*Journal of Structural Geology*, Vol. 5, No. 6, pp. 619 to 622, 1983
Printed in Great Britain

0191-8141/83 $03.00 + 0.00
Pergamon Press Ltd.

# Brevia

## SHORT NOTES

# Progressive refolding in ductile shear zones

J. P. PLATT

Department of Geology and Mineralogy, University of Oxford, Parks Road, Oxford OX1 3PR, U.K.

(*Received 8 July* 1983; *accepted in revised form* 29 *August* 1983)

**Abstract**—The foliation formed in a ductile shear zone can become folded by continuing shear in the zone without the foliation having to enter the shortening sector of the flow field. If the foliation lies parallel to the shear plane, infinitesimal variations in the rate of shear strain cause compensating rotations that may amplify into folds. Small increases in the rate cause backward rotations (counter to the sense of shear): these do not amplify. Small decreases cause forward rotations that do amplify.

A COMMON feature in ductile shear zones are folds that deform the principal foliation formed in the zone, but which appear themselves to be related to continued deformation along the zone. In some cases the folds have been refolded in turn, one or more times. The phenomenon is most clear-cut where the shear zone is bounded by completely undeformed rocks, so that the folding is clearly restricted to and related to the shear zones, rather than being caused by a later unrelated deformation. Folds of this type have several distinctive characteristics.

(1) The folds are restricted to the shear zone. They die out along their axial surfaces in both directions, defining a lens-shaped folded domain in profile.

(2) The axial-plane foliation in the fold hinges is microstructurally similar or identical to the folded foliation (Carreras *et al.* 1977, Bell 1978). In the Cabo de Creus example, described by Carreras *et al.* (1977), a crystallographic preferred orientation pattern related to the main mylonitic foliation is also rotated around the folds, but in the tighter folds a similar fabric is developed with a symmetrical relation to the axial plane.

(3) The folds vary continuously in style from open and nearly concentric, to virtually isoclinal 'similar' folds. The open folds have axial planes at fairly high angles to the plane of the shear zone; the tighter and more flattened folds are inclined at lower angles (Carreras *et al.* 1977). This suggests that the folds may have nucleated continuously or episodically during deformation, and became progressively tightened, flattened, and rotated towards the shear plane. Fold hinges also appear to have been progressively rotated towards the bulk elongation direction, and in some cases the folds are deformed into sheaths (Bell 1978, Quinquis *et al.* 1978).

It is not immediately clear why these folds form. Theories of buckle-folding of layers of contrasting viscosity require the layering to lie in the shortening sector of the instantaneous flow field [note that 'inverse folds', initiated when the layering lies in the extensional sector, do not amplify significantly (Smith 1977)]. But if the foliation forms parallel or close to the principal plane of finite strain, we should expect it always to lie in the extensional sector. In an 'ideal' shear zone, parallel sided and effectively of infinite extent (Ramsay & Graham 1970), deforming by progressive simple shear, the principal plane of finite strain should rotate towards the plane of shear (the plane of zero finite and instantaneous elongation), but never reach it. To get into the instantaneous shortening sector, the foliation would have to pass through the shear plane.

The problem may, of course, be an artefact of the assumptions in the previous paragraph. The idea that the foliation in shear zones represents a principal plane of finite strain stems mainly from the observation that it is defined in part by deformed grains that were initially approximately equant. It clearly does not apply if there was a pre-existing fabric in the rocks, or if the foliation becomes active during the deformation. The latter may happen fairly rapidly: the foliation (S) is a plane of anisotropy, and deformation by slip along S becomes progressively easier as it develops and rotates towards the shear plane. In the extreme case, the deformation, while it may remain bulk simple shear, becomes divided into three components: slip along S, coaxial extension along S, and rotation (spin) of the system (Lister & Williams 1979, 1983). In this situation, S will rotate towards the shear plane more rapidly than the principal plane of finite strain, but it should nevertheless not pass through the shear plane.

We could also argue that the deformation in shear zones locally departs significantly from progressive simple shear, so that the foliation temporarily finds itself in the instantaneous shortening sector. Bell (1978) suggests that the folds in the Woodroffe mylonite were initiated by flow heterogeneity around 'islands' of undeformed rock in the mylonite zone. This is also the obvious explanation for the folds commonly developed around porphyroclasts in shear zones: the porphyroclast rotates bodily at up to twice the vorticity rate; the foliation near it is rotated also, and comes to lie in the shortening sector. A similar (but distinct) concept has been developed by Hudleston (1977), to explain recumbent similar folds formed in the zone of shear at the base of glaciers and ice sheets. Irregularities in the bed of the glacier cause local rotations of the foliation through the shear plane; and subsequent heterogenous simple shear then causes fold development. In Hudleston's concept the foliation is not an active surface in the deformation. Related models of passive amplification of folds initiated by instabilities in simple-shear flow have been developed by Cobbold & Quinquis (1980).

The abundance of folds in some shear zones, and their common lack of association with porphyroclasts, 'islands' or marginal irregularities on a scale comparable with that of the folds, suggests that they may not depend for their origin on large-scale inhomogeneities. Carreras *et al.* (1977) suggest that the folds may initiate as 'internal' buckling instabilities in the foliation. Internal buckles (Biot 1965, Cobbold *et al.* 1971) are controlled by the presence of a mechanical anisotropy, and do not depend on a spatial variation in material properties. They are caused by the change in effective rheology of an anisotropic material as it rotates with respect to the flow field (Cobbold 1976). A rotated domain therefore deforms at rates different from those in the surrounding material, and this may cause it to amplify into a fold. Internal buckles will not produce tight folds, however, if the foliation lies in the extensional field. The only structures that are likely to form in this situation are foliation boudinage, and very open folds and crenulations associated with shear bands (Cobbold *et al.* 1971, Platt & Vissers 1980).

A possibility remains that internal buckling may occur if the strain is so high that the foliation has been rotated until it is effectively parallel to the shear plane. In this position it undergoes neither extension nor shortening. This is not an adequate condition for 'ordinary' buckle-folding of layers with contrasting mechanical properties, but as discussed below, a type of internal-buckling instability can arise in this situation and amplify into folds. Biot's (1965) theory was developed for infinitesimal deformation in elastic materials. He suggested that it could be extended, by analogy, to viscous and plastic behaviour, but the significance of concepts such as initial stress and elastic strain energy in a viscous situation is obscure. A much simpler kinematic analysis of a simple anisotropic viscous material is therefore adopted here.

*Kinematic analysis*

I make the following assumptions, and the analysis is, of course, limited by their validity.

(1) The shear zone is planar, parallel sided, and of effectively infinite extent, so that it conforms to the criteria established by Ramsay & Graham (1970) for a zone of simple shear.

(2) The flow field is progressive simple shear at the scale of discussion: grain scale heterogeneity is ignored.

(3) The material in the zone is mechanically homogeneous at the scale of discussion; but is anisotropic with, in two dimensions, reflection lines of symmetry normal and parallel to the foliation ($S$).

(4) The material obeys a linear viscous flow law such that

$$T_{ij} = C_{ijrs} D_{rs},$$

where the $T_{ij}$ are deviatoric stresses, $D_{rs}$ are strain rates, and $C_{ijrs}$ the 36 viscous moduli (tensor notation and symbol conventions after Malvern 1969). The viscous moduli for a two-dimensional incompressible anisotropic material can be reduced to two, if the plane of anisotropy is used as a reference frame. These are a compressive modulus $N$ and a shear modulus $Q$, both measured parallel to S (Cobbold 1976).

If the principal rate of elongation $D_1$ is oriented at $\alpha$ to S, then the principal tensional deviatoric stress $T_1$ will be oriented at $\theta$ to S, such that

$$\tan 2\theta = Q/N \tan 2\alpha$$

(Cobbold 1976). In the case under discussion, of simple shear parallel to S, $D_1$ must be at 45° to S, and $T_1$ is parallel to $D_1$. (The material therefore satisfies one of Biot's orthotropy conditions, but this is only true if S is parallel to the plane of simple shear.)

Consider now a localised perturbation in the flow field. The simplest sort of perturbation in this situation is an infinitesimal variation in the rate of shear strain. For example, such variations are very likely as a result of grain-scale inhomogeneities. But if we postulate that the perturbation is infinitesimal, we need not concern ourselves with its cause: we need only consider whether it will amplify. We consider a volume that completely encloses the perturbation, and treat it as an element in the deformation field. This element is continuous with its surroundings, and must therefore obey the flow compatibility relations (Platt & Vissers 1980). Taking Cartesian coordinates $x_1$, $x_2$, with $x_1$ parallel to S, the flow field is described by the velocity gradient matrix $\mathbf{L}$

$$L_{ij} = \frac{\partial v_i}{\partial x_j} = \begin{bmatrix} 0 & \Gamma \\ 0 & 0 \end{bmatrix} \tag{1}$$

where $\Gamma$ is the rate of shear along S. $\mathbf{L}$ can be divided into a strain rate tensor $\mathbf{D}$, comprising rates of elongation and shear-strain, and a vorticity tensor $\mathbf{W}$, which superimposes angular velocities on the components of $\mathbf{D}$. For our purposes, the vorticity can be expressed as a single angular velocity vector $\mathbf{w}$ about the $x_3$ axis. $\mathbf{w}$ can be expressed in terms of the components of $\mathbf{L}$ (Platt & Vissers 1980) such that

$$\mathbf{w} = \tfrac{1}{2}\left(\frac{\partial v_1}{\partial x_2} - \frac{\partial v_2}{\partial x_1}\right). \tag{2}$$

Therefore from (1)

$$\mathbf{w} = \tfrac{1}{2}\Gamma. \tag{3}$$

From (2)

$$\frac{\partial \mathbf{w}}{\partial x_1} = \tfrac{1}{2}\left(\frac{\partial^2 v_1}{\partial x_1\,\partial x_2} - \frac{\partial^2 v_2}{\partial {x_1}^2}\right). \tag{4}$$

But from (1),

$$\frac{\partial v_1}{\partial x_1} = \frac{\partial v_2}{\partial x_2} = 0, \tag{5}$$

therefore

$$\frac{\partial \mathbf{w}}{\partial x_1} = 0. \tag{6}$$

The vorticity must therefore remain constant along S. This means that to maintain compatibility, the vorticity of the perturbation as a whole ($\mathbf{w}^{\mathrm{p}}$) must equal the vorticity outside it ($\mathbf{w}^{\mathrm{e}}$) along the foliation direction. Within the perturbation, $\mathbf{w}^{\mathrm{p}}$ may be partitioned into vortical ($\mathbf{w}^{\mathrm{pv}}$) and spin ($\mathbf{w}^{\mathrm{ps}}$) components (Platt & Vissers 1980, Means *et al.* 1981) such that

$$\mathbf{w}^{\mathrm{p}} = \mathbf{w}^{\mathrm{pv}} + \mathbf{w}^{\mathrm{ps}} = \mathbf{w}^{\mathrm{e}}. \tag{7}$$

The vortical component expresses angular velocities relative to the principal axes of the strain-rate tensor, **D**. The spin component is the angular velocity of the principal axes of **D** relative to the reference frame. If the perturbation consists of a small change in the rate of shear strain, $\Delta\Gamma$, then the vortical component $\mathbf{w}^{\mathrm{pv}}$ will change by $\tfrac{1}{2}\Delta\Gamma$, and from (2), the spin component will have to change by $-\tfrac{1}{2}\Delta\Gamma$ to compensate. This means that the perturbation, together with the included foliation, will rotate in the opposite sense to the local change in $\Gamma$. The effect of this is illustrated in Fig. 1, which shows that a domain where $\Gamma$ increases will rotate backwards (against the sense of shear), whereas a domain where $\Gamma$ decreases will rotate forwards. Rotation of S can therefore arise from small variations in $\Gamma$.

The next question is whether these perturbations will amplify. As shown above, $T_1$ is at 45° to S, so the resolved shear stress on S is maximal. Any rotated domain will therefore experience a lower resolved shear stress along the local S than elsewhere (Fig. 2). This will cause the local value of $\Gamma$ to decrease, so that the domain, whatever its initial sense of rotation, will tend to rotate forwards. Forward-rotating zones will therefore amplify, whereas backward-rotating zones will disappear. The forward-rotating zones could amplify indefinitely, and develop into the type of folds found in mylonite zones. Other compatibility constraints will arise during amplification, but their solution depends on the degree of anisotropy, and whether volume is conserved during deformation.

It is worth noting that the process of geometric softening or rotation-softening (a decrease in strength caused by rotation of a plane of anisotropy in a stress field), which has been suggested as a possible cause of strain localization, is not in itself a sufficient condition for the amplification of a perturbation. Both forward- and backward-rotating zones described above are zones of rotation hardening: one amplifies, the other disappears.

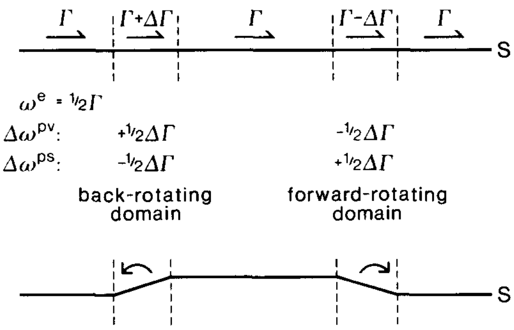


Fig. 1. Diagram to show how small variations in the rate of shear strain, $\Gamma$, in a shear zone lead to rotations of a plane of anisotropy S oriented parallel to the shear plane. Top: two perturbations, one positive and one negative, in the rate of shear strain. The vorticity of the flow, $\mathbf{w}^{\mathrm{e}} = \tfrac{1}{2}\Gamma$. The vortical component, $\mathbf{w}^{\mathrm{pv}}$, of the vorticity in the perturbation varies with $\Delta\Gamma$, and the spin component, $\mathbf{w}^{\mathrm{ps}}$, varies in the opposite sense to compensate. This causes the perturbations to rotate as a whole (bottom).

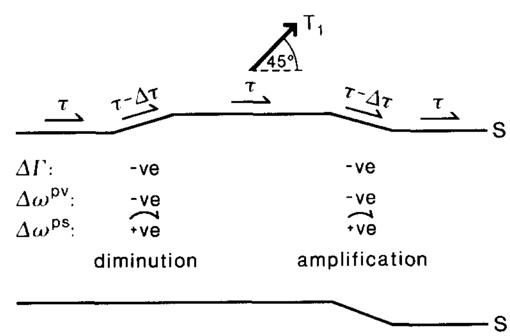


Fig. 2. Diagram to show how the ambient stress field in a shear zone controls the amplification of a rotational perturbation. The principal tensional deviatoric stress $T_1$ is oriented at 45° to S. The resolved shear stress, $\tau$, on S is therefore diminished by $\Delta\tau$ on both forward- and backward-rotating domains. This causes similar variations in $\Gamma$, $\mathbf{w}^{\mathrm{pv}}$, and $\mathbf{w}^{\mathrm{ps}}$ in both types of perturbation, leading to the amplification of forward-rotating domains and the suppression of backward-rotating domains.

*Acknowledgements*—I thank Jan Behrmann, Geert Konert, Gordon Lister, and Reinoud Vissers for discussion.

## REFERENCES

Bell, T. H. 1978. Progressive deformation and reorientation of fold axes in a ductile mylonite zone: the Woodroffe thrust. *Tectonophysics* **44**, 285–321.

Biot, A. 1965. *The Mechanics of Incremental Deformation.* Wiley, New York.

Carreras, J., Estrada, A. & White, S. 1977. The effects of folding on the *c*-axis fabrics of a quartz mylonite. *Tectonophysics* **39**, 3–24.

Cobbold, P. R. 1976. Mechanical anisotropy during large finite deformations. *Bull. Soc. géol. Fr.* **18**, 1497–1510.

Cobbold, P. R., Cosgrove, J. W. & Summers, J. M. 1971. The development of internal structures in deformed anisotropic rocks. *Tectonophysics* **12**, 23–53.

Cobbold, P. R. & Quinquis, H. 1980. Development of sheath folds in shear regimes. *J. Struct. Geol.* **2**, 119–126.

Hudleston, P. J. 1977. Similar folds, recumbent folds, and gravity tectonics in ice and rocks. *J. Geol.* **85**, 113–122.
Lister, G. S. & Williams, P. F. 1979. Fabric development in shear zones: theoretical controls and observed phenomena. *J. Struct. Geol.* **1**, 283–298.
Lister, G. S. & Williams, P. F. 1983. The partitioning of deformation in flowing rock masses. *Tectonophysics* **92**, 1–33.
Malvern, L. E. 1969. *Introduction to the Mechanics of a Continuous Medium.* Prentice-Hall, New Jersey.
Means, W. D., Hobbs, B. E., Lister, G. S. & Williams, P. F. 1981. Vorticity and non-coaxiality in progressive deformations. *J. Struct. Geol.* **2**, 371–378.
Platt, J. P. & Vissers, R. L. M. 1980. Extensional structures in anisotropic rocks. *J. Struct. Geol.* **2**, 397–410.
Poirier, J. P. 1980. Shear localization and shear instability in materials in the ductile field. *J. Struct. Geol.* **2**, 135–142.
Quinquis, H., Audren, C., Brun, J. P. & Cobbold, P. R. 1978. Intense progressive shear in Ile de Groix Blueschists and compatability with subduction or obduction. *Nature, Lond.* **273**, 32–45.
Ramsay, J. G. & Graham, R. H. 1970. Strain variation in shear belts. *Can. J. Earth Sci.* **7**, 786–813.
Smith, R. B. 1977. Formation of folds, boudinage, and mullions in non-Newtonian materials. *Bull. geol. Soc. Am.* **88**, 312–320.

*Note added in proof*

Dr. M. Casey has pointed out that the compatibility constraint in equation (6) also applies to all the components of the strain-rate tensor **D**, none of which may vary along $x_1$. This means that any perturbation in Γ such as I propose must be surrounded by a region of compensating deformation, such that there exists a closed volume the bulk deformation of which conforms to equation (1). These compensating deformations could involve volume changes, strain-rates and rotations. A couplet of adjacent forward and backward rotating zones could partly compensate for each other. As mentioned in the text, these additional compatibility constraints become significant as the amplitude of the perturbation increases.