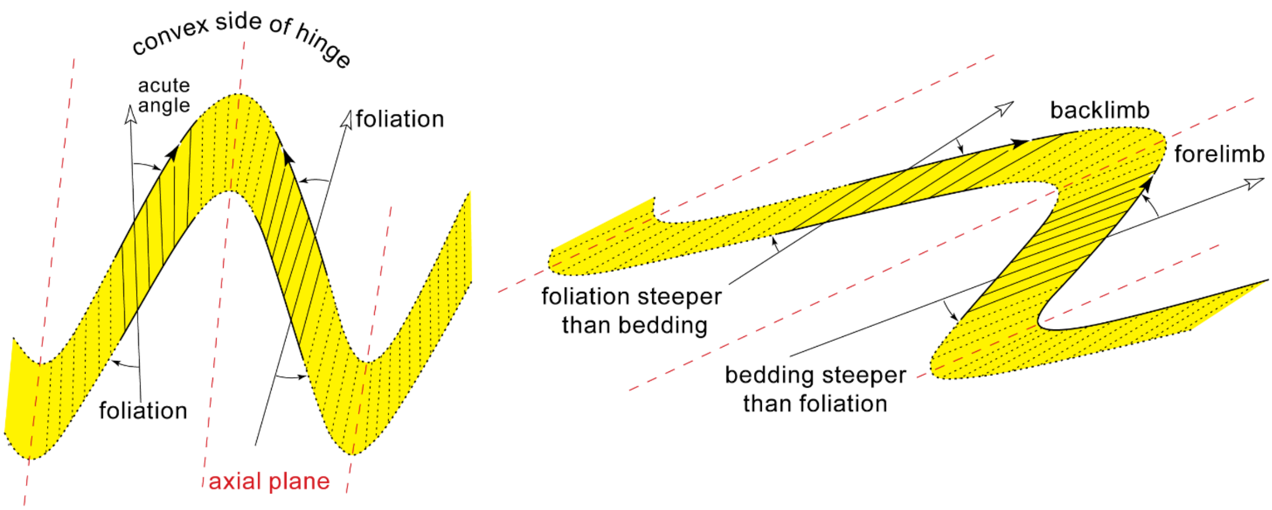
A transecting foliation is no later foliation superimposed on an earlier generation of folds. Transection is generally attributed to a rotating strain field during folding, with eventually prefolding initiation of foliation and fold axes developing oblique to the bulk flattening direction. The subsequent **hinge-migration** while layers roll around the fold axis, which is oblique to the vorticity vector of the same deformation event, accounts for the abnormal relationship between foliation and axial planes.

Use of axial plane foliation in geometrical analysis

The association between folds and axial plane foliation patterns is so systematic that the **foliation-bedding angular relation** is routinely used to determine which limb of an incompletely exposed fold a particular outcrop belongs to, even if hinges cannot be seen.

A sketched fold with associated foliation shows how. Four relations allow locating folds larger than the outcrop scale:

- The foliation on the limbs is inclined relative to bedding towards the convex side of the hinge. In one limb the acute angle between bedding and foliation opens to the left on the upper side of the bed; on the other limb, the acute angle opens to the right.
- Bedding and foliation lie at a nearly right angle to each other at the hinge of the fold. The foliation attitude can be used to infer directly the orientation of axial planes.



Angle relationship between bedding and foliation in fold profiles

- The geometry of bedding and foliation is very powerful in determining whether the beds are overturned or right side up in an outcrop of recumbently folded rocks. In the normal limb, the foliation is steeper than the bedding dipping in the same direction; in the overturned limb the bedding dips steeper than the foliation.
- The smaller the average cleavage-bedding angle, the tighter the fold.

Note that these rules are consistent with the distribution of S, M or Z minor folds around a major fold, the foliation plane remaining axial plane to both scales of folds.

Non-axial plane secondary planar elements

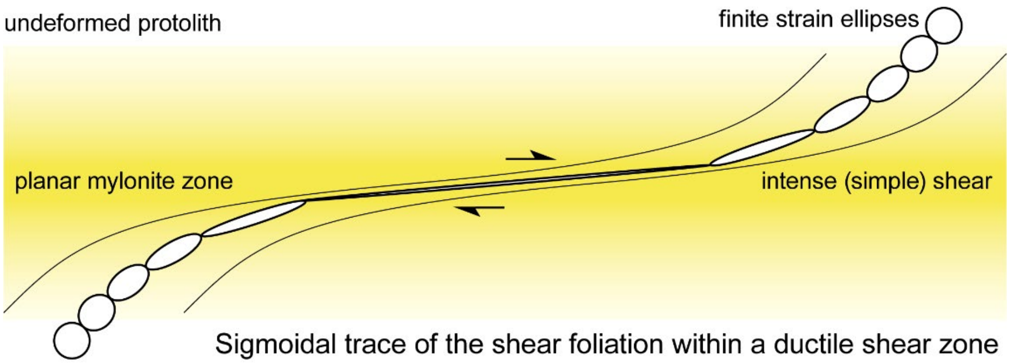
Not all foliations are associated with folding.

Bedding foliation

In sediments, a single set of foliation planes parallel to the bedding may exist, although there is no ostensible phase of folding. This **bedding foliation** is attributed to vertical compaction of the sediments under the static load of overlying strata. The resulting diagenetic foliation results essentially from oriented crystallization of diagenetic minerals.

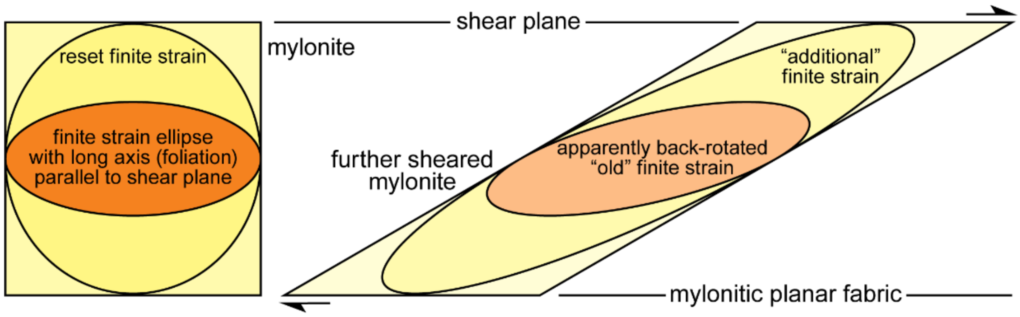
Shear foliation

When a ductile shear zone develops in granular rocks like granites, the foliation that initially develops from the undeformed rock is not associated with a synchronous system of folds. The foliation, which tracks the plane of maximum finite shortening, typically shows a progressive rotation along with an increase in intensity (depicted by foliation planes that come progressively closer) from the undeformed rock towards a strongly foliated, planar zone (the mylonite). Typically, also, this rotation is symmetrical on both sides of the planar zone, conferring to the shear foliation a sigmoidal shape, which is a most reliable structure to readily define the relative movements involved.



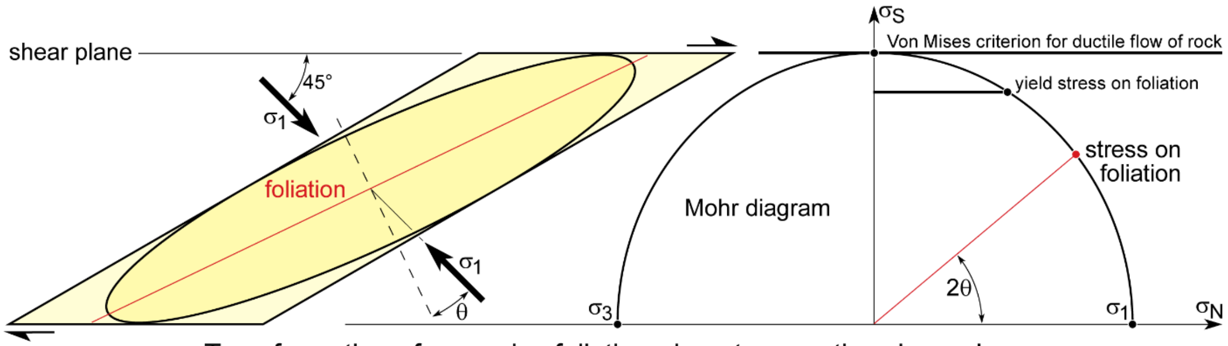
Sigmoidal trace of the shear foliation within a ductile shear zone

Note that owing to simple shear geometry, the plane of maximum finite shortening becomes rapidly (after a shear strain of about 10) nearly parallel to the shear plane. At that stage, further simple shear distorts the finite strain ellipsoid so that its plane of maximum flattening does not stay parallel to the shear plane. However, the planar fabric, i.e. shear foliation visually remains parallel to the shear plane. The rule equating the $\lambda_1\lambda_2$ plane of the finite strain ellipsoid with foliation does not strictly apply.



Instability of the finite strain ellipsoid in mylonites
once the plane of maximum flattening (foliation) is parallel to the shear plane

Departure from the $\lambda_1\lambda_2$ plane of finite strain ellipsoid is even more consequent when the foliation is transformed from a passive strain marker into an active slip surface. Indeed, foliation as a plane of mechanical anisotropy is a plane of weakness along which the shear stress for failure is smaller than that of the rock. Therefore, shear foliation planes will tend to lose cohesion and slip once they have been rotated to a near shear plane attitude. At that stage, they become microfaults that are no longer related to the plane of maximum flattening.



Transformation of a passive foliation plane to an active shear plane

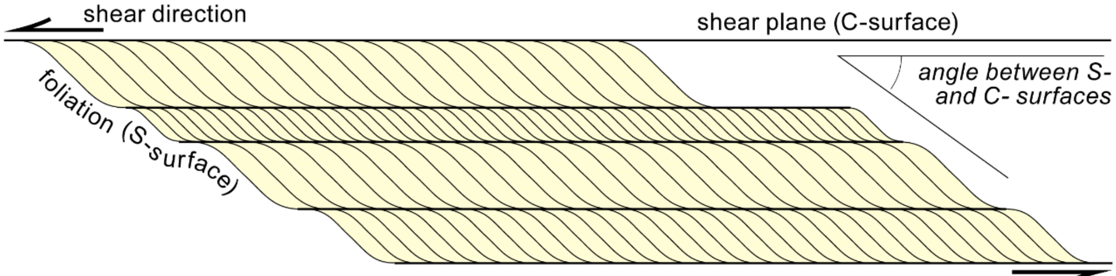
Shear foliations and their deterioration into shear planes are common in high-grade gneisses and deformed igneous rocks that have suffered intense shear deformation.

Shear bands - Extensional crenulation cleavages

Extensional crenulation cleavage describes planar fabric elements that look like an asymmetric and usually mesoscopic crenulation cleavage in strongly sheared rocks. The "crenulation cleavage" is not parallel to axial planes of coeval folds. The "crenulation" aspect is due to the "older" foliation being bent in a systematic way into discrete planes or zones across which markers are displaced, which implies some movement (strain slip) between adjacent microlithons. Therefore, "extensional crenulation cleavage" refers to regularly distributed and parallel **micro-shear-zones**. They form parallel to a plane of high resolved shear stress and not parallel to the $\lambda_1\lambda_2$ plane of the finite strain ellipsoid. They usually are at a small angle (<30°) to the host mylonitic fabric and their movement is that of normal faults with respect to the trace of this mylonitic fabric (hence the term extensional). The acute intersection angle and the sigmoidal shape of the mylonitic fabric between these shear zones (or bands) show the direction of shear.

S-C fabrics

Large displacements along shear zones curve shear foliation S at a shallow angle to the shear plane. When deformation is sufficiently strong, penetrative S-surfaces display a sinusoidal shape that rotates into parallelism with discrete and regularly spaced zones of concentrated shear parallel to the gross shear plane: the C-surfaces.

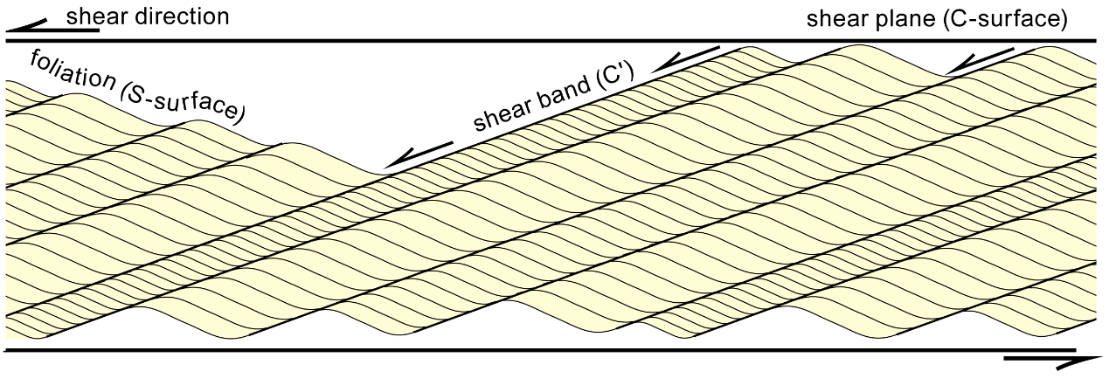


S-C fabric in the XZ-plane of finite strain ellipsoid

Each C shear zone is relatively planar and develops its curved foliation pattern on a small scale so that the sense of deflection of S into C is the same as the general sense of shear.
The rocks where ductile shear zones are sufficiently abundant to constitute a fabric have **S-C structures** (or **fabrics**). S stands for the foliation (schistosity) planes, defined by the skewed grain shapes of the rock. S is typically deflected into the smaller grain size C (from the French *cisaillement* = shear) shear zones or planes. The foliation S is leaning over in the direction of shear and has an acute bulk inclination to the C-surfaces. Parting of the rock is easier along the C planes.
Note that owing to additional and localized displacements on C surfaces, the foliation planes S do not record the total shear-strain of the rock. Note also that C-surfaces are parallel to surfaces of no finite shear strain.
S-C fabrics are most common in granular rocks, in particular porphyroclastic granitoids.

C' (and C'')-surfaces

In strongly foliated rocks, one or more sets of secondary, spaced planar elements may appear systematically oblique to both the early foliation and the shear plane and shear zone boundary. They are essentially small-scale shear zones oblique to a pre-existing foliation, such that the displacement on this new spaced "foliation" results in net extension parallel to the earlier planar anisotropy.



S-C' fabric in the XZ-plane of finite strain ellipsoid

These micro shear zones are usually less continuous than C planes and delineate an extensional crenulation cleavage termed **C' planes**. S-C'-structures may indicate intense, non-coaxial and partitioned flow. Conjugate sets may arise to accommodate extreme flattening.

C' shear bands seem to develop at a late stage of shear activity after the strong planar anisotropy represented by the S-foliation has already been established. Like C-surfaces, they are related to the amplification of finite strain of the foliated host rock.

C'' designate an additional set of spaced micro shear zones that may occur oblique to C' planes, in the same way as C' are oblique to C.

Formation of foliations

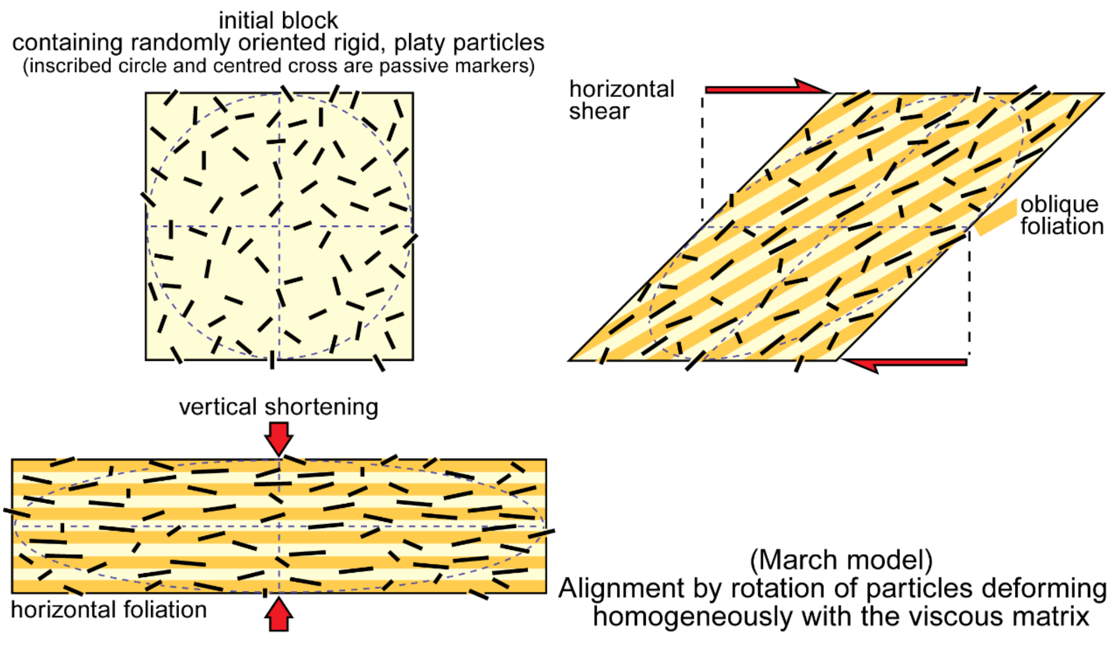
Four processes operating either separately or in conjunction may achieve preferred dimensional orientation of grains or produce rock foliations. They are:

1. Shape-controlled mechanical rotation of pre-existing, non-equant grains or fabrics.
2. Modification of grain shape and volume through pressure-solution.
3 Modification of grain shape by within-crystal slip or diffusion.
4. Growth of non-equant grains in a preferred dimensional orientation.

Shape-controlled grain rotation

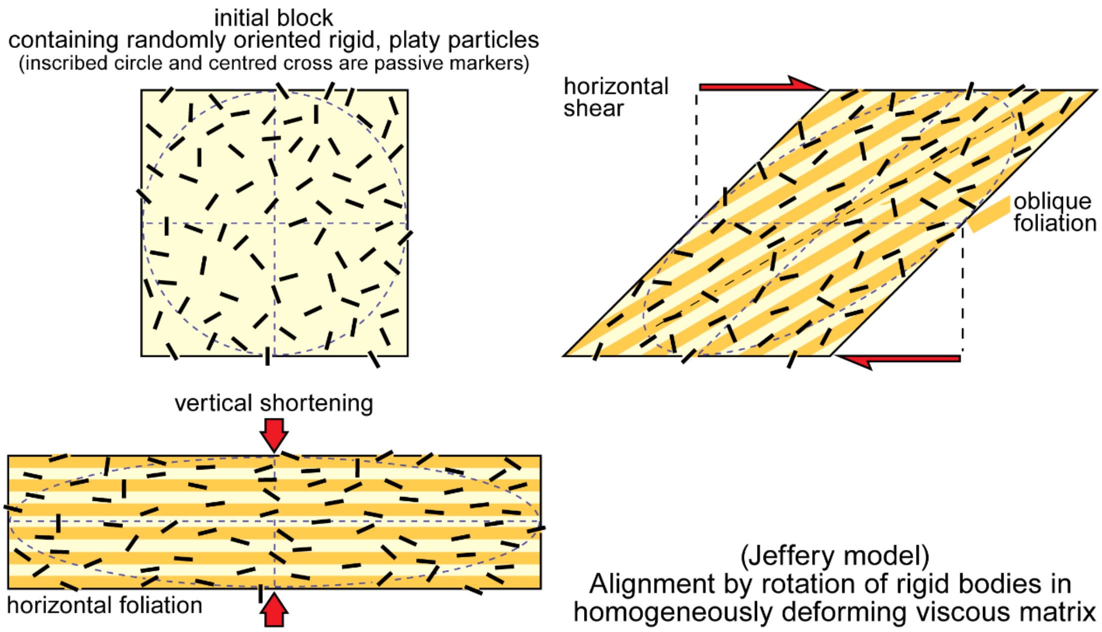
There are three principal mathematical models applicable to the rotation of pre-existing platy or needle-like minerals (i.e. grains with high aspect-ratio) with an initial random angular distribution:

The "**March analysis**", is concerned with the rotation of passive marker planes during the shortening of a homogeneous body.

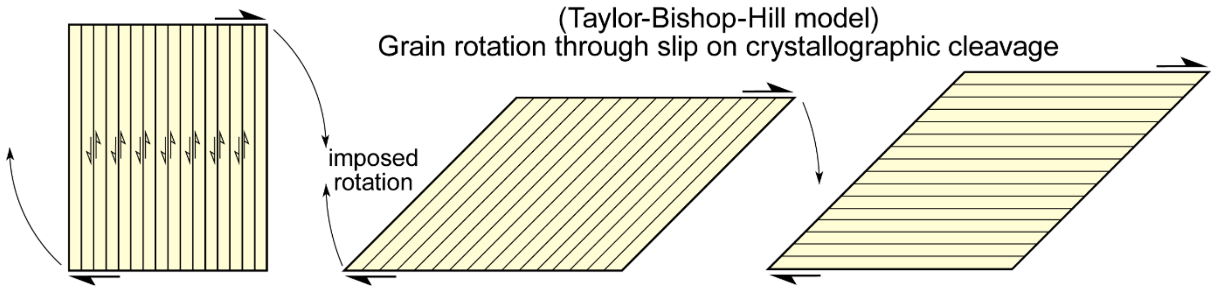


(March model)
Alignment by rotation of particles deforming homogeneously with the viscous matrix

The **"Jeffery model"** is concerned with the rotation of ellipsoidal, rigid bodies in a viscous fluid. Rigid rotation of initially randomly distributed, clastic, tabular or elongate grains towards the plane of flattening defines statistically a foliation parallel to that plane. This process is particularly applicable to foliations formed during diagenetic compaction and/or tectonic dewatering of incompletely lithified mudstones. It is also applicable to foliations formed during the emplacement of an unconsolidated magma. Slip along grain boundaries usually accompanies rotation.



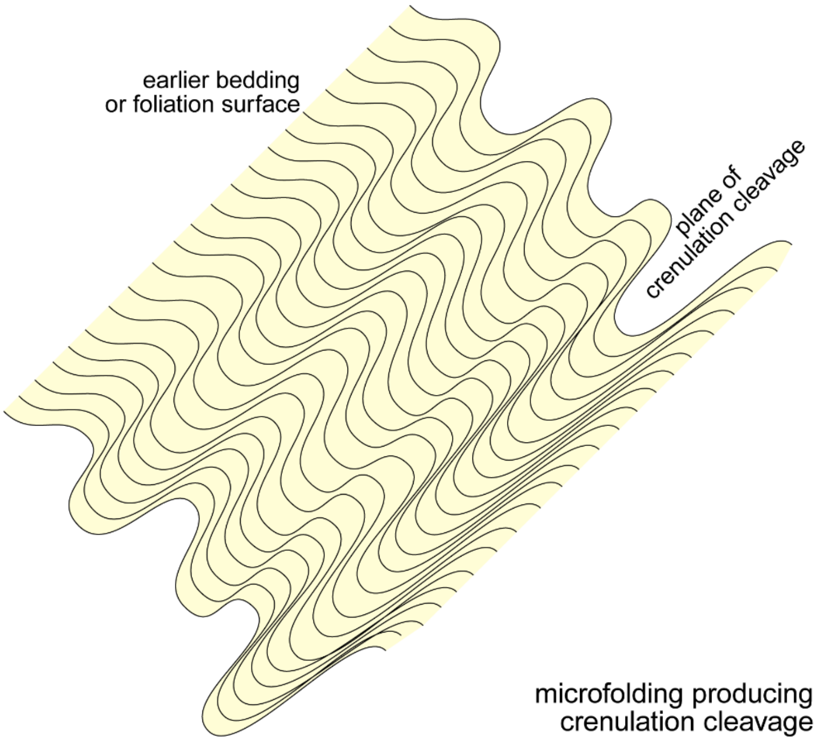
The progressively developed preferred orientations predicted by both, Jeffery and March models are essentially the same, Jeffery's analysis differing from that of March by a factor that describes the shape of the rotating particles. Furthermore, the Jeffery rigid particles define in simple shear regime a stable statistical orientation that cannot be achieved in the March model.
Experiments on aggregates containing layer silicates or other platy crystals have shown that two kinds of fabric may develop by rotation. The first is a homogeneous fabric characterized by the general preferred orientation of the platy grains with their planar dimensions parallel to the $\lambda_1\lambda_2$ plane. The degree of preferred orientation intensifies with strain, but the rotation patterns of individual rigid particles are complicated due to particle-to-particle interactions. The second is a heterogeneous or domainal fabric in which reorientation is localized along narrow shear zones.
The third model (**"Taylor-Bishop-Hill model"**) implies the rotation of a single mineral grain due to internal shear on a unique set of crystallographic slip planes so that the final grain shape is compatible with the deformation imposed by the surrounding matrix.



The principal limitation of these models, however, is that they do not explain the domainal fabric observed in many slates. Nevertheless, they provide one possible model for the development of preferred orientation in low-grade rocks.

Buckling - Microfolding

The formation of crenulation cleavage involves periodic, small-scale and intense buckling and/or kinking of an earlier planar fabric. Buckling and/or kinking give rise to microfolds.



As the microfolds become more closely compressed, the limbs become progressively thinned out and parallel while the fold hinges become relatively thicker. The new crenulation cleavage is parallel to the aligned limbs of stacked microfolds. Micas within the limbs of crenulations remain approximately parallel to the earlier fabric. They are still parallel to the earlier foliation but have been rotated toward parallelism with the new foliation.

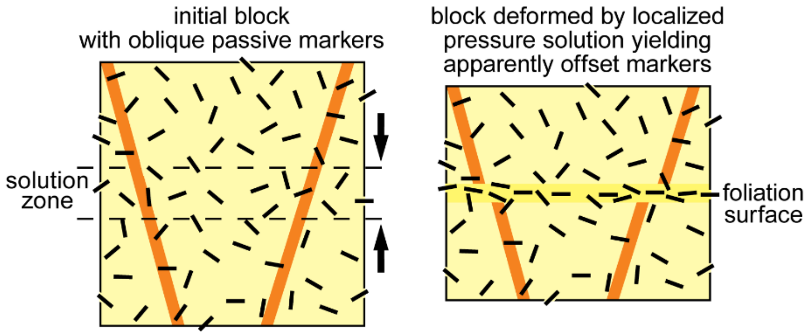
In this manner, the development of crenulation cleavage likely involves the mechanical rotation of existing grains accompanied by chemical processes such as modification of grain shapes and sizes by diffusive processes and growth of new grains with an orientation and shape compatible with the local strain history.

Solution transfer

Several foliation types involve compositional layering. This layering (called **banding** in two-dimension observations) is attributed to some metamorphic differentiation (or **segregation**) during the foliation development. The solution, mass transfer and re-deposition of material (**pressure-solution**) cause segregation, which is part of a diffusion process through interstitial, aqueous fluids flowing along grain boundaries. The process can result in significant volume loss if the dissolved material is transported out of the system.

Pressure solution

Dissolution (removal) occurs on grain-to-grain or layer boundaries in porous rocks under non-hydrostatic stress at a rate controlled by the magnitude of normal stress across the boundary. Boundaries perpendicular to the direction of the greatest compression dissolve into the aqueous pore fluid most rapidly. The dissolved material reprecipitates, often as fibrous minerals on low-stress intergranular boundaries and opening veins. Truncated fossils and veins, the missing parts of which have been dissolved and not sheared along discrete surfaces, are classical examples showing that foliation planes may occupy a zone of lost volume. **Stylolitic foliations** are familiar examples in limestone.



Reorientation of platy minerals due to removal of their matrix by solution

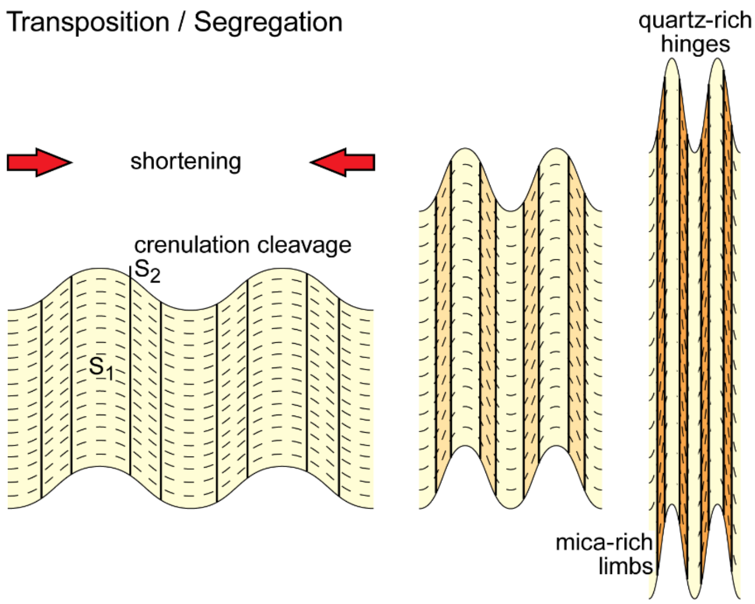
The juxtaposition of quartz-rich domains and cleavage domains with insoluble and densely-packed minerals suggests that pressure solution is involved in the formation of most foliation types.

Differentiation

Migration of dissolved material (**solution transfer**) away from high-solubility sites occurs down a stress-induced chemical potential gradient (i.e. created by variations in the magnitude of normal stress at grain boundaries) to nearby sites of low solubility. Material precipitation may take place at sites of lower normal stress, commonly in extension veins and **pressure fringes** with fiber-growths at extremities of flattened grains, while the insoluble residue (secondary minerals and oxides) remains within the foliation planes. Rocks have become **differentiated**, which means that there is a preferential redistribution of minerals in the rock.

From differentiated crenulation cleavage to transposition

In fine-grained metamorphic rocks, a crenulation cleavage develops along the limbs of microfolds deforming an earlier planar fabric. Quartz and feldspar may dissolve under pressure solution in the highly compressed limbs and be reprecipitated at the hinges where pressure is lower. As the process continues, the new foliation aligns itself perpendicular to maximum shortening and bands of micas or sheet silicates (limb sites) alternating with bands of quartz or feldspar (hinge sites) define a differentiation layering parallel to the new foliation.
Fold limbs and hinges may completely disappear when strain-induced solution transfer is extreme. The resulting bands constitute a **transposition** of the old structures.



Metamorphic segregation in gneiss is important either by enhancing a deformational layering or by forming a new layering by pressure solution for example in the case of melt production.

Directed metamorphic (re)-crystallization

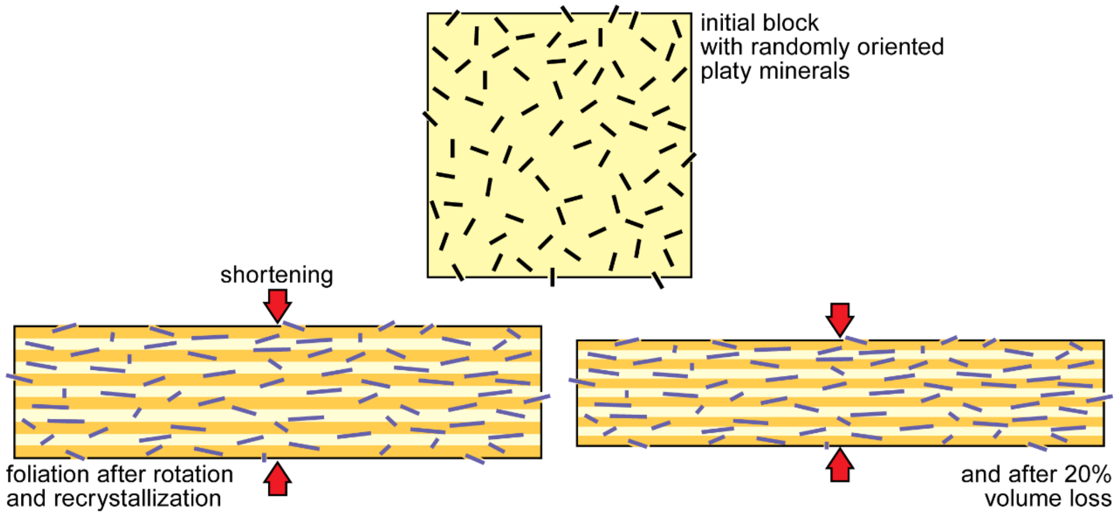
Foliations are often marked by the preferred orientation of minerals that grew directly parallel to the foliation plane in response to new metamorphic conditions. Some or all of the grains may be syntectonic and result from chemical reactions. Along with metamorphic crystallization, strained crystals tend to recrystallize into unstrained, larger crystals via dynamic recrystallization processes (governed by the release of internal strain and surface energies). Synmetamorphic foliations result from two mechanisms:

1) preferred dimensional growth and
2) within-crystal slip or diffusion.

Metamorphic grains that grow before or during the deformation may be rotated by strain in the same way as detrital grains.

Oriented growth

Metamorphic reactions consume some minerals and produce others that modify the rock fabric. Factors such as stress and strain influence the growth orientation of anisotropic minerals. For example, fibers grow as elongate grains parallel to the stretching direction on the foliation plane, a mechanism that also produces a crystallographic preferred orientation. Micas grow with their (001) sheets perpendicular to the direction of local shortening, hence producing a foliation. Oriented growth enhances the flattened appearance of syn-metamorphic foliations.



Foliation enhancement due to metamorphic recrystallization

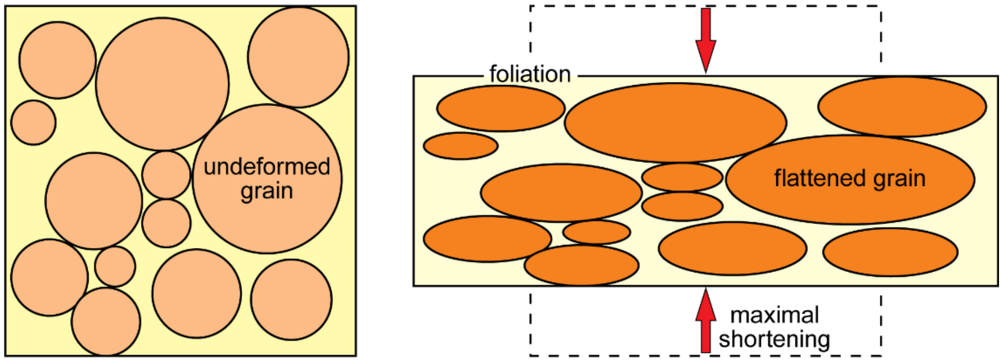
In some metamorphic rocks, the orientation of new crystals may be governed by, and is parallel to, the orientation of pre-existing grains or aggregates. Their shape fabric **mimics** the older fabric and sometimes inherits the shape of previous crystals, so defining a **mimetic foliation**. Growth is **mimetic**.

Crystal plastic deformation

Foliation can simply form when grains in the rocks become flattened. Two mechanisms may control this change in shape: dislocation creep and, usually at a higher temperature, solid-state diffusion. These processes are important in producing a crystallographic preferred orientation.

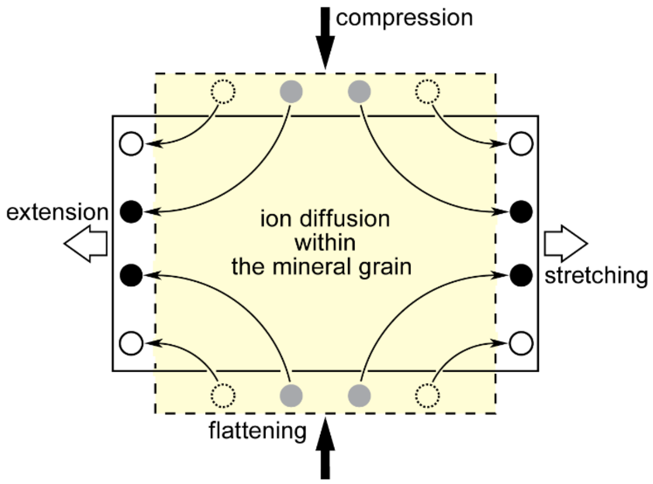
Dislocation creep

The change in shape results from lattice distortion through the movement of **dislocations** or **twinning** or **kinking**. In **coherent** (also **dynamic**) **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 (**clasts**) 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. Dislocation creep is most active in mylonites. In **reconstructive recrystallization,** the old crystal breaks down, and a new crystal forms (a **blast**) that generally has a different composition.



Solid-state diffusion

A purely diffusion-controlled mass transfer can flatten grains: The change in shape results from ionic diffusion within grains (**Nabarro-Herring creep**) or along grain boundaries in the absence of a fluid phase (**Coble creep**). Highly strained (facing compression) grain boundaries are favorable for degradation by transfer of their components towards less pressured (facing extension) boundaries, resulting in a net increase in the proportion of grains elongated towards the X-axis of finite strain.



Change in shape of a grain due to diffusion-controlled mass transfer

Transposed foliations

Transposition is a mechanical transformation of layers from an initial orientation into another orientation. It gives a striped appearance to the rock.
An essential part of the **mechanical transposition** process is the rotation of a pre-existing plane by tight folding into an orientation approximately parallel to the axial plane of the resulting isoclinal folds. Extreme flattening, development of discontinuities parallel to axial surfaces, development of axial plane foliation, elimination of fold closures and segmentation of marker beds or layers achieve transposition.
Severe shear deformation in gneisses produces layering partly by flattening and elongation of large crystals, and partly by rotation of veins and other heterogeneities into the plane of flattening, towards the plane of shearing. As a result, foliation-parallel layers may become alternately enriched in granular and in micaceous minerals. A similar process involves boudinage of competent layers, with

veins in various structural positions such as tensions gashes and necks. Their rotation and translation in the incompetent matrix can also produce a transposed sequence.
In some instances, recrystallization is accompanied by metamorphic differentiation along foliation planes: thus, foliation domains may become enriched alternately in segregated material, such as quartz, and micaceous minerals. This is a **chemical transposition**.

Summary

Foliation is a planar fabric element that originates from sedimentary and magmatic processes (primary fabric) and ductile deformation (secondary fabric). The latter provides clues to the geometry of large-scale structures, kinematics, strain, and conditions of deformation. Foliations are systematically associated with tectonic deformation and are common in all grades of metamorphic rocks. Ductile flattening and the parallel alignment of platy minerals are believed to be the principal cause of axial plane foliations.
The type of foliation depends on the composition of the deformed rock and varies in morphology between classified end members that refer to deformation processes. Mechanical rotation, solution/precipitation, crystallization, and recrystallization are involved in the development of diverse foliations. Differentiated types are increasingly obvious as grade increases but are also ordinary in the lower grade rocks.
All of these mechanisms tend to produce a preferred dimensional orientation of non-equant grains and/or aggregates of grains that define a planar structure parallel to the $\lambda_1\lambda_2$ (i.e. XY) plane of the strain ellipsoid.

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