Chapter 15

Shear zones and mylonites

Strain, and shear strain in particular, tends to localize into zones or bands. We have already looked at some types of strain localization structures, such as shear fractures and faults that form in the brittle regime. Localization also occurs in the plastic regime, where foliations and sheared markers tend to show continuity across the zone. Such classic shear zones form an important end-member in a spectrum of shear zones in which both microscale deformation mechanism and ductility vary. On the other end of this spectrum are faults with a measurable thickness. Shear zones can be many kilometers wide, but they also occur on the scale of a hand sample. We will look at shear zones and their internal structure and strain pattern in this chapter, going from a discussion of definitions via the ideal shear zone to different and more complex types of high-strain zones. The last part is devoted to kinematic structures, structures that can reveal the sense of movement in a shear zone, and shear zone growth.

15.1 What is a shear zone?

Faults and shear zones are closely related structures, and Figure 15.1 illustrates the general perception of shear zones as the deep counterpart or extension of faults. Both shear zones and faults are strain localization structures, both involve displacement parallel to the walls, and both tend to grow in width and length during displacement accumulation. Here is a fairly general and simple definition of a **shear zone**:

A shear zone is a tabular zone in which strain is notably higher than in the surrounding rock.

Some would add that a shear zone should contain at least a component of simple shear, but if we want a terminology that is consistent with that of fractures, stylolites and deformation bands (Figure 7.37), a zone of localized coaxial strain also should to be regarded as a shear zone. This opens up a classification of shear zones according to deformation type, for example **pure shear zones**, **subsimple shear zones** and **simple shear zones**.

A shear zone is bounded by two margins or **shear zone walls** that separate the shear zone from its **wall rock** (Figure 15.2). This fact allows us to apply the hanging-wall–footwall terminology from Chapter 8. Once the margins of a shear zone are defined, we can measure its thickness, and sometimes its displacement. Displacement can be directly measured if the zone offsets a marker, and can also be estimated from internal structures in some cases, as discussed later in this chapter.

The above definition of a shear zone is wide, and embraces faults as well as classic **ductile shear zones**, where markers can be traced continuously through the zone. However, there are differences between faults and classic ductile shear zones, in terms of displacement distribution, anatomy and deformation mechanisms, that are worth noting. Let us first make the observation that any shear zone has a thickness that is significant relative to its displacement. Hence, the millimeter-thin cataclasite associated with a single frictional slip surface of several meters offset is too thin to be regarded as a shear zone. More well-developed faults tend to show a two-fold anatomy, with a central high-strain core and a low-strain damage zone (Figure 8.10). We can see from Figure 15.3 that, for a given displacement, fault cores tend to be thinner than ductile shear zones. However, the combined fault core and damage zone thickness corresponds well with the thickness of more ductile shear zones. A difference between faults and ductile shear zones lies in the distribution of strain within the zone, with strain

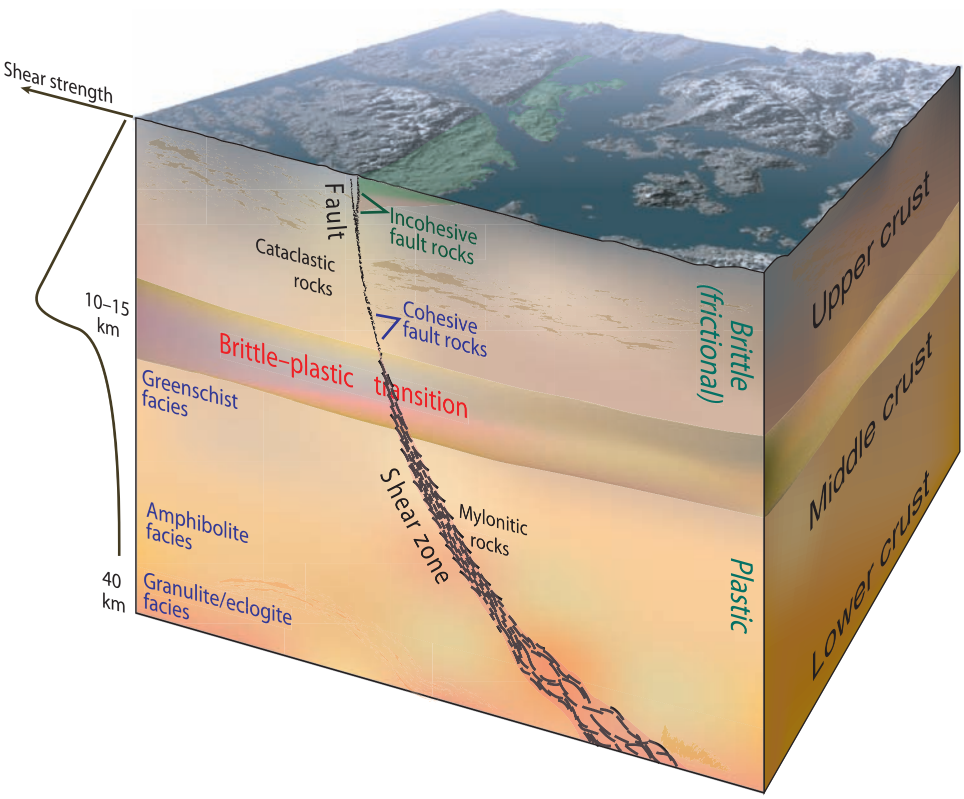

Figure 15.1 Simplified model of the connection between faults, which normally form in the upper crust, and classic ductile shear zones. The transition is gradual and known as the brittle–plastic transition. The depth depends on the temperature gradient and the mineralogy of the crust. For granitic rocks it normally occurs in the range of 10–15 km.

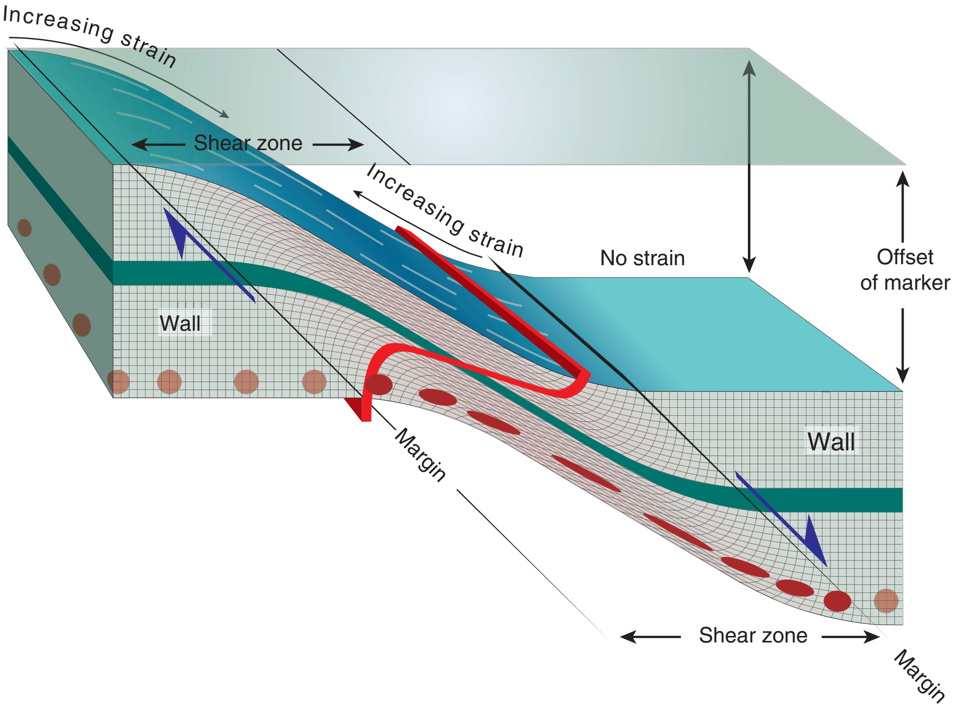

Figure 15.2 Ideal shear zone deforming a grid with two planar markers and circular strain markers. Note how the grid squares change shape and the planar markers change orientation and thickness across the zone. The strain is at its maximum in the central part of the shear zone.

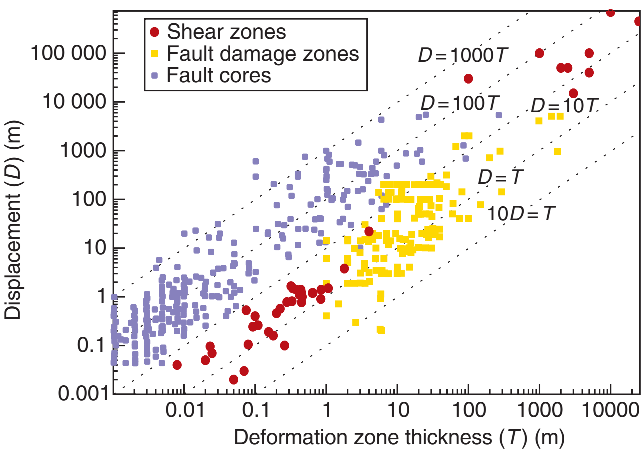

Figure 15.3 Shear zone thickness plotted against displacement for ductile shear zones dominated by plastic deformation mechanisms, faults damage zones/cores, and just fault cores (data from Figure 8.12).

variations being more gradual for ductile shear zones. Another difference is that ductile shear zones can involve both plastic and brittle deformation mechanisms, while faults are totally dominated by brittle mechanisms.

In summary, faults are non-ductile shear zones, forming a subclass of shear zones with its own characteristic features. Other subclasses of shear zones can be defined on the basis of kinematics, microscale deformation mechanisms (plastic or brittle), metamorphic grade, tectonic significance, and so on. Depending on the tectonic regime, shear zones may show normal, reverse, strike-slip or oblique shear, just like faults. Extensional and contractional shear zones tend to be low-angle (<30° dips) while those dominated by strike-slip movement tend to be steep. Shear zones occur at almost any scale in any tectonic regime and form at any depth, although most commonly in the plastic regime.

Kinematic classification

Just like faults and fractures, shear zones can be classified on the basis of relative movement between the two walls, i.e. on the basis of kinematics (Figure 15.4). Although most shear zones tend to be dominated by simple shear, there may also be dilation, compaction and/or pure shear involved. Shear zones in sand and sandstones show the whole range from compaction zones (compaction bands) through simple shear zones (simple shear bands) to dilation zones (dilation bands). Pure shear is also possible in fault cores where clay and sand can more freely be extruded or injected along the fault.

Even though most shear zones may be simple shear-dominated, there is a full 2-D kinematic spectrum from compaction zones via simple shear zones to dilation zones.

Plastic shear zones can show significant deviations from simple shear. However, the ideal shear zone as seen from a strain perspective is one where the deformation is simple shear with or without additional compaction or

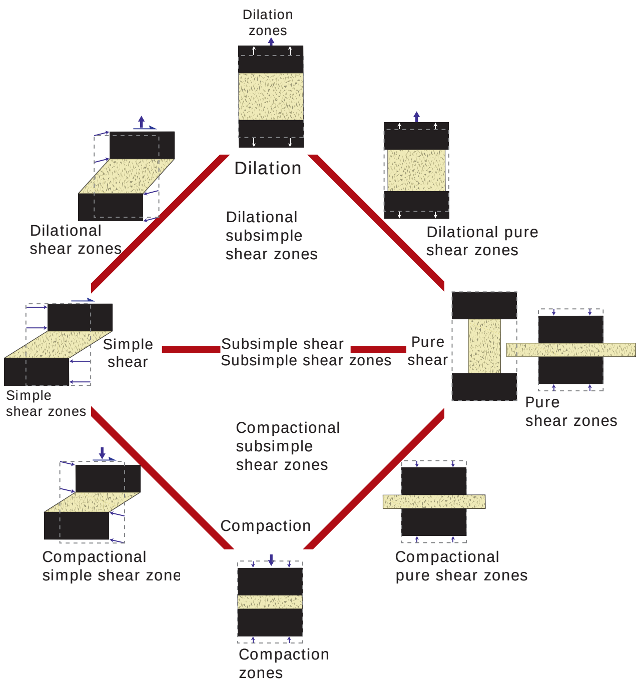

Figure 15.4 Kinematic classification of shear zones (plane strain). Note that pure shear may involve shear zone-perpendicular shortening as well as extension.

dilation. We will have a special look at the ideal shear zone before proceeding with more complex ones. Before that we need to discuss another way of classifying shear zones, which is related to the somewhat ambiguous terms brittle, plastic and ductile deformation.

Brittle versus plastic shear zones

Our general definition of shear zones at the beginning of this chapter makes no restrictions regarding microscale deformation mechanisms. Hence, a shear zone can contain some elements (mineral grains, lenses, layers) that deform plastically and others that deform brittlely at the same time, or everything may deform either plastically or brittlely. The deformation mechanism(s) activated depends on temperature, pressure, metamorphic reactions, cementation, strain rate and amounts of fluid available, in addition to the properties and distribution of rocks and minerals in the zone. It is also worth noting that deformation mechanisms can vary through the life of a shear zone due to changes in physical conditions.

In the overall brittle or frictional regime of the upper crust (Figure 15.1), brittle deformation mechanisms dominate so that deformation in the zone is by cataclastic flow. As discussed in Chapter 7, cataclastic flow involves microfracturing, frictional sliding on grain boundaries and rigid rotation of grain fragments. In sand and poorly cemented sandstone at shallow burial depths, shear zones may develop that deform by granular flow, which involves frictional grain reorganization without grain fracturing. As long as this deformation occurs in a zone of finite width, it is possible to classify the zone as a shear zone, or shear deformation band if occurring on the scale of a hand sample. Shear zones forming predominantly by brittle deformation mechanisms are called **brittle shear zones** or **frictional shear zones**. While an individual slip surface may be too thin to be called a shear zone, more well-developed faults with cores and damage zones can be thought of as end-members in a spectrum of shear zones, even though most of us give preference to the term fault or fault core when referring to brittle discontinuities that sharply cut off structures in the wall rocks.

Deeper in the crust, in the plastic regime, the plastic deformation mechanisms discussed in Chapter 10 come into effect. Where plastic deformation mechanisms dominate we get **plastic shear zones**. In the brittle–plastic transition zone, which can be quite wide if polymineralic rocks such as granites are involved, **brittle–plastic shear zones** form in which both brittle and plastic deformation mechanisms are important. **Semi-brittle shear zone** is another commonly used term in this context, applied to brittle shear zones influenced by plastic deformation mechanisms. Most plastic shear zones contain some brittle elements, such as fractured feldspar or garnet porphyroclasts, unless the temperature is very high.

Ductile shear zones

The term ductile shear zone is a popular term, but also an ambiguous one, because some geologists use the term ductile to imply plastic deformation mechanisms. Equating ductility with plasticity is not recommended, even though many ductile shear zones are indeed plastic. Instead we relate the term ductility to continuity of originally continuous marker elements in the shear zone. A **perfectly ductile shear zone** contains no internal discontinuities, so that marker layers crossed by the shear zone can be traced continuously through the zone at the mesoscopic scale (Figure 15.2). This type of deformation is sometimes named continuous deformation or continuous strain, and most plastic and some brittle shear zones preserve continuity of passive markers. The shear zone shown in Figure 15.5 is an example of a ductile shear zone formed by brittle deformation mechanisms at near-surface conditions. Hence, a ductile shear zone can deform by means of brittle as well as plastic microscale deformation mechanisms.

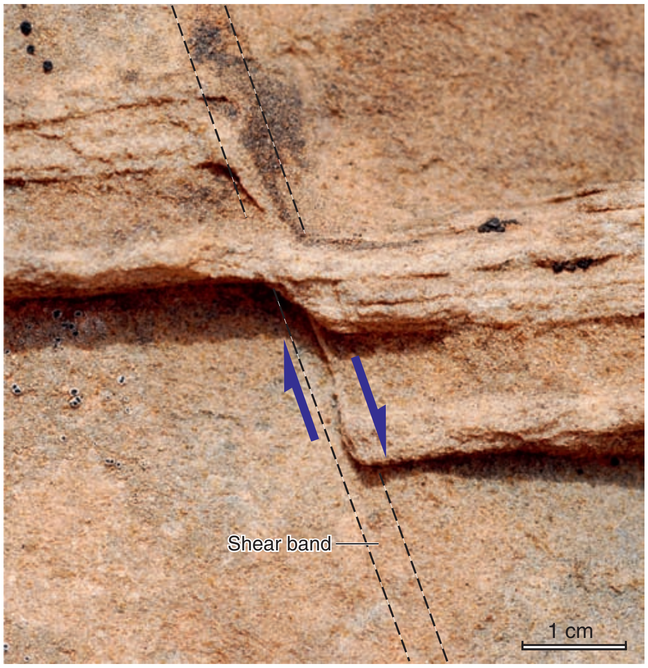

Figure 15.5 Thin shear zone (disaggregation band) in the Navajo Sandstone (Arches National Park, Utah) showing continuity of sand laminae across the zone. The shear band formed prior to lithification.

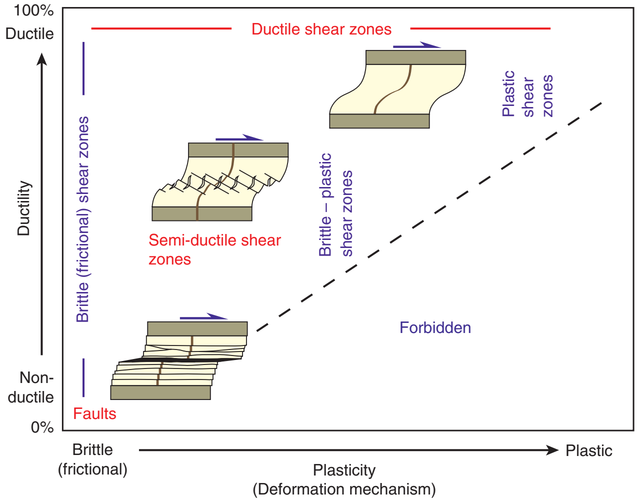

Figure 15.6 Simple classification of shear zones based on deformation mechanism (horizontal axis) and mesoscopic ductility (continuity of markers).

Passive markers can be traced continuously through a perfectly ductile shear zone.

Field observations show that many shear zones, including those dominated by plastic deformation, show internal sharp discontinuities in the form of slip surfaces, extension fractures, veins or pressure solution seams. Such shear zones can be termed **semi-ductile**. Discontinuities are found in many plastic shear zones, but are more characteristic for brittle (frictional) shear zones.

Many faults show little or no ductility, with a very rapid transition from nearly unstrained wall rock to intensely sheared breccia, fault gouge or cataclasite in the fault core. However, other faults show fault drag along the core, and thus display a combination of continuous and discontinuous deformation. A fault-propagation fold ahead of a fault can be considered as a ductile shear zone (Figure 8.30 and Box 8.3). Where or when the fault enters the fold, the structure becomes affected by discontinuous deformation and becomes semi-ductile. Hence the deformation in a shear zone can change in space and time from being continuous (ductile) to (semi-)discontinuous.

In summary of what has been discussed in the last two sections, shear zones can be classified according to ductility (continuity of markers) and plasticity (degree of plastic versus brittle deformation mechanisms). This is illustrated in Figure 15.6, which indicates that the term brittle shear zone may have two meanings, either frictional deformation at the microscale (no plasticity), the use that is preferred here, or complete discontinuity of structures (no ductility). In both understandings of the term, brittle shear zones plot in the lower left corner of the diagram shown in Figure 15.6.

15.2 The ideal plastic shear zone

The **ideal shear zone** is limited by two perfectly planar (straight in cross-section) boundaries separating it from completely undeformed wall rocks (undeformed by the shear-zone forming deformation – an earlier fabric may exist). Ideal shear zones are also ductile, so that slip surfaces or other discontinuities are non-existent. In ideal plastic shear zones, which we will focus on in this section, we can see a foliation, a lineation and evidence of strain variations that contain important information about the zone.

For closer analysis, a coordinate system is placed with the horizontal axis along the shear zone margin and parallel to the shear direction, as shown in Figure 15.7. We can now move the walls while keeping the distance between them constant. The result is a perfect simple shear (Figure 15.8b). If we wish we could also increase or decrease the distance between the walls by adding a component of dilation or compaction. This would result in a change in volume (Figure 15.8d) that can take place before, during or after the simple shearing; the end result will, in principle, be the same.

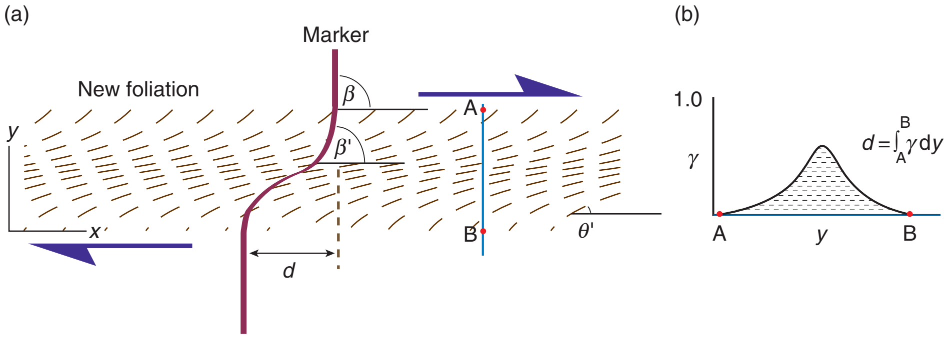

Figure 15.7 (a) Shear zone with genetically related foliation. The foliation makes ∼45° with the shear zone along the margins. This angle is reduced as strain increases toward the center of the zone. θ' is the angle between the shear zone and the foliation. (b) The displacement can be found by measuring or calculating the area under a shear strain profile across the zone if the deformation is simple shear.

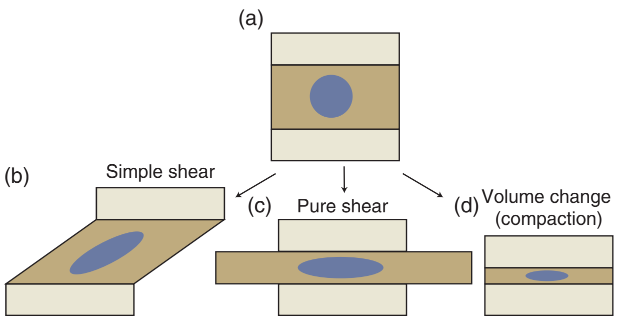

Figure 15.8 The difference between simple shear, pure shear and isotropic vertical volume change.

Ideal shear zones are perfectly ductile and involve simple shear with or without additional compaction/dilation.

Any deformation that deviates from simple shear (with or without additional compaction or dilation across the zone) requires discontinuities to form and thus contradicts the restrictions set for an ideal shear zone (Figure 15.8c). Such deviating deformation will cause extrusion of material along the zone and create structural incompatibilities. By incompatibility we mean that different parts of the shear zone do not fit together in a continuous framework, causing overlaps, gaps or discontinuities to arise. At the advanced level, tests of compatibility are necessary steps in the analysis of strain in shear zones.

Classic ideal plastic shear zones are found in more or less isotropic rocks, particularly in magmatic rocks such as the one shown in Figure 15.9. In these cases the deformation matrix for simple shear (Equation 2.16), or the one for combined simple shear and dilation, can be used to model the deformation. Using these matrices, shear strain, shape and orientation of the strain ellipse and displacement can be calculated at any point in the zone. Let us look at the structures forming within ideal shear zones and what they can give us.

Figure 15.9 Ductile shear zone in greenstone where feldspar aggregates (amygdales) get progressively more strained into the shear zone (downward). White line indicates how the shear zone-related foliation rotates toward parallelism with the zone. Photo: Graham B. Baird.

Foliation and strain

Characteristic structures develop as displacement and strain accumulate in plastic shear zones. These structures have orientations and geometries that depend on both the type of shear zone and the amount of strain. If a shear zone develops in a fairly isotropic rock, then a faint foliation will appear at low shear strains. We can see such a faint foliation along the margins of the shear zones portrayed in both Figures 15.9 and 15.10. The angle θ

between this initial foliation and the margins will be close to 45° for a simple shear zone. In most cases it will be slightly less, because it takes a certain strain to form a visible foliation, and during that first strain increment the foliation will rotate a certain amount.

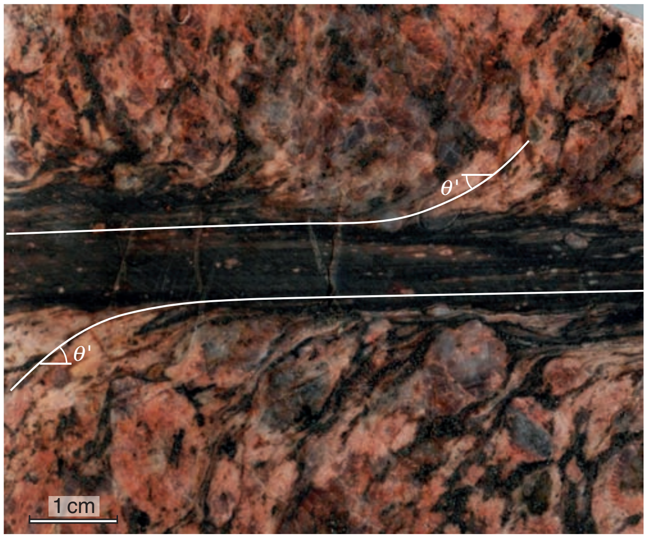

Figure 15.10 A shear zone in the Diana Syenite in Harrisville, New York State, showing profound strain increase toward the central ultramylonitic part of the zone. This is expressed by the change in orientation of the foliation and a marked decrease in grain size in the zone. Photo: Graham B. Baird.

The faint initial foliation is also seen along the low-strain shear zone margins of a well-developed shear zone, and the angle it makes with the zone (now termed θ') decreases as we trace the foliation into the more strained central parts of the zone. The foliation initiates perpendicular to the fastest shortening direction (ISA_3), but because simple shear is non-coaxial the foliation will constantly rotate toward parallelism with the shear plane. For low and moderate shear strains (up to 10–15) the foliation is a flattening foliation that represents the orientation of the *XY*-plane of the strain ellipse. This close relationship between the orientation of the shear zone foliation and strain can be very useful, and helps us to map strain in shear zones. For a simple shear zone, where θ' denotes the angle between the foliation and the shear zone, the relationship is given by the formula

$$\theta' = 0.5 \tan^{-1}(2/\gamma) \tag{15.4}$$

Note that this relationship only holds for a simple shear zone where the foliation under consideration formed during the formation of the shear zone in a previously unstrained rock, and in a section perpendicular to the foliation in the direction of shear.

In some cases the shear zone contains strain markers so that the aspect ratio $R = X/Z$ of the strain ellipse can be estimated. We can then relate the orientation of the foliation to both the shape of the strain ellipse and γ, as shown graphically in Figure 15.11.

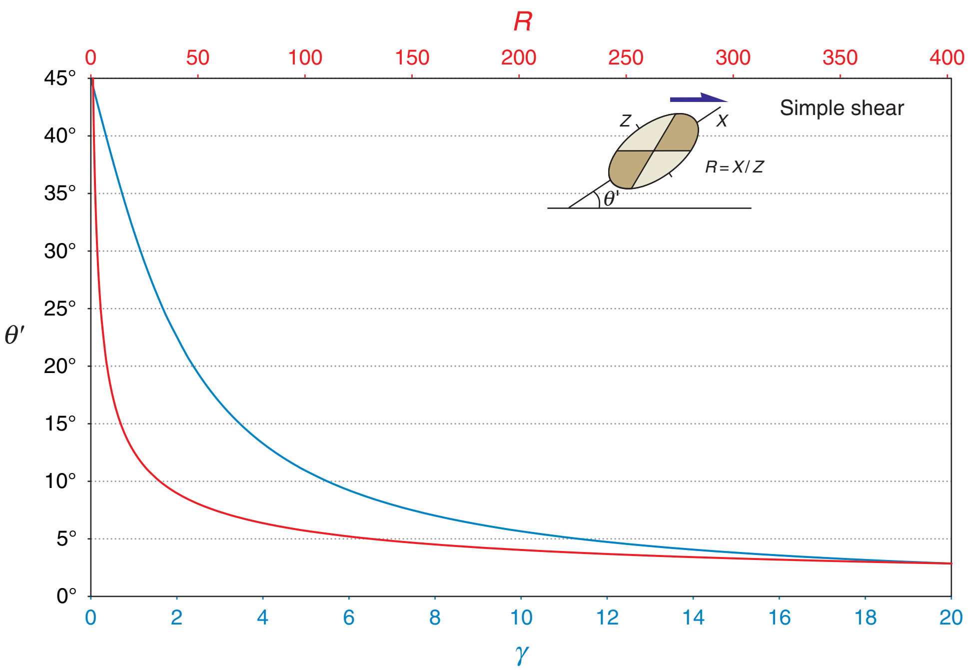

Figure 15.11 Curves showing how the angle θ' between the maximum finite strain axis (X) and the shear plane decreases with increasing strain for simple shear. Blue curve represents shear strain (γ) and red curve represent the aspect ratio $R = X/Y$ of the strain ellipse. Note different scales for γ and R (top and bottom).

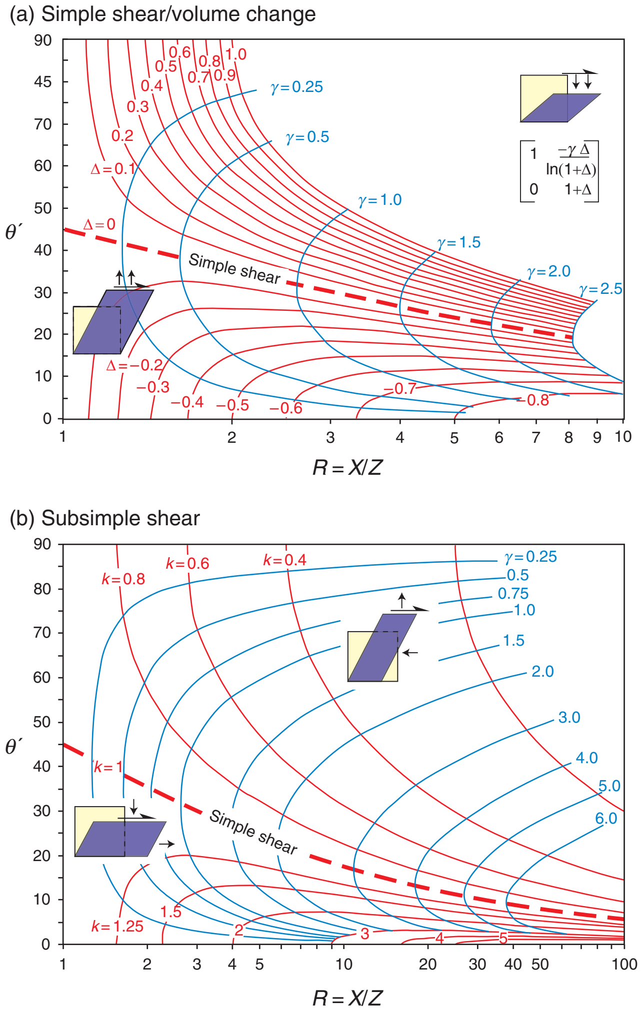

Figure 15.12 R–θ' plot for (a) simple shear and compaction and (b) simultaneous simple and pure shear (both plane strain). The simple shear curve starts at $\theta = 45°$. Note that the horizontal axes are logarithmic.

If there is a component of compaction (or dilation) across the simple shear zone, then even the faint foliation along the margins will make an angle θ with the shear zone that is less (or more) than 45°. How much less depends on the amount of compaction. This means that we cannot use Figure 15.11 if the shear zone has thinned by compaction, but expand it with contours accounting for the compaction or dilation, as shown in Figure 15.12a.

There is a simple relationship between shear strain, the orientation of the foliation and strain in an ideal shear zone.

Lineations

A stretching lineation develops along with the foliation in plastic shear zones. Ideally, in a simple shear zone the lineation indicates the X-axis of the strain ellipsoid. The stretching lineation lies in the foliation surface and thus defines the same angle θ' with the margin or shear plane as does the foliation (Figure 15.11). The projection of the lineation onto the shear plane will indicate the shear direction (Figure 15.12), which is also the flow apophysis for the simple shear.

Offset and deflection of passive markers

There is a connection between the total offset and finite strain in a shear zone. For simple shear this connection can be portrayed graphically by a shear strain profile across the zone. The length of the curve will be the thickness of the zone and the shear strain value will represent strain at any point across the zone. By integrating across the zone or by finding the area under the shear–strain curve, the total offset d is found (Figure 15.7b).

Sometimes shear zones are found to cross markers such as dikes, veins and layering (Figure 15.7a). Such markers show the offset across the shear zone directly, and the result can be compared to the strain integration method illustrated in Figure 15.7b. Remember that the strain integration method assumes that the XY-plane is represented by the foliation and that the shear zone represents simple shear deformation. A significant discrepancy between the two methods indicates that the deformation deviates from simple shear. Perhaps it can be explained by additional shortening or dilation across the zone, or perhaps other deformation types are involved. Before going into other types of deformation we will explore passive rotation of markers that cross a simple shear zone.

A line marker that is initially perpendicular to the shear zone is easy to handle, since its angle of rotation is the angular shear. More generally, if a marker originally makes an angle β to the shear zone (Figure 15.5a) then its new angle β' after the deformation depends on the type of deformation in the zone. For simple shear the relationship between these angles and the shear strain γ is given by the formula

$$\gamma = \cot\beta' - \cot\beta \qquad (15.5)$$

Remember that $\cot\beta = 1/\tan\beta$. We could also use the deformation matrix **D** to find the new orientation of markers for appropriate matrixes **D**. In a section (two-dimensional analysis) the marker can be represented by a unit vector **l**, and the new orientation becomes

$$\mathbf{l}' = \mathbf{D}\mathbf{l} \qquad (15.6)$$

where $\mathbf{l}'$ is the vector representing the new orientation. For simple shear $\mathbf{D}$ is the simple shear matrix from Equation 2.16. It is then easy to show that if $\mathbf{l} = (x, y)$, then its new orientation becomes $\mathbf{l}' = (x + \gamma y, y)$:

$$\mathbf{l}' = \begin{bmatrix} 1 & \gamma \\ 0 & 1 \end{bmatrix} \begin{bmatrix} x \\ y \end{bmatrix} = \begin{bmatrix} x + \gamma y \\ y \end{bmatrix} \tag{15.7}$$

For deformation types other than simple shear, different matrices apply.

One of the advantages of using matrix operations is that any line orientation (not just those in the plane perpendicular to the shear plane and along the shear direction) can easily be handled. In such three-dimensional analysis, planes are treated by means of their poles. In this case the strain takes the normal vector $\mathbf{p}$ to a new orientation represented by the vector $\mathbf{p}'$ according to the equation

$$\mathbf{p}' = \mathbf{p}\mathbf{D}^{-1} \tag{15.8}$$

Since line and plane rotations are determined by the deformation type, and hence the deformation matrix $\mathbf{D}$, we can easily model how lines and planes rotate during the deformation history by means of incremental strain. This shows that simple shear produces rotations along great circles, which differ from those caused by other deformation types (Figures 2.30 and 18.21). Passive linear and planar structures that are increasingly deflected toward the center of a shear zone can be plotted and their rotational paths may reveal important information about the deformation type. In practice such analyses are done by plotting the orientation of markers outside and inside differently strained parts of the shear zone, and then comparing to numerically modeled rotation paths using Equations 15.6 and 15.8.

Before moving to more general shear zone types we will summarize some facts about simple shear zones, compaction zones and the combination of the two.

Simple shear zones

A simple shear zone has the following characteristics:

Plane strain with $W_k = 1$.
No shortening or stretching along or normal to the zone.
ISA_1 oriented at 45° to the shear plane (walls).
The strain ellipsoid X-axis initiates at 45° to the walls and progressively rotates toward parallelism with the shear plane according to the formula $\theta' = 0.5 \tan^{-1}(2/\gamma)$.
Knowing θ', the local shear strain is given by $\gamma = 2/\tan(2\theta')$.
Passive markers at an initial angle β to the shear direction obtain a new angle β' after the deformation according to the formula $\cot \beta' = \cot(\beta) + \gamma$.
The vorticity (w) is identical to the shear strain rate: $w = \dot{\gamma}$.

Dilation/compaction zones

A compaction zone or dilation zone is a deformation zone that has experienced pure compaction or dilation perpendicular to its walls (Figure 15.8d). Such deformation zones form end-members in the spectrum of shear zone kinematics illustrated in Figure 15.4 (left-hand side) and are therefore useful to explore. Thick zones of pure compaction or dilation are rare, but millimeter- to decimeter-thick zones occur in porous rocks and are a special type of deformation band known as compaction/dilation bands. For such zones we have:

Planar and non-coaxial deformation with $W_k = 0$.
Shortening or extension perpendicular to the zone.
ISA oriented parallel and perpendicular to the zone.
X initiates and remains parallel to the zone.
Passive markers at an initial angle β with the shear vector end up at a new angle β' according to the formula:

$$\cot \beta' = \frac{\cot \beta}{1 + \Delta} \tag{15.9}$$

where Δ is the volume change across the zone (see Section 2.12).
Vorticity $w = 0$.

Dilational/compactional shear zones

Shear zones that involve a combination of simple shear and dilation or compaction are known as **dilational** and **compactional shear zones**. Such zones, where simple shearing and dilating/compacting are simultaneous, are probably common and are characterized by:

Plane strain with $0 < W_k < 1$.
Generally non-coaxial deformation.
Shortening (compaction) or extension (dilation) perpendicular to the shear zone.
No shortening or extension along the shear zone.
ISA_1 oriented obliquely to the shear zone (<45° for compactional shear, >45° for dilational shear) (Figure 15.12a).

X forms obliquely to the zone and rotates according to the formula

$$\tan 2\theta' = 2\frac{(1+\Delta)\Delta\gamma/\ln(1+\Delta)}{1-[\Delta\gamma/\ln(1+\Delta)]^2-(1+\Delta)^2} \quad (15.10)$$

For compactional shear $\theta < 45°$ and θ' is always less than for simple shear for a given shear strain. For dilational shear zones the orientation of X can show a range of values, depending on strain and amount of dilation, but always with a higher θ than simple shear for a given shear strain.

Passive markers at an initial angle β with the shear direction end up at a new angle β' according to the formula

$$\cot\beta' = \frac{\cot\beta + \Delta\gamma/\ln(1+\Delta)}{1+\Delta} \quad (15.11)$$

All of these formulas can be extracted from the deformation matrix.

15.3 Adding pure shear to a simple shear zone

Many deformation zones deviate from the conditions of ideal shear zones: walls may be non-parallel, slip surfaces or other sharp discontinuities may occur within or along the zone, walls may be deformed and displacement may vary along the zone. A shear zone that deviates from the ideal shear zone model in any of these ways is called a **general shear zone**.

A general shear zone is one that deviates from the ideal ductile shear zone.

Few natural shear zones have perfectly parallel and planar walls because of lithologic heterogeneities or variations in the internal conditions of the zone during deformation. Non-parallel and curved walls create local deviations from the simple shear flow pattern.

Sharp discontinuities or slip surfaces are seen in many shear zones. Such internal discontinuities allow for a pure shear component to occur. If the shear zone walls are undeformed, a pure shear with shortening across the shear zone will thin the zone, and material will extrude along the shear zone as shown schematically in Figure 15.8c. On the contrary, if the pure shear component involves shortening parallel to the zone, the zone will widen. Flow of rock into the widening part of the zone can then occur.

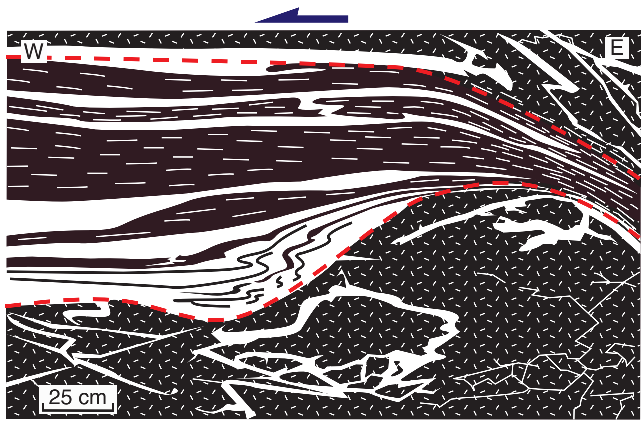

Figure 15.13 Folds formed on the lee side of a tectonic lens. Note the lee side thickening, the thinning on the opposite side and the vergence of the folds. Red stippled lines outline the shear zone. Modified from Fossen and Rykkelid (1990).

We expect to see layer thickening, imbrication and folding of layers where a shear zone widens. Local thinning of shear zones is associated with thinning of layers and possibly the formation of extensional shear bands. Figure 15.13 shows both effects: thinning of layers where the shear zone thins on the left side of the rigid lens, and thickening and folding on the lee side of the lens (central part of the figure). Furthermore, along-strike variations in coaxial deformation can lead to a shear zone of varying thickness and with internal slip surfaces of local extent. An example is shown in Figure 15.14, where a predominantly plastic shear zone changes character from completely ductile in the lowermost part to semi-ductile in the central and upper part.

If frictional slip occurs in shear zones, a slip-related lineation can form. Such lineations form on slip planes, which may or may not be parallel to the shear zone, and should not be confused with the stretching lineation illustrated in Figure 15.15.

Slip surfaces and other brittle expressions in predominantly plastic shear zones commonly form in shear zones that exhume and cool during shearing. They can also occur in the overall plastic regime of the lower crust for reasons other than changing temperature and pressure. Slip can occur where platy metamorphic minerals align to form a mechanically weak plane and where strain rate locally and temporarily increases. Narrowing shear zone walls or disturbance of the flow by a large rigid inclusion in the zone may be what it takes to increase the local shear-strain rate. External conditions forcing the shear zone to accumulate displacement at a higher rate are also possible. Finally, the role of fluids is important.

Figure 15.14 Plastic shear zone with a local discontinuity along its left margin, suggesting that the deformation may deviate from simple shear.

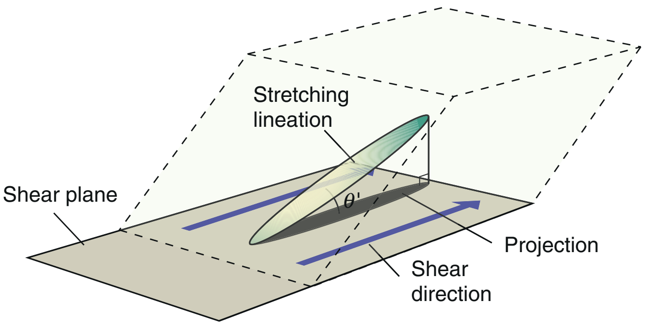

Figure 15.15 Illustration of the role of the stretching lineation in a plane strain shear zone. Ideally, its projection onto the shear plane indicates the shear direction. Deviations occur if there is a viscosity contrast between the linear elements and the matrix.

Dry rocks are much more prone to respond to stress in a brittle manner, so variable "wetting" of the deforming rock can control its rheologic behavior even in the lower crust. Hence, several factors can control the formation of transient sharp discontinuities in shear zones. It is also worth noting that transient formation of slip surfaces, pseudotachylyte and other brittle elements during otherwise plastic deformation implies that such structures are

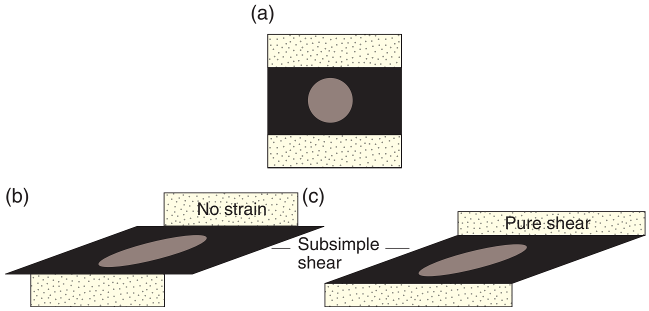

Figure 15.16 A shear zone that involves coaxial strain (b) has compatibility problems along its margins that are solved if the wall rocks also deform by the same amount of coaxial strain (c).

likely to be obscured by subsequent plastic shearing and metamorphic recrystallization.

Some shear zones have walls that were coaxially strained during the shearing in the zones. In this case strain compatibility is preserved if the coaxial strain is homogeneous throughout both the shear zone walls and the shear zone (Figure 15.16c). This means that there will be no need for the discontinuities seen in Figure 15.8c and 15.16b – the walls and the zone will stretch together (Figure 15.16c).

Subsimple shear zones

It is obvious from the discussion above that a whole spectrum of plane strain shear zones exists where simple shear and pure shear are combined simultaneously in the same plane. We know this type of deformation from Chapter 2 as subsimple shear, and such shear zones are therefore called **subsimple shear zones**.

Subsimple shear zones cover the plane strain spectrum of simultaneous simple shear and pure shear.

For such zones we will see an initial angle θ between the foliation and the margin that is $>45°$ if the zone is thickening, and $<45°$ where the pure shear causes the zone to thin and extend. The close relationship between θ', strain and the type of deformation can be used to analyze natural shear zones. This is done by plotting the orientation of the foliation against strain in diagrams such as Figure 15.12b. This particular diagram is constructed for subsimple shear, so if there is additional volume change the diagram changes. In any event, data that plot off the simple shear curve must be explained by means of additional volume change or coaxial strain. To separate the two, evidence for or against volume change

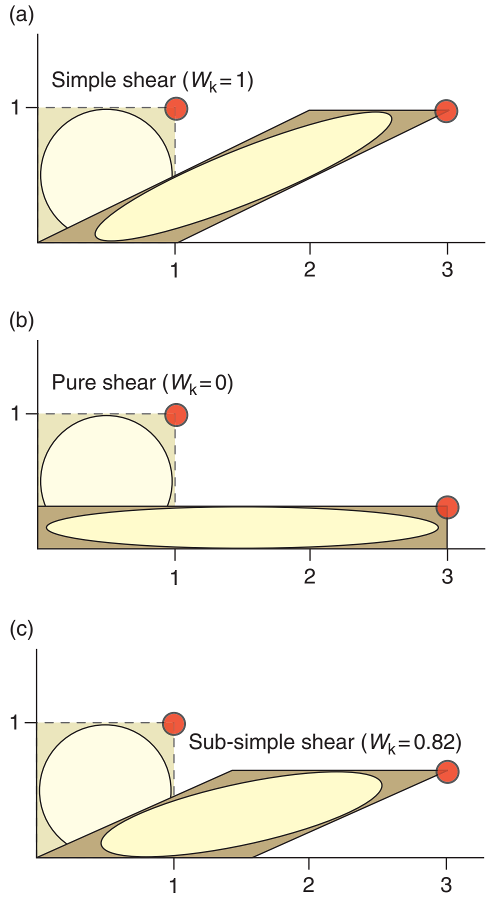

Figure 15.17 Comparison of shear-zone parallel offset and strain. A square with a circle is deformed under simple shear, pure shear and subsimple shear so that the horizontal displacement of the upper right corner is the same in all cases. The resulting strain is greatest for pure shear and least for the chosen subsimple shear. Based on Fossen and Tikoff (1997).

should be searched for. Such evidence could be stylolites, vein material, new minerals and concentrations of immobile minerals in the shear zone.

Stretching lineations in subsimple shear zones will, as for simple shear zones, reflect the shear direction when projected onto the shear plane (Figure 15.15). Note that if the simple shear component is small or zero the lineation indicates the stretching direction caused by pure shear, and can no longer be called the shear direction.

Let us take a closer look at the connection between strain and offset. Consider the upper right corner (1, 1) of a quadrant in a shear zone, as illustrated in Figure 15.17. If we apply a simple shear so that this corner is displaced to a new position (3, 1), then this offset will correspond to a strain ellipse with a particular axial ratio (Figure 15.17a). If instead we apply a pure shear to obtain the same horizontal displacement of this corner, then we will find that the pure shear has generated a higher strain (compare the ellipses in Figure 15.17a, b). Some subsimple shear deformations require less strain than simple shear to produce the same shear-zone parallel displacement (Figure 15.17c). In fact, the subsimple shear that requires the least strain is the one with $W_k = 0.82$. This means that a subsimple shear of $W_k = 0.82$ is the most strain-efficient way to create shear-zone parallel offset.

It takes less strain to reach a given offset by means of subsimple shear than by pure or simple shear.

It is worth noting the nature of the horizontal displacement along the zone for different deformation types. For simple shear the displacement across the zone is constant along the zone. For subsimple and pure shear, however, the displacement changes along the zone.

15.4 Non-plane strain shear zones

We have so far been discussing shear zones from a plane strain perspective, in which any coaxial (pure shear) strain acts in the same plane as the simple shear. Any plane strain (without anisotropic volume change) produces strain ellipsoids that plot along the diagonal of the Flinn diagram. However, strain measurements in many shear zones show significant deviations from plane strain (Figure 15.18), implying extension or shortening related to a coaxial and/or simple shear component in the third direction (Y-axis). This opens up a wide class of general shear zones that produce three-dimensional strain. In such general shear zones, foliations and lineations may form with new orientations and rotate in different ways, and lineations do not relate to the direction of shear in a simple way.

Non-plane strain shear zones have flattening or constrictional strain geometries and lineations that may not represent the transport direction.

It is important to find out if a given shear-zone strain is plane or not. We could look for field evidence of thickness variations and flow in more than one direction. If we have strain markers, such as in the sample shown in Figure 15.9, we can section it and plot the strain data in the Flinn diagram, as done in Figure 15.18. The data in this figure come from shear zones of the type shown in Figure 15.9, and even though this shear zone may look like a simple shear zone at first glance, the strain data shown in Figure 15.18 reveal non-plane strain. This

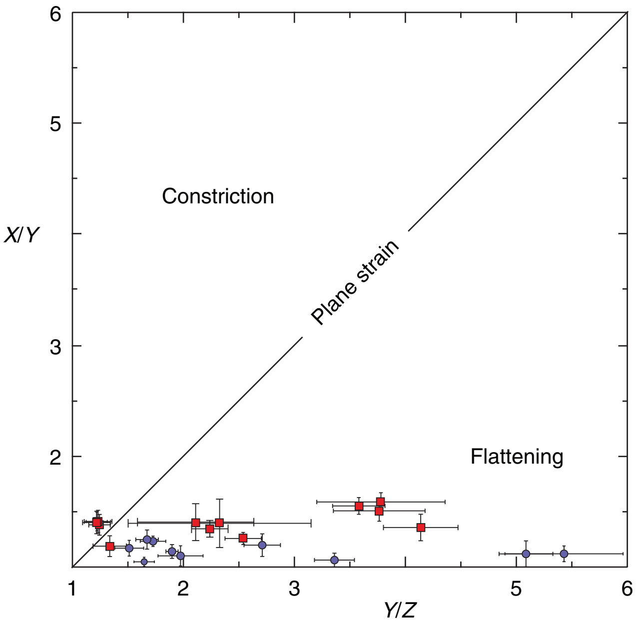

Figure 15.18 Flinn diagram with strain estimations from two shear-zone samples similar to that shown in Figure 15.9, cut parallel and perpendicular to the shear direction. The strain data clearly tell us that the two shear zones are affected by flattening and therefore deviate from simple shear. Data from Bhattacharyya and Hudleston (2001).

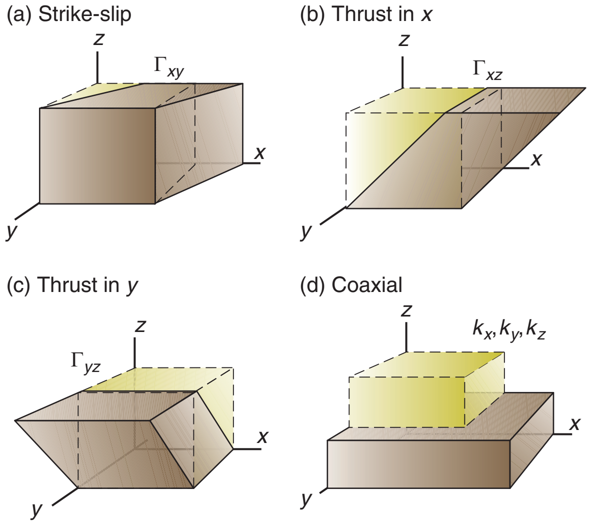

Figure 15.19 Orthogonal deformation components that can be combined into the three-dimensional strain matrix of Equation 15.12. Based on Fossen and Tikoff (1997).

means that the shear zone shown in Figure 15.9 cannot be a simple shear zone.

If we find that our shear-zone strain is three dimensional we are challenged by the fact that three-dimensional considerations open up a wealth of possible deformation types. We can still think in terms of pure and simple shear, but now we need to combine elements of simple and pure shear in more than one plane. Figure 15.19 shows how we can combine a three-dimensional coaxial strain with up to three orthogonal simple shear systems. The deformation matrix then becomes three-dimensional:

$$\begin{bmatrix} k_x & \Gamma_{xy} & \Gamma_{xz} \\ 0 & k_y & \Gamma_{yz} \\ 0 & 0 & k_z \end{bmatrix} \tag{15.12}$$

This matrix can be used to model such deformations, but calculations generally require computer programs. There are also cases where the coaxial strain acts at an oblique angle to the shear component(s), in which case things get more complicated. The many possibilities make three-dimensional analyses of non-plane strain shear zones challenging, and we usually look for the simplest model that can explain our data. Transpression and transtension are two quite popular and closely related classes of simple three-dimensional shear-zone deformations, discussed in the last part of Chapter 18.

15.5 Mylonites and kinematic indicators

Mylonites

In the central parts of some plastic shear zones, for example the one shown in Figure 15.10, strain can get so high that preexisting textures and structures are totally flattened and transposed. The rock becomes strongly banded and is called a **mylonite** – a word that was coined by earlier workers in the Scottish Moine Thrust Zone (see Box 15.1) The characteristics of mylonites vary with temperature, pressure, mineralogy, grain size, presence of fluids and strain rate. In general, mylonites are more fine-grained than their host rock (very apparent from Figure 15.10), with well-defined foliation and lineation.

Mylonite zones or belts can be up to several kilometers thick, particularly in Precambrian shield areas and eroded collisional orogens. In these cases the walls and shear plane may be difficult to define, and there will be many heterogeneities within the zone. Not only will finite strain vary, with lenses and slivers of less-deformed rocks, but also the deformation type will vary across the zone. The result can be a zone containing a rich variety of deformation structures such as folds, cleavages, foliations, lineations and other structures discussed later in this chapter. It is still a shear zone in the general sense of the word, but certainly more complicated and varied than ideal shear zones dealt with in the first part of this

BOX 15.1 MYLONITE VERSUS CATACLASITE

The term **mylonite** had at least two different meanings through the last century. The word stems from the Latin word for milling or crushing into fine pieces. This was also the process that once was assumed to have formed the mylonites along the famous Scottish Moine Thrust Zone – the place where the term mylonite was first applied. With the great aid of optical and electron microscopes, we now know that mylonites are formed mainly by plastic deformation mechanisms.

The term mylonite is now being used for strongly deformed rocks that have been exposed to grain size reduction due to plastic deformation, while the related term **cataclasite** is used where cataclastic flow dominates. Some cataclasis can occur during mylonitization, for instance when feldspar is crushed in a matrix of plastically deforming quartz.

Mylonites are separated into three subgroups, depending on how much of the original groundmass is still intact (not recrystallized):

Protomylonite: <50% matrix (new grains)
Mylonite: 50–90% matrix
Ultramylonite: >90% matrix

Mylonites are common in thrusts, extensional shear zones and steep basement shear zones.

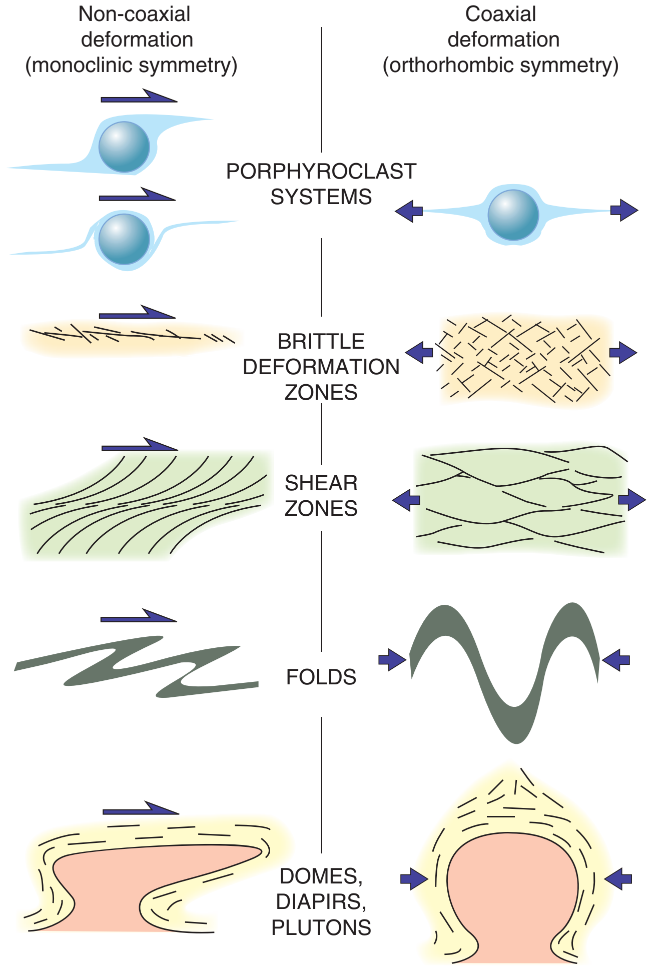

Figure 15.20 Asymmetric structures (left side) characterize non-coaxial deformations, while coaxial deformations tend to result in more symmetric structures. Based on Choukroune *et al.* (1987).

section. This makes thick mylonite zones both exciting and challenging.

Strongly sheared rocks of the kind found in high-strain shear zones have been defined in several ways through the past century, and the most common classification is shown in Boxes 8.1 and 15.1. This classification distinguishes between protomylonites, where original grains still dominate (90–50%), mylonites (50–10% original grains) and ultramylonites, which have less than 10% of the original grains intact. This change from protomylonite through mylonite to ultramylonite is recognizable in many shear zones, but is made more difficult if rocks have been exposed to previous phases of strong deformation.

Remains of the original texture or original minerals can be found as large lenses or fragments wrapped in the mylonitic foliation. Where such lenses are tens of centimeters or more in length they are called **protolithic lenses**. Fragments of single crystals or minerals are called **porphyroclasts**. Feldspar typically forms porphyroclasts in deformed granitic rocks during greenschist facies shearing, where temperature is too low for feldspar to deform by crystalloplastic deformation mechanisms.

Kinematic indicators

Our understanding of microstructures as kinematic indicators in shear zones and mylonite zones increased considerably during the 1970s and 1980s. Understanding the connection between structural asymmetry and kinematics represented an important breakthrough in the study of strongly sheared rocks. The key point is that many mylonites contain structures that show monoclinic (low) symmetry, simply referred to as **asymmetric structures** by most geologists. The asymmetry is related to the rotational component or non-coaxiality of the deformation, or the fact that objects rotate in a preferred direction, as shown in Figure 15.20. It is

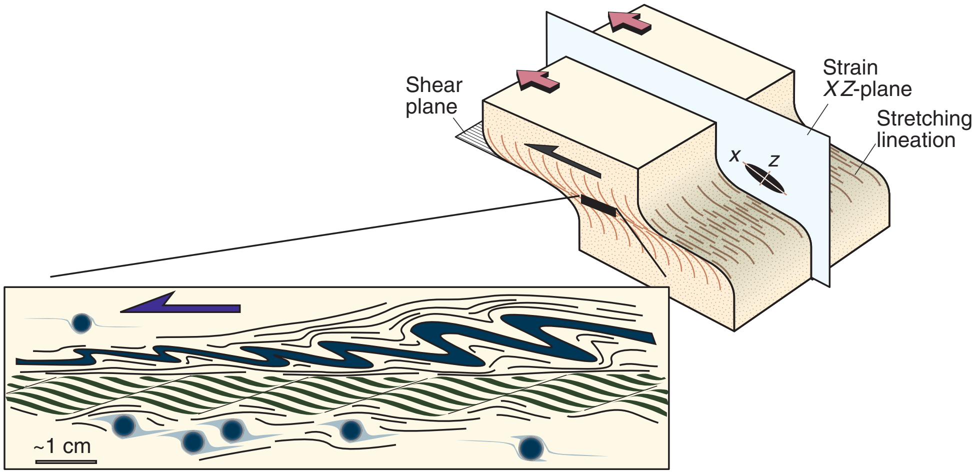

Figure 15.21 Illustration of the section of observation for shear-sense determination, and some common asymmetric structures that consistently indicate sense of shear.

also reflected in the particle paths, which for non-coaxial deformations are asymmetric. Coaxial deformations produce structures whose geometries show a higher (orthorhombic) degree of symmetry, which is related to their symmetric particle paths. The connection between symmetry and coaxiality is shown schematically for a variety of structures in Figure 15.20. Structures formed in coaxial flow are thus sometimes referred to as orthorhombic.

The (a)symmetry of mylonite structures can be used to evaluate the sense of shear and sometimes also the degree of coaxiality of a mylonite zone.

It is the monoclinic structures that give information about the sense of displacement or sense of shear in mylonite zones and we will therefore focus on monoclinic or asymmetric structures in this section. In a shear zone we primarily study the section perpendicular to the foliation that contains the lineation (Fig. 15.21), although the section perpendicular to the lineation can also be of interest for evaluation of three-dimensional deformation.

Deflected markers

We have already looked at how pre-existing markers (linear or planar) become rotated into shear zones (Figure 15.2). Even if we do not see the shear zone margins, rotation of planar markers from an area of low strain to an area of high strain provides a very reliable criterion for sense of shear determination, for instance along the margins of tectonic lenses in the shear zone.

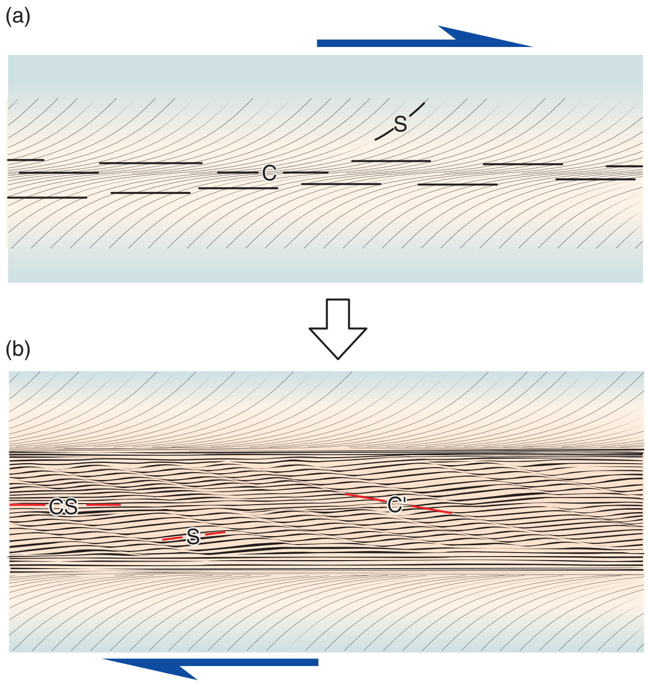

Figure 15.22 Schematic illustration of the development of S-C structures in a shear zone in a magmatic rock. (a) The new-formed foliation (S) is cut by shear surfaces (C) that parallel the shear zone margins. (b) Continued deformation rotates S into close parallelism with C, together referred to as a CS-foliation. New and oblique shear bands (C′) form and back-rotate the CS-foliation, which then is called S.

Mylonitic foliation and shear bands (S-C structures)

The foliation that develops in a shear zone is usually thought to trace the *XY*-plane of the strain ellipsoid. The sense of rotation of the foliation from the margin into the shear zone is generally a safe kinematic indicator.

As strain accumulates, a set of slip surfaces or shear bands commonly forms parallel to the walls of the shear zone (Figure 15.22a). These shear bands are called C (French "cisaillement" for shear, which relates to the

(a)

(b)

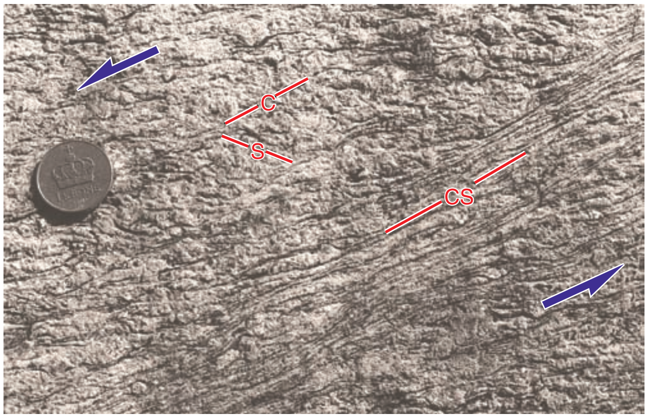

Figure 15.23 (a) Undeformed granite. (b) Sheared version of the same rock. Two sets of planar surfaces (S and C) are developed. S rotates to become subparallel with C in the zone of high strain. The combined foliation is named CS.

movement of scissors) and the foliation is named S (for schistosity or "schistosité"). C-surfaces are not really surfaces, but small-scale shear zones that affect the foliation within the main shear zone. In detail, the foliation curves into and out of the C-surfaces, and the sense of deflection shown by the curving foliation reflects the sense of shear of the entire shear zone. C-surfaces are particularly common in shear zones in magmatic rocks (Figure 15.23).

If the shear strain is high, say above 10, the angle between the shear zone and the foliation becomes indiscernible. We then have a composite foliation consisting of rotated foliation and C-surfaces. In addition, heterogeneities in the deforming rock as well as slip on micaceous elements in the foliation perturb the flow enough to break down the simple picture seen at lower strains.

Such high-strain complications promote the formation of new structures that can reveal the kinematics of the flow. Notably, a new set of shear bands, oblique to the shear zone margins, form when strain gets high (Figure 15.22b). These shear bands are designated C′ when their obliquity to the shear zone can be demonstrated, and are particularly common in mylonites rich in platy minerals. C′ surfaces are thus similar to the Riedel shears (R) in brittle shear zones and can only be separated from shear-zone parallel C-surfaces when their orientation with respect to the shear zone boundaries is known. This information is not always available, and the structural geologists Gordon Lister and Arthur Snoke suggested in 1984 that any kind of structure that is composed of two planar structures formed during one progressive shearing event be called S-C structures. Furthermore, mylonites that show S-C structures are called **S-C mylonites**.

S-C mylonites are made up of two sets of planar structures, a foliation (S) and shear bands (C) that obliquely transect and often back-rotate the foliation.

The angle between S and C can vary, but is typically about 25–45°. The angular relationship between the foliation and shear bands (C) (Figures 15.24 and 15.25) is a reliable shear indicator if the angular relation is consistent. The higher the degree of coaxiality, the lower the consistency, and coaxial deformation is expected to produce sets of oppositely dipping shear bands. However, a consistent angular relationship between shear bands and the general foliation is very common in mylonite zones. It is, however, important to remember that shear bands may form relatively late during the evolution of a shear zone, thus reflecting the last part of the deformation history.

Microscale foliations

Foliations that are oblique to the main foliation or the mylonitic banding occur in mineral aggregates that dynamically recrystallize during the deformation. Quartz aggregates or deformed quartz veins are typical examples. The aggregates themselves form part of the main mylonitic foliation, while the long axes of its deformed grains define an oblique foliation (Figures 15.26 and 15.27). There are two competing processes going on in the aggregates. One is the strain produced by elongation of grains. The other is the dynamic recrystallization and recovery that erases the strain recorded by those grains. The oblique foliation within the aggregates reflects only the last increment of deformation, while the main foliation is a result of the entire deformation history and therefore represents an orientation that is close to the

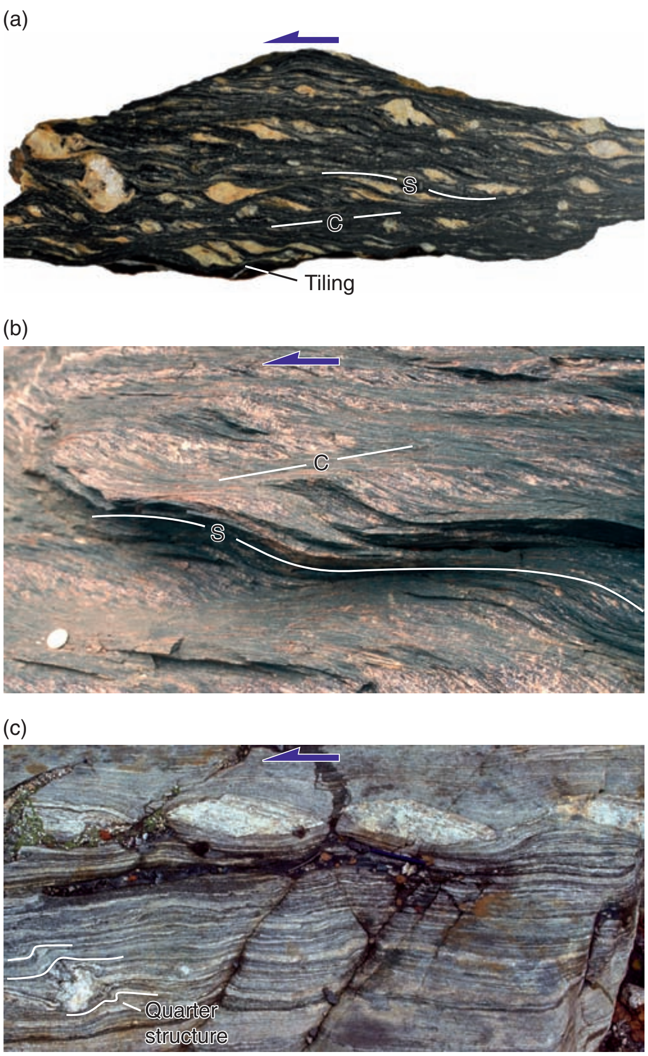

Figure 15.24 (a) S-C structure in protomylonitic granite, Antarctica. Note tiling of feldspar porphyroclasts. (b) Shear bands in phyllite, together with asymmetric folds. Caledonian basal décollement. (c) Asymmetric boudins in granitic gneiss. Asymmetric folds around inclusion represent quarter structure. Caledonides east of Bergen.

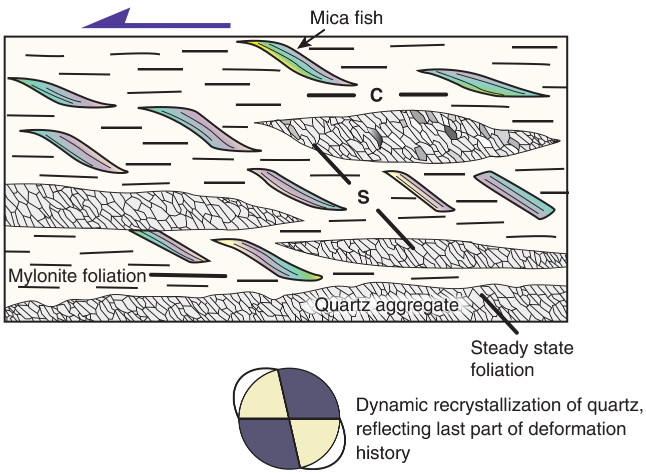

Figure 15.26 Typical S-C structures in quartz–mica-dominated mylonites.

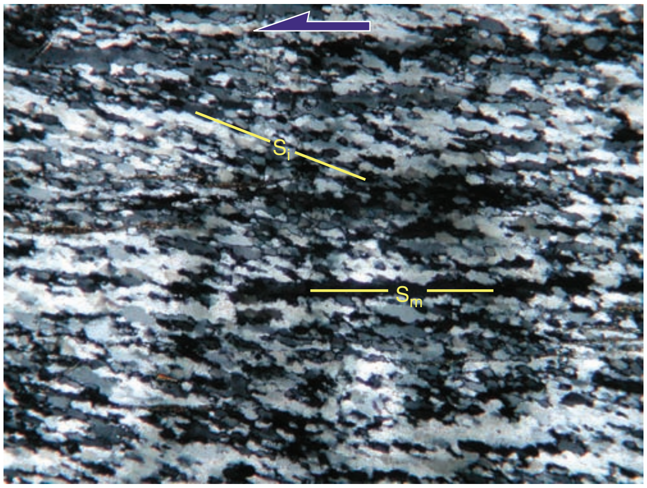

Figure 15.27 Thin section of a mylonite with a horizontal mylonitic foliation (S_m). The quartz grains are stretched in a direction S_i oblique to S_m, and the angular relations are consistent with top-to-the-left sense of shear. We could also use S-C terminology, where S_m represents C and S_i corresponds to S.

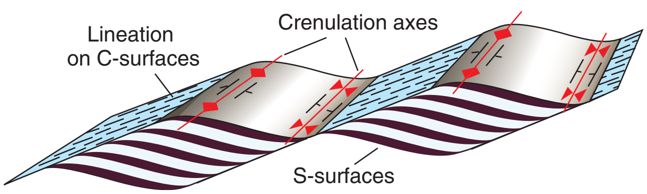

Figure 15.25 Schematic illustration of the geometry of shear band-type S-C structures. The crenulation axis is typically at a high angle to the sense of shear, which is reflected by a lineation that may appear on the shear bands (C-surfaces). Slip may or may not occur on the S-surfaces.

shear plane. The angle between the mylonitic foliation and the grain shape fabric within mineral aggregates indicates that the deformation is non-coaxial, reveals the sense of shear, and even indicates the degree of non-coaxiality or W_k of the deformation. If the grains within the aggregates are only slightly strained, then the angle should be close to 45° for simple shear ($W_k = 1$). In terms of S-C terminology, both S and C are foliations in this case (Figure 15.26) and are also termed S_i (internal) and S_m (mylonitic) (Figure 15.27).

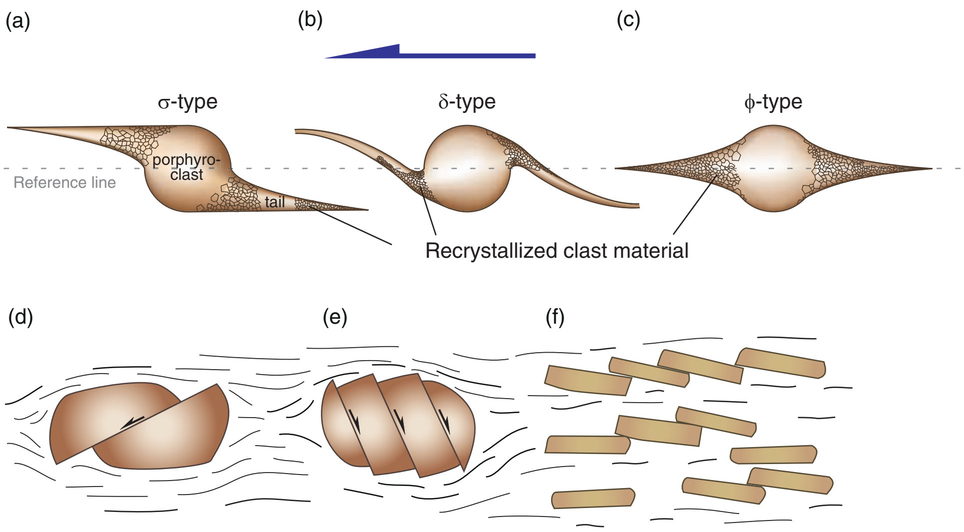

Figure 15.28 Porphyroclast systems. (a–c) Porphyroclasts with recrystallized tails. σ-type porphyroclasts have tails that do not cross the reference line, while the δ-type have tails that do. (a) and (b) show monoclinic symmetry (with the rotation axis being perpendicular to the page). The ϕ-type is symmetric about the reference line. (d) Fractured porphyroclast with synthetic fracture. (e) Antithetic shear fractures. (f) Tiling (imbrication) of porphyroclasts. All structures (except (c)) consistent with sinistral shear.

Mica fish

Mica porphyroclasts in mylonitic rocks tend to have tails that systematically curve away from the general orientation of the porphyroclasts, as shown in Figure 15.26. Such microstructures are known as **mica fish**, and the resulting asymmetry indicates the sense of shear. Mica fish are commonly seen to be confined by shear bands, and can be regarded as a type of S-C structure.

Foliation fish and foliation boudinage

Parts of strongly foliated mylonites sometimes back-rotate with respect to the shearing direction. The result is structures called **foliation fish** that look like mica fish but occur on a larger (for example meter) scale (Figure 14.12). These structures are also known as **asymmetric foliation boudinage**.

Boudinage

Boudinaged competent layers can be used as kinematic indicators, provided that one knows the approximate orientation of the layers before deformation started. If such a layer is located in the field of active stretching during a non-coaxial deformation, then the boudins can rotate against the shear direction, as shown in Figure 15.24c (see also Chapter 14).

Porphyroclasts

Porphyroclasts of feldspar, quartz, mica or other minerals can develop a mantle of recrystallized material that also forms tails, as illustrated in Figure 15.28a–c. Coaxial deformations produce tail geometries that are symmetric with respect to the general mylonitic foliation (ϕ-type; Figure 15.28c). For non-coaxial deformation types, an asymmetric geometry tends to arise, and the final shape depends on W_k and the thickness of the mantle versus the core (rate of recrystallization), among other things. Asymmetric structures are classified into σ-types, where the tails have a stair-stepping geometry, and δ-types, where the tails are thinner, strongly curved and influenced by porphyroclast rotation (Figure 15.29). A characteristic difference between the two is that the tails of the σ-type are located on each side of the foliation reference line while those of the δ-type cross this line.

Figure 15.29 δ-porphyroclast indicating rotation during top-to-the-left shearing, consistent with the asymmetry of small-scale folds (right).

Many porphyroclasts show growth of quartz or other minerals in what are called porphyroclast pressure shadows or **strain shadows**. Strain shadows are asymmetric when formed in non-coaxial deformation types and show similarities with σ-type porphyroclasts (Figure 15.30).

The garnet porphyroclast in Figure 15.30 also shows an inclusion pattern. Such patterns may represent a crenulated foliation that has been overgrown by the garnet, but may also form during synkinematic growth of porphyroclasts. In the latter case the pattern indicates the sense of rotation and thus the sense of shear for non-coaxial deformations.

Folds and cleavage

Asymmetric folds and related axial planar cleavage can give information about the sense of shear in strongly deformed rocks if we know the approximate orientation of the folded layer before deformation started. In mylonite zones we often find that it is the mylonitic foliation itself that has been folded during the deformation history. Hence, the foliation must have rotated through the shear plane and into the contractional field. In such cases the vergence of asymmetric folds indicates the sense of shear (Figure 15.29). All it takes to make mylonitic folds is for the foliation to be rotated into the field of instantaneous contraction. A rotation of a few degrees is sufficient and easily occurs where the foliation is perturbed around rigid inclusions or lenses. Alternatively, asymmetric folds can develop from pre-deformational features such as cross-beds or cross-cutting dikes that initially lie within the field of contraction. In such cases vergence must be interpreted with care, particularly if the initial orientation of the marker is unknown. Asymmetric folds can also indicate the "wrong" sense of shear in special cases, as illustrated in Figure 15.31.

Asymmetry and strong non-cylindricity (strongly curved hinge lines) are characteristic features of folds in mylonite zones (Figure 15.32 and 15.33). Sometimes the inverted limb of such asymmetric folds is displaced by small reverse or thrust faults (Figure 15.32), which helps determine the sense of shear.

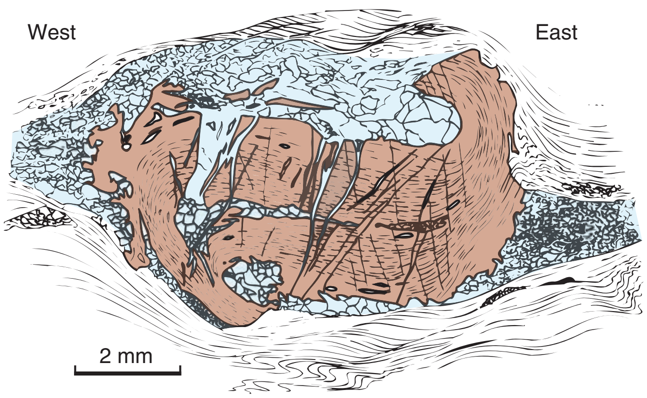

Figure 15.30 Rotated garnet with quartz tails (blue) in the strain shadows. The inclusion pattern indicates top-to-the-E sense of shear. The younger shear bands and extension fractures in the garnet indicate the opposite sense of shear. Modified from Fossen (1992).

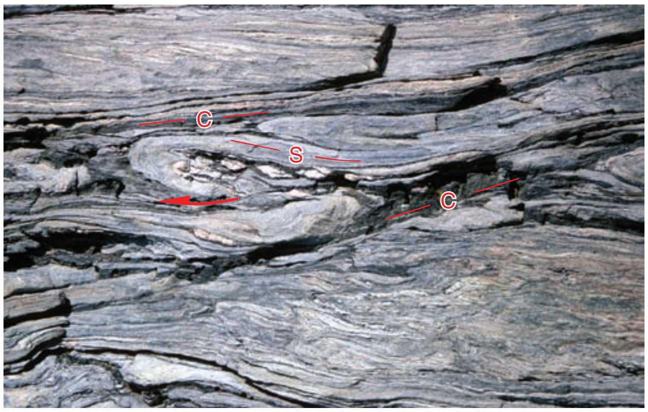

Figure 15.32 Folds formed in non-coaxial deformation in gneisses. The vergence of the asymmetric folds indicate top-to-the-left transport. A small thrust-like structure (arrow) cuts off the inverted fold limb. The complete structure can be considered to be an S-C structure where the intrafolial fold train is caught between two C-bands.

Figure 15.31 Progressive development of folds in simple shear (strain increasing to the left). This particular initial orientation can result in a vergence apparently inconsistent with the sense of shear.

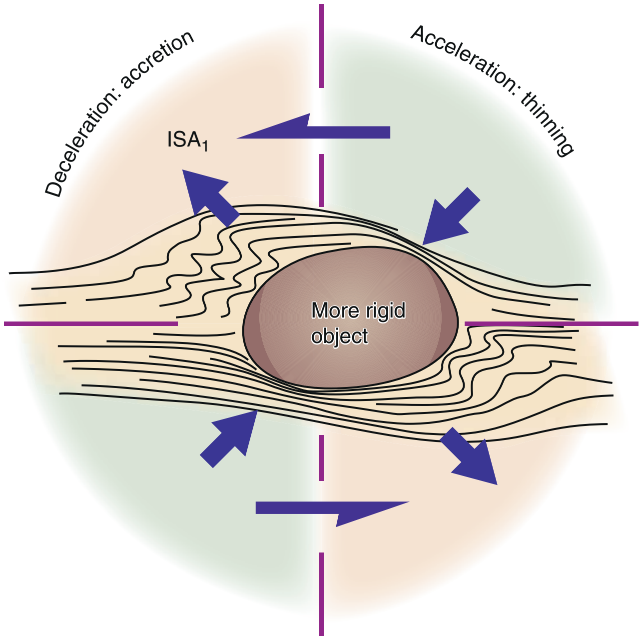

Figure 15.33 Sectors (quarters) of layer thinning/thickening around a rigid object in a mylonite zone indicating sense of shear. The structures are related to particle acceleration/deceleration and are called quarter structures.

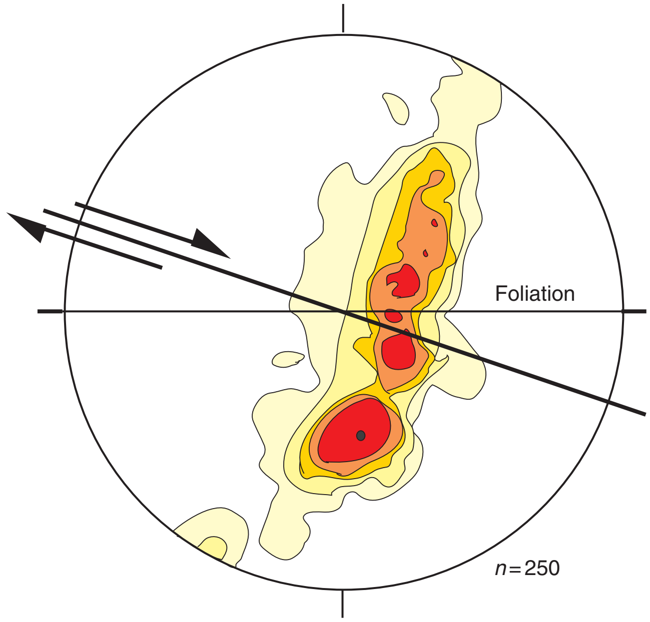

Figure 15.34 Two hundred and fifty quartz *c*-axes measured on the U-stage and plotted in the stereonet. An asymmetrical pattern with respect to the foliation (trending E–W in the plot), such as the one shown here, indicates the sense of shear. From Fossen (1993).

Quarter structures

The association of contraction and extension structures around tectonic lenses and rigid objects in a mylonite characterizes the sense of shear. As shown in Figure 15.33, the area around such objects can be divided into quarter sectors of accretion and stretching/thinning of layers. The structures vary according to how the foliation is rotated around the object. In the quarters of thinning, the layers have to thin, as particles experience acceleration to bypass this corner of the object. In addition to layer thinning we may find evidence of dissolution (mica concentrations). In the accretion sectors, layers thicken and fold as particles experience deceleration.

Crystallographic orientation

The orientation pattern of the optical *c*-axis of quartz can sometimes be used to determine sense of shear in quartz-rich mylonites. A number (*c*. 150–250) of orientations are measured and plotted in a stereonet with the foliation being vertical and the lineation oriented E–W in the net. The pattern, or girdle, that arises is typically asymmetric for non-coaxial deformations. The asymmetry will indicate the sense of shear, as shown in Figure 15.34. The actual girdle will depend on the crystalline slip systems that were active and therefore the temperature during the deformation, and on the strain geometry (prolate versus oblate strain ellipsoids).

Tiling of objects

Rigid and elongated crystals in a deforming rock can be imbricated or tiled in a mylonite zone (Figure 15.28f). Such tiling requires a high density of crystals and is common in porphyritic magmatic rocks, particularly if they have not completely solidified at the onset of deformation.

Shear transfer structures

Heterogeneous rocks can experience strain partitioning during deformation in the sense that simple shear or slip is localized in weak, mica-rich layers. In some cases shear is transferred from one layer to another, particularly at the termination points of weak layers. The transfer or overlap zone can be contractional or extensional, depending on the arrangement of the weak layers and the sense of shear. Contractional structures are portrayed as fold trains and/or imbrication structures, while extensional structures are dominated by shear bands. Figure 15.35 shows how the shear direction can be determined from such structures.

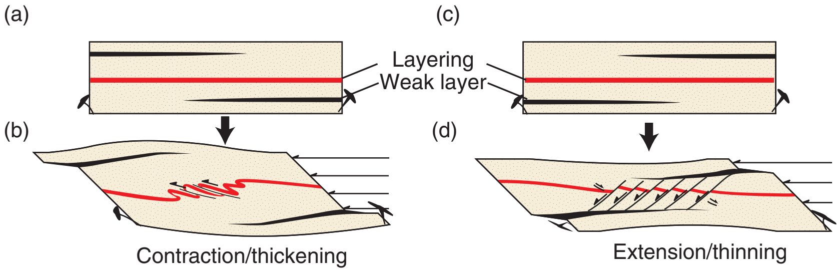

Figure 15.35 Transfer of localized shear strain from one weak layer to another can give contractional or extensional structures. Based on Rykkelid and Fossen (1992).

Microfaulted mineral grains

Mineral grains that deform in a brittle manner in a plastically deforming matrix may show sharp intragranular shear fractures. The most common example is brittle feldspar in a ductile quartz-rich matrix deforming under greenschist facies conditions. The orientation and sense of slip on such fractures relate to the sense of shear. The challenge is that such shear fractures can be antithetic as well as synthetic with respect to the shear direction (Figure 15.28d, e). Their orientation depends not only on sense of shear and W_k, but also on the shape of the grain, its orientation and that of any crystallographic cleavage or other plane of weakness that the grain may possess. Fractured mineral grains must therefore be used with care in kinematic analysis.

Fibers and veins

The orientation of extensional veins (Figure 15.36a) indicates the sense of shear in mylonites, and sometimes also W_k and the orientations of the ISA. If the veins are fibrous the fibers can give a better definition of the stretching direction.

Veins forming under non-coaxial deformation will rotate from the moment they form. This results in a sigmoidal geometry that can be used to determine the sense of shear, as illustrated in Figures 15.36b and 15.37.

Special folds can arise around veins in mylonites where the sense of shear from the fold geometry appears to oppose the sense of displacement on the vein. The geometry arises due to shear on the vein and is illustrated in Figure 15.38. This example shows that care should be exercised when determining shear sense from folds associated with veins in mylonitic rocks.

In general, populations of veins and dikes that have a variety of orientations are useful structures for reconstructing strain history and kinematics. Their appearance indicates whether they are located within the field of stretching, shortening, or stretching followed by shortening. For instance, veins that form boudinaged folds have a

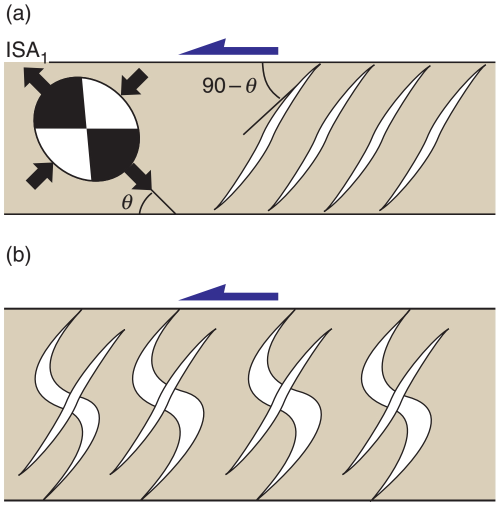

Figure 15.36 En-echelon-arranged extension veins in a shear zone. The vein tips are oriented perpendicular to ISA$_1$. They are sheared into sigmoidal shapes and may be cut by younger veins.

Figure 15.37 En-echelon-arranged extension veins defining a shear zone. Note the foliation that is oriented at a high angle to each vein.

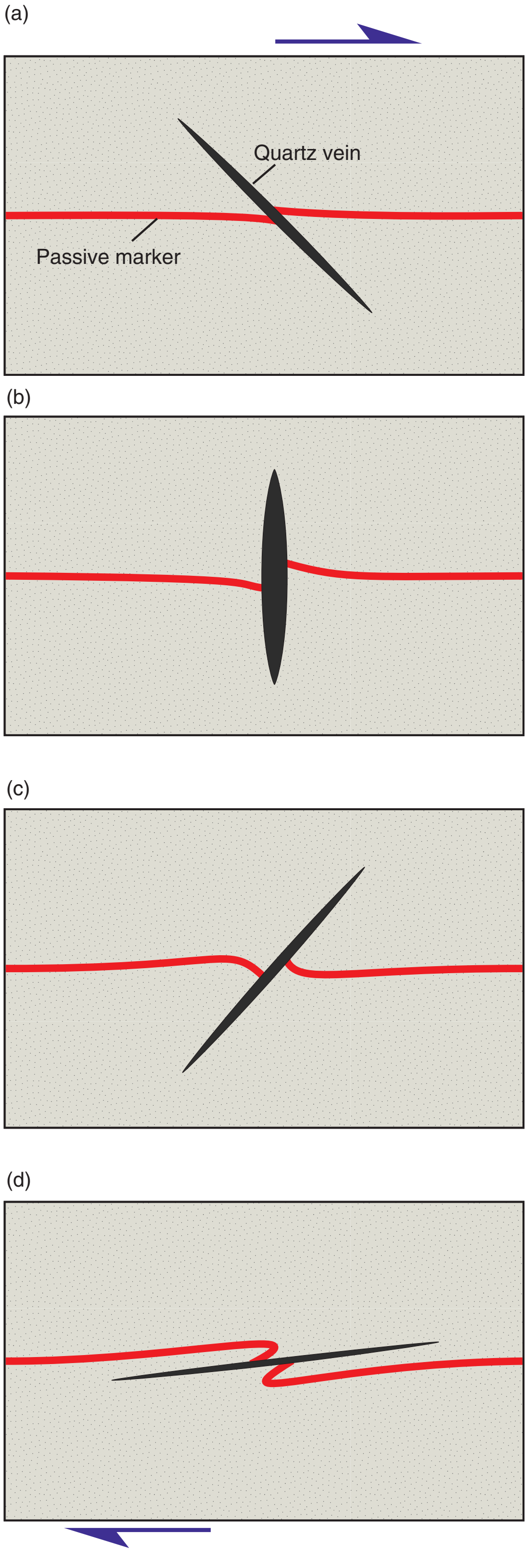

Figure 15.38 The development of an extension fracture with apparent wrong sense of displacement. (a) An extension vein forms in a shear regime. (b) The fracture opens, causing weak folding of the marker horizon. The fracture then rotates during continued shearing and the displacement along the fracture is opposite to that of the sense of shear (c, d). Based on Hudleston (1989).

history of shortening followed by extension. As discussed in Chapter 2, the sizes and distribution of these fields reflect W_k and kinematics.

15.6 Growth of shear zones

Many shear zones are simple and well-defined structures, yet our knowledge and understanding of how they form and develop is far from complete. For simplicity, it is often assumed that they form in homogeneous rocks such as granite. In practice, shear zones form at the weakest point or along the weakest layer in the rock, such as micaceous layers, partly molten zones, veins, fractures, fine-grained layers, dikes etc. This will have to be addressed separately for each shear zone.

Once a shear zone forms, we can imagine several histories of development leading to different types of shear zones, each illustrated schematically in Figure 15.39:

Type I shear zones are strain hardening shear zones where the deformation in the central part slows down as the zone thickens. The central part thus records the first part of the deformation history while the marginal parts only record the last part. Type I shear zones develop wide and plateau-type displacement profiles.

Type II shear zones are strain softening zones that quickly establish a certain thickness, but after a while the deformation localizes to the central part. Thus, the margins become inactive and the active part of the shear zone gets thinner. The result is thin shear zones with high shear strain gradients.

Type III shear zones develop a fixed thickness, and the entire zone keeps deforming without any sign of internal localization. The shear zone maintains its thickness, which equals its active thickness. This model is perhaps the least realistic one, but it may work for some types of kink bands.

Type IV shear zones have the same development as Type I zones, but the entire shear zone remains active throughout the deformation history. In other words, the zone grows in width and the strain increases from the margins toward the center. The margins only record the last increment of shearing while the central part has experienced the entire shearing history.

In general, shear zones show a positive correlation between displacement or length and thickness (Fig. 15.3). This observation indicates that most shear zones widen as deformation progresses and thus favors Type I and IV shear zones. If so, this means that the margin records the last increment of strain. In order to find out how they have developed, we could compare related shear zones

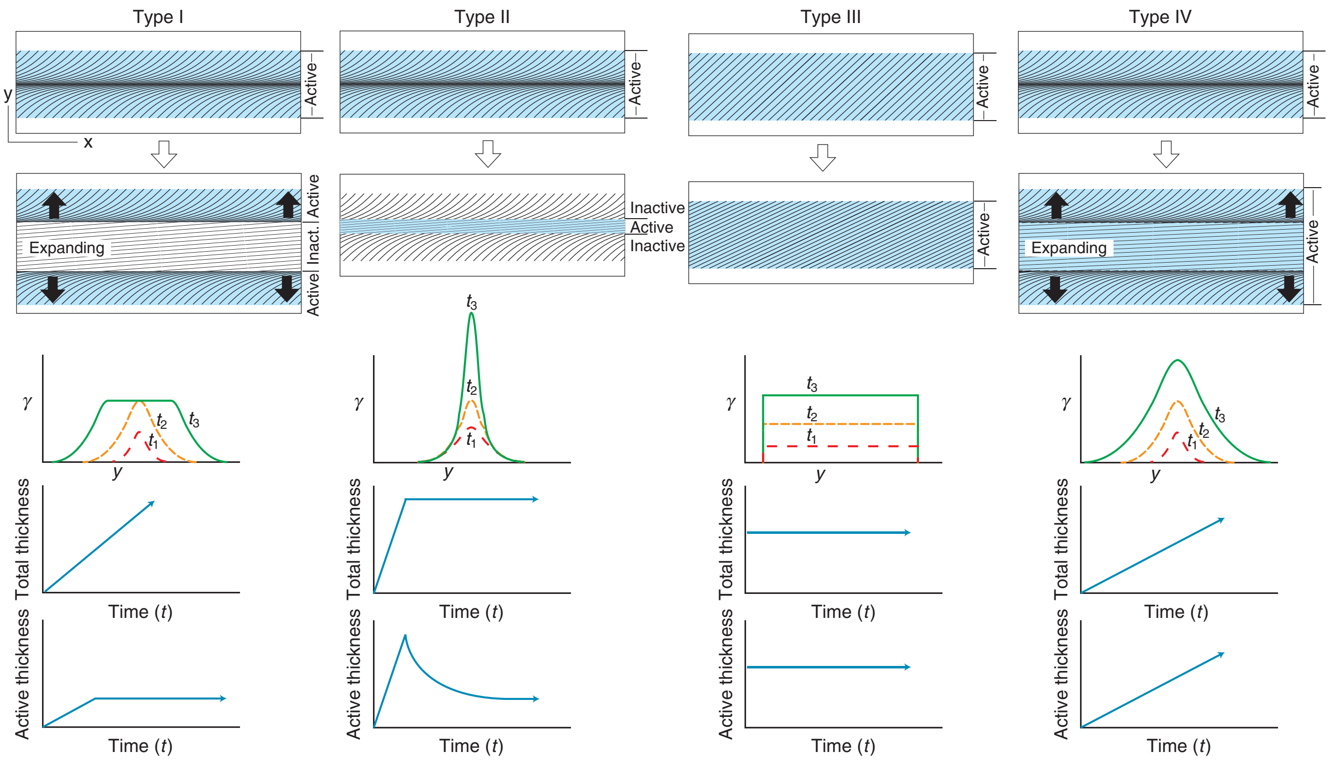

Figure 15.39 Four models for the growth of shear zones.

that show different displacements, i.e. shear zones at different stages of development.

Some shear zones are uplifted through their history, moving from the plastic regime to the plastic–ductile transitional regime and eventually into the brittle regime. This generally results in localization of shear into one or more relatively narrow zones, i.e. Type II, except that strain does not necessarily localize to the central part. Many kilometer-thick shear zones have gone through synkinematic exhumation and developed a late and narrow (semi-)brittle shear zone in their upper part. An example of this is the metamorphic core-complex-forming shear zones discussed in Box 17.3. Type II shear zones leave behind sheared rocks that formed at early or intermediate stages of shearing, which is very useful when unraveling the *P–T* history of a shear zone, as discussed in Chapter 21.

Shear zones can develop quite differently, depending on rock properties, fluids, deformation mechanisms and metamorphic reactions, but generally go through an early phase of widening.

Summary

Shear zones are important structures that usually contain internal structures that reflect their deformation type and deformation history, which may be important for the understanding of the tectonic development of an area. What shear zones do not directly tell us is whether they formed in an extensional, contractional or strike-slip regime. Simple shear, subsimple shear and any other type of shear zone can all form in any tectonic regime, and we have to know their orientation at the time of deformation in order to make such inferences. In other words, the shear zones have to be placed into a tectonic context. In the next three chapters we will see how shear zones and faults occur in the three main tectonic regimes. First some important points from the present chapter:

- Shear zones are zones where strain is higher than in the surrounding rock.
- Shear zones are subdivided based on ductility (ductile or brittle) and deformation mechanism (brittle/frictional or plastic).

- Ductile shear zones preserve the original continuity of passive layers.
- Plastic shear zones develop a foliation whose orientation relates to strain.
- If the deformation type (e.g. simple shear) and strain across the zone are known, these can be used to calculate the offset.
- Asymmetric structures indicate sense of shear.
- Always consider as many sense-of-shear indicators as possible to establish the shear sense.

Review questions

1. What makes a shear zone different from a fracture?
2. Can you draw the upper shear zone margin on Figure 15.9? Is it easily definable?
3. What type of data support the idea that shear zones grow in width as they accumulate displacement, and where would we look for the last increment of strain?
4. What assumption do we have to make for the last increment of strain to be representative of earlier strain increments in the zone?
5. What is meant by the term S-C structures?
6. What do you think are the most and least reliable shear sense structures described in this chapter?
7. How could you get information about the strain path (strain evolution) of a shear zone?

E-MODULE

The e-learning modules called *Shear zones* and *Kinematic indicators* are recommended for this chapter.

FURTHER READING

General

Passchier, C. W. and Trouw, R. A. J., 2006, *Microtectonics*. Berlin: Springer Verlag.

Ramsay, J. G. and Huber, M. I., 1983, *The Techniques of Modern Structural Geology. Vol. 1: Strain Analysis*. London: Academic Press.

Images and descriptions

Snoke, A. W., Tullis, J. and Todd, V. R., 1998, *Fault-related Rocks: A Photographic Atlas*. Princeton, NJ: Princeton University Press.

Ideal shear zones

Ramsay, J. G., 1980, Shear zone geometry: a review. *Journal of Structural Geology* **2**: 83–99.

Complex shear zones

Passchier, C. W., 1998, Monoclinic model shear zones. *Journal of Structural Geology* **20**, 1121–1137.

Kinematic structures

Berthé, D., Choukroune, P. and Jegouzo, P., 1979, Orthogneiss, mylonite and non-coaxial deformation of granites: the example of the

South Armorican Shear Zone. *Journal of Structural Geology* **1**: 31–42.
Dennis, A. J. and Secor, D. T., 1990, On resolving shear direction in foliated rocks deformed by simple shear. *Geological Society of America Bulletin* **102**: 1257–1267.
Lister, G. S. and Snoke, A. W., 1984, S-C mylonites. *Journal of Structural Geology* **6**: 617–638.
Passchier, C. W. and Williams, P. R., 1996, Conflicting shear sense indicators in shear zones: the problem of non-ideal sections. *Journal of Structural Geology* **18**: 1281–1284.
Platt, J. P., 1984, Secondary cleavages in ductile shear zones. *Journal of Structural Geology* **6**: 439–442.
Simpson, C., 1986, Determination of movement sense in mylonites. *Journal of Geological Education* **34**: 246–261.
Wheeler, J., 1987, The determination of true shear senses from the deflection of passive markers in shear zones. *Journal of the Geological Society* **144**: 73–77.

Strain in shear zones

Bhattacharyya, P. and Hudleston, P., 2001, Strain in ductile shear zones in the Caledonides of northern Sweden: a three-dimensional puzzle. *Journal of Structural Geology* **23**, 1549–1565.
Hudleston, P., 1999, Strain compatibility and shear zones: is there a problem? *Journal of Structural Geology* **21**: 923–932.
Simpson, C. and De Paor, D. G., 1993, Strain and kinematic analysis in general shear zones. *Journal of Structural Geology* **15**: 1–20.

Rigid objects and porphyroclasts

Bjørnerud, M., 1989, Mathematical model for folding of layering near rigid objects in shear deformation. *Journal of Structural Geology* **11**: 245–254.
Passchier, C. W. and Simpson, C., 1986, Porphyroclast systems as kinematic indicators. *Journal of Structural Geology* **8**: 831–843.
Passchier, C. W. and Sokoutis, D., 1993, Experimental modelling of manteled porphyroclasts. *Journal of Structural Geology* **15**, 895–909.

Growth of shear zones

Means, W. D., 1995, Shear zones and rock history. *Tectonophysics*, **247**: 157–160.
Sibson, R. H., 1980, Transient discontinuities in ductile shear zones. *Journal of Structural Geology* **2**, 165–171.

Soft-sediment shear zones

Lee, J. and Phillips, E., 2008, Progressive soft sediment deformation within a subglacial shear zone: a hybrid mosaic-pervasive deformation model for middle Pleistocene glaciotectonised sediments from Eastern England. *Quaternary Science Reviews* **27**, 1350–1362.
Maltman, A. J. and Bolton, A., 2003, How sediments become mobilized. In P. van Rensbergen, R. R. Hillis, A. J. Maltman and C. K. Morley (Eds.), *Subsurface Sediment Mobilization*. Special Publication **216**, London: Geological Society, pp. 9–20.