

Shear Zones

John G. Ramsay, ETH Zürich, Switzerland

Abstract

Shear zones may be classified into Brittle-, Brittle-ductile, and Ductile-shear zones. The geometry and displacement boundary conditions of these zones are established. The geometric characteristics of ductile shear zones relevant to geological studies are described:- orientation and values of principal finite strains, rotation, deformation features of pre-existing planar and linear structures. Ductile shear zones show a fabric (schistosity and lineation) related to the finite strain state. The methods for determining strains and displacements from field studies are described.

The geometric features of ductile shear zones suggest that strain softening mechanisms play a special role, probably grain size reduction and chemical transport are important. Increasing obliquity of the progressively strengthening fabrics during deformation lead to a reduction in the deformability of the highly strained parts of ductile shear zones.

Shear zones often occur in conjugate sets. but the two differently oriented sets do not seem to be able to operate synchronously. The angular relationships of conjugate ductile shear zones are different from those of brittle shear zones.

The termination of all types of shear zones pose complex compatibility problems, some solutions are suggested.

A synthesis of shear zone geometry in regions of crystal contraction and crustal extensions is made, and ideas on how deep level ductile shear zones relate to high level brittle shears and gliding nappes are put forward.

Discussion points

At the meeting the geometric discussions and conclusions of this paper will be supported by geological illustrations of the phenomena. It should be clear that my basic philosophy is that the geometric constraints offered by the structures seen in naturally deformed rocks (which can be mathematically expressed in strain compatibility) offer some of the strongest tools we have to unravel the complex deformation features of the earth's crust. As this contribution is meant to provide a basis for open discussion at the meeting it seems a good idea to list a number of points which I think worthy of discussion:

1. Do all shear zones occur in conjugate sets?
2. If so, can these develop synchronously?
3. What happens where two conjugate shear zones intersect?
4. What controls the angle between conjugate shear zones?
5. Is the Anderson stress model of brittle faults correct, and does the geometry of fault sets accord with the orthorhombic symmetry of the stress tensor?
6. What is the reason for strain softening in ductile shear zones?
7. How does progressively developing anisotropy in a ductile shear zone control the development of that zone?
8. What is the significance of a finite strain profile across the shear zone in terms of rock rheology?
9. How do volume changes take place in shear zones, how is the material removed and where does it go?
10. Are you convinced in the proposed geometric relationships between deep level and high crustal level shear zones proposed in the last section of this paper?

Types of shear zones

Generations of geologists studying the natural deformations of the earth's crust have noticed that high deformations are often localised in narrow, sub-parallel sided zones, and these have been loosely termed shear zones.

Faults or Brittle shear zones are special variety of shear zone where a clear discontinuity exists between the sides of the zone, and where the shear zone walls are almost unstrained or perhaps brecciated (fig. 1, A). Such fault zones are generally attributed to brittle failure controlled by the limiting elastic properties of the rock under orogenic stress. Other fault-like features show some ductile deformation in the walls, and are perhaps best termed Brittle-ductile shear zones. The walls may show permanent strain for a distance of up to 10 metres on either side of the fault break (fig. 1, B). In the past these distortions have often been attributed to localised "drag" effects as the adjacent walls of the fault moved past each other, and the distortion of marker layers of rock entering this zone of flow have been used to determine the approximate movement sense of the displacement or even to specify the exact direction of the displacement vector. This latter deduction is not always possible, because the geometry of the deflected marker surfaces in this zone is controlled by the line of intersection of the initial surface and the shear zone, and not by the movement vector (fig. 2). In Brittle-ductile shear zones it is quite possible that the ductile part of the deformation history formed at a different time from that of the fault discontinuity. Another type of Brittle-ductile shear zone is the extension failure type (fig. 1, C). The deformation zone shows an en-echelon array of extension openings, usually filled with fibrous crystalline material, the openings usually making an angle of 45 degrees or more with the shear zone and sometimes showing a sigmoidal form. The rock material between the extension fissures generally shows a certain amount of coherent permanent deformation, and it appears that the extension fissures developed at some stage of the deformation when a certain limit to coherent flow in the zone was reached. The third type of shear deformation is that of the Ductile shear zone. Here the deformation and differential displacement of the walls is accomplished entirely by ductile flow, and on the scale of the rock outcrop no discontinuities can be seen (fig. 1, D). Marker layers in the country rock can be traced through the shear zone, they are deflected and may change their thickness, but they remain unbroken. Ductile shear zones are extremely common in deformed crystalline basement rocks (granites, gabbros, gneisses) which have been deformed under metamorphic conditions of greenschist, blueschist, amphibolite and granulite facies. They seem to be the dominant deformation mode whereby large masses of physically rather homogeneous rock can change shape under medium to high grades of metamorphism. The mineralogy of the rock in the ductile shear zone usually shows the characteristics of the metamorphic facies under which it developed. It is historically interesting as well as pertinent to this last point to note that the first understanding of the geological significance of hornblende schist was made from a study of shear zones in the Lewisian rocks of NW Scotland by Teall (1885). He was able to trace an unmetamorphosed basaltic dyke with normal igneous mineralogy through an amphibolite grade shear zone where it was transformed into a hornblende-biotite-oligoclase rock with a strongly schistose fabric.

The main aim of this contribution is to establish the geometrical features of shear zones, especially those of the ductile and brittle-ductile varieties because in many regions these seem to be the deep level counterparts of brittle shear zones and faults at higher levels seen at higher levels in the crust.

Relationships of displacement and strain

At the outset of this contribution it seems best to set out the mathematical notation to be used and to specify the geometric constraints of continuum mechanics as it relates to geological structures. Displacement specifies the relations between any initial point in space defined by orthogonal coordinate axes x, y and z . The initial point of coordinates (x, y, z) is transformed to a final position (x_1, y_1, z_1) . No restrictions are placed on the numerical values of these numbers, i.e. the theory is appropriate to large displacement and therefore to large strains. The displacement vector joins (x, y, z) to (x_1, y_1, z_1) and has three components u, v and w (eg. $u = x_1 - x$ etc.). If the material deforms as a continuum (the material is coherent; neighbouring points remain neighbours; no holes or discontinuities develop) then it is possible to set up three displacement equations (a Lagrangian description) specifying how all points in the displaced body relate to their initial coordinates:-

$$\begin{aligned} x_1 &= f_1(x, y, z) \\ y_1 &= f_2(x, y, z) \\ z_1 &= f_3(x, y, z) \end{aligned} \quad 1.$$

or an Eulerian description specifying how all points in the initial body are related to their final coordinates:-

$$\begin{aligned} x &= f_4(x_1, y_1, z_1) \\ y &= f_5(x_1, y_1, z_1) \\ z &= f_6(x_1, y_1, z_1) \end{aligned} \quad 2.$$

The functions f_1, f_2 etc. must be single-valued and continuous. If these functions are linear in the three variables then a homogeneous state of finite strain is set up, that is all initial unit cube markers with edges parallel to the coordinate axes become transformed into identically shaped parallelepipeds. If the functions are non-linear then the body is heterogeneously strained. For the case of homogeneous strain an initial sphere of unit radius becomes transformed into the finite strain ellipsoid. The directions of the three orthogonal ellipsoid axes can be derived from the eigen vectors of the matrix of the linear equation, these finite strain axes are specified $X_f \gg Y_f \gg Z_f$. Their semi-axis lengths $1+e_1, 1+e_2, 1+e_3$ respectively define the three principal strains e_1, e_2 , and e_3 , and are the eigen-values associated with the respective eigenvectors. The finite strain state at any point in a heterogeneously strained body can be derived from the non-linear displacement equations from the eigenvectors and eigenvalues of the following matrix:-

$$\begin{pmatrix} \frac{\partial f_1(x,y,z)}{\partial x} & \frac{\partial f_1(x,y,z)}{\partial y} & \frac{\partial f_1(x,y,z)}{\partial z} \\ \frac{\partial f_2(x,y,z)}{\partial x} & \frac{\partial f_2(x,y,z)}{\partial y} & \frac{\partial f_2(x,y,z)}{\partial z} \\ \frac{\partial f_3(x,y,z)}{\partial x} & \frac{\partial f_3(x,y,z)}{\partial y} & \frac{\partial f_3(x,y,z)}{\partial z} \end{pmatrix}$$

3.

after substituting specific coordinate values for (x,y,z) .

The basic theory for all these relationships can be found in Ramsay and Graham, 1970, 787-798.

Shear Zone Geometry

The boundary conditions for the geometrically simplest shear zones are first, that the shear zone is parallel sided, and second, that the displacement profiles along any cross sections of the zone are identical. This second condition implies that the finite strain profiles and the orientations and appearance of small scale structural features across profiles are also identical - these being more easily recognised boundary conditions in practical geological terms. Although these conditions can never be completely realistic because all shear zones have to come to an end (their sides must eventually come together, and their displacement profiles must change near their termination), it is my experience that very many shear zones often approximate closely to these over quite large zone lengths.

From the nature of the general displacement equations it can be shown that these boundary conditions constrain the types of displacement field (Ramsay and Graham, 1970, 796-798). If the walls of the shear zone are unstrained, then the following displacement fields are possible:-

- (i) Heterogeneous simple shear (fig. 3,A)
- (ii) Heterogeneous volume change (fig. 3,B)
- (iii) Combinations of (ii) and (i) (fig. 3,C)

For the mathematical description of the third of these possibilities one should note that the simple shear component should be specified first, followed by a volume reduction even though the two effects may have been geologically synchronous. In all three types the intermediate or Y_1 axis of the strain ellipsoid is always contained in a plane parallel to the shear zone walls. The types of strain ellipsoid produced from these displacement field is illustrated in the graphical plot of fig. 4. Simple shear (fig. 4,A) is clearly plane strain ($e_2 = 0$). Heterogeneous volume reduction leads to progressive flattening ellipsoids (fig. 4, B) and the combination leads to a deformation field in the apparent flattening sector of the diagram.

If the walls of the shear zone are themselves heterogeneously strained, then the displacement plans can be expressed mathematically as the same as (i) to (iii) above but with the addition of a homogeneous strain component inside and outside the shear zone, i.e.:-

- (iv) Homogeneous strain combined with simple shear (fig. 3, D)
- (v) Homogeneous strain combined with volume change (fig. 3, E)
- (vi) Homogeneous strain combined with simple shear and volume change (fig. 3, F)

With these three types the Y-axis of the strain ellipsoids will, in general, not have any specific relationship to the shear zone plane and all principal strain axes will have a variable orientation across a profile of the zone. The types of strain (i.e. apparent flattening or apparent constriction) have no general constraints, and all types are possible depending upon the orientations of the various "componental" strains.

Although all six basic types are clearly geometrically separable, it should be noted that their overall visual effect on marker surfaces crossing the shear zone is remarkably similar (see fig. 3). This means that the field geologist has to be especially careful in avoiding jumping to premature conclusions as to the displacement field if he only observes deformed planar markers. In my experience natural shear zones accord very closely to these six models. Some geologists might object to the inclusion of models (ii) and (v) (fig. 3, B and E) which involve differential volume loss under the category of shear zones, but I think it is right to include them for two reasons. First their basic geometric boundary conditions accord with those of more "orthodox" shear zones, second their effects on oblique planar markers are geometrically very close to those of shear zones with differential shear displacement parallel to the zone walls. In fact, the amount of volume change across a shear zone can be calculated using the displacement of initially plane markers. The differently oriented surfaces are necessary to completely solve the general geometric problem - one has to compare the shape of the triangle of intersection of the three surfaces on one side of the shear zone with that on the other (fig. 5). If only differential volume changes are taking place normal to the shear zone surface (types (ii) and (v)), then the amounts of volume loss can be calculated from the deflections of a single surface. Deformation zones involving only volume loss are sometimes called pressure solution zones or solution-seams or ~~stripes~~. These zones are usually deficient in certain soluble mineral species compared with their walls. In low to medium grade metamorphic environments, where such zones are often common, quartz or calcite are usually removed from the rock leading to a proportional increase in relatively insoluble components, such as clays, chlorite and micas. These changes in relative proportions of mineral species leads to a characteristic colour striping.

The deformation zones involving heterogeneous simple shear without differential volume change (types (i) and (iv), fig. 3, A and D) are often isochemical and the proportions of different mineral species across a zone profile is nearly constant. However, the mineral grain sizes and the rock fabric and texture do usually change with the strain variations across the zone.

The **zones** involving combinations of simple shear and volume change are usually variable in mineralogy, in mineral proportions and in texture and fabric.

The simple shear component

The basic component in practically all shear zones is that of heterogeneous simple shear. **It** is therefore appropriate to investigate the geometric properties of this type of displacement, to establish how strains are related to displacement and how the structural features we see in deformed rocks within a shear zone can be related to displacement and strain.

A zone with heterogeneous simple shear can be considered to be made up of a number of infinitesimally small elements showing homogeneous simple shear. In such a small homogeneously strained element **it is** mathematically convenient to relate the x coordinate direction parallel to the shear direction, and the z-axis normal to the plane of shear (xy) - see fig. 6. **If** the top face of an initially cube element is displaced by a distance s , the shear strain γ (more exactly γ_{zx}) is related to the angular shear strain ψ produced by the deflection of the lines initially parallel to z such that

$$\gamma = \tan \psi \quad 4.$$

$$\text{and so } s = z \tan \psi = z\gamma \quad 5.$$

The displacement matrix is then given by:

$$\begin{vmatrix} 1 & 0 & \gamma \\ 0 & 1 & 0 \\ 0 & 0 & 1 \end{vmatrix} \quad 6.$$

A circle of unit radius drawn on the xz face is deformed into a strain ellipse with principal semi-axis lengths along X_f of $1+e_1$ and along Z_f of $1+e_3$ (fig. 7). The values of these strains is given by

$$(1+e_1) = \frac{1}{2} [2 + \gamma^2 + \gamma(\gamma^2 + 4)^{1/2}] \quad 7.$$

$$e = 0 \quad 8.$$

$$(1+e_3) = \frac{1}{2} [2 + \gamma^2 - \gamma(\gamma^2 + 4)^{1/2}] \quad 9.$$

The orientations θ' of these principal strains measured from the x coordinate direction is given by

$$\tan 2\theta' = \frac{\gamma}{\gamma} \quad 10.$$

These two directions originated in two other orthogonal directions θ in the unstrained state given by

$$\tan 2\theta = \frac{-\gamma}{\gamma} \quad 11.$$

The strain is clearly a rotational strain, the finite rotation ω can be derived from

$$\tan \omega = \tan(\theta' - \theta) = \frac{-\gamma}{2} \quad 12.$$

Any passive plane marker is displaced by the shear such that, if its trace on the xz plane makes an initial angle α with the x -direction before deformation and an angle α' after deformation (fig. 7), then

$$\cot \alpha' = \cot \alpha + \tau \quad 13.$$

The mathematical relationships refer to a special choice of coordinate reference frame such that x is parallel to the shear direction. It is sometimes more useful to refer to the displacements with the shear direction at some angle ϕ to the x -direction. For example one may be interested in computing the effects of subjecting an element to superposed shear zones at differing angles (we will discuss this problem later).

If the shear zone is oriented at an angle ϕ to the x -direction, then from the geometry shown in fig. 8 an initial point (x, z) is displaced to a new position (x_1, z_1) according to the displacement matrix:

$$\begin{array}{ccc|c} 1 - \gamma \sin \phi \cos \phi & 0 & \gamma \cos^2 \phi & \\ 0 & 1 & 0 & \\ -\gamma \sin^2 \phi & 0 & 1 + \gamma \sin \phi \cos \phi & \end{array} \quad 14.$$

The values of the principal strains and rotations are identical to those of 7, 8, 9 and 12, but the orientation of the principal strains is given by

$$\tan 2\theta' = \frac{2 + \gamma \tan 2\phi}{\gamma + 2 \tan 2\phi} \quad 15.$$

The positions of the principal strain axes for the first small increment of simple shear are from eq. 10 oriented at 45° and 135° to the x -coordinate axis. The incremental strain axes X_i and Z_i are identically oriented and coincide with the principal strain rate axes. The principal strain rates \dot{e}_1 and \dot{e}_3 are related to the shearing strain rate $\dot{\gamma}_{zx}$ according to

$$\begin{array}{l} \dot{e}_1 = -\dot{e}_3 = \\ \dot{e}_2 = 0 \end{array} \quad 16.$$

and the rate of rotation, or vorticity $\dot{\omega}$ is

$$\dot{\omega} = \frac{\dot{\gamma}}{2} \quad 17.$$

During the first stages of deforming an initially homogeneous isotropic body the principal axes of stress σ_1 , σ_2 and σ_3 will coincide with the incremental strain axes X_i , Y_i and Z_i respectively, and the magnitudes of the strain rates will be some function of the magnitudes of the stresses. As deformation proceeds, however, it is very common for the deforming mass to become strongly anisotropic as a result of the large finite strains which build up with large values of shearing strain γ . During these mature stages of deformation, although the principal strain rate and incremental strain axes X_i and Z_i must remain at 45° and 135° to the x-direction because of geometric constraints of the process, the principal stress axes σ_1 , and σ_3 will almost certainly be influenced by the increasing obliquity of the anisotropic fabric to these strain axes, and are therefore unlikely to coincide with the strain increment axes.

Fabrics produced by strain

As simple shear displacement gets larger the angle θ' between the principal finite elongation X_f and the finite plane of flattening $X_f Y_f$ make a progressively smaller angle with the shear direction x and shear zone walls (fig. 9). In many naturally deformed rocks in ductile shear zones a statistically preferred orientation of the minerals is developed in this finite flattening plane, giving rise to schistosity or cleavage. Within this planar tectonically induced fabric there is often a strain related linear orientation of the mineral components parallel to the finite greatest direction X_f on the schistosity plane. Progressive shearing leads to a progressive intensification of these planar and linear fabrics and a change in their orientation. Because ductile shear zones generally show a maximum displacement gradient in the zone centre, the gradient decreasing towards the margin it follows that the tectonically induced planar fabrics of shear zones generally show a characteristic sigmoidal form as shown in fig. 10.

Because the orientation of the $X_f Y_f$ surface with its coincident schistosity is a function of shear strain, it follows that the angle between schistosity and shear zone walls can be used to measure the shear zone parallel shear strain, and shape of the strain ellipsoid at that point. This technique, originally developed by Ramsay and Graham (1970), can be extended so as to integrate successive finite shear strains across a shear zone profile and, in so doing, to calculate the total differential displacement across a zone with heterogeneously developed shear strain (fig. 11). The results of the application of this technique on a regional scale are often very striking. For example, Beach (1974) showed that the group of shear zones making up a large movement zone in the frontal part of the Precambrian Laxfordian orogenic belt has a minimum displacement of 25 km, which can be resolved into a horizontal component of 18 km and a vertical component of 16 km.

The main problem of using this technique occurs where the shear displacement gradients are high. For example angles between schistosity and shear zone wall of 5.7° and 2.9° imply a difference of displacement of 10γ and 20γ respectively. With such high displacement gradients (and high finite strains) very small observation errors can lead to gross errors in computing the total displacement across the shear zone.

Deformation of pre-existing planar features

It was pointed out above (eq. 13 and fig. 1) that pre-existing planar features have their orientation modified where they pass through a ductile shear zone. It should be stressed that this change of angle only applies to the angle made by the original plane measured in the xy profile of the shear zone. Although it does not refer directly to the dihedral angle between the two planes, the modification of such dihedral angles can be easily computed (Ramsay 1967, 504-508). Where a set of sub-parallel planes crosses a shear zone with a varying displacement gradient profile, variations in the amount of local shear imposed on the layers sets up folding with a characteristic similar style (fig. 12). The axes of these folds is controlled by the line intersection of the initial plane structure and the xy plane of the zone. Their axial planes are parallel to the zone, and are located at positions dependent upon the local changes in shear gradient and the orientation of the plane before shearing (Ramsay 1967, 508-509). Variations in orientation of the layers can be used to compute finite strain values at different positions in the zone (according to the relationships of eq. 12) and can be used to determine the total shear across a zone. The practical application of this technique for calculating displacements has been shown in a beautiful study of basic dykes passing through shear zones along the Nagssugtoqidian orogenic front of west Greenland (Escher et al. 1975). Here it has been shown that shear zones have led to an average 66 per cent crustal shortening of the deep-level basement rocks of this orogen over a distance of about 100 km. Clearly ductile shear zones in this region have a very great structural significance.

The change in orientation of pre-existing planar features is accompanied by other geometric modifications. If there is no ductility contrast between the layer and its matrix it may either increase (fig. 13; Ai, ii) or decrease (fig. 13; Bi, ii) in thickness as a result of passive shearing. The conditions for increases or decreases in thickness are specified in fig. 13. The change of thickness (t to t') is given by

$$t' = \frac{\sin \alpha'}{\sin \alpha} t \quad 18.$$

If the angular conditions are such that $\alpha > 90^\circ$ and $\alpha' > 180^\circ - \alpha$ the progressively sheared layer first thickens, then thins to its original width (where $\alpha' = 180^\circ - \alpha$) then subsequently becomes thinner than its original value. The reverse sequence - thinning followed by thickening - can never occur in progressive simple shear. If the layers undergoing shear are not passive, but show a competent contrast with their matrix, then mechanical instabilities are set up in the competent layer which lead to the formation of new structures. If the layer is shortened during deformation (equivalent to the passive layer undergoing thickening), buckle folds develop with wavelengths characteristic of the thickness of the competent layer, and the extent of the ductility contrast (fig. 13; A, iii). The orientation of the axes of these folds depends upon the obliquity of the layer to the shear zone. It depends upon the direction of maximum shortening within the surface of the contracting competent layer and not simply relate to the principal displacement features of the shear system (as with the similar folds produced by passive layer deflection). For example, the fold axes

will not generally form parallel to y or Y_f and they need not lie in the xy shear plane or the $X_f Y_f$ plane of the finite strain ellipsoid (fig. 14).

If the competent layer is oriented in a position such that it becomes extended by the shearing, then pinch and swell or boudinage structure may be developed (fig. 13; B, iii). The necking directions of these boudins will be controlled by the maximum extension directions in the layer being sheared and will not in general, be related to the xy plane of the shear zone. In fact the two dimensional strains within the deflected competent layer are likely to involve both extensions and contractions. Both buckle folding and boudinage then develop together, probably one of these being dominant over the other (fig. 14).

One further possible complication that can develop in competent layers occurs where the initial orientation and value of shear lead to layer shortening followed by extension. Here we have the possibility of layers developing buckle folds, which subsequently become unfolded or boudinaged (fig. 13, C, iii).

Deformation of pre-existing linear features

Most pre-existing line elements (e.g. pre-shear zone fold axes) become deflected and take up new orientations in the shear zone. These lines come to lie closer to the direction of the shear (x) and move on a plane locus which connects their initial orientation with the x -direction (fig. 15) (Weiss, 1959, Ramsay, 1960, 1967). In projection they move on a great circle locus (fig. 15, $p p' p''$ and $q q' q''$). The only exceptions to this general rule are those lines which lie in the plane of the shear zone xy ; these show no deflection (fig. 15, $r r' r''$). The axis of principal finite elongation X_f also moves on the xz plane towards the x -direction (fig. 15 $X_f' X_f''$). It is sometimes stated that all lines move towards the X_f direction during simple shear. In a sense this is geometrically correct, but in reality the X_f direction itself moves, and so the best fixed line element for reference is the x -direction.

The geometric effect of the drawing together of initially variably oriented linear directions is often very striking at high shear strains. Particularly interesting is the effect described by Cobbold et al (personal com. 1976) where folds outside a shear zone showing slight variations in their axial plunges are deformed in the shear zone so that their axes show extreme variations in plunge. The interlimb angles of these folds are drastically modified in the shear zone according to the principles discussed in the previous section. This effect is shown diagrammatically in fig. 16.

Terminations of shear zones

Ductile shear zones can, and in some cases do, come to a stop within the ductile environment. The boundary conditions for the ends of shear zones are much more complicated than for those of simple parallel sided zones with constant displacement profiles used to set up the six basic types described earlier. As far as I am aware there are no complete mathematical solutions for the possible finite strain and displacement variations where a shear zone terminates. However, one can proceed intuitively, and try and arrive at geometric resolutions which fit field observations and natural situations.

There appear to be an unlimited range of possible models: Alison and Ramsay (1979) have suggested that this range has two end members (fig. 17). One is a plane strain model, whereas the second is a solution where all displacements take place normal to the shear zone (in the y direction). Both give rise to characteristic strain fields, and the non-plane strain model sets up constrictive and flattening deformations on either side of the shear zone tip. If the boundary deflections seen in the plane strain model are constrained, then there will be a tendency for the tips of right handed shear zones to bend and propagate in a clockwise sense with respect to the main shear zone, and left handed shear zones to propagate in an anticlockwise sense (fig. 18). This effect could be the reason for the crossing or merging of shear zones of similar displacement sense that have been recorded in some regions.

Conjugate shear zones

Naturally occurring ductile shear zones generally occur in conjugate sets. A set of parallel shear zones is unable to change the overall shape of a rock mass into all possible new configurations, whereas deformation proceeding by two differently oriented sets of zones of the six basic types of shear zone can accommodate any overall regional shape change.

The Anderson theory of conjugate brittle shear zones (1951) is well established in the structural geology literature, and is usually given as the standard explanation of fault sets in most text books (including my own!). However, the more experience I get from observing naturally deformed rocks the less satisfactory I find this theory. The Anderson view is that conjugate faults are related to the directions of principal stress at the time of rock failure: faults develop along surfaces of high shear stress, but not along the two where maximum shear stress acts and which are oriented at 45 degrees to the σ_1 and σ_3 directions (notation $\sigma_1 \gg \sigma_2 \gg \sigma_3$ in tensile values). Anderson's view was that the faults formed along surfaces containing σ_2 , oriented at 30 degrees to the maximum compressive stress, because of internal friction in the rock. This theory predicts that faults at initiations should be symmetrically paired, with the greatest compressive stress at failure bisecting the acute angle between the faults sets. Anderson pointed out that, because the earth's surface is effectively a plane of zero shearing stress, only three possible orientations of principal stress axes can occur, and therefore only three types of faults are found near the surface: σ_3 normal to the earth's surface giving rise to high angle normal faults, σ_2 normal to vertical strike-slip faults, and σ_1 normal to low angle thrusts (fig. 19). Several observations worry me about the validity of this theory. The first is that faults do not generally occur in conjugate sets as predicted from this theory. Along orogenic fronts all the main thrusts seem to show displacements in one sense, and so-called "back thrusts" are regarded by most field geologists as something exceptional. In the classic regions of strike-slip faulting (the Great Glen fault system of Scotland, the Jura Mountains of Switzerland and France, the San Andreas systems - although this might be a transform fault situation) everyone agrees on the direction and movement sense of one set of faults, whereas the conjugate set is either absent, or its

location is at best open to discussion. Perhaps the only fault systems which occur in more or less equally developed conjugate sets are the normal fault, horst-and graben-structures of the Rhine valley, the African **rift** systems or the Basin and Range region. However, even these show certain regions where the **faults** are almost exclusively of one sense, and others where they have the opposite inclination and sense of throw. **It** is a point of interest to discuss the possibility of synchronously developing crossing conjugate faults; is **it** possible for them to act at the same time? The reason of the improbability of synchronicity is concerned with compatibility: synchronous faulting leads to large volume increases beneath the down-dropped keystone wedge (fig. 20, A). **If** two conjugate faults alternate in their activity, first one moves, then the other, the geometry becomes extremely complex at their intersection. **When** the old fault is reactivated its original counterpart on the other side of the conjugate fault is no longer in line, **so** new unfaulted rock has to be broken. This surely requires more work than is required by just continuing slip on one of the faults. The geometry of repeated activation of conjugate faults is a real "mess" - Although **it** is not completely disorganised. I would pose a question to those who have a practical experience of fault intersections - do you really see geometric situations like that of fig. 20, B?

These discussions of the geometry of brittle shear zones suggest that the orthorhombic **symmetry** of the stress tensor is not reflected by the faults produced by that stress, and that we might with profit query the Anderson theory.

Ductile shear zones of types **(i)(iii)**, (iv) and (vi) usually occur in conjugate sets. One of the shear zone sets has a right handed shear displacement, the other is left handed. One of the most striking features of the angular relationships of the two sets of shears, which is in contrast to **that** of brittle shear zones, is that **it** is the obtuse angle (generally $90^\circ - 130^\circ$) between the shears which faces the greatest shortening direction of the system (fig. 19, Bi). At the intersection of the two conjugate sets one finds complex but not chaotic structures with much higher finite strains than are found in either of the two individual zones. The geometric features of the intersectional generally indicate that one shear zone is later than the other and displaces **it** (eg Fig. 19, Bi, the l-hand shear is later than the r-hand shear). **Why** this is **so** is a point **worth** discussion. I think **it** is connected with compatibility problems similar to those existing at the cross-over positions of brittle shear zones. The geometric problem is shown **diagrammatically** in fig. 21. This shows the strains in the sector where the displacements of the two simple shears (of equal shear strain) are superposed. The strain state produced by shearing on zone a is followed **by** shearing **on** zone b) (fig. 21, A) is not the same as that where shearing on zone a follows that of zone b (fig. 21, B). The strain ellipses have the same ellipticity, but their orientations differ. This may be verified mathematically for all angles of shear zone intersets using the displacement matrix of eq.14 first with values of γ, ϕ followed by $-\gamma, \phi' = 180^\circ - 2\phi$ and then reversing the displacement order: the two **total** displacement matrices are different.

The overall structural pattern of a region deformed by conjugate ductile shear zones is very characteristic: lozenge shaped areas of relatively undeformed material are bounded by shear zones of right and left handed aspect. Such a pattern will be familiar to any geologist who has mapped "**basement**" terrain for it is the normal deformation mode. The dimensions of the lozenge-shaped areas can vary from the centimetric to the **kilometric** scale.

The geometric effect of developing conjugate ductile shear zones on a series of parallel passive layer marker horizons depends upon the orientation of this layering to the bulk strains produced by the conjugate shears.

If the layering is sub-perpendicular to the maximum bulk shortening of the whole mass the crossing shear zones give rise to a thinning of the layers, a feature which is especially noticeable where the layers pass through the intersection region of two conjugate shears (fig. 19, **Bii**). The pronounced **semi-symmetric** thinning that occurs at the shear zone intersection has been termed internal boudinage (Cobbold et al 1971). Internal boudinage, in contrast to normal boudinage, is not dependant upon competent contrasts between the boudinaged layer and its surrounding matrix, the neck zone being located only by the chance location of the intersecting shears. Internal boudinage is developed in rocks which are rheologically uniform or practically uniform. This structure might be considered to be the ductile equivalent of normal faulting (fig. 19, cf. **Aii** and **Bii**).

If the layering is subnormal to the greatest extension of bulk strain then conjugate folds result. Such conjugate folds need not be genetically controlled by the layering, or by the thicknesses of individual beds within the multilayer packet although the planar anisotropy might be the instigator of the shear zones (Cobbold et al. 1971). This conjugate fold structure might be considered as the ductile homologue of brittle shear zone thrust faults (fig. 19, cf. **Aiii** and **Biii**).

Mechanical aspects of ductile shear zones

Ductile shear zones can form in a variety of rock types which can have anisotropic or isotropic properties. In anisotropic media (eg rock multilayers, or rocks with planar anisotropy such as schistosity or cleavage) the features of the anisotropy may control the angular relationships of the conjugate zone system (Cobbold et al 1971), but the fact that shear zones of identical aspect occur in anisotropic granites and gabbros shows that an initial anisotropy is not an essential **feature** of shear zone development.

Ductile shear zones presumably initiate at some point in the rock and propagate from that point. The rapidly developing higher strains at the initiation point must trigger the system. The spacing of adjacent shear zones depends upon the numbers of such initiation points. Their left or right handed aspect is probably controlled by the shape of the strain irregularity at their initiation point and by the nature of the displacement changes taking place along the "boundary" of the rock mass.

The shear zones will propagate until they meet an obstruction (eg a competent layer surface), and they can be "reflected" off this surface (Cobbold, 1976). In making a "reflection" they change their displacement sense.

Ductile shear zones are characteristically without discrete rupture surfaces which might mechanically weaken the shear zone. At first sight **it** is surprising that the later ductile **shearing** displacements in the zone appear to continue to act along the zone of previously deformed rock. For example, during the deformation of metal by plastic flow processes **it** is usual for the metal crystals to become more difficult to deform as they become strained. Such strain hardening is generally attributed to the interlocking and piling up of dislocations inside the deformed crystals, a process which inhibits the free movement of the new dislocations necessary for the crystals to continue to deform. With ductile shear zones we see just the reverse of ~~this~~ hardening process. The strained rock becomes progressively easier to deform, a feature which one might term strain softening (Ramsay and Graham, 1970). The usual plastic field stress-strain curve with positive slope that is so often produced in laboratory deformations of metal and rock are clearly not appropriate to explain the rheological processes of shear zones, and **it** seems likely that the mechanisms of deformation appropriate to shear zones are not those of normal crystal plasticity. The rocks of shear zones commonly have a smaller grain size than the parent undeformed material outside the zone. This reduction in grain size could result from sub-grain growth in deformed crystals containing dislocations. **It** is quite well established that the deformability of crystalline aggregate increases as the grain size decreases (Schmid et al., 1977), so this feature might explain the observed strain softening. **It** also seems likely that in many environments chemical transfer of ~~material~~ **have** more influence than physical plasticity. Many shear zones show a markedly different mineralogy from their host rocks, and mineral reactions clearly imply a redistribution of elements in the shear zone. These changes of metamorphic facies ~~may~~ be isochemical, but shear zones often show a pronounced increase in the amounts of certain elements above that of the unsheared rock. For example, the ductile shear zones of the Laxfordian complex of North West Scotland cut through dry, potassium-poor granulite facies gneisses, yet these shear zones are characterized by biotite, hornblende and K-felspar, all indicating an enrichment in the shear zone of potassium and water. These shear zones appear to have acted as channelways for certain elements and ions, and **it** may be that fluid phases played a significant role in their formation and propagation. The amount of fluid passing through a shear zone can sometimes be shown to be very large. Fyfe (1976) has argued from the known solubilities of gold in brines that in shear zones where gold is concentrated enormous volumes of fluid must have flowed along the zone.

Does strain softening continue indefinitely; does **it** continue even when very high finite strains have been accomplished? A number of shear zones show shear strain-distance profiles which have a flat top to the profile curve (e.g. Ramsay and Graham, 1970- fig. 16). The first reaction one has on seeing these curves is that the flat top to the curve is just a result of the inability to distinguish shear strain values accurately when the shear strain is high. **It** has previously been pointed out

that an angular difference of schistosity planes of only three degrees separates strains of 10 from 20 γ . However, the validity of a number of these flat topped profiles is convincing, because the intensity of the schistosity fabric through the zone where the shear strain profile is flat is constant. The strain ellipsoids at shear strains of 10 and 20 γ have X_f/Z_f ratios of 102:1 and 402:1 respectively, and this difference should be seen as a difference in intensity of schistosity. If one accepts that these flat topped profiles are real and not an artifact of the measurement technique, it follows that the deformation might be held back in very highly strained rocks. One explanation of this holding back of continued deformation in the regions of high finite strains might relate to orthodox strain hardening. Another possible explanation concerns the progressive obliquity of the rock fabric in these highly strained regions to the orientation of the incremental strain axes. It has been shown that the geometry of the strain increments during simple shear is fixed; the principal incremental strains are always located at 45° and 135° to the shear zone walls. In contrast, the finite strain $X_f Y_f$ plane with its genetically related schistosity progressively takes up a position at a lower angle to the shear zone according to eq.10. If we assume that a line which already has been strained becomes correspondingly easier to strain in proportion to the finite strain (admittedly a somewhat naive view of strain softening) it is possible to compute the resistance to the incremental changes offered by the body as it changes its finite strain state (fig. 23). During the early stages of shear zone development, when the finite strain axes lie fairly close to the fixed incremental principal strains, the rock becomes progressively easier to deform. However, there comes a stage at which this progressive increase in deformability reaches a peak, and at increased finite strains the deformability decreases. It should be stressed that this conclusion does not result from changing the conditions of strainsoftening, but only depends on the increasing obliquity of the finite strains with the increments necessary for the zone to continue to develop. This result, although only mathematically and not mechanically described, perhaps gives a hint how the highly deformed parts of ductile shear zones might "lock up" without infringing the basic ideas of strain softening. It suggests that, once a critical level of finite strain has been reached, further displacement would be most easily accomplished by continued deformation of the less deformed marginal parts of the shear zone, and perhaps by widening the zone by deformation of the undeformed walls.

Those shear zones developed with volume change must involve chemical transfer. Where is the material which is removed from the zone deposited? In deep seated ductile shear zones in high metamorphic environments the extent of volume change is probably quite small, probably 1 per cent is a typical value - so there are no great problems. However, in low and medium grade metamorphic environments volume changes can be very large, 20-50 percent is not uncommon. The material removed from the shear zone may have some local sink, perhaps even in a nearby part of the shear zone of its source. For example, fig. 23 illustrates a shear zone in a greywacke layer situated in a slate belt environment where removal and redeposition of material appears to be on a very local basis. The shear zone consists of a series of sigmoidally curved, volume-loss, pressure solution stripes with an array

of en-echelon quartz filled extension fissures. These fissures appear to be the sink for the material removed from the pressure solution stripes. Because the removal of material was more or less synchronous with the deposition of the vein quartz, and because the two structures are at an angle to each other the pressure solution stripes "cut into" and displace the vein material.

In some pressure solution stripes there is no such obviously structurally controlled local sink for the removal material. In porous sediments one possible site for the deposition of material removed from solution zones is the pore space between the clastic grains, the new material being deposited as grain overgrowths. The brittle-ductile shear zones of en-echelon extension fissure arrays that are found in medium to high crustal levels appear to show an overall volume increase in the shear zone as a result of vein filling, and these might provide the sink for material removed from volume-loss shear zones at deeper levels.

Regional development of shear zones: relationships between deep level ductile zones and high level brittle zones.

Of particular interest to structural geologists is the behaviour of regional shear zones which pass from ductile types at deep levels to brittle types at higher crustal levels. This problem is of major importance in discussions of basement-cover relations in orogenic belts and in zones of regional crustal extension.

1. Contraction zones

Most, if not all, orogenic belts show an overall shortening of the sedimentary sheets of cover strata by means of thrusting and nappe gliding. These cover strata are often of continental shelf facies known to have been laid down on crystalline continental type basement. Where this basement is exposed in the orogenic zone it is usually seen to contain ductile shear zones. Reconstructing the overthrust and nappe piled sheets of sedimentary cover rock one concludes that shortenings of the order of hundreds of kilometers are general (Rocky Mountains, Appalachians, Alps are good, well documented examples). By considering the relations of these thrust sheets as a problem of geometric compatibility it can be shown that many of the folds and overall geometry of the nappes are related to the movement of the sheet over step-like ramps in the underlying irregular sub-thrust surface, these ramps being controlled by lithological guides of the more competent strata (fig. 24).

Where the basement has suffered shortening, but does not form true fold nappes (eg. in the external crystalline alpine massifs of the Aar, Mt. Blanc and Belledonne) the zones of ductile shear are usually steeply inclined (50° to 80°) towards the more internal parts of the orogenic zone. Typically these zones form a single set **and are** not conjugate. Where the basement forms cores to large scale, recumbent fold nappes (eg in the Alpine Pennine zone) the shear zones are generally more gently inclined. The relationships of deep level ductile shear zones with higher

level thrust that accord best with my own experience are shown schematically in fig. 25. The shortening of the basement leads to an uplift of the internal part as a result of the crustal thickening. The ductile shear zones pass upwards across the basement-cover unconformity, and the ductile shear that is common to both basement and cover here accounts for many of the changes in angular relationships between the two series of rocks. Probably many of the "slate belts" of the world occur in these regions of highly deformed lower cover strata. At higher levels the ductile shear zone is transformed into a brittle-ductile zone with the upper side of the zone showing less ductile features than the lower. At still higher levels the internal deformation of the cover sediments becomes less marked and is mostly related to displacements associated with flexural slip movements in incompetent layers within the succession. The cover shortening that is not taken up by these folds is accomplished by low angle thrusts passing into strata guided "glide sheets". The thrusts below the glide sheets move to progressively higher stratigraphic levels outwards from the orogenic zone, making high angle steps where they cut across competent layers. The brittle shear zone eventually reaches the surface at the thrust toe.

In the more internal parts of the orogenic belt the basement is more thoroughly deformed by ductile shear zones, generally of conjugate types. Recrystallisation processes accompany the development of the shear zones. The ductile shear zones pass from basement to cover so that folds of similar style found in the basement penetrate large distances into the cover strata. The ductile shearing is often so intense that the original basement-cover unconformity is often completely obliterated, and can pass almost unrecognised on geometric criteria. In the internal regions there are often a succession of ductile shear zones developed during the orogenic history, leading to the formation of superimposed folding. The latest shear zones are formed in conditions of declining metamorphic grade and generally produce crenulation of previously acquired schistosity fabrics. These late shear zones may be inclined in the opposite direction to the earlier main shear zones, and this leads to so-called backfolding relationships (eg South Gotthard Massif and the wellknown Mischabel **Backfold** in the Alps, Ayrton and Ramsay, 1974). The transition from ductile to brittle shear zones often occur high in the stratigraphic cover succession.

2. Extension zones

The structural features of the **upper crustal** levels are quite well known from studies of graben and **rift** systems, but the structural relationships of these high level phenomena to those of the underlying basement are less well known. Moderately to steeply inclined normal faults (45° to 60°) predominate at the high levels, the fault plane geometry being guided by strata competence, being steepest in the most competent layers. Normal fault displacements over the irregular fault surfaces lead to geometric features related to compatibility problems in the same way that geometric problems arise from pushing low angle thrust sheets **across** their underlying ramps and flats (fig. 26).

At deeper levels the rheological contrasts of the cover layers become less marked, and we pass through a lower angle brittle-ductile shear zone transition into conjugate low angle ductile shear zones. These shear zones enable the deep crustal levels to extend: they can give rise to internal boudinage or to conjugate folding of the layering in the basement depending upon the orientation of this layering. Conjugate pairing of the shear zones is necessary for a complete horst-graben tectonic pattern at higher levels. Basement-cover angular unconformities will be geometrically modified where crossed by ductile shear zones (see fig. 26).

References

- Allison, I. and J.G. Ramsay, 1979, Structural analysis of shear zones in a deformed granite from the Pennine zone, Swiss Alps. In Press. Schweiz. min. pet. Mitt.
- Anderson, E.M., 1951, The dynamics of faulting. Oliver and Boyd Ltd. Edinburgh.
- Ayrton, S.N. and J.G. Ramsay, 1974, Tectonic and metamorphic events in the Alps. Schweiz. min.pet.Mitt., 54, 609-639.
- Beach, A. 1974, The measurement and significance of displacements on Laxfordian shear zones, North-west Scotland. Proc. Geol. Assoc. London, 85, 13-21.
- Cobbold, P.R., J.W. Colgrove and J.M. Summers, 1971, Development of internal structure in deformed anisotropic rocks. Tectonophysics, 12, 23-53.
- Escher, A., J.C. Escher and J. Watterson, 1975, The reorientation of the Kangâmiut dicke swarm, West Greenland. Can.Jour. Earth.Sci. 12, 158-173.
- Ramsay, J.G., 1960, The deformation of lineations in areas of repeated folding. Jour.Geol. 68, 75-93.
- 1967, Folding and fracturing of rocks, McGraw Hill, N.Y.
 - and R.H. Graham, 1970, Strain variation in shear belts. Can.Jour.Earth Sci. 7, 786-813.
- Teall, J.J.H., 1885, The metamorphosis of dolerite into hornblende schist. Quart.Jour.Geol.Soc.London 41, 133-145.
- Weiss, L.E. 1959, Structural analysis of the basement system at Turska, Keny, Overseas Geol. Mineral Resources (Gt.Brit.) 7, 3-35.

Fig.1

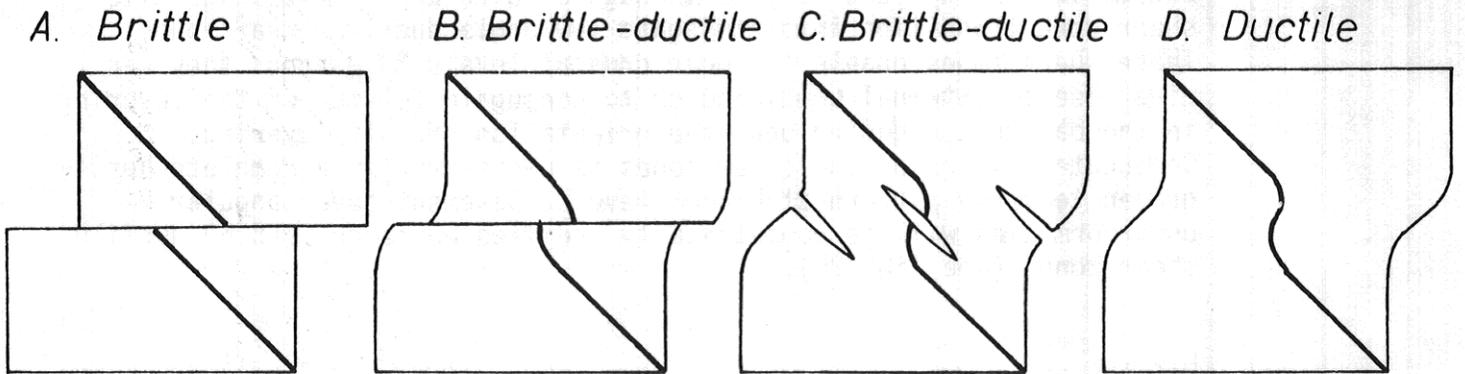


Fig.2

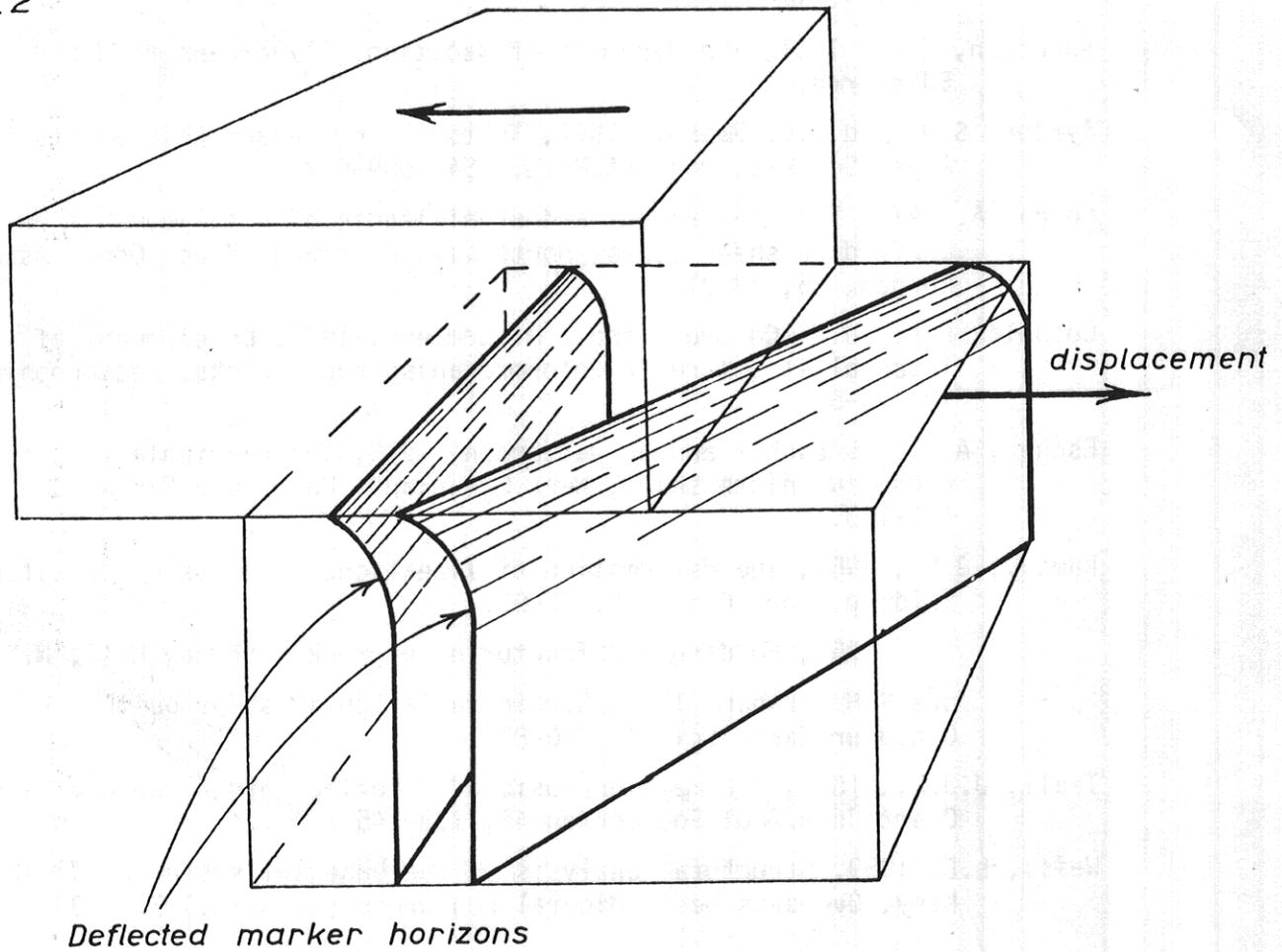


Fig. 3. Displacement fields of ductile shear zones

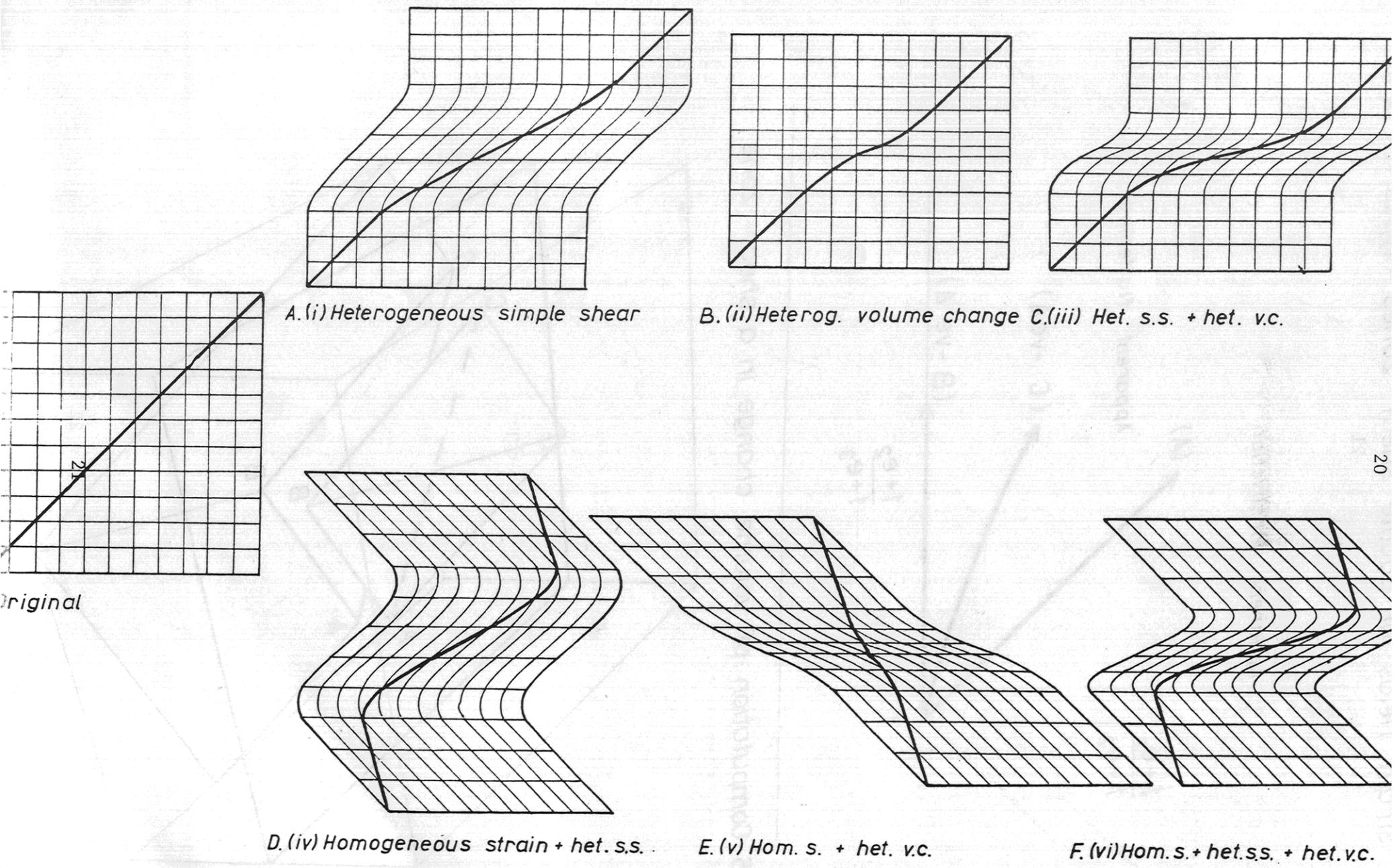


Fig.4 Strain fields of ductile shear zones - types I, II & III

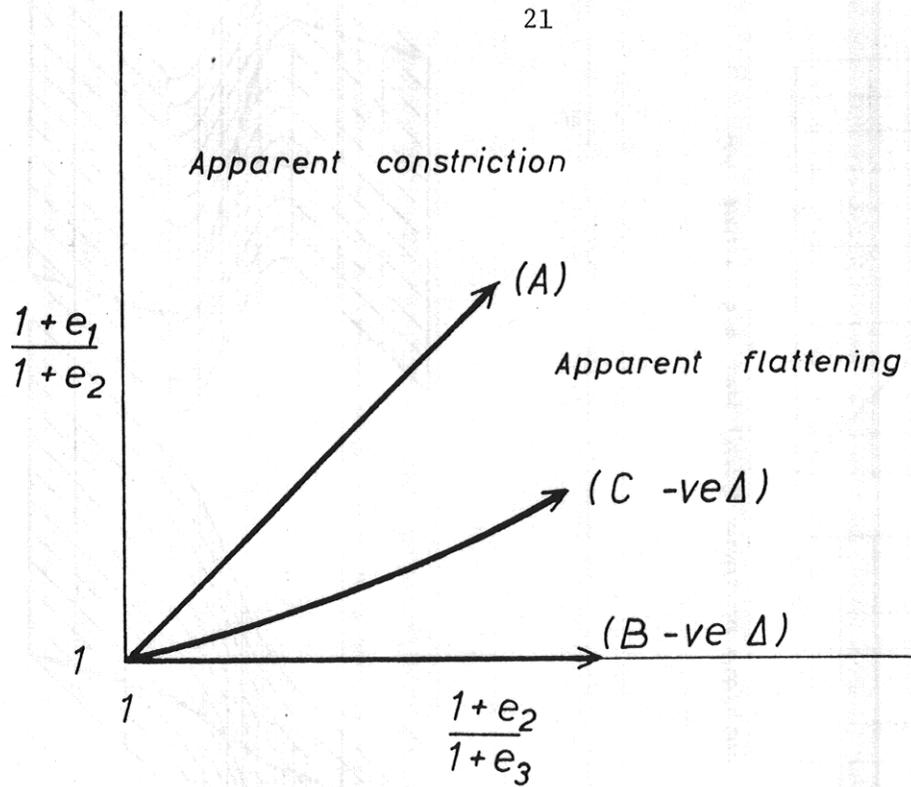


Fig.5 Computation of volume change in a shear zone

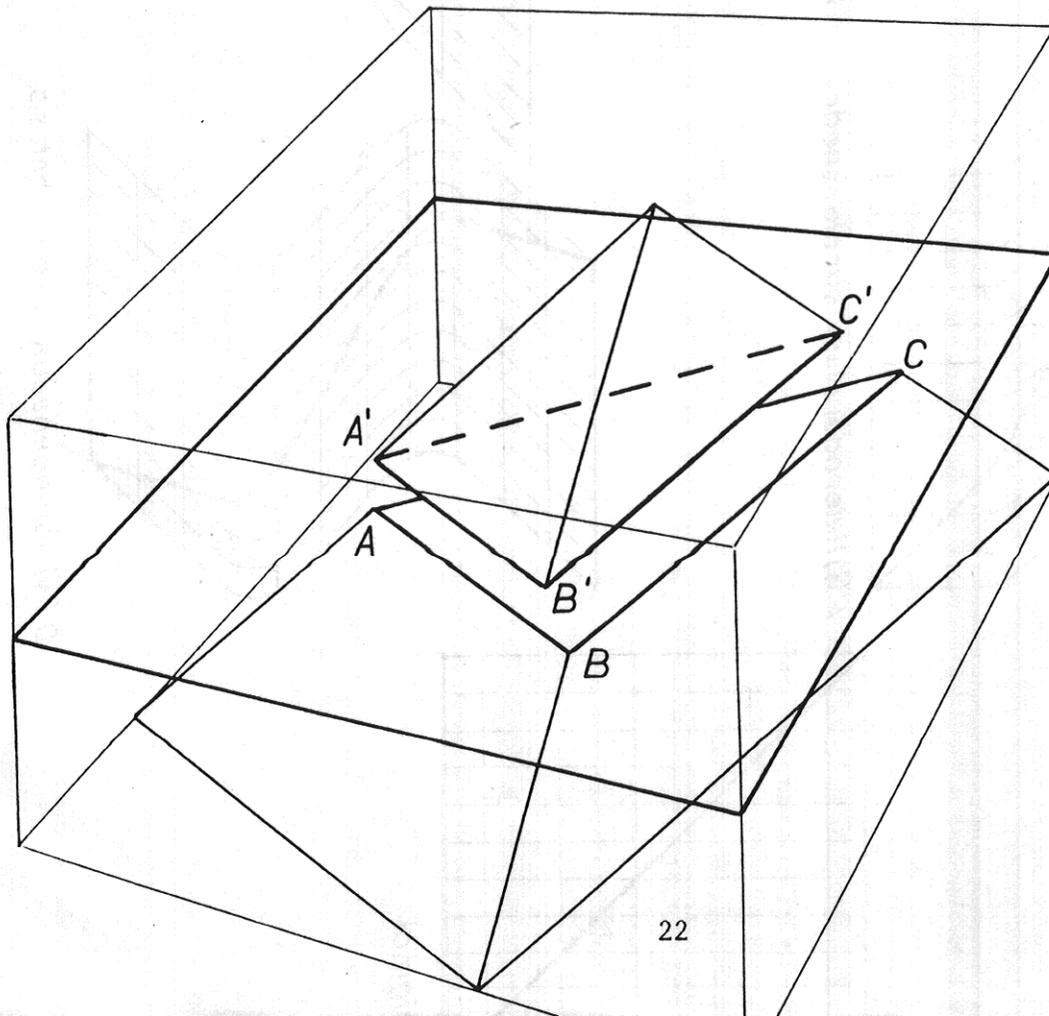


Fig. 6 Simple shear geometry 22

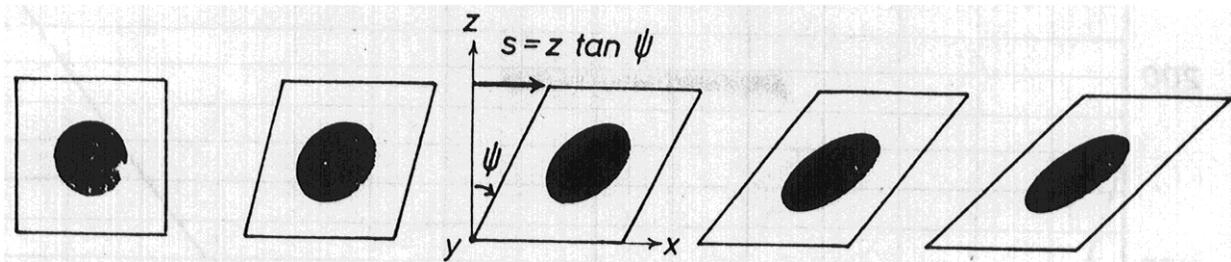


Fig. 7

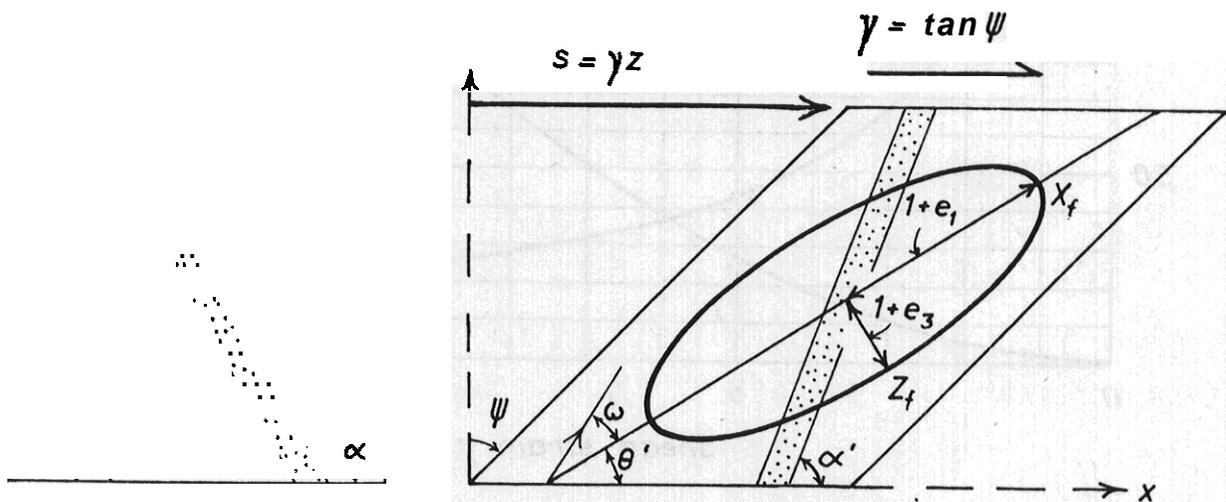


Fig. 8 Simple shear at an angle α to the co-ordinate frame

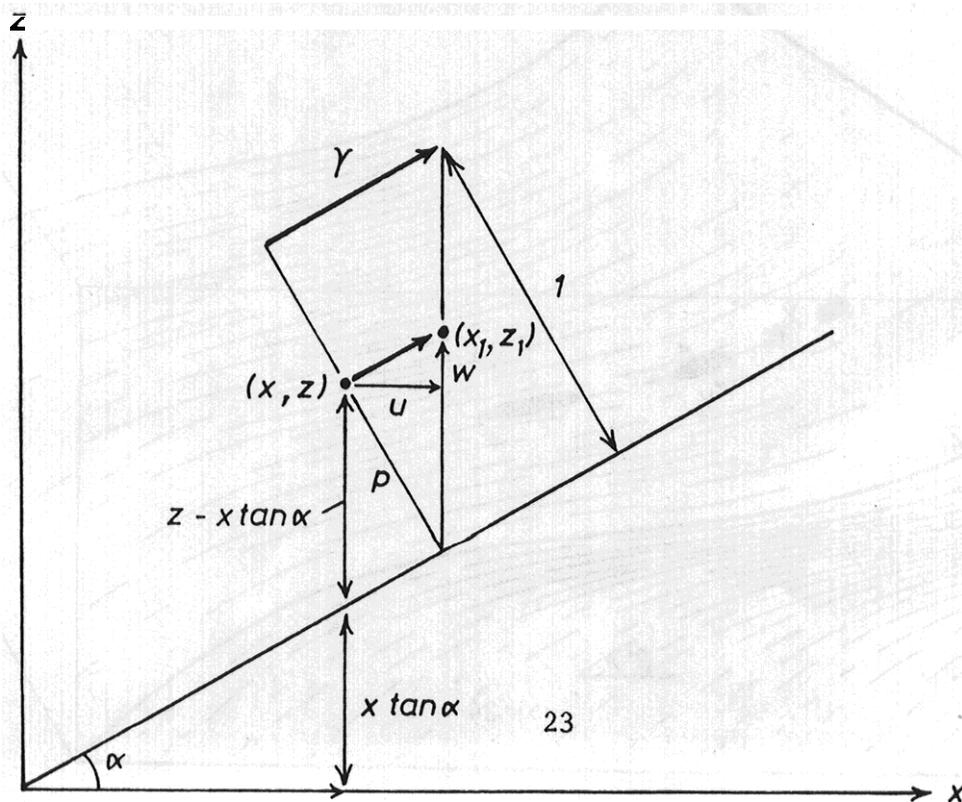


Fig.9 Simple shear, values and orientations of principal finite strains

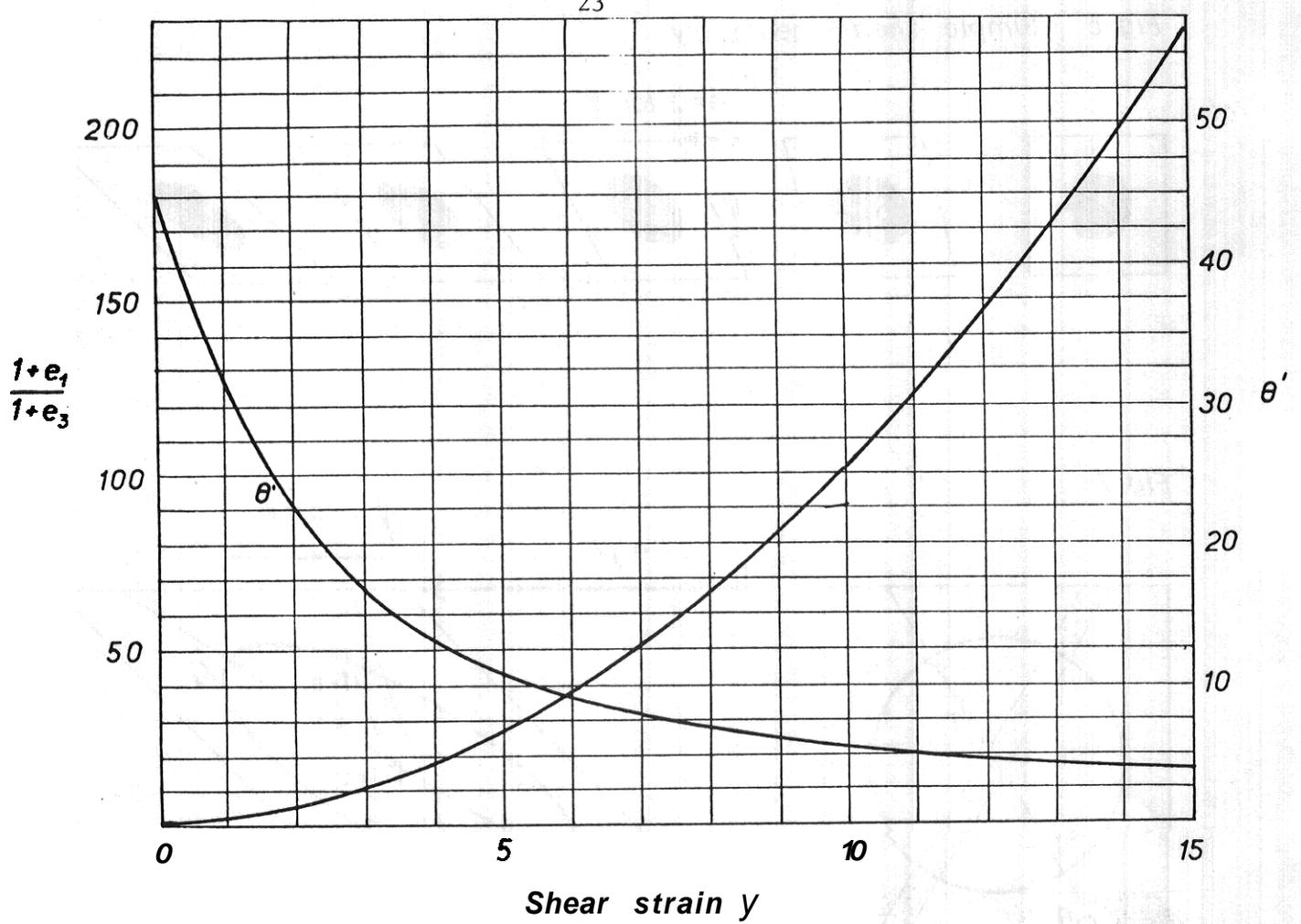
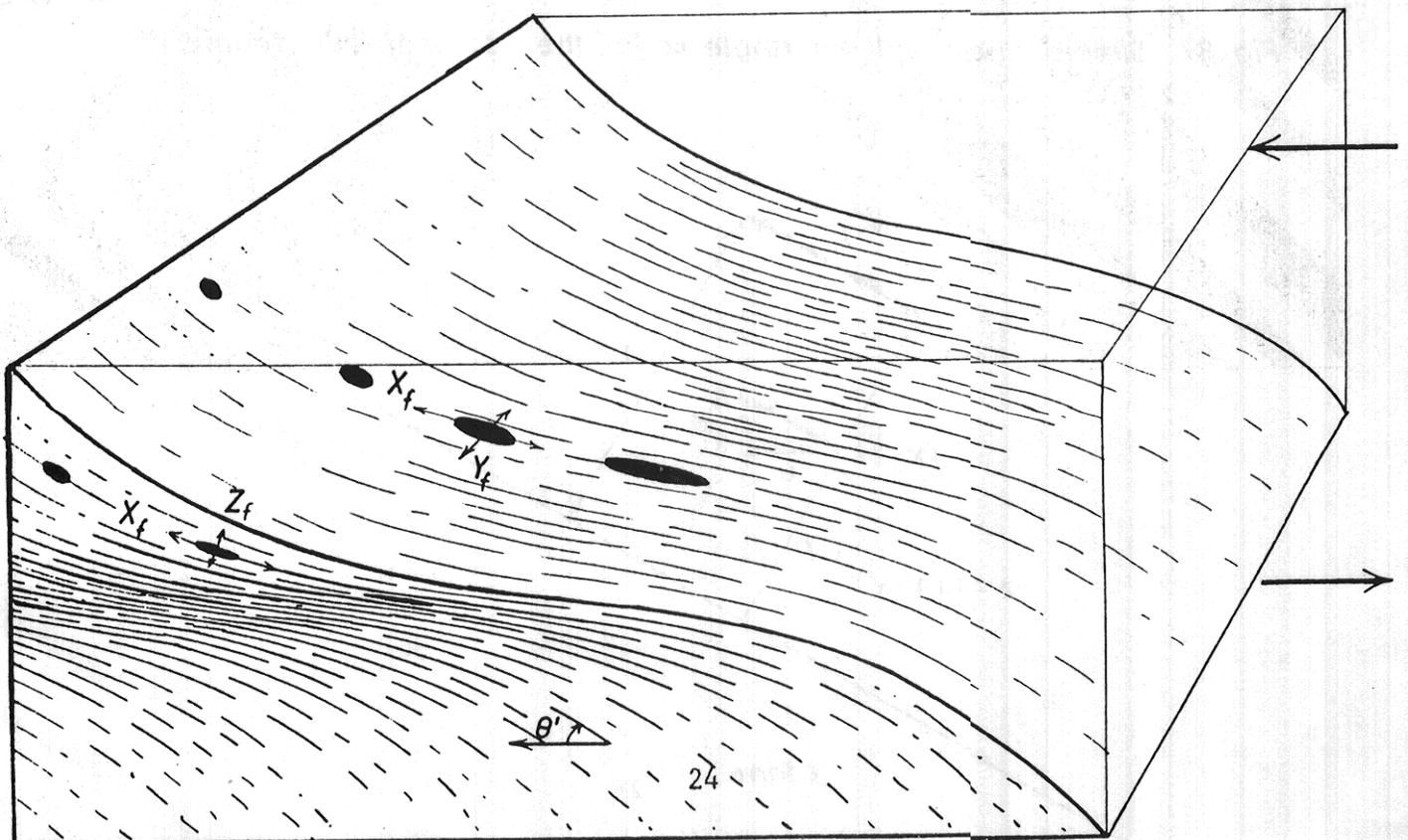
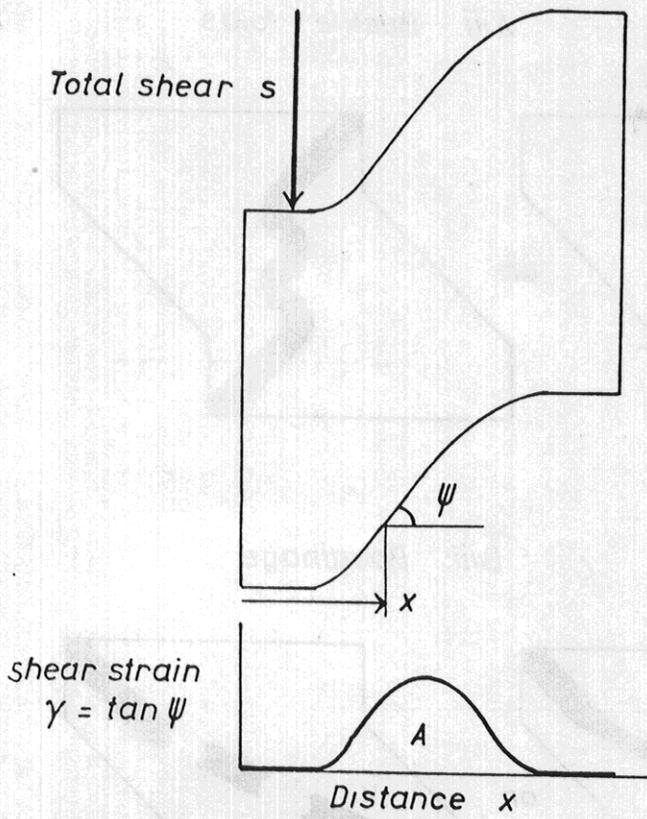


Fig.10 Fabric in a ductile shear zone





$$s = \int_0^x \gamma dx \quad \text{which is the area } A \text{ under the strain/distance curve}$$

Fig. 12 Folds of 'similar' style formed in passive layers

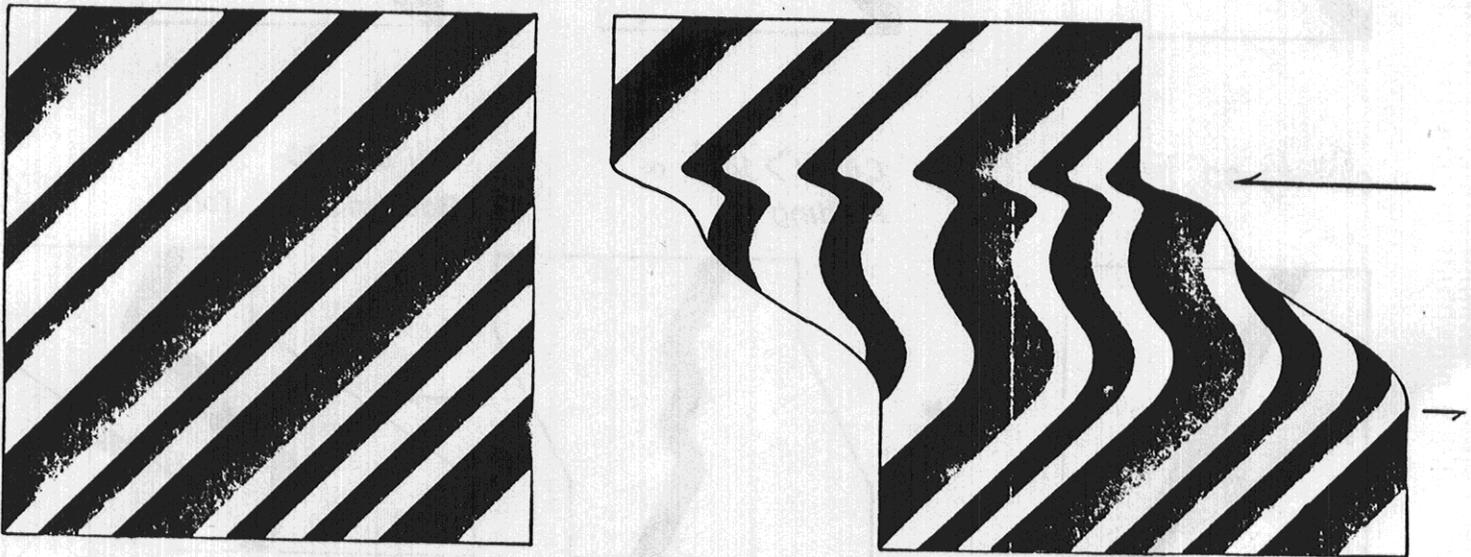


Fig.13 Effects on layers deflected in shear zones

25

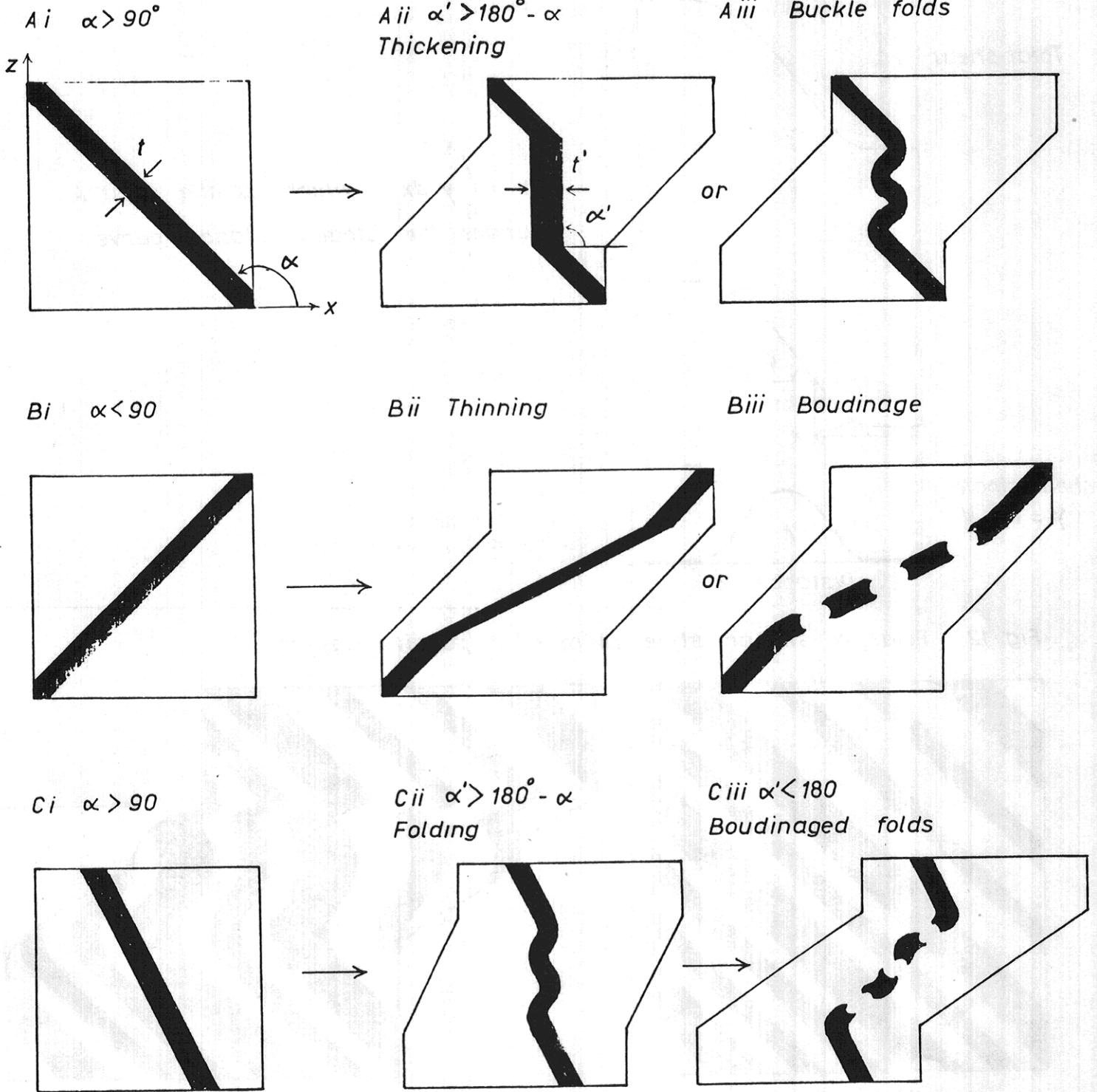
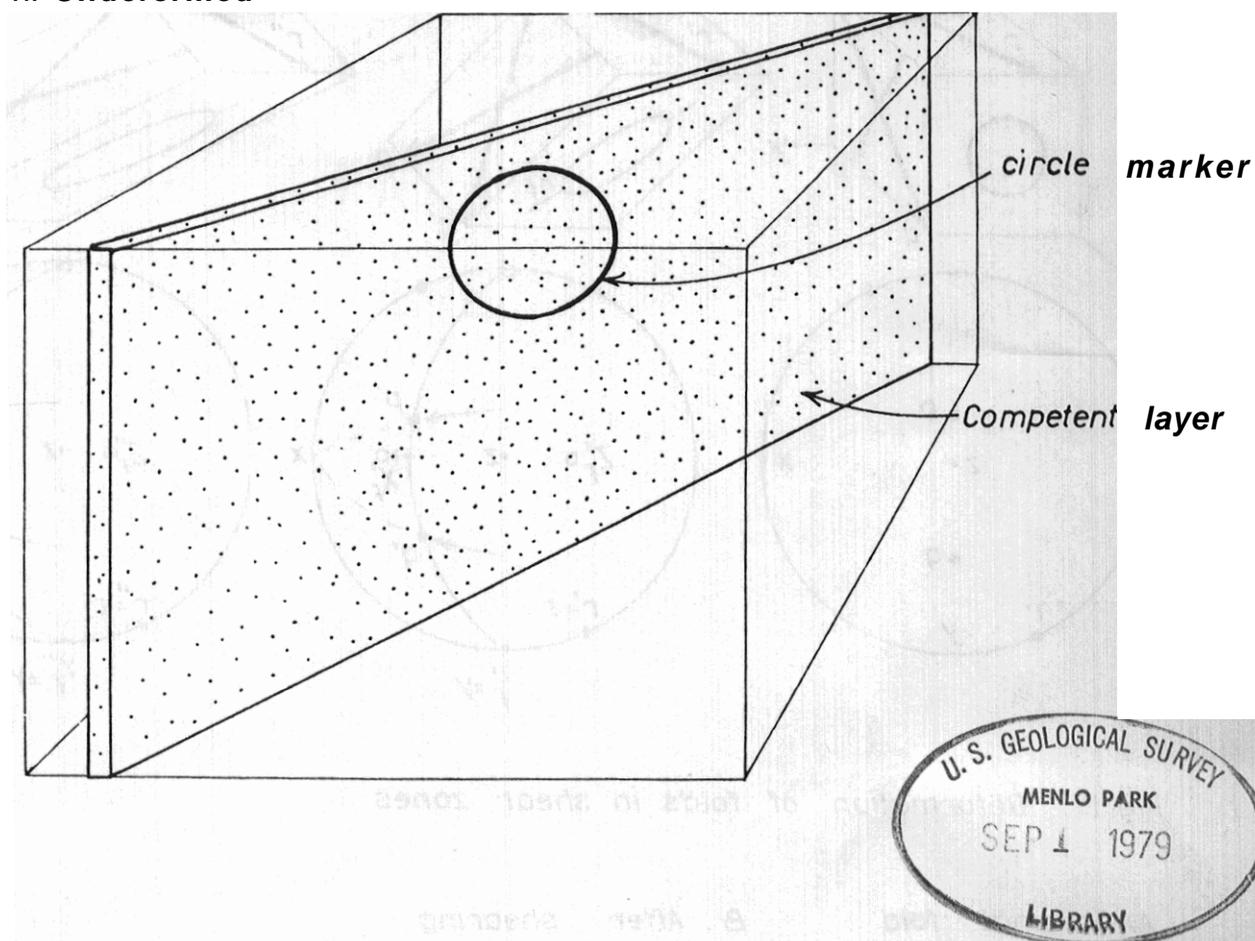


Fig. 14 Deformation of competent layers

A. Undeformed



B. Deformed development of buckle folds & boudins

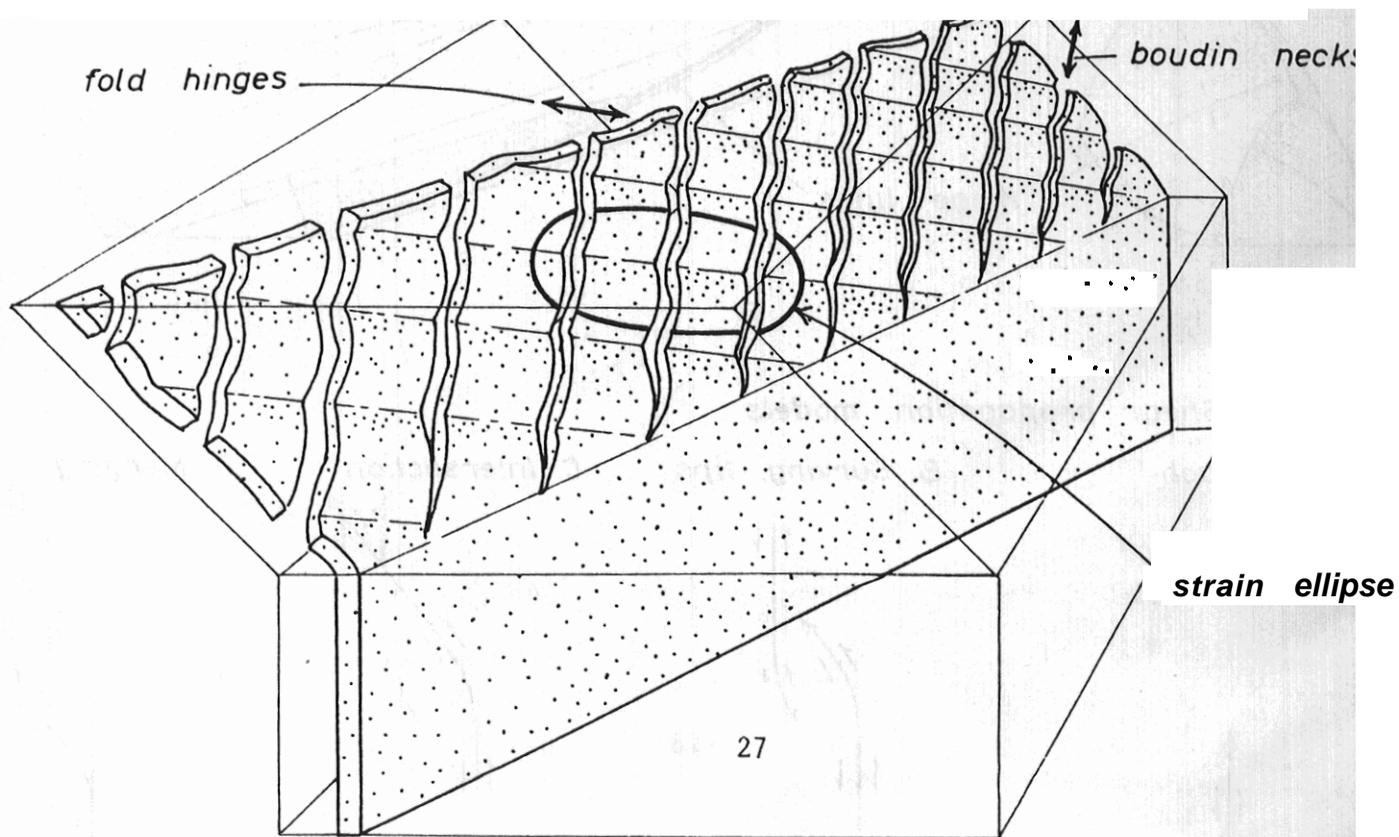


Fig. 15. Deformation of lineations in ductile shear zones

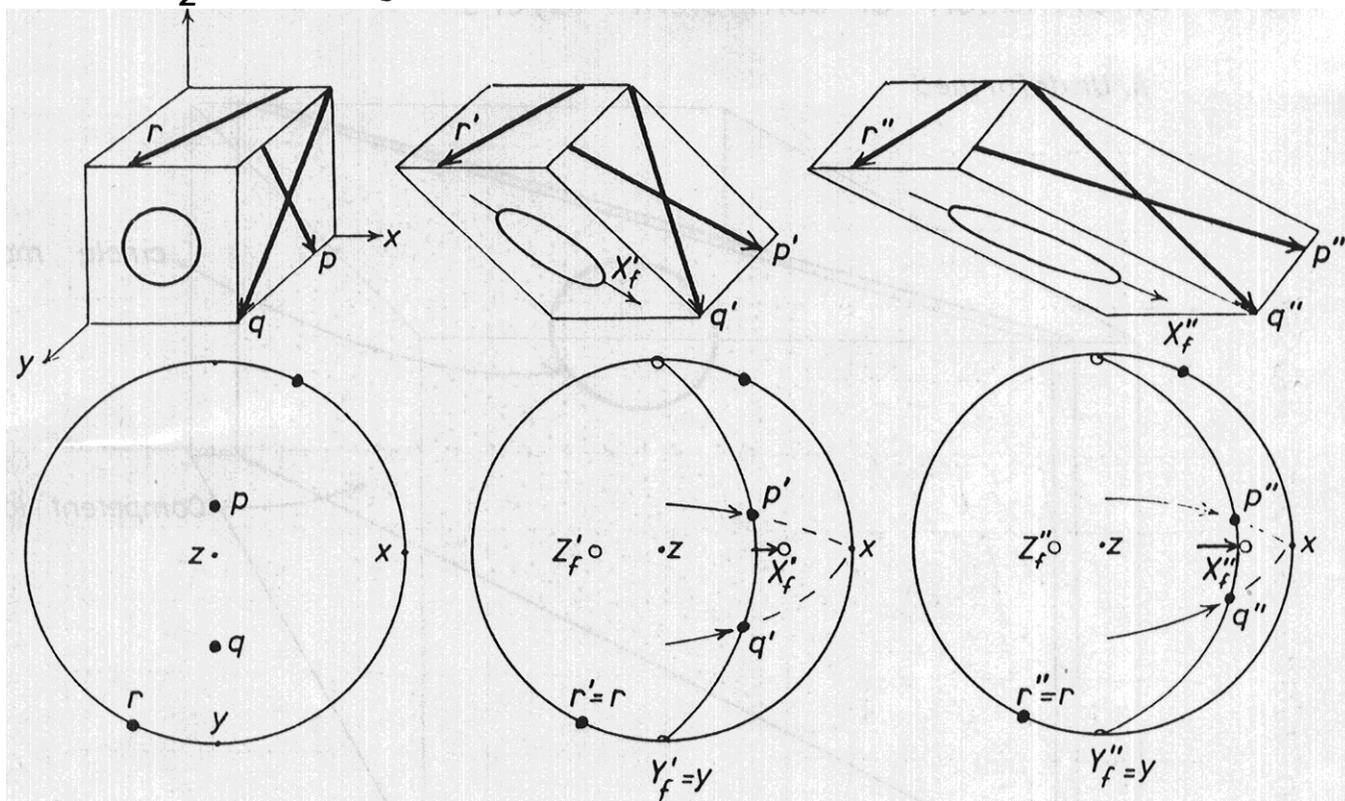


Fig. 16. Deformation of folds in shear zones

A. Original fold B. After shearing

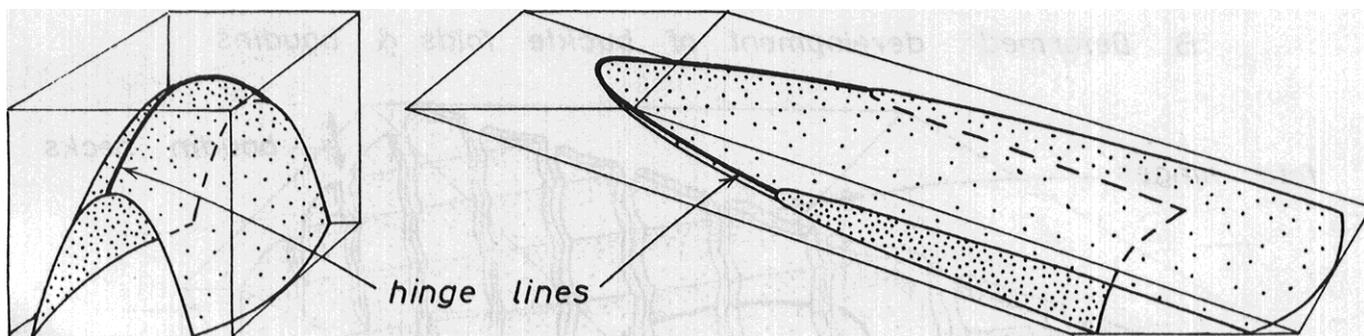
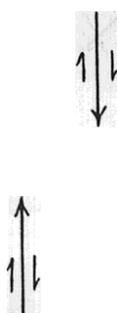
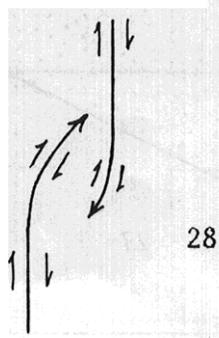


Fig. 18. Shear propagation models

A. Approach



B. Curving tips



C. Intersection

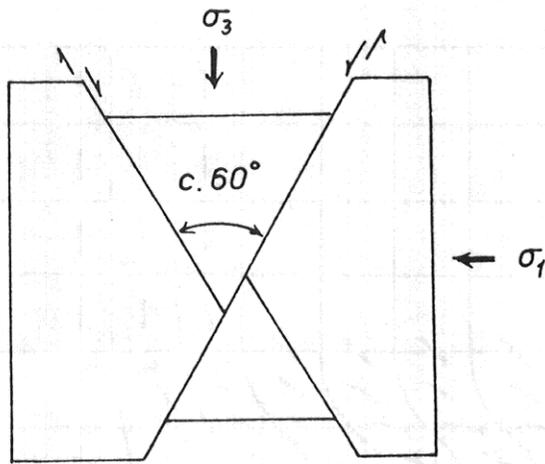


D. Merging

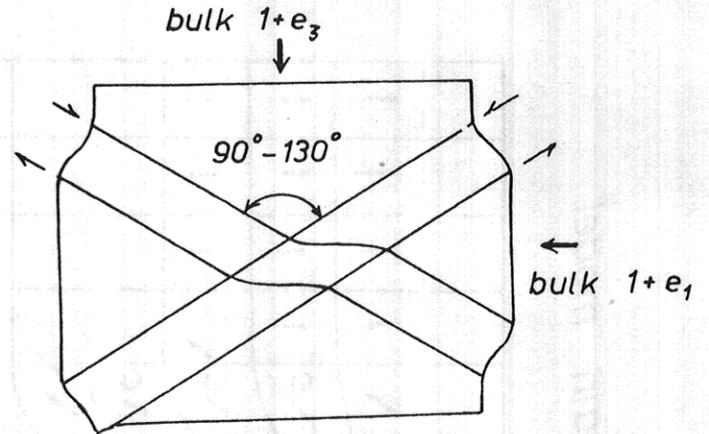


Fig.19 A comparison of brittle- & ductile shear zones

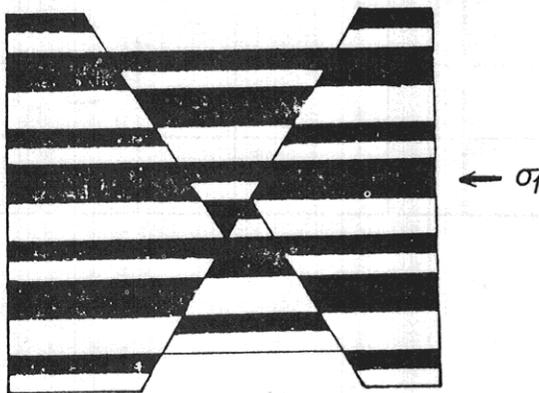
Ai Anderson's brittle faults



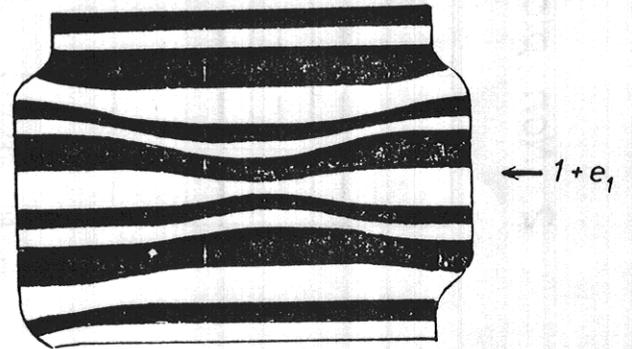
Bi Ductile shear zones



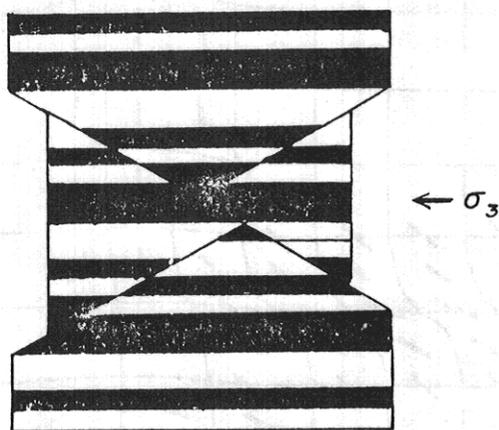
Aii Normal faults



Bii Internal boudinage



Aiii Thrust faults



Biii Conjugate folds

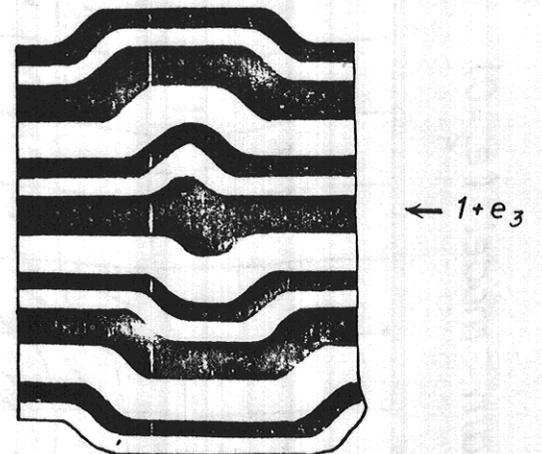
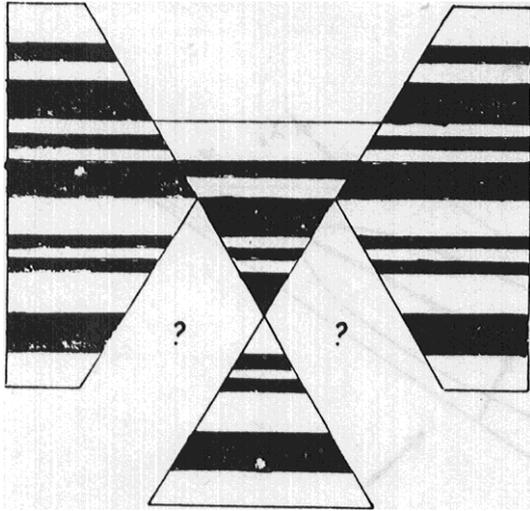


Fig.20. Space problem with conjugate normal faults

A Synchronous faults



B. Alternating fault activity

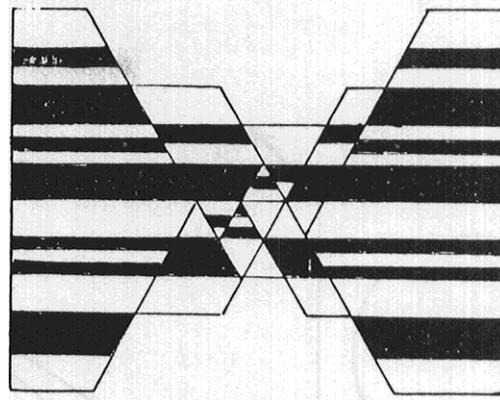


Fig.21. Finite strain at the intersections of ductile shear zones

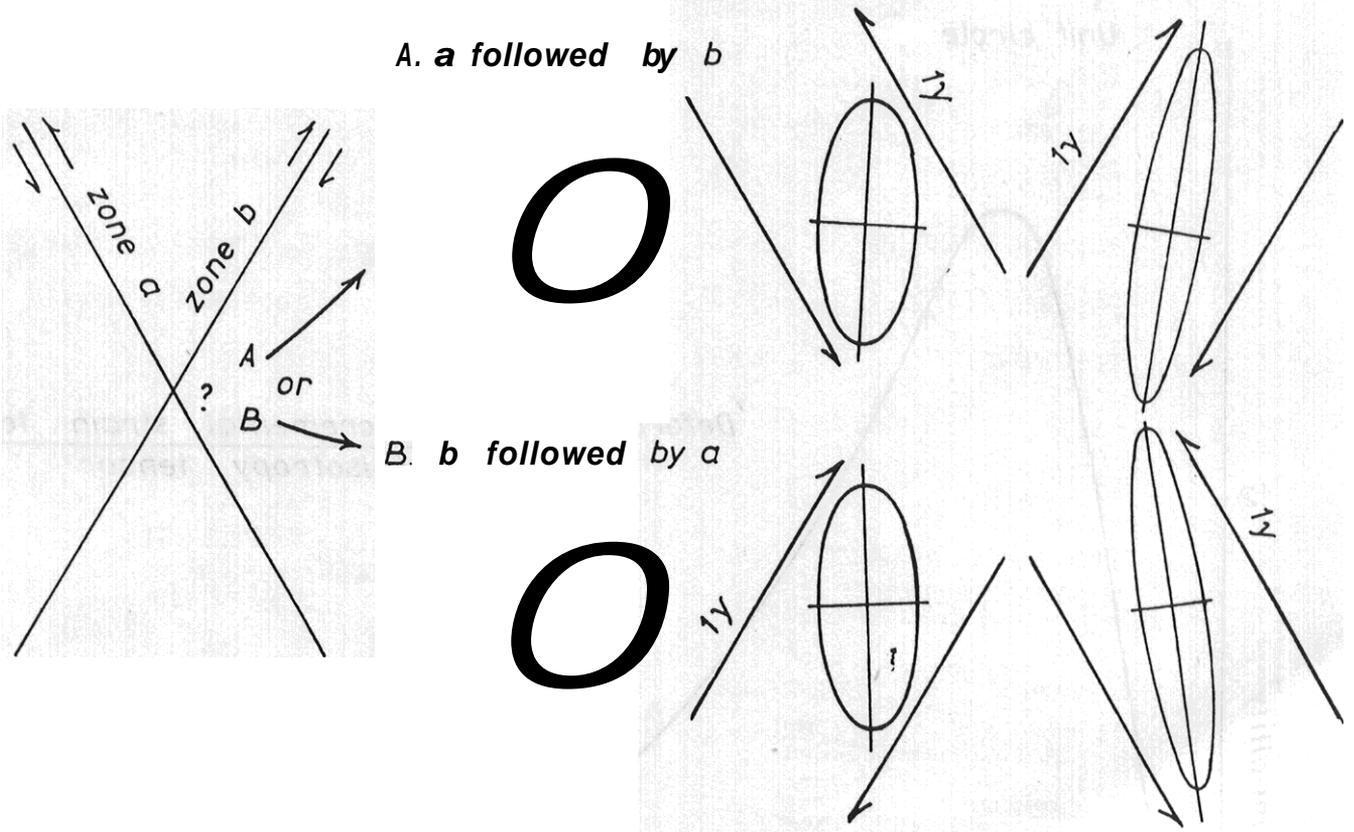


Fig.22 Chemical redistribution in a shear zone (pressure solution)

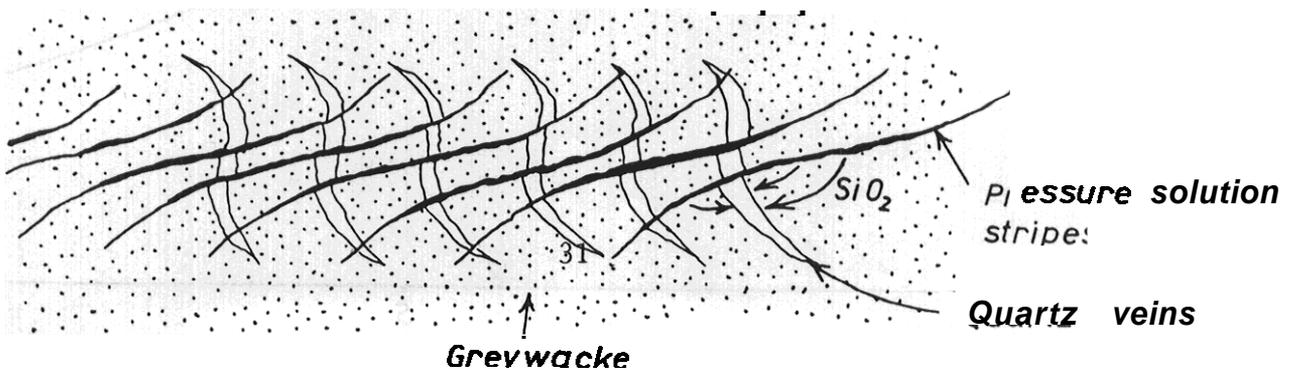


Fig. 23. Finite strain and shear zone deformability

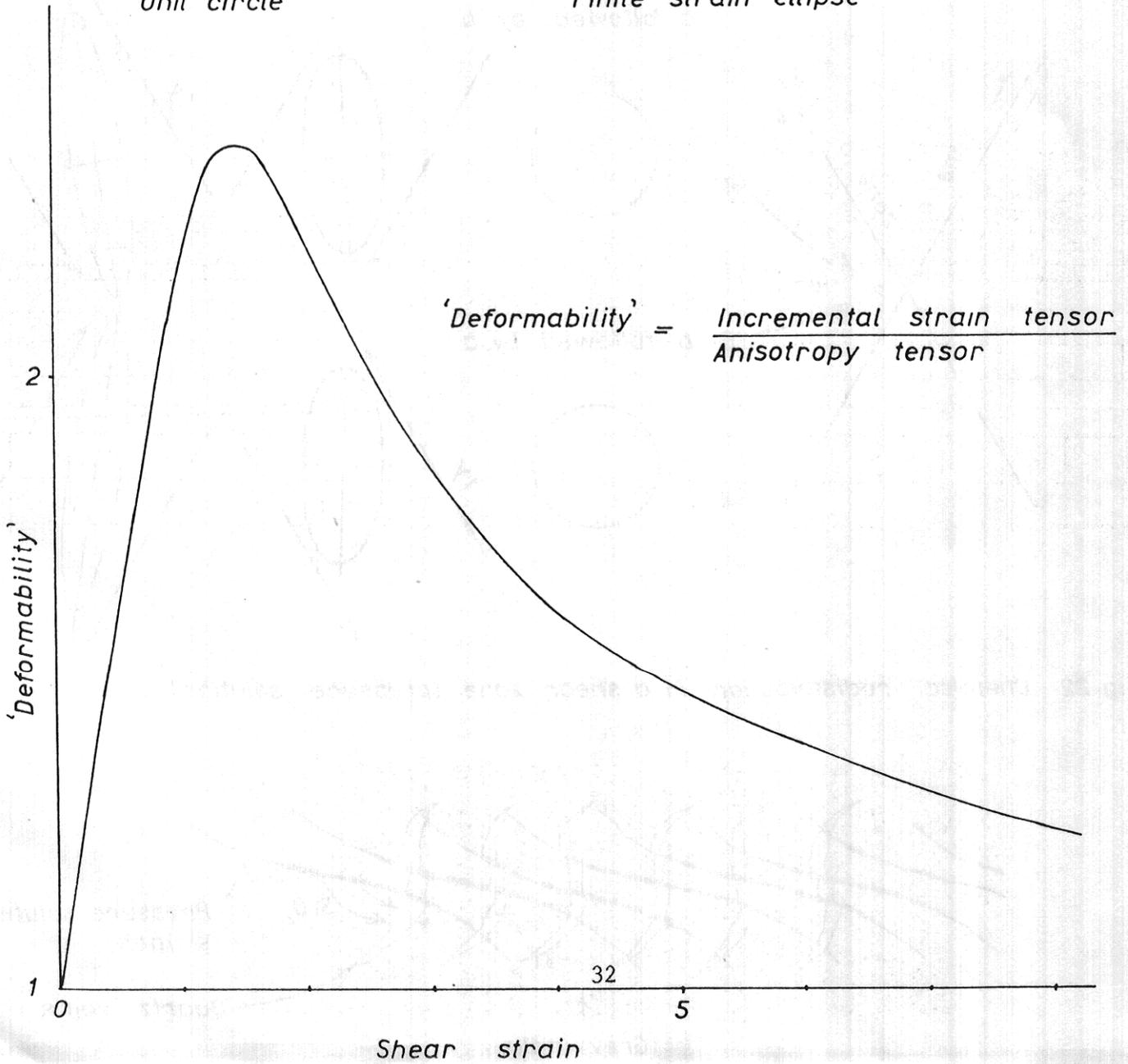
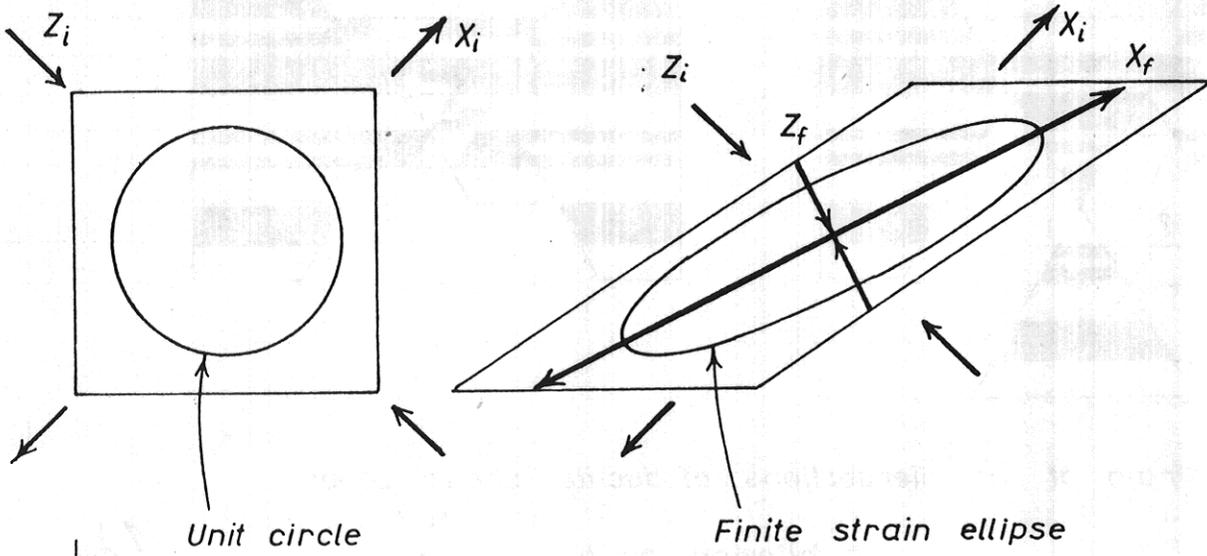


Fig. 24. Characteristic fault and fold geometry- high level nappes.

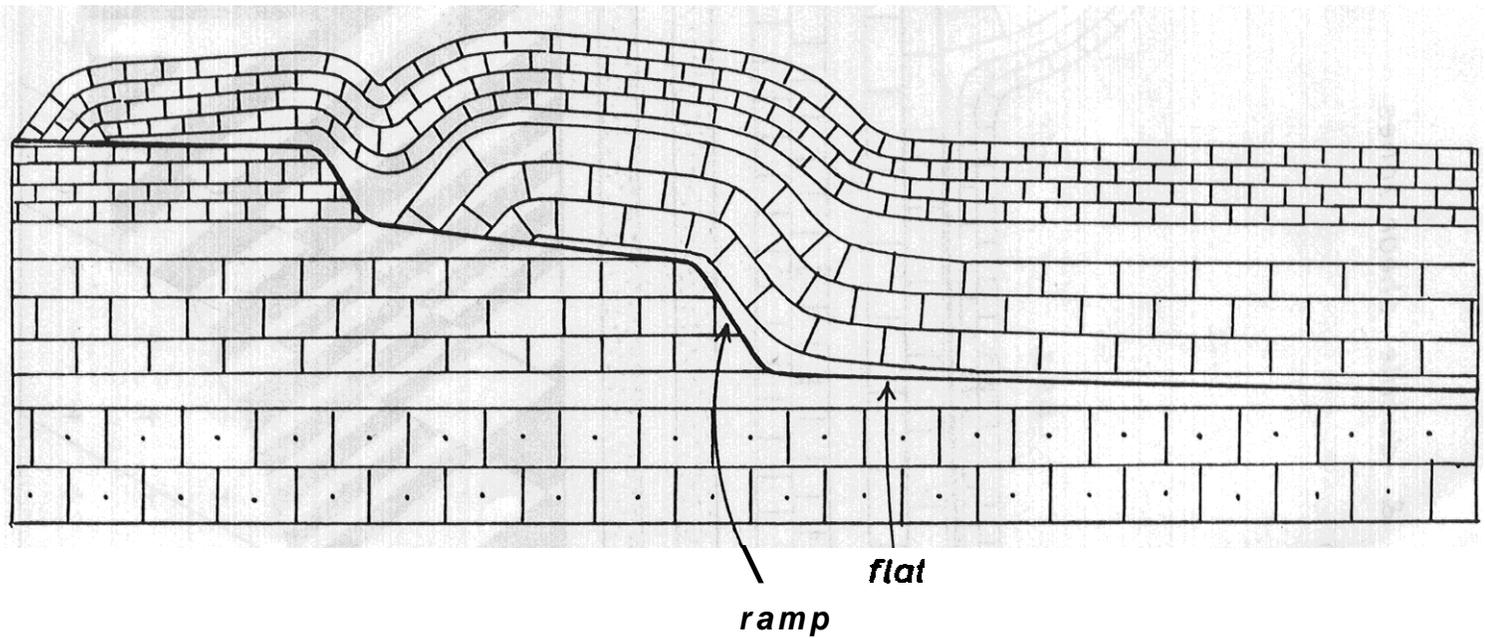
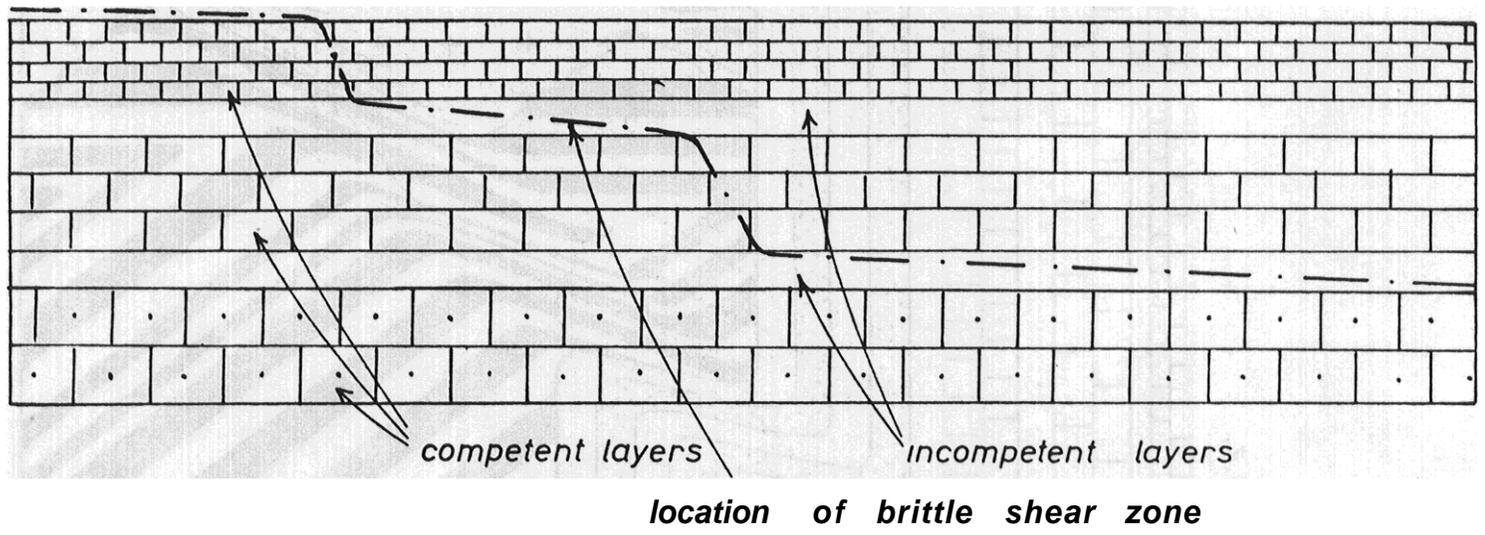


Fig.25 Relationship of brittle & ductile shear zones

1. Crustal contraction

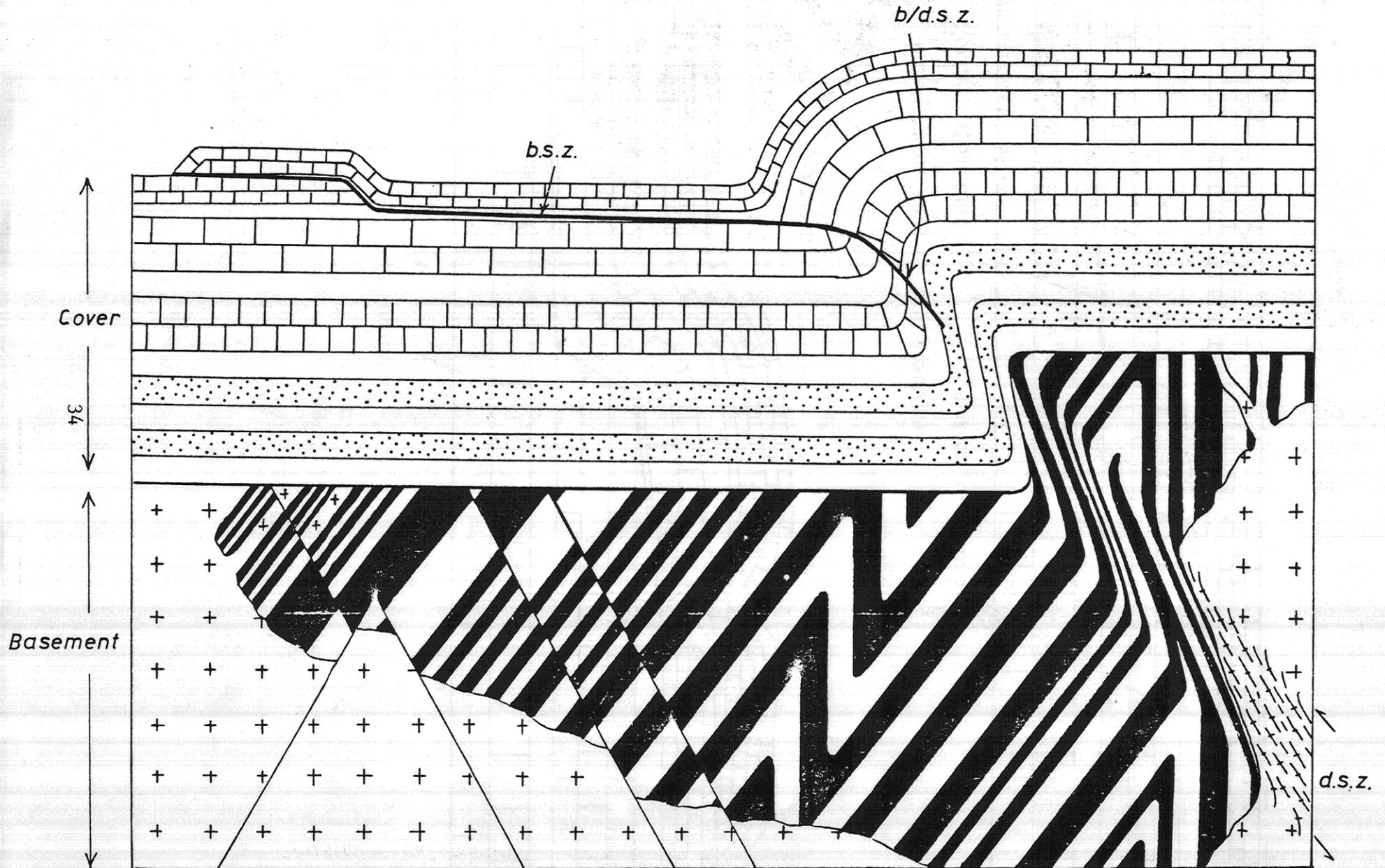


Fig. 26. Relationship of brittle & ductile shear zones
2. Crustal extension

