DUCTILE FAULTS: SHEAR ZONES

Under appropriate temperature, pressure and/or fluid conditions rocks flow by ductile creep, accommodated at the grain scale by motion of dislocations and/or diffusion processes. Relative displacement between adjacent rock domains is however commonly concentrated in planar zones consisting of intensely sheared rock bordered on both sides by strain gradients. Accordingly, ductile shear zones are frequent in metamorphic rocks. They range in width from infinitesimal to several kilometres. Shear strain intensity is nil or low in the wall rock, progressively increases across the gradients and is strongest at the contiguity plane between both gradients.
Ductile faulting is a process resulting in offset across a localized velocity gradient in distributed flow. This is a simplified statement but, like brittle fault zones, ductile shear zones usually contain a number of small-scale structures that indicate the sense of shear. Often also, they transport fluid, dilate and may host mineralization.

Definition

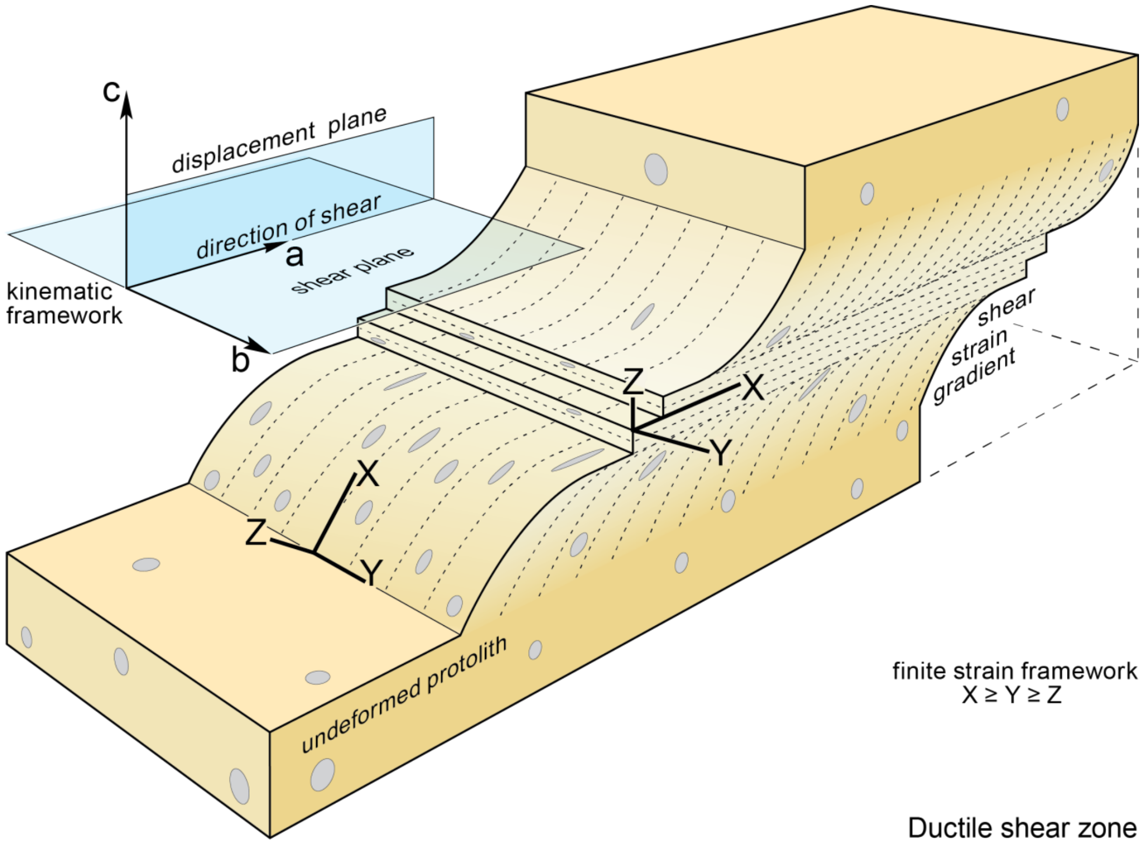
Ductile shear zones are long and narrow zones of relative displacement. They are analogous to faults but without fracture planes (unless they are reworked) because dominantly ductile deformation has caused the concentration of large strain into the shear zones. The formation of a ductile shear zone is commonly associated with a drastic reduction of grain size and the development of a well-banded and lineated rock called **mylonite**. Ductile shear zones generally record a non-coaxial deformation and may range from the grain scale to the scale of a few hundreds of kilometres in length and a few kilometres in width. The strain gradients from mylonite to undeformed rock are criteria to distinguish large-scale shear zones from regional deformation. Localization of deformation into such narrow zones reflects continuous but heterogeneous strain in rock.

Morphology

Ideally, a ductile shear zone is contained between two parallel and imaginary boundaries, the **shear zone walls** outside of which the rock is unstrained. Ideal shear zones are produced by plane strain, simple shear deformation. Accordingly, there is no stretch along the intermediate, Y axis of finite strain, perpendicular to the plane of strain. The structural study of shear zones is thus carried out in the XZ plane of finite strain (i.e. orthogonal to the foliation plane and parallel to the stretching lineation), which is also the ac kinematic plane since the lineation is parallel to the displacement direction c. Extending the fold terminology, this plane can be called the **profile** of a shear zone.

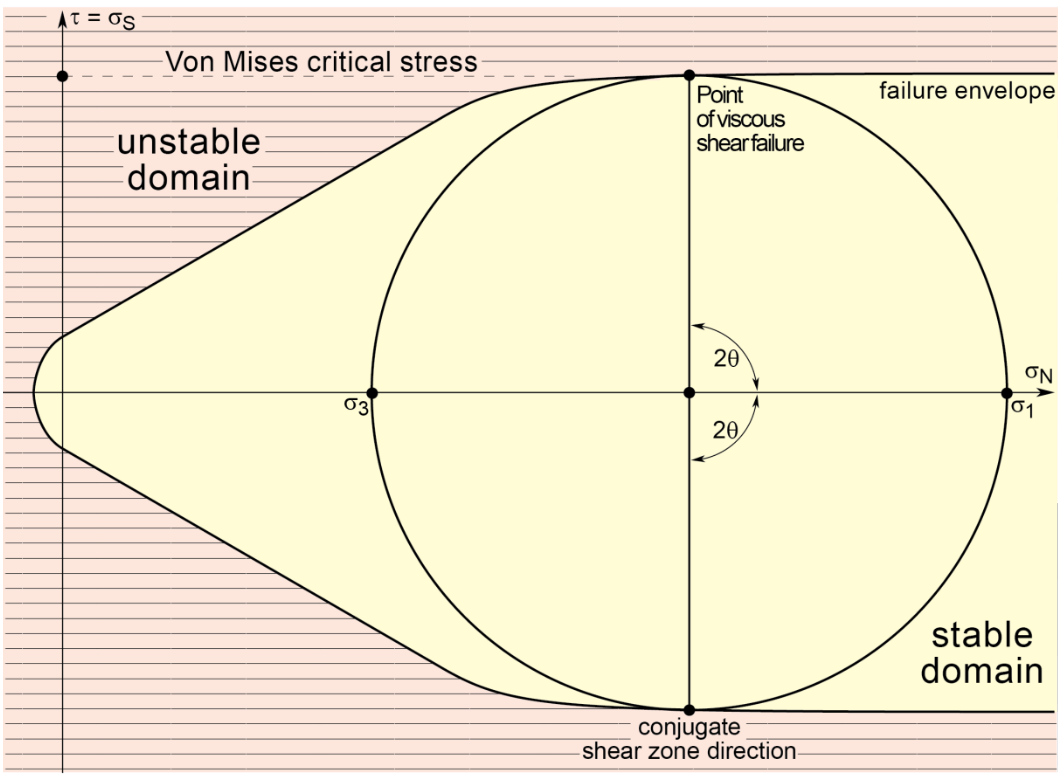
Single shear zone

In initially isotropic rocks, platy and flattened minerals may become aligned to form a foliation (labelled S) that makes an angle of about 45° to the shear zone at its weakly deformed boundaries and rotates progressively towards the shear plane to become essentially subparallel to the shear zone boundaries at large shear strain. In the strongly deformed domains, the stretching lineation can be equated with the shear direction. The curved or **sigmoidal pattern of the foliation** in the XZ sections of rocks defines the **sense of shear**. The bulk acute angle of the foliation to the shear zone walls is always sympathetic to the sense of shear.
Ideally also, strain gradients are continuous and antisymmetric from the shear zone walls to the medial, highest strain plane. Most commonly, the two sides differ in shape and size. One can directly infer the sense of relative displacement from the curved shape of the new shear foliation and from deflected pre-existing markers. The curvature of the sigmoidal foliation trace, comparable in shape (and shape only) to drag folds, is a direct indicator of the sense of shear. Continuity is maintained across ideal shear zones. However, the strong mechanical anisotropy created by the new shear foliation and fine grained mylonites make them prone to brittle reactivation or failure in discontinuous shear zones.



Conjugate shear zones

In contrast to conjugate brittle faults, a pair of conjugate ductile shear zones is ambiguous in terms of the positions of the maximum and intermediate principal stresses.



Orientation of new shear zones fitting the Von Mises criterion in a Mohr diagram

Ductile shear failure obeys the Von Mises criterion, viscous flow taking place under constant stress, independent of differential stress and pressure. The criterion approximately corresponds in a Mohr diagram to the part of the failure envelope which is parallel to the normal stress axis (i.e. constant shear stress). The tangency point where the stress circle can reach the envelope to trigger shear failure readily shows that ductile shear zones, in theory, initiate at 45° to σ_1 (the angle $2\theta = 90°$).

Conjugate shear zones are contained in a viscous material that may admittedly deform less rapidly than the shear zone mylonites, yet still deforms to respond to the regional stress field. Under these conditions, rotation of the shear zones may be imposed by bulk flattening of the country rock, which opens the angle containing the flattening direction between the conjugate shear zones. Consequently, obtuse wedges may contain the shortening direction. However, like for brittle faults, the intermediate principal stress coincides with the line of intersection of the two conjugate ductile shear zones.
The Griffith theory is inapplicable to ductile faults.

Multiple shear zones

Shear zones self-organize in specific patterns that are not fully understood. Spacing and orientation depend on strain regime, material and loading parameters, in particular the number and potency of initiation sites, externally applied strain rate, and stress state.

Spacing

Shear zones may appear as regularly spaced, high strain planar structures. Two possibilities have been proposed to explain this, both derived from considerations validated for brittle faults: (i) a diffusion mechanism, (ii) a perturbation mechanism.

Diffusion

The rapid loss of strength across a developing shear zone forces the wall rock to unload. Unloading is communicated outward by momentum diffusion and/or elastic wave propagation. The minimum separation between independently nucleating zones arises from the distance traveled by the diffusive unloading front as strain localization occurs.

Perturbation

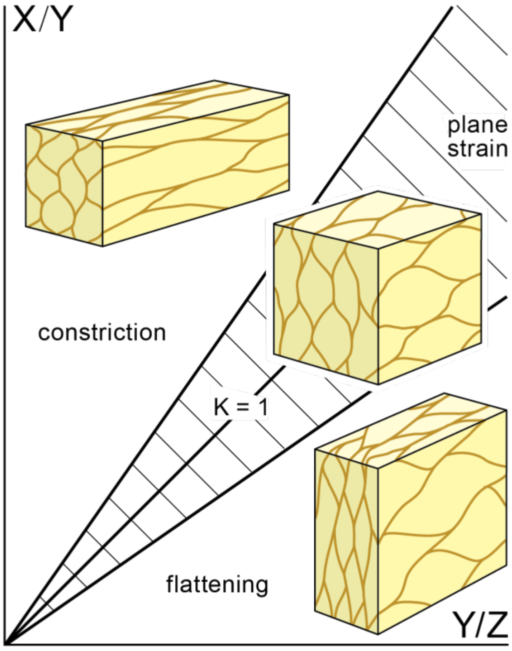
The idea is that shear zones grow from small heterogeneities in an otherwise uniform rock. Like in folding, the disturbance wavelength with the highest rate of amplitude dominates and will determine the shear zones spacing.

Patterns

Shear zones often anastomose around lenses of less deformed country rock. The shape of the bulk finite strain ellipsoid, representing the regional deformation regime, controls the three-dimensional pattern of anastomosed shear zones, which in turns defines the shape of lower strain rock lenses. Three qualitative patterns are identified:

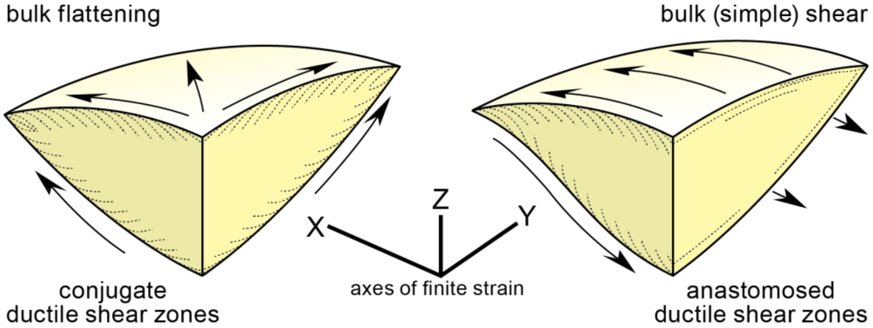
- Flattened rock lenses indicate the flattening field of bulk finite strain.
- Lozenge-shaped lenses indicate near plane strain.
- Rod-shaped lenses indicate constriction.

Array of shear zones
with respect to bulk strain regime
adapted after Gapais *et al.* 1987 *J.Struct.Geol* **9**(5-6) 635-646



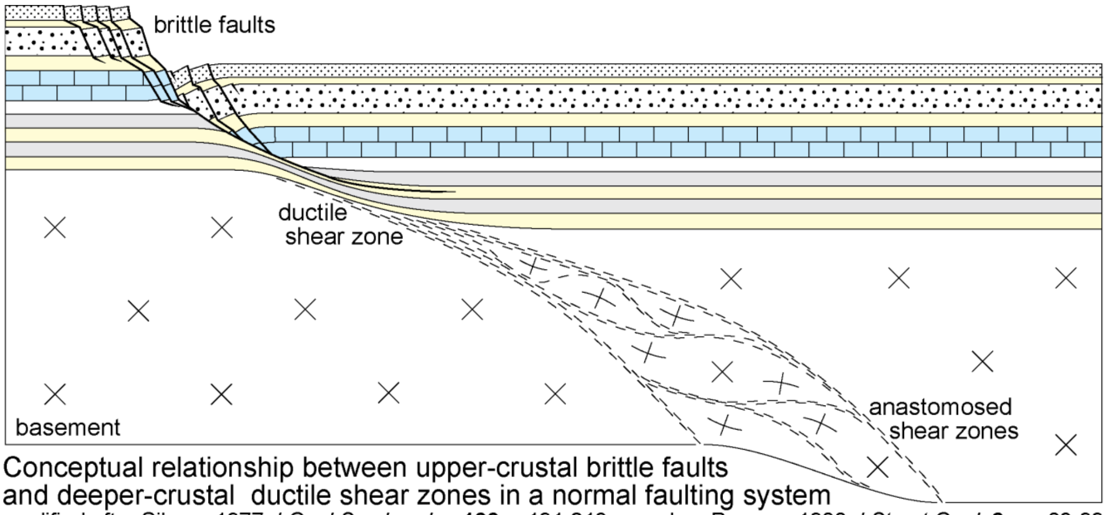
The relationship between the shape of low-strain lenses and the strain regime is a bulk scale, geometrical information. The kinematics of the shear zones that wrap around individual lenses provides additional information. Conjugate shear zones indicate bulk coaxial deformation; shear zones with identical sense of shear denote bulk non-coaxial deformation.

Relationship between shear zone pattern, kinematics and bulk finite strain



Relationship of deep shear zones to near surface faults

With temperature and pressure increasing with depth, discrete planes and narrow zones of brittle displacement in the upper crust are transformed into wider zones of ductile displacement in the middle and lower crust.



Conceptual relationship between upper-crustal brittle faults and deeper-crustal ductile shear zones in a normal faulting system
modified after Sibson 1977 *J.Geol.Soc.London* **133(3)** 191-213 and Ramsay 1980 *J.Struct.Geol.* **2(1-2)** 83-99

Strain in shear zones

Strain distribution and the associated, symmetrical displacement profiles are perhaps the most important characteristics of shear zones. The rock seen in the XZ plane of finite strain is subjected to a shearing couple, which instigates progressive rotation and coeval flattening with maximum finite shortening in one direction being accompanied by maximum finite extension in the perpendicular direction within the XZ strain plane. Particles move parallel to the shear sense, parallel to the walls of the shear zone so that the displacement and finite strain profiles in any shear zone cross section along the c kinematic axis are identical. These conditions are not fulfilled near the terminations of shear zones, but are adequate for shear zone segments away from these terminations. All arguments developed from now concern the XZ plane of finite strain in ideal shear zones.

Shear strain

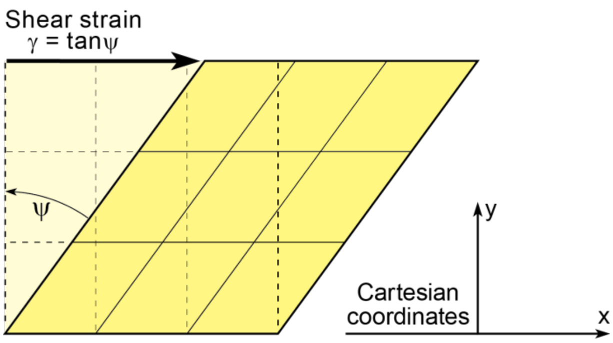
Imagine a shear zone with walls parallel to the ordinate, x axis. Assuming simple shear with the shear direction parallel to the shear zone walls and homogeneously distributed shear strain γ throughout the shear zone, then γ is related to the displacement Δx across the shear zone of width y according to the equation:

$$\gamma = \Delta x / y \qquad (1)$$

This happens to be the tangent to the angular shear ψ (the angle that describes the rotation of a line initially parallel to the y ordinate axis and remaining anchored to the origin):

$$\gamma = \tan\psi \qquad (2)$$

Note that shear strain ignores the changing length of this rotating line, i.e. its longitudinal strain. Shear strain refers only to change in angles.



Foliation trajectory

The non-linear decrease of strain with distance from the most intensely sheared shear zone centre is documented by sections perpendicular to the walls, along the c kinematic axis.

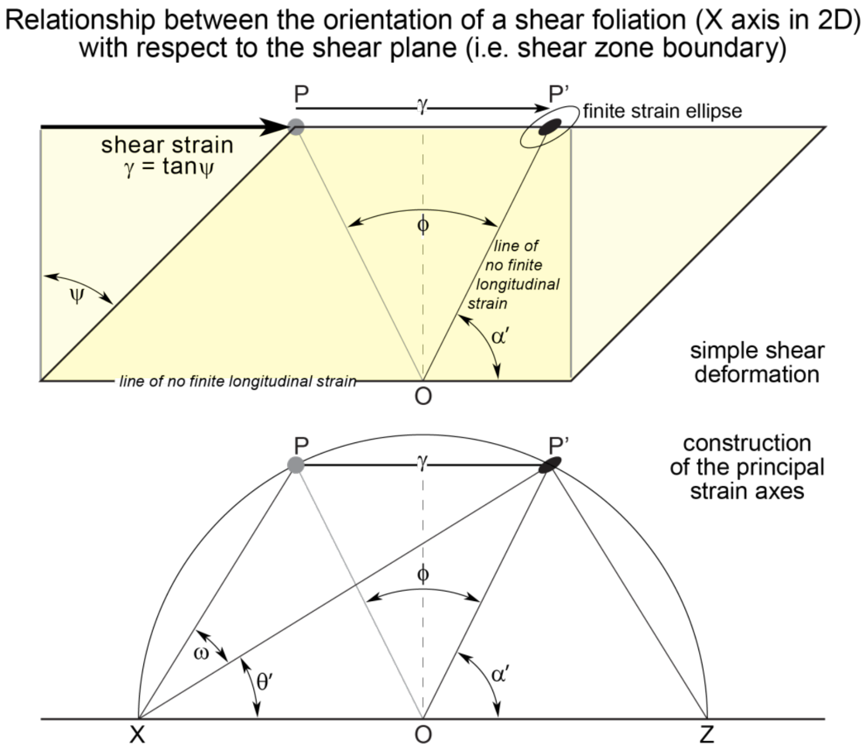
Orientation of foliation with respect to the shear zone orientation.

Ductile rocks obey the Von Mises failure criterion, the horizontal part of the failure envelope in a Mohr diagram. Accordingly the shear zone in theory initiates at 45° to σ_1 and contains an initial foliation (first incremental XY plane of finite strain) developing perpendicular to σ_1, i.e. inclined at 45° to the shear zone boundaries. In effect, at any given stage during the “simple” shearing, the instantaneous plane of flattening is at 45° to the shear zone boundaries (i.e. orthogonal to the maximal principal stress). Therefore, the occurrence of the new shear foliation along the immaterial walls of a shear zone is always formed at 45° to the shear zone boundary, with the inclination of foliation to shear zone in consistency with the sense of shear. This has been verified for shear zones developed in intact, homogeneous and isotropic host rocks.
This orientation control applies to the opening of tension gashes and fractures perpendicular to the instantaneous principal stretch or to the growth of a foliation perpendicular to the instantaneous principal shortening (flattening).

Strain gradient

From wall to medial mylonite, strain gradients are assumed to result from heterogeneous simple shear. Describing the trace of foliation across the shear gradient consists in tracking the rotation of the principal axes X and Z of the strain ellipse about the Y axis of constant length. The description consists in measuring the rotational component of deformation, the angle ω between the line of maximum stretch before and after a shearing increment.
The angle between the foliation within the shear zone and the walls, parallel to the shear plane, decreases with increasing shear according to the relation:

$$2/\gamma = \tan 2\theta' \qquad (3)$$



The mathematical demonstration is complex. One can employ geometrical constructions to reach it. The basis is that in simple shear, one line of no finite longitudinal strain is parallel to the shear direction (i.e. the shear zone boundaries, the non-rotating sides of a rectangle sheared to a parallelogram). A point P on the shifted side of the rectangle (for example the top one) comes to P' after simple shearing by the amount γ.

Draw a line orthogonal to the shear plane (in that case parallel to the top and bottom sides of the rectangle transformed into a parallelogram) through the mid-point of the segment PP'. The drawn line (vertical in this example) cuts the bottom side at a point O. OP represents a pre-deformation line that did not change length through the simple shearing to OP', since OP = OP', but rotated by an angle ϕ so that $\tan(\phi/2) = \gamma/2$. Therefore, OP' is the second line of no finite longitudinal strain in the strain ellipse after simple shearing by the amount γ. This second line of no finite longitudinal strain makes an angle α' (' denotes the deformed state) with the shear direction, i.e. with the other line of no finite longitudinal strain. This angle is directly related to shear strain:

$$\alpha' = 90 - (\phi/2) = 90 - \arctan(\gamma/2)$$

The directions of the two principal strain axes are then the two lines bisecting the acute (for X) and the obtuse (for Z) angles between the two lines of no finite longitudinal strain at P'. These two bisecting lines cut the basis line at points that we can name X and Z, respectively. Indeed, these points are also on a circle circle centered in O and passing through P and P'. This construction is reminiscent of the Mohr construction for strain, with points X and Z representing the stretches along the respective directions.

The line XP represents XP' before deformation, and the angle between these two lines is the angle of rotation ω due to shearing. This inscribed angle intercept the same arc PP' as the centred angle ϕ so that $\omega = \phi/2$. Hence

$$\tan\omega = \gamma/2$$

For the same geometrical reason the angle θ' that makes the X axis of finite strain (i.e. the trace of the foliation in the XZ plane of observation) with respect to the shear direction (the shear zone boundaries) is equal to $\alpha'/2$ (inscribed angle is half the centered angle). Then we have: $2\theta' = 90 - \arctan(\gamma/2)$ or $\tan(90 - 2\theta') = \gamma/2$

Using the trigonometric identity $\tan(90 - \beta) = 1/\tan\beta$ yields:

$$\tan 2\theta' = 2/\gamma \qquad (3)$$

Thus, θ' decreases with increasing shear strain. The structural consequence is that foliation is progressively curved from 45° at the margins (where $\gamma = 0$) to practically parallelism to the shear zone boundaries within the central mylonite, if shear strain is intense enough.

Exercise

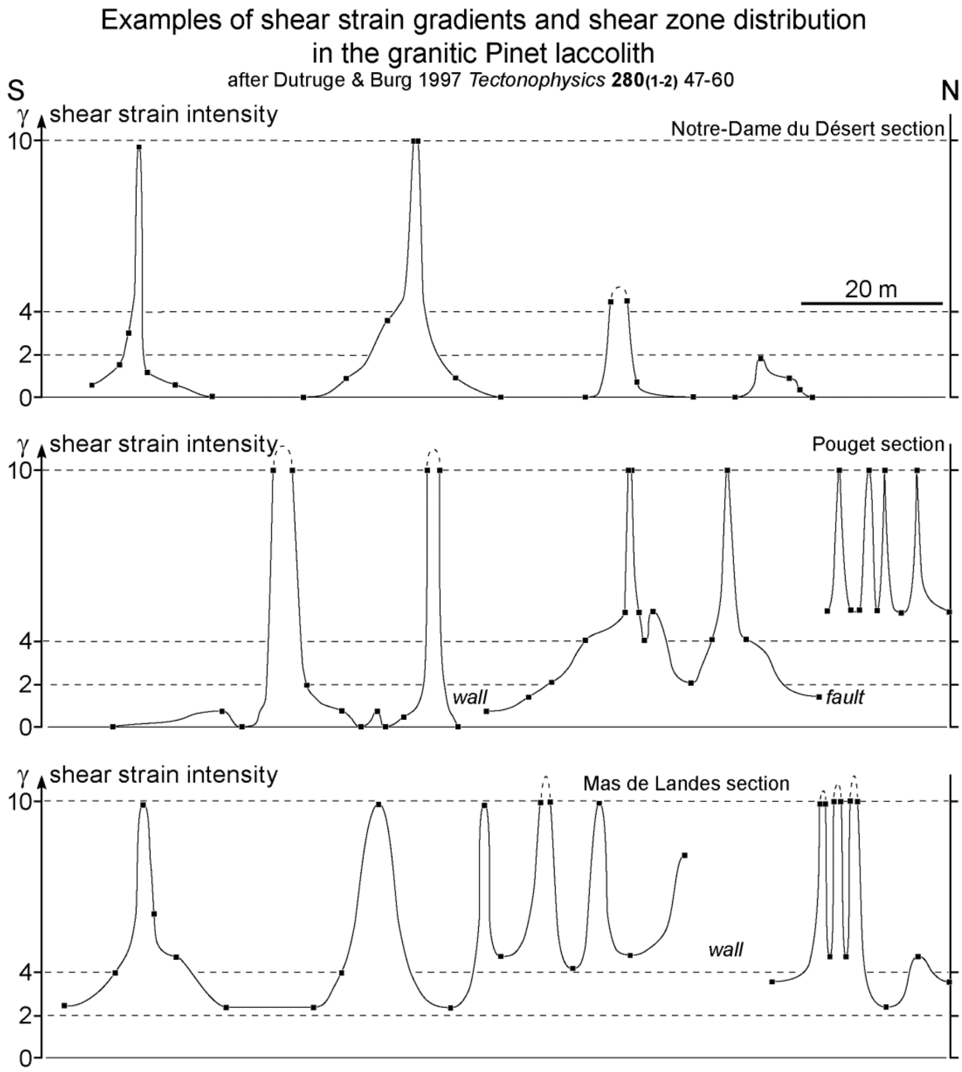
Shear Zone Foliation and Strain: Assuming simple shear, calculate shear strain from inclination θ of foliation to the shear zone boundaries. Accept that an angle of 2° is no longer distinguishable from parallelism. What is the shear strain reached when foliation is quasi-parallel to the shear plane (and therefore to the shear zone boundaries).

The magnitude of the principal stretches is straightforward.

In the X direction the stretch $S_X = XP'/XP = XP'/ZP' = 1/\tan\theta'$ and, under constant area conditions, $S_Z = 1/S_X$.

Exercise

Show that the aspect ratio of the strain ellipse $(X/Z) \approx \gamma^2$ when $\gamma > 3$.



Effects of volume change

A volume change ΔV adds a component of flattening or thickening to simple shearing, which affects both magnitude and orientation of the principal strain axes. These components, acting perpendicular to the shear plane, change the relationship between θ' and γ where, incorporating volume change:

$$\tan 2\theta' = \frac{2\gamma(1+\Delta V)}{1+\gamma^2-(1+\Delta V)^2} \qquad (5)$$

Exercise

Generate curves of θ' with respect to volume loss and volume gain. Compare and discuss changes with simple shear conditions.

Displacement across shear zones

The shear displacement δd over an infinitesimally narrow shear zone element of width δy measured perpendicular to the shear zone boundaries is given by (equation 1):

$$\delta d = \gamma\,\delta y$$

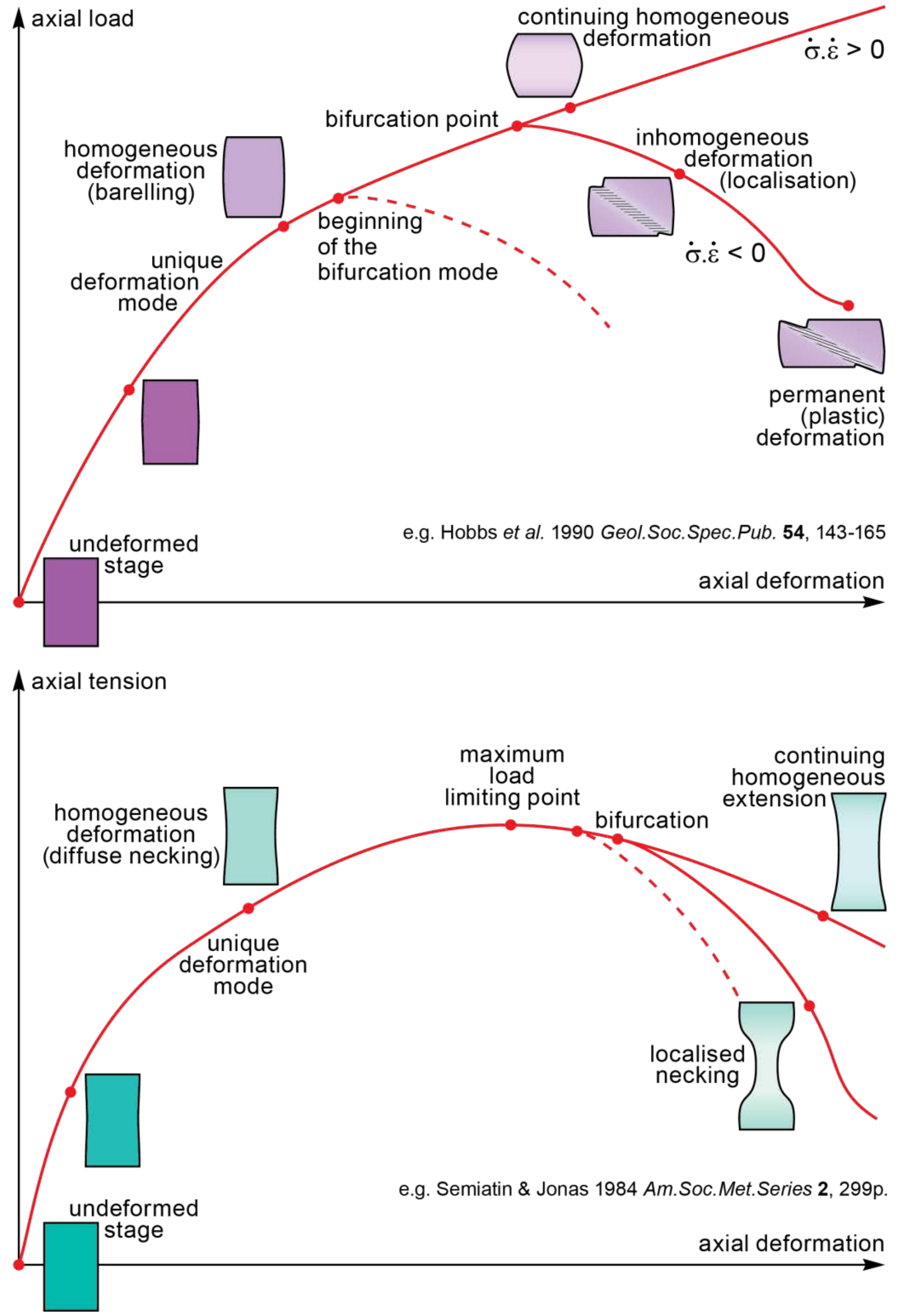
The total shear displacement across the whole shear zone is the sum of all elements:

$$d = \int_0^y \gamma\, \delta y$$

This integral represents the area under the curve shear strain/width across the shear zone.

Strain localisation

The commonness of shear zones indicates that localisation efficiently dominates the macroscopic deformation behaviour of ductile rocks. This behaviour demonstrates that, under some conditions, rocks are given two possibilities to solve the bulk deformation problem: diffuse homogeneous vs. localised heterogeneous strain. These conditions are a **bifurcation** point in physical and mathematical terms. The choice between homogeneity and heterogeneity exists both temporally and spatially, which means that switching from one mode to the other is offered not only along the deforming path at each incremental strain, but also at every point within the body.



For ductile, ideally viscous and homogeneous deformation the stress at the yield point remains perfectly constant for increasing strain. The dislocation creep failure criterion indicates that ductile deformation is fundamentally governed by mineral composition, temperature and strain-rate dependent:

$$\dot{\varepsilon} = \frac{d\varepsilon}{dt} = A\left(\sigma_1 - \sigma_3\right)^n \exp\frac{-Q}{RT} \quad (6)$$

where $\dot{\varepsilon}$ is the strain rate, A, n and Q are constants depending on rock type, T is absolute temperature and R the gas constant. Generally $\sim 3 < n <\sim 5$

The dislocation creep failure criterion involves processes through grain interiors and boundaries, thus depends on the grain size d:

$$\dot{\varepsilon} = Bd^m \left(\sigma_1 - \sigma_3\right)^n \exp\frac{-Q}{RT} \quad (7)$$

where $n \approx 1$ and $\sim 2 < m <\sim 3$. Since dislocation and diffusion creep are independent, the total strain rate of a ductile material is simply the sum of equations 5 and 6 for given pressure and temperature conditions. For a given grain size, dislocation creep dominates at high stress and diffusion creep at low stress.

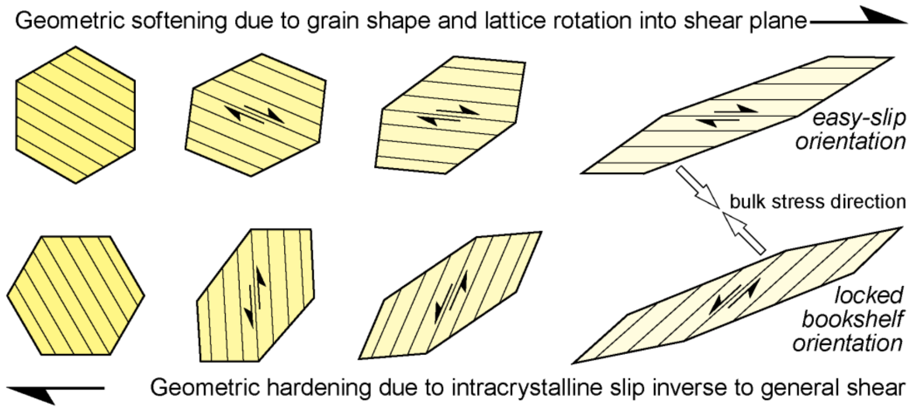
From these considerations and equations, in geology like in material physics, shear zones are thermo-mechanical instabilities. They are formed when combined rates of geometrical softening and thermal softening overcome the rate of strain hardening.

It is generally assumed that localising deformation requires strain softening (i.e. deformation is allowed to proceed at a decreasing level of stress). However, softening is relative. All parts of a rock system may harden but strain localisation will take place within the less hardening parts. Accordingly, both heterogeneous softening and hardening may lead to strain localisation. In any case, the strain rate is faster inside the zone of localisation than outside so that the rock within a shear zone remains easier to deform than its surrounding. Strain localisation indicates a transition from higher-stress to lower-stress deformation mechanisms at a constant strain rate. This does not imply that rocks bounding shear zones are rigid. While deformation occurs in shear zones, outer parts may also deform, yet following another rheology, therefore registering different structures. Juxtaposed structures may be simultaneous responses to significantly different stress and strain rates. In rheological terms, while shear zones are following a softening course, rocks in between may still be hardening. Shear zones may represent localised bursts in deformation in their country rocks undergoing slower and steadier strain. The site where localisation begins raises questions on the role of any sort of imperfection.

Several mechanisms are discussed, which might combine to see the strength of the deforming rock to decrease with strain and produce shear zones. Some mechanisms involve no change in material constitutive equations, and bifurcation towards localisation has geometrical causes. This is the case for **geometric softening**, dynamic recrystallisation and viscous/**shear heating**. Other mechanisms introduce activation of an additional process, for example microboudinage of grains, **grain boundary sliding**, chemical reactions and fluids.

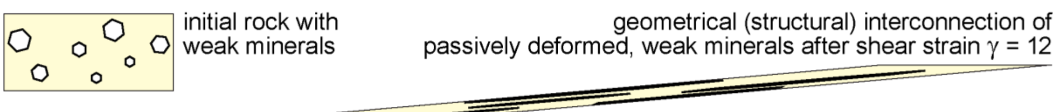
Geometric softening

Ductile deformation changes the texture (grain shape and lattice orientation) of the rock. Grain lattices progressively rotate with grain shapes towards parallelism with the shear plane, which induces geometrical softening. The process is envisioned as follows: Intra-granular slip systems are microfaults. The resolved shear stress required on a given crystallographic **slip plane** decreases as this plane rotates into parallelism with the shear plane and as the intra-crystalline slip direction (Burger's vector of dislocations) rotates into parallelism with the stretching lineation.



Modelled evolution of grain shape and crystalline orientation during two-dimensional dislocation flow - Effect of initial grain orientation on geometrical softening / hardening
modified after Etchecopar 1977 *Tectonophysics* **39(1-3)** 121-139

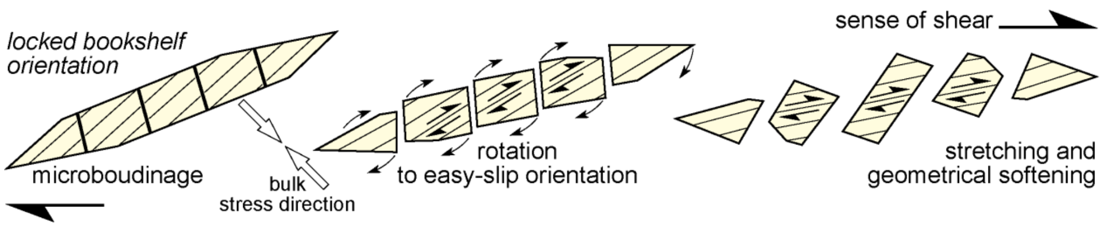
Grains with lattices aligned with this slip direction remain in this easy slip situation and the increasing abundance of grains that reach this stable attitude develops a **lattice preferred orientation**. The stress for deformation is then at a minimum. Geometric softening is most pronounced in minerals with few slip systems. The extent of softening depends on the starting fabric and its orientation to the shear plane and shear direction, but any amount of softening significantly enhances strain localization.
Even at low (ca 10%) volume fractions, deformation-driven interconnection of initially randomly dispersed weak minerals is also an important softening mechanism in polymineral rocks.



Structural softening of a polyphase rock

Grain size reduction

In the ductile regime, grains grow to reduce surface energy while subgrain rotation and grain boundary bulging produce new small grains (dynamic recrystallization). Hence, the average grain size evolves towards the equilibrium value for which the two competing processes balance each other. Continual recrystallization leading to small grains weakens rocks since, in a deforming mylonite, new strain-free grains are permanently formed, which enhances diffusion creep and reduces the dislocation density.



Modelled softening due to boundary processes involving grain fragmentation and rotation
modified after Etchecopar 1977 *Tectonophysics* **39(1-3)** 121-139

Weakening from coarse- to fine-grained rocks may also reflect a change in deformation mechanism from dislocation creep to lower-stress, grain boundary processes.

Reaction softening

Reaction softening is a complex process involving micro-mechanical and chemical processes.

Syn-kinematic metamorphic reactions generate new minerals, usually growing in the local flow plane and direction, hence under easiest-slip conditions. This is particularly true for new mica on foliation planes. Hard mineral phases may be converted to softer assemblages (e.g. feldspar replaced by quartz and sericite). The new, strain-free (hence weaker) mineral phases grow from minute nuclei whose very small grain size promotes strain localization. In all cases, mineral reactions produce a loss of cohesion across grain boundaries, thus further promote weakening and grain size reduction arising from both brittle and ductile processes. Many metamorphic reactions release fluids that also enhance softening by facilitating grain-boundary sliding.
In addition, metamorphic reactions involve localised diffusion. Chemical softening results from changes in elemental content of minerals. For example bound water is known to weaken quartz and olivine.

Fluid-related softening

As in brittle deformation, pore fluids can significantly reduce the effective normal stress, hence the cohesive strength of grain boundaries and assist microfracturing. Besides this pore pressure contribution, infiltration of aqueous fluids into the shear zone may lead to retrograde hydration reactions. Fluid circulation is also efficient at dissolving grains that resist ductile deformation and changing dislocation or diffusion flow to dominantly diffusive mass transfer deformation mechanisms. Importantly, solution / precipitation processes with softening potential strongly depend on fluid composition as much as on temperature and pressure. If pressure solution creep may weaken some rock domains, strain concentration may be further promoted by neighbouring domains that are strengthen through precipitation and subsequent cementation.

Temperature softening (shear zone is hotter)

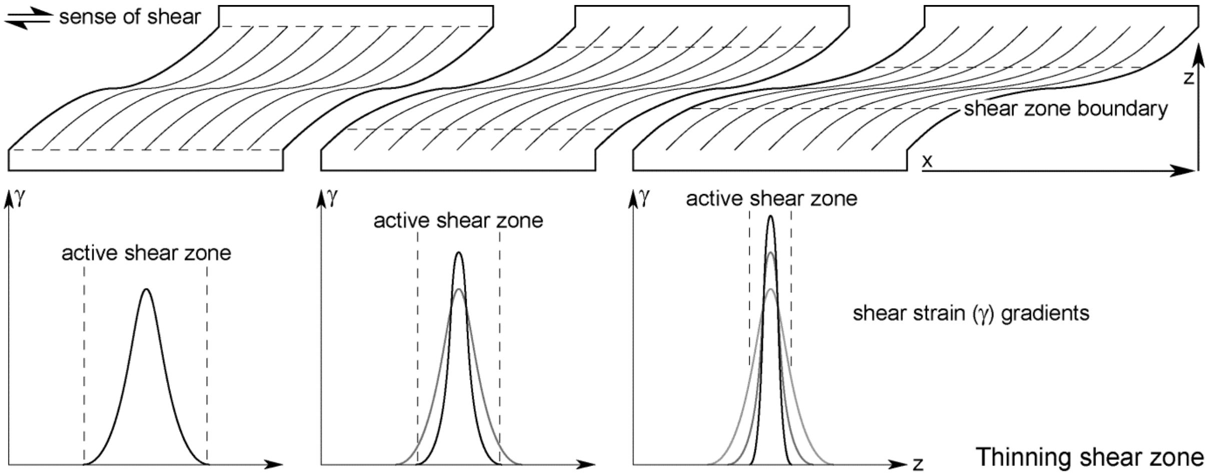
Thermal effects alone do not cause strain localization. But once deformation is confined to relatively thin zones, strain dissipates heat. Irreversible conversion of mechanical work into thermal energy is termed **viscous heating**. Dissipated heat adds to the temperature term of the rock temperature-dependant viscosity, hence weakens the shear zone with respect to its surrounding, which activates further localization etc… the local feedback straining-heating is working in loops whose consequences are continuous strain-weakening. The interplay between strain-generated heat and heat diffusion through the country rock has effects on the velocity profile.

Thickness changes

Equations used for ideal shear zones are approximations assuming homogeneous simple shear and no change in shear zone width. Variation of the thickness (width in the two-dimensional XZ plane of finite strain) of a shear zone during a progressive and continuous shearing event impels adapting displacement estimates from structural features.

Thinning (narrowing) shear zones

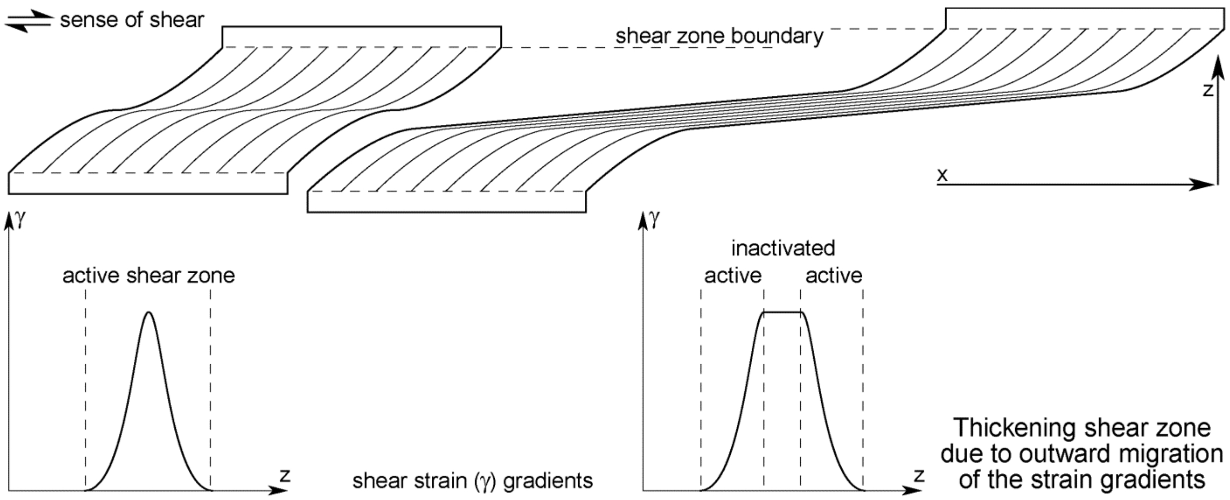
Strain softening may enhance strain localisation in the median part of the shear zone. The relationship between foliation and shear zone boundaries has a transient significance, being frozen in the earlier stage, once the shear zone boundary has started to migrate inward towards highly strained mylonite and ultramylonite.



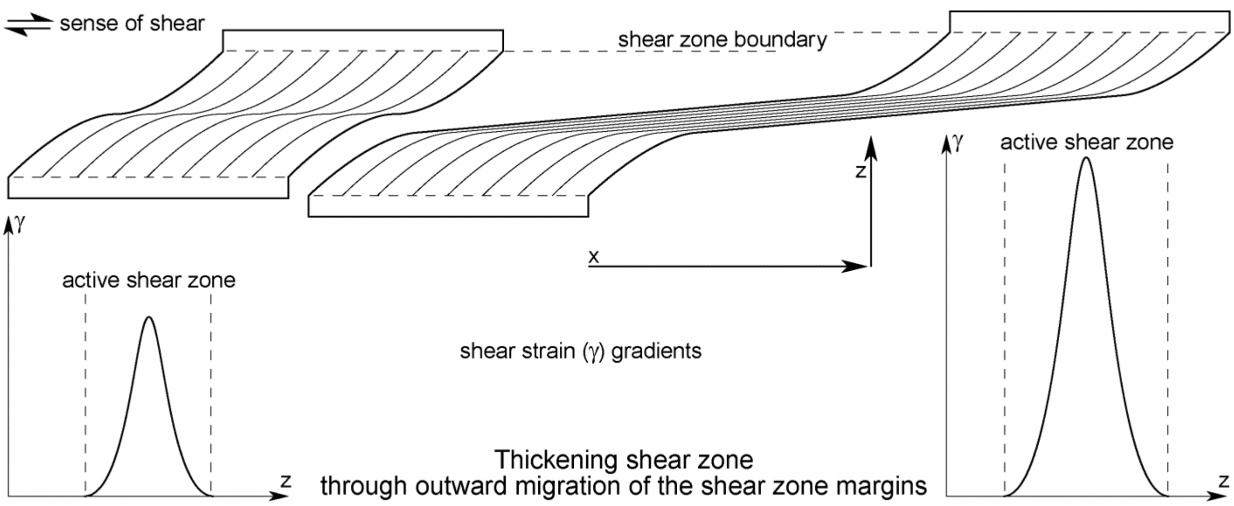
Thickening (widening) shear zones

Widening of the shear zone during shearing generates a new foliation at 45° to the outward migrating shear zone margins. This requires that wall rocks become easier to deform than the deformation zone. There are two main possibilities according to the behaviour of the central mylonite.

The mylonite widens with the shear zone boundaries because the mylonite hardens at some level of shear strain so that it does not deform further. The shear zone undergoes strain hardening.



Alternatively, the whole shear zone remains active. The shear zone is hardening with respect to the country rock so that deformation integrates a continuously thicker deformation zone.



Rocks of ductile shear zones

Since shear strain is heterogeneous, rocks with variable grain size and textural characteristics reflect the variable intensity of ductile deformation across shear zones. These cohesive rocks (meaning they kept their continuity) form the **mylonite series**.
Mylonitization is a gradual process of grain size refinement associated with a very large stretching along the newly formed foliation. The fine-grained product of a large grain is drawn out as a thin sheet and can often be recognised as tails of adjacent to larger fragments or relict minerals, the **porphyroclasts**. This is one of the reasons why mylonites often show a fine colour-banding parallel to the foliation. Because of the extreme stretching, mylonites also acquire a prominent lineation.

Classification

The variety of mylonitic microstructures and the different stages of their evolution are best documented when the parent rock is coarse grained, as in granite. Three stages are distinguished on the relative proportion of porphyroclasts to fine-grain, syn-kinematically recrystallized matrix.

* A **protomylonite** is a rock in the early stages of mylonitisation, containing more than 50% porphyroclasts. With the onset of deformation, a protomylonite shows a mortar texture with a very fine-grained matrix surrounding large residual grains of the parent rock (**mantled** grains). The rock has a coarse foliation and rather weak lineation that mould the porphyroclasts (**Augen** for feldspars in deformed granites).

* A **mylonite** is a foliated and lineated rock which has undergone a drastic reduction in grain size by dominantly crystal-plastic processes. A mylonite contains 10-50% porphyroclasts, i.e. 50 to 90% of matrix.

* An **ultramylonite** is hard, flint-like, and dark, which is the visual result of extreme grain size reduction and dynamic recrystallization. The surviving, commonly tiny porphyroclasts constitute less than 10% of the rock.

* At very high temperature and low strain rate, minerals tend to extensively recrystallize. The static, postkinematic recrystallization is called **annealing**. These metamorphic rocks have coarse grains with straight grain boundaries. Minerals grown during and after deformation are called **porphyroblasts**. The result is a **blastomylonite**. Blastomylonite may have lost part of their mylonitic, mesoscopic fabric. Note that the term is ambiguous since it is difficult to know a blastomylonite from a normal metamorphic rock like gneiss.

*** Phyllonite** is a mica-rich mylonite that has the mesoscopic appearance of schist.

Shear zones sometimes contain a dark glassy rock or **pseudotachylite** derived from local melting of the rock by frictional heating. Pseudotachylites may occur either as concordant veins parallel to the shear zone walls or as discordant veins injecting the surrounding rocks. Under the microscope it shows angular fragments of rock and minerals occurring in a glassy matrix. The matrix may have undergone various degrees of recrystallization but, unless it is formed at a later stage, it does not show a preferred orientation of the recrystallized grains. The original glassy character of the matrix can often be identified by the presence of microlites, spherulites and devitrification structures.

Paleopiezomety

There is an empiric, non-linear relationship between the dynamically recrystallized grainsize d and subgrain size formed during dislocation creep and the steady-state differential stress σ in monomineral isotropic rock:

$$\sigma = Bd^{-n}$$

where B is an experimentally determined material constant, and usually $0.5 > n > 0.75$ according to the considered mineral.
The **piezometers** are therefore applicable to minerals, grainsizes and recrystallization mechanisms for which they were experimentally calibrated. Major limitations to apply the theory to geological cases concern the average grain size, which is necessarily larger than the smallest grain size and

perhaps much larger if annealing caused grain growth while temperature was lasting long after deformation. Thus natural mylonite grain size likely yields too low stress estimates.

Shear sense indicators

For ductile shear zones like for brittle faults, it is important to determine the orientation of the displacement vector and the sense of shear; unfortunately, it is often impossible to measure the magnitude of the displacement because of huge uncertainties in shear strain assessment where foliation is subparallel to shear plane (i.e. for $\gamma > 10$). **Kinematic indicators** are meso and microscopic structures from which geologists can infer the sense of relative displacement and/or ductile flow in rocks. This is facilitated in ductile shear zones because kinematic indicators have a systematic asymmetry reflecting the sense of **vorticity**.

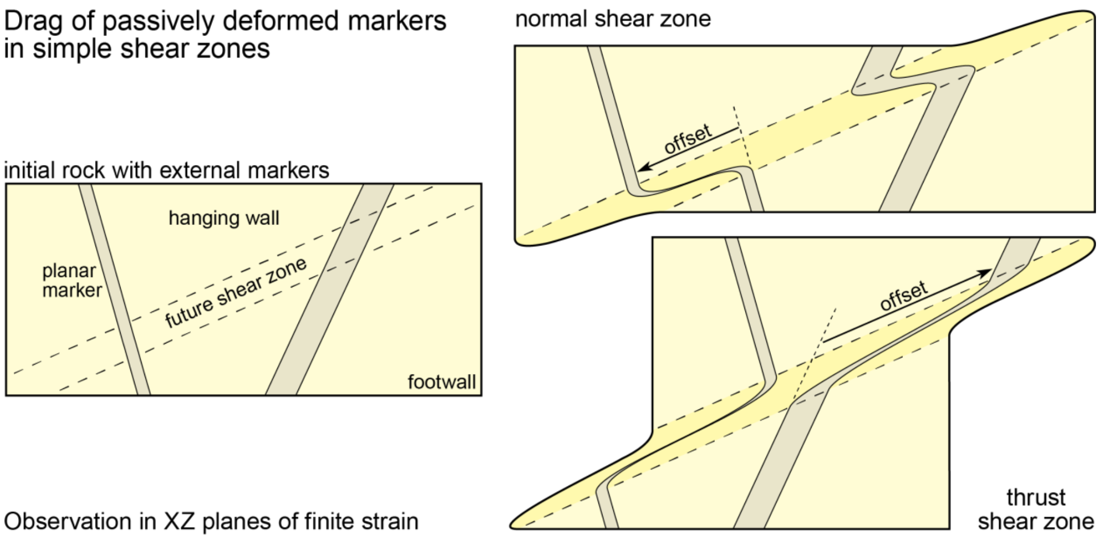
Shear sense indicators are structures that indicate the sense of shear in a progressively non-coaxially deformed material.

Reference Frame

The basic premise is that in intense shear deformation the bulk flow plane is parallel to the average foliation and the shear of the flow is parallel to the lineation. Sense of shear indicators must then be observed in XZ sections cut perpendicular to the foliation plane and parallel to the stretching lineation. Such shear zone profiles are sense-of-shear (SOS) planes. A thorough study of kinematic indicators from a shear zone involves recording many kinematic observations at the mesoscopic and microscopic scales. Shear zones with defined sense of shear accept the same terminology as for faults, normal, thrust, strike-slip, etc.

Offset markers

The gradual deflection of a planar, pre-existing marker such as a vein, a dyke or an old foliation initially oblique to the shear zone is often a direct reflection of the sense of relative displacement within the shear zone. The change in orientation of the external marker is generally a passive rotation in accordance with the shear sense in the zone, and this deflection is often referred to as **drag**.



En échelon veins

Common minor structures associated with shear zones are mineral-filled **tension veins**, in particular where semi-brittle deformation is involved. Veins are often arranged *en échelon* within the shear zone, each vein opens along the direction of maximum instantaneous extension, which enables to determine the extensional quadrants from which the relative sense of shear can be inferred. With progressive deformation, the veins may rotate but continue to grow along the principal compressive stress, at 45° to the shear zone margin. This results in the formation of curved or sigmoidal veins.

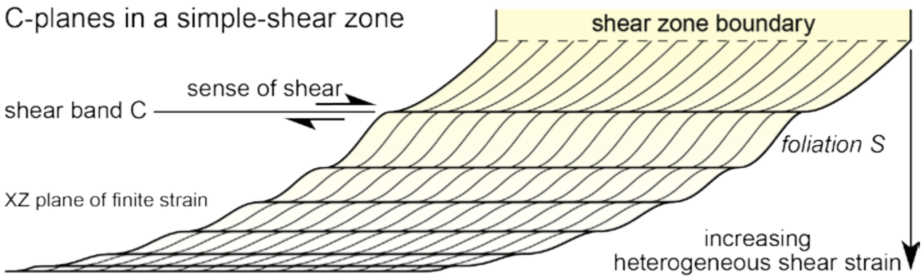
Earlier-formed veins that have been rotated and sheared into sigmoidal shapes may be cross cut by younger veins forming in accordance with instantaneous strain ellipse.

Foliation and shear bands

The systematic, sigmoidal curvature of the shear foliation across shear zones is the most direct sense of shear indicator. The simple shear foliation leans in the sense of shear and turns from ca 45° along the shear zone boundaries towards parallelism to the shear zone boundaries in the central mylonite.

C-S structures

The foliated fabric may eventually become unstable with respect to the principal compression and the strain gradient loses its regular curvature. Heterogeneous deformation generates a composite fabric defined by S-surfaces deflected into parallelism with, spaced, straight and narrow zones of concentrated shear parallel to the main shear zone (i.e. parallel to the flow plane of the shear deformation): the **C-shear bands** also designated as **C- surfaces**. These discrete shear zones are visibly less penetrative than foliation planes S. The composite C-S fabric usually weakens outward (larger spacing between C-planes towards the shear zone boundaries) and ultimately disappears outside the shear zone, which suggests its direct affiliation to shearing. The C planes maintain a constant orientation but the S planes rotate with increasing shear strain, so that the angle between the two sets reduces to close parallelism in highly sheared mylonites.



In general, 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. This angular relationship is a frequent and reliable indicator of the relative sense of shear, with the acute angle between C and S planes indicating the direction of shear displacement.

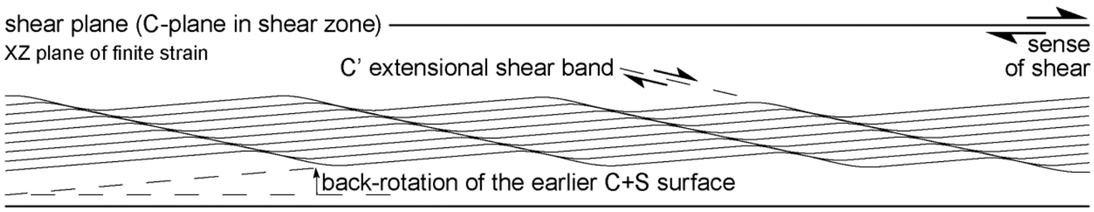
Note that owing to additional and localized displacements on C surfaces, the foliation planes S do not record the total shear-strain of the rock. Note also that C-surfaces are parallel to surfaces of no finite shear strain.

S-C fabrics are most common in granular rocks, in particular porphyroclastic granitoids.

Alternating shear zone domains of strong and weaker shear deformation may impart to the rock a banding mostly contained between C-planes.

C'- (and C'')-shear bands

As high shear strain is approached, the C and S surfaces become sub-parallel, so that their separate identity is lost. A new set of shear bands may form oblique to the shear zone boundaries, dipping towards the shear direction, in synthetic orientation reminiscent of Riedel shears in brittle fault zones, at about 15-25° to the bulk shear (flow) plane. Such fabric elements, systematically oblique to both the earlier foliation and to the shear plane, are known as extensional shear bands C’, which result from some localised slip in a component of extension within the bulk flow plane. C’ shear bands are generally more discontinuous than C-planes. They may occur in a single set or in multiple sets (C”-, etc.) at low angles to each other. S-C’-structures may indicate intense, non-coaxial and partitioned flow. Conjugate sets may arise to accommodate extreme flattening. C’-shear bands accommodate normal displacement disclosing net extension parallel to the dominant mylonitic fabric. Some back-rotation of the earlier foliation may accompany slip on these shear bands. The sense of shear is such that C’ surfaces tend to “crenulate” the older planar fabrics. The acute intersection angle and the sigmoidal shape of the mylonitic fabric between these shear zones (or bands) show the direction of shear.



Extensional shear bands C' in strongly foliated mylonitic rock

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

Note

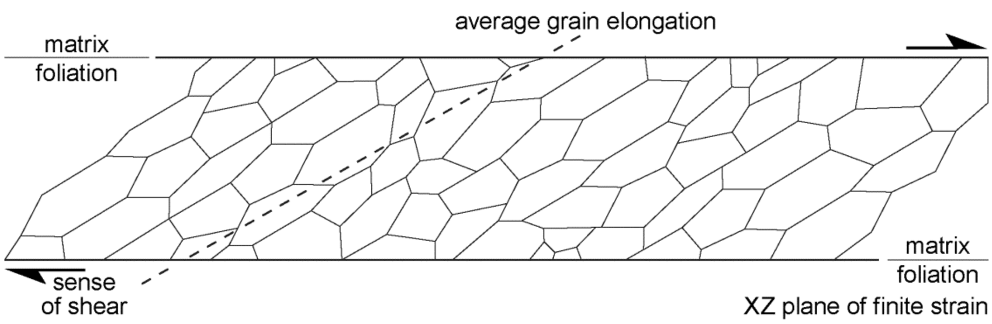
C and C' type shear bands may be difficult to distinguish but this difficulty has no kinematic consequence since their obliquity to foliation yields the same sense of local motion.

Multiple foliations

Progressive rotation due to shearing rotates and/or rolls earlier structures while flattening of the rock grains is perpendicular to the instantaneous shortening axis. The interplay between finite and instantaneous strain axes results in multiple orientations of structures (e.g. grain shape fabrics) generated during a single continuum of deformation. However, the consistent shearing couple, throughout, impels systematic asymmetry that reveals the bulk sense of shear.

Oblique grain shapes

In mylonites, grains can be strained into very long **ribbons** parallel to the main foliation plane. Elongated, recrystallized grains within the ribbons may develop a **grain-shape alignment** oblique to the main foliation in the rock. This oblique, grain shape foliation indicates deformation faster than recrystallization, which tends to revert grain shapes to equidimensional. The obliquity is consistent with the bulk sense of shear.



Grain shape preferred orientation oblique to a monomineral ribbon

Oblique minerals

Coexisting, multiple foliations may be coeval, in particular in magmatic fabrics and in partially molten rocks. This complexity is often related to shape-dependent rates of rotation, with different minerals defining different preferred shape orientations oblique to each other. Acute angles show the direction of flow. Often also, magmatic or migmatitic banding is lying stable on the shear plane whereas platy minerals in between define the oblique foliation(s).

Folds

Local, mechanical instabilities cause, at places, folding of the mylonitic foliation

Sheath folds

Ductile faults also characteristically exhibit extended tube-shaped **sheath** folds. Sheath folds contain the stretching lineation along the length of the tube or tongue. Hence, the long dimension of these folds is approximately parallel to the direction of slip on the ductile fault.

Asymmetric folds

The local and systematic **asymmetry of intrafolial fold trains,** whose fold axes lay at a high angle to the lineation, may indicate the sense of shear in strongly deformed shear zones. However, care must be taken to check whether fold trains are not small parasitic folds whose asymmetry can be inverted around folds too big to be seen on outcrop. Therefore their regionally consistent asymmetry is a key verification to be made to infer shear flow from fold asymmetries. Otherwise, inferring a regional transport (shear) direction from fold asymmetry is litigious.

Asymmetry of intrafolial fold train as sense of shear indicator

> Warning
> If fold axes are parallel to the shear direction, these “**curtain folds**” cannot be used as sense of shear criterion.

Quarter folds

More reliable, disharmonic, asymmetrical **scar folds** *(or* ***quarter folds****)* often form in diametrically opposed parts of the ductile matrix adjacent to stiff inclusions. Their distribution and asymmetry defines the sense of shear.

Hard textural elements

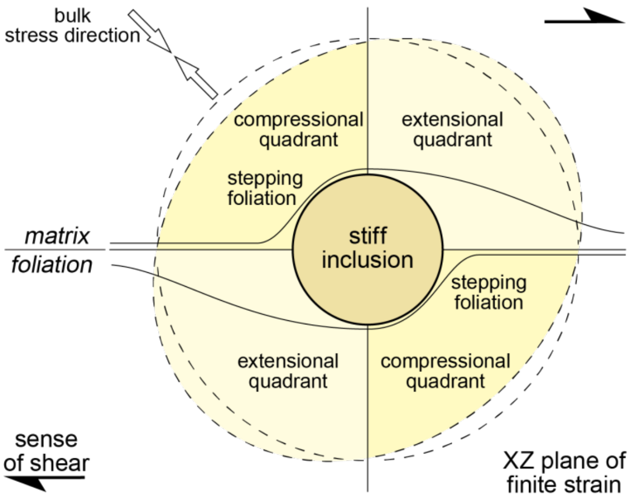
Rigid elements react to non-coaxial deformation by developing asymmetric structures in the XZ-plane of finite strain. Three main structural groups are distinguished: rotated rigid objects with or without mineral tails, broken objects and newly recrystallized mineral grains.

Asymmetric grain features

Many rocks in ductile shear zones contain large crystals. Some are relict crystals, or **porphyroclasts** (after the Latin word porphyry, which means purple, and the Greek word *klastos*, which means broken), that survived the shearing and reduction in grain size from the original rock. Others are **porphyroblasts** (after porphyry and the Greek word *blastos* meaning growth), which are mineral grains that grow to a relatively large size in a rock during metamorphism and deformation. (The use of porphyry for a rock with large crystals comes from large feldspar phenocrysts in statues of Roman emperors carved from purple volcanic rock).

Deformation quadrants

A stiff inclusion perturbs the flow in the soft matrix immediately adjacent to it. The perturbation area, wherein the instantaneous stretching axes lie at different orientations than those of the bulk flow, can be divided into four quadrants separated by the bulk foliation plane and its normal at the inclusion centre. Two quadrants are extensional; they alternate with the other two shortening quadrants. In bulk non-coaxial flow these quadrants of perturbed local flow are asymmetrically disposed about the stiff inclusion with respect to the main foliation, which is equated with the shear plane in strongly sheared rocks. This asymmetry directly reflects the shear-sense of the bulk flow in the matrix.



Incremental deformation quadrants
in the perturbed flow of a soft matrix around a hard object

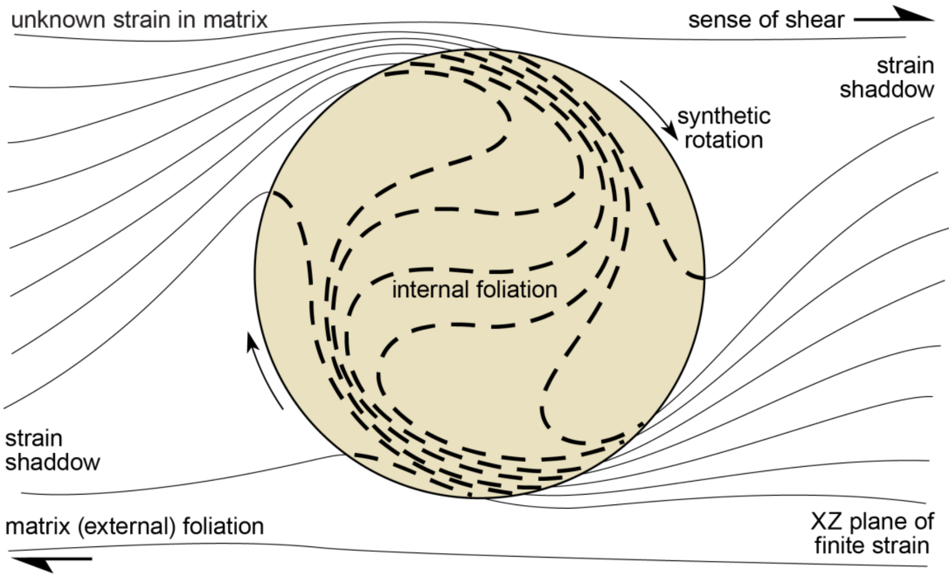
Stair-step structures

The shear sense can most often be read directly from the apparent deflection of foliation asymmetrically wrapping around in a **stair-step** manner any inclusion such as porphyro- blasts and -clasts.

Rotated objects

Snowball structures

Minerals that form porphyroblasts include all metamorphic crystallisations of any metamorphic grade, in particular relatively stiff minerals such as chloritoid, garnet, staurolite and Al-silicates. Porphyroblasts do not deform with the rest of the rock but roll as rigid grains during ductile shearing of the matrix. As they grow, they enclose adjacent minerals from the matrix, which results in an **internal foliation** or **helical trail of inclusions** representing segments of matrix foliation overgrown by, and included in the porphyroblast. Typical examples are **snowball porphyroblasts** that have a spirally oriented internal foliation from which the sense the grains have been rotating in the matrix can be deduced, consistently with the vorticity. The relationship between the orientation of the internal foliation and the **external** matrix foliation further defines the sense of rotation of the blasts and thereby the sense of shear in the rock. Mineral inclusions indicating the sense of shear provide additionally a relative timing of metamorphic growth. Inclusions in the core of porphyroblasts are older than those in the rim.

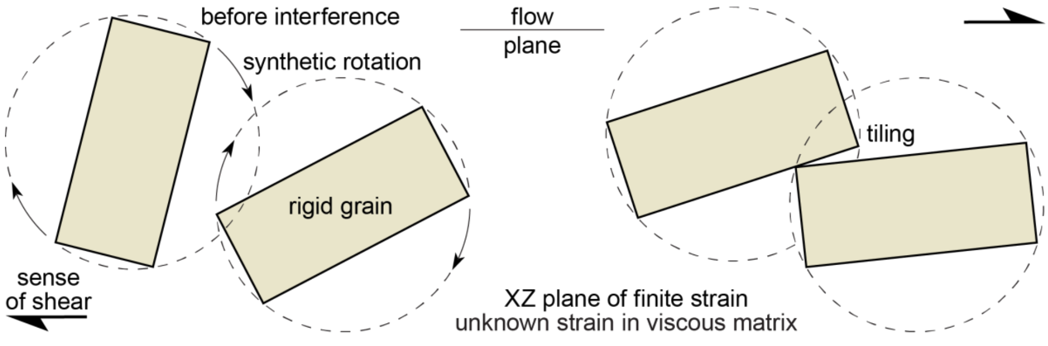


Helical trails of inclusions in a garnet pophyroblast rotated by 180°
String-cylinder model of Schoneveld 1977 *Tectonophysics* **39(1-3)** 453-471

Interpretation of shear sense from inclusion trails that feature small amounts of rotation, however, is unreliable because such trails may actually preserve crenulations of the original external foliation rather than a record of porphyroblast rotation.

Tiling structures

In plutonic rocks, stiff and elongate (platy) porphyroblast and clasts in the viscous matrix rotate and develop a shape distribution oblique to the flow plane. Mutually interfering inclusions may block each other's rotation, which develops overlaps termed **tiling structures**. In reality, tiling indicates rotation of individual crystals but not necessarily a non-coaxial character of the bulk flow. The tiling direction must be statistically largely predominant to indicate the bulk rotation (vorticity), hence the general sense of shear. In reality, elongated objects can undergo reverse rotation when their long axis lies initially at an appropriate angle to the shear direction.



Tiling (imbrication) due to rigid rotation of crystals in a very viscous matrix

<u>Oblique stable position</u>

The rotation rate of porphyroclasts depends on their shape ratio and the sense of rotation depends on both this shape ratio and the initial orientation with respect to the shear plane. For sake of simplicity, objects with long axes at a high angle to the shear plane tend to rotate faster than similar objects initially lying near the shear plane. Clast populations statistically organise themselves near a stable orientation, oblique to the shear direction. The rotation of each porphyroclast stops when the long axis reaches the stable attitude. On the contrary, objects with shape rtios smaller than a critical value rotate permanently. The average inclination of the stable position indicates the sense of shear.

Asymmetric tails

In some cases, concentration of relatively insoluble material is observed diametrically opposed and adjacent to stiff inclusions. These **mats** formed in the quadrants that undergo shortening; their distribution directly reflects the shear sense of the bulk flow in the matrix.

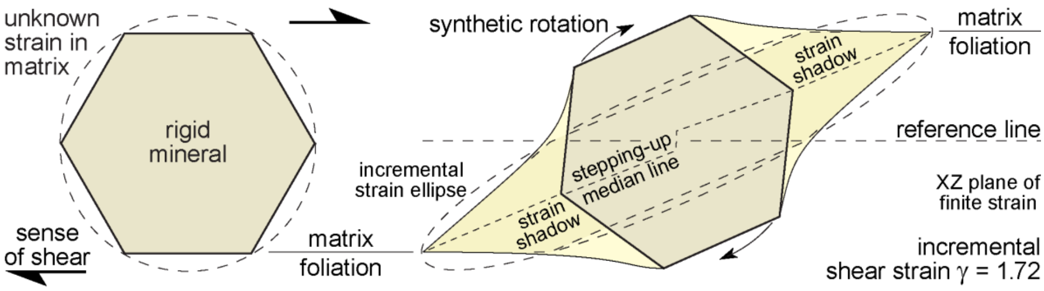
Asymmetric tails

Porphyro-clasts and -blasts in mylonites have **"tails"** of tapering, very fine grain aggregates that are recrystallized from the edges of the clast/blast itself. These tails are attached to the edges of the clast/blast and are installed in the the regions where the general flow, hence the rock foliation, is disturbed by the relatively rigid grain. In effect, rigid objects tend to rotate in their ductile matrix since the bulk deformation is non-coaxial. The disturbed foliation around the objects reflects matrix strain and object rotation. If the tails have the same mineral composition as the neighbouring clast, the central grain+tails system is a mantled, **winged** porphyroclast. Tails consisting of other minerals as the clast or blast are **strain shadows**. The morphology of a pair of tails is defined with respect to a **reference line**, which is the imaginary line parallel to the bulk mylonitic foliation through the centre of the porphyro-clast or -blast in XZ planes of rock. In practice, because of the high shear strain, this line is parallel to the shear zone boundaries, i.e. to the shear plane. The asymmetry is defined by the **stair-stepping** of an imaginary line running along the middle of the tails, called the **median line**, and joining the tails across the reference line through the clast. The sense of asymmetry of the tails defines the sense of shear in the deformed rock.

<u>Winged grain tails</u>

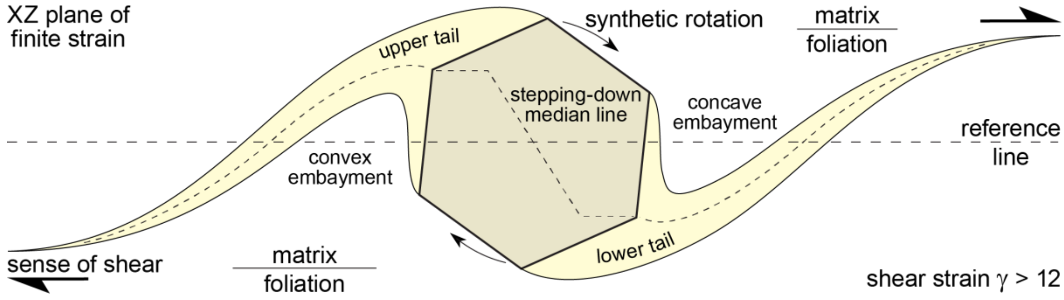
There are two tail morphologies, the **σ-type** and the **δ-type** clast-tail systems.

- On σ–clasts, the tails extend parallel to the foliation and wedge out from each side of the grain in the "downstream" direction of the relative shear in the matrix. The tails are essentially on opposite sides of the reference line. The stepping up direction of the median line defines the sense of shear. This type of clast-tail system characterises recrystallisation rates higher than clast rotation rates.



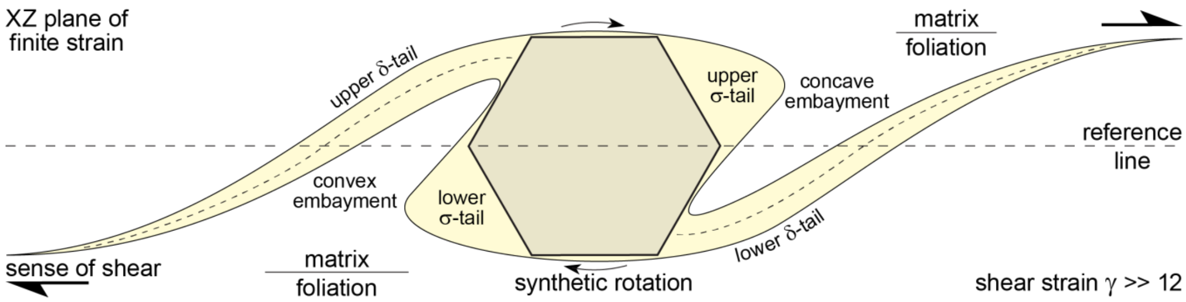
Morphologic features of a sigma (σ-) clast-tail system

- The δ-type is derived from the σ-type by rotation of the clast. The clast entrains and coils the tails in a sense consistent with the bulk shear to produce embayed shapes. Consequently, the folded wings wrap around the clast and cross the reference line in the clast or blast. δ-tails are characteristic of regions of high shear strains where recrystallisation rates are lower than rotation rates.



Morphologic features of a delta (δ-) clast-tail system

- Extreme shear strain may lead to complex structures combining σ- and δ-type tails. δ-and complex $\delta+\sigma$ clast systems are also described as **rolling structures**.



Morphologic features of a complex, delta+sigma (δ+σ) clast-tail system

- Symmetrical tails, classified as θ-type, denote pure shear.

Pressure shadows

Pressure shadows are tails containing mineral(s) different from the rigid porphyroclast ~~in~~. The pressure shadows (or pressure fringes) are formed between foliation planes where rock grains are decoupled from larger, more rigid grains during deformation. The resultant wedge-shaped voids are "shielded" from strain on opposite sides of the clast, where the foliation wraps around the rigid object. Fibrous minerals whose long axis is parallel to the direction of incremental stretching in the matrix are commonly filling pressure-shadow zones. The fibre geometry depends on (1) whether the fibres grew from the rigid grain or from the surrounding rock, (2) whether the fibres deformed during growth and (3) the local strain history during growth.

Asymmetric contact areas

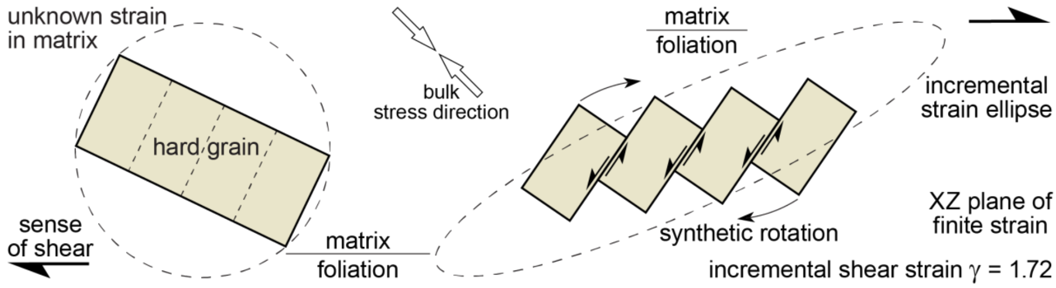
The foliation is more narrowly spaced in compressional than in extensional quadrants. In some cases, concentration of relatively insoluble material is observed diametrically opposed and adjacent to stiff inclusions. These **mats** formed in the quadrants that undergo shortening; their distribution directly reflects the shear sense of the bulk flow in the matrix.

Fractured Objects

Some porphyroclastic minerals, such as mica and feldspar, tend to shear on discrete fractures or crystallographic planes to accommodate ductile deformation in the surrounding matrix. The individual crystal fragments may be rotated in the direction of shear like a collapsing set of books, hence the term "**bookshelf sliding**". However, microfractures may be antithetic or synthetic to the general sense of shear, which makes then litigious sense-of-shear indicators.

Antithetic fractures

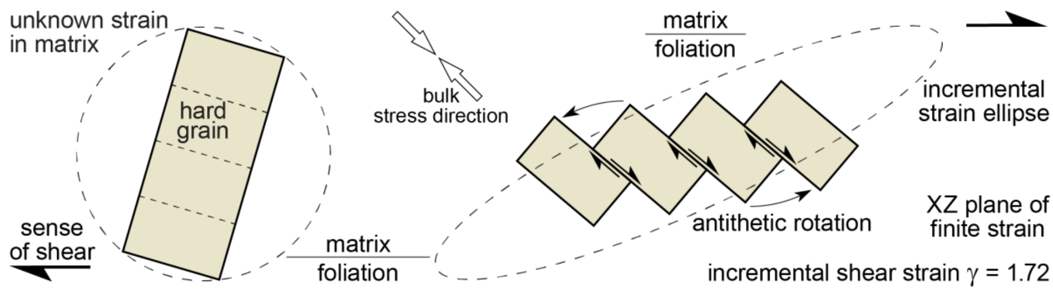
Slip on antithetic fractures is opposite to the bulk sense of shear. This is possible if the fractures or mineral cleavage initially make a high angle with the shear plane but are at an angle to the bulk principal stress that satisfies micro-faulting in a sense opposite to the general shear direction. Mineral fragments between these micro-fractures are thus rotating in consistency with the rotation in the surrounding matrix.



Morphologic features of a bookshelf structure after activation of antithetic microfractures

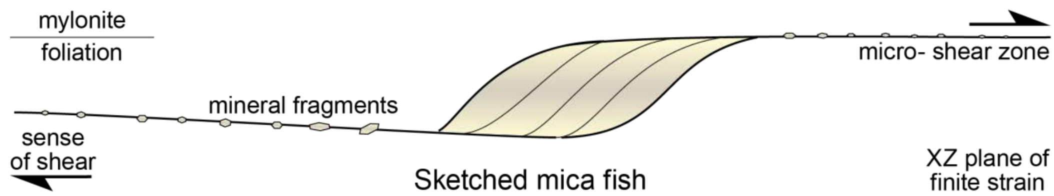
<u>Synthetic fractures</u>

If fractures or mineral cleavage initially make a relatively small angle with the shear plane, then the shear sense on the fractures is the same as it is in the matrix, implying some **back-rotation** of the fragments. Further shearing motion can lead to separation of the individual fragments and **displaced crystals** show displacements consistent with the bulk sense of shear.
Note that clasts and, on a larger scale, boudins may undergo back-rotation in ductile, non-coaxial flow.
Note also that similarity in shape with bookshelf structures does not allow defining a general sense of shear from broken mineral alone. Antithetic and synthetic fractures largely depend on the initial orientation of the clast with respect to the bulk stress direction and are identified once the general shear is established from more reliable kinematic indicators.



Fragmented porphyroclast with displacement of fragments along synthetic microfaults

If crystals possess an easy-glide cleavage, crystal-plastic deformation / brittle deformation / fracturing parallel to this cleavage lead to **mineral fishes.** Mineral fishes (most commonly **mica fishes**) are clastic grains or clusters pinched at their corners and connected by micro-shear zones delineated by strings of minute mineral fragments.



Sketched mica fish

The lens-shaped clasts are oblique to the dominant foliation in XZ planes of finite strain. The sense of shear is indicated by their overall asymmetry and inclination. Their tips and stepping up direction indicate the relative sense of shear, as σ-type tails would.

Recrystallized grains

Crystallographic orientations

Plastic flow of crystals leads to a variety of preferred alignments patterns of the optical axes of the crystals, depending on several parameters among which glide system is active. Asymmetric fabrics may be a useful indicator for strongly deformed rocks. They are mentioned here but deserve a whole chapter by themselves.

Non-simple shear zones

Shear zones were stated to have undergone simple shear.
For obvious reasons, combined thrust or normal simple-shear zones with any strike-slip simple-shear zone will produce a transpressional or transtensional simple-shear zone.

Exercise
Show that adding 2 simple shear deformation matrix yields a simple shear matrix.

However, contractional or extensional pure shear and/or volume change may be combined to simple shear during the same deformation event.
The first obvious consequence is that the angular relationship between foliation and shear zone boundary is no longer 45°.

Exercise
Generate the angular change between foliation S with respect to shear plane in a shear zone with different values of flattening.

Summary

Homogeneous ductile deformation is an unstable response to stress. Under some conditions, small fluctuations in strain can grow rapidly from geometrical, fluid and/or temperature heterogeneities. Consequently, deformation localizes into narrow zones of intense shearing separated by non- or less deforming blocks. Rock continuity and cohesion is maintained across gradients of heterogeneous shear strain in idealized shear zones. Localisation of shear generates fine- grain, recrystallized mylonites that typify ductile fault zones along which the intervening blocks move with respect to each other. Patterning of high strain zones may help defining the bulk strain regime on a regional scale. Small shear fluctuations produce sets of shear bands in sheared rocks; C-shear bands are parallel to the bulk shear plane in rather low shear strain conditions, extensional C'-shear bands where severe localization is reached. The simple shear component of deformation impels asymmetry to a set of structures which, along with relative movement along C- and C' shear bands, provide the local kinematic information.

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