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Large strain deformation: report on a Tectonic Studies Group discussion meeting held at Imperial College, London on 14 November 1979

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THE ESTIMATION of strain in rocks has become second nature to most structural geologists and is highlighted by the intuition that has been displayed when selecting strain markers (see contributions in Ramsay & Wood 1976). Although there may sometimes be difficulties in obtaining accurate estimates (see the abstract by Coward), the studies have shown that regional strains of greater than 1 (natural or true strain) are not uncommon during tectonism with local strains often being higher than this, especially in shear zones (Ramsay & Graham 1970, Carreras *et al.* 1975, Coward 1976, Burg & Laurent 1978 and the abstracts of Cobbold, and Watts & Williams). These estimated strains are much larger than those obtained during the laboratory deformation of rocks and minerals. A cursory look at the literature on experimental deformation of rocks and minerals (see Nicolas & Poirier 1976) shows that most tests stop at natural strains between 0.1 and 0.3. In addition, in an attempt to approach geologically realistic strain rates in the laboratory, stress relaxation tests are being used in which the strains obtained are an order of magnitude less (Rutter *et al.*.1978, Rutter & White 1979).

In spite of this marked strain difference, experimental data is playing an increasingly important role in the interpretation of deformation mechanisms and of microstructure, fabric and foliation development during tectonism. However, there has been little questioning of the validity in extrapolating from small to large strains. The recent use of deformation mechanism maps (Stocker & Ashby 1973, Rutter 1976, White 1976, Twiss 1977, Schmid *et al.* 1977) serves as an example; deformation fields are plotted for any combination of stress, strain rate, temperature and grain size with strain being neglected. It is tacitly assumed that deformation mechanisms and related textural features are the same at low and high strains for any selected combination of the above parameters. However, recent studies of large strain deformation in metals (Fargette & Whitham 1976, Duggan *et al.* 1978, Willis & Hatherly 1978, Dillamore *et al.* 1979, Malin & Hatherly 1979) and of the microstructures of mylonites (White *et al.* in press) indicate that the above inference may not always be correct.

This topic was recently discussed at a meeting of the Tectonic Studies Group (affiliated to the Geological Society of London) held on 14 November 1979 at Imperial College, London. The meeting brought together metallurgists and geologists with a common interest in large strain deformation. This report presents the abstracts of the contributions given and summarizes the main conclusions of the meeting; most of these emerged during the discussions which followed the contributions.

The metallurgical contributions gave an overview of the large strain deformation of polycrystalline metals. They covered; (1) cold and warm working (Willis, Ridha & Hutchinson), (2) creep and hot working (Humphreys), and (3) superplasticity (Hammond). Geologists who expected to find solutions to their particular problems were to be disappointed for it emerged that apart from (3) large strain deformation has been largely neglected by metallurgists and materials scientists. Even in the case of superplasticity, it became evident that the responsible deformation mechanisms are not known with any certainty and that the form of the constitutive equation is debatable. It would appear that, at present, any geologist plumping for a given metallurgist's model does so at his own risk.

The complexities that arise during the cold and warm working of polycrystalline metals (deformation below 0.4 of the homologous *T*) was a revelation to most geologists present. Not only can deformation mechanisms change with increasing strain but also the deformation can change from being homogeneous to inhomogeneous and, in some cases, back to homogeneous. The inhomogeneous stage is marked by the development of shear bands which form a foliation, similar to a crenulation cleavage, oblique (at about 35°) to the rolling, or flattening, foliation. Changes in microstructure and fabric with increasing strain were emphasized and highlighted the problems that can arise if an attempt is made to interpret them as the products of a single mechanism and/or a homogeneous deformation.

Although it appears that deformation is more likely to remain homogeneous through large strains at high temperatures, it tends to concentrate into discrete shear zones with increasing strains at intermediate temperatures. The main softening processes responsible for the concentration appear to be continuous recrystallization and geometric (fabric) softening, that is the grains became orientated for easy slip. It was demonstrated that low symmetry analogue materials can provide an insight into shear zone development during large strain deformation. A film illustrated the processes of continuous recrystallization and geometric softening during experimental deformation of camphor.

The geological contributions concentrated on deformation in shear zones in naturally deformed rocks. Emphasis was placed on softening mechanisms that are associated with the formation of shear zones (see abstract by White) and on the strains within shear zones (see abstracts by Coward, Cobbold, and Watts & Williams). An insight into the complexities that can occur during a large strain natural deformation emerged. It was argued that most ductile shear zones form because of the inability of the country rock to accommodate a homogeneous deformation. Of the several possible softening mechanisms discussed, it was thought that the most important were geometric (fabric) softening and continuous recrystallization, similar to that described for metals, along with reaction enhanced ductility. Geometric softening in shear zones arises when the dominant slip planes in a given mineral rotate into parallelism with the edge of the shear zone and subsequently remain in this position. The opposite occurs in an axial symmetric deformation in which the slip planes rotate towards the *XY* plane of the strain ellipsoid, that is away from the planes of maximum resolved shear stress.

At high temperatures (amphibolite and granulite facies) deformation within shear zones remains homogeneous apart from sheath folding. In contrast, shear bands, reflecting inhomogeneous deformation at lower temperatures, develop within the mylonites and may destroy the mylonitic foliation, replacing it with an oblique foliation — a shear band cleavage, which has no relationship to existing sheath folds. Usually only one cleavage develops and shows a relationship with the mylonite foliation and the shear direction which is illustrated in Fig. 1. It was felt that when two directions were formed (Fig. 2) they reflected significant flattening within the shear zone.

The above temperature dependent behaviour is broadly similar to that in metals and analogue materials, and suggests that metals and analogues can be of great assistance when interpreting mylonite microstructures and fabrics.

The contribution of Watts & Williams showed that brittle minerals may fracture during ductile flow in a mylonite matrix. This occurs after significant ductile strains, and it follows that pulled-apart fractured minerals and associated fibrous infills in mylonites, and possibly in other rocks, convey no information about deformation mechanisms in the ductile matrix and are unlikely to give meaningful estimates of the ductile strains in the matrix. Finally, Cobbold demonstrated that sheath folds are more common than previously thought in mylonites and that they can be used to estimate strains and flow directions in mylonite zones.

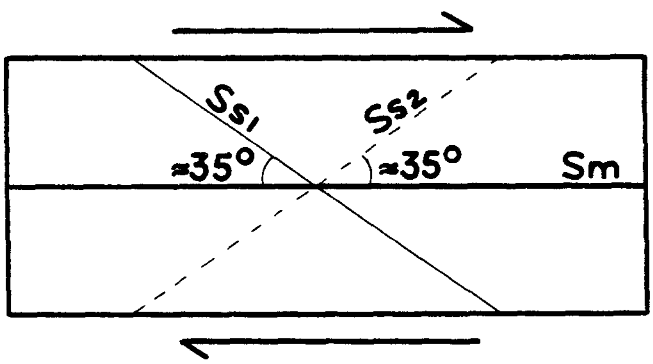

Fig. 2. Sketch showing the orientations of two coexisting shear band foliations S_{s1} and S_{s2} which exist in some mylonites. S_{s1} is normally the better developed. The sketch shows the shear bands intersecting S_m, the mylonite foliation, at about 35°; a commonly encountered angle in mylonites. The angle is always less than 45°.

Summarizing, from a geological point of view, the meeting highlighted the difficulties associated with strain analyses using traditional markers in large strain environments, but offered a new and more promising technique, namely the analysis of sheath folds. It stressed the need for work on softening processes in rocks and the care that must be exercised when extrapolating laboratory results to the field. The assumption that deformation processes are independent of strain is not justified in geological environments. The meeting highlighted the shortfalls in present experimental data for rocks and minerals and showed that some insight into large strain geological deformation could be gained from metals, and especially from analogue materials. However, there is no substitute for real rocks.

ABSTRACTS OF PAPERS PRESENTED

Substructure development in cold-worked metals. D.J. Willis, Department of Metallurgy, University of Leeds, England.

A metal is considered to be cold-worked if the homologous temperature of deformation is less than 0.4. Besides homologous temperature, three other parameters are important in controlling substructure development in metals, namely crystal structure, stacking fault energy, and deformation mode. The most common deformation modes which have been investigated are tension, compression, rolling and wire drawing.

Wire drawing of iron for true strains up to 6 (99.8% reduction in area) produces a substructure of ribbon shaped cells elongated parallel to the wire axis (Langford & Cohen 1969). The transverse cell size decreases from about 0.5 μm at $\epsilon_T = 0.2$ to 95 nm at $\epsilon_T = 6$, and the misorientation between adjacent cells increases from about 1° to 10° (Langford & Cohen 1975). Although the cell size decreases with strain it does so more slowly than the wire diameter, due to dynamic recovery effects which are believed to be mechanically, rather than thermally, activated.

Cold rolling has been the most extensively investigated deformation mode, due to its industrial importance, but it is only since the use of edge-section thin foils that the substructure development has been clarified. If attention is confined to single phase materials the following four metals cover the range of observations thus far.

(1) *Ti, h.c.p.* (Blichorski *et al.* 1979)

0 - 40%	twinning with some accommodation slip, probably
40–90%	involving <c + a>, <c> and <a> slip dislocations
>90%	shear bands

(2) *Cu-30%Zn, f.c.c., low s.f.e.* (Duggan *et al.* 1978)

0–50%	octahedral slip but no cell structure, dislocations in planar arrays
40–60%	microtwin bundles form, with twin thickness 7–30 nm
(50–80%	microtwin bundles rotate to become parallel with the rolling plane)
60–95%	shear bands
>85%	return to octahedral slip in the refined substructure

(3) *Cu, f.c.c., medium s.f.e.* (Malin & Hatherly 1979)

0–10%	octahedral slip and cell formation
10–65%	mircoband formation, and rotation of the microbands parallel to the rolling plane
>65%	shear bands

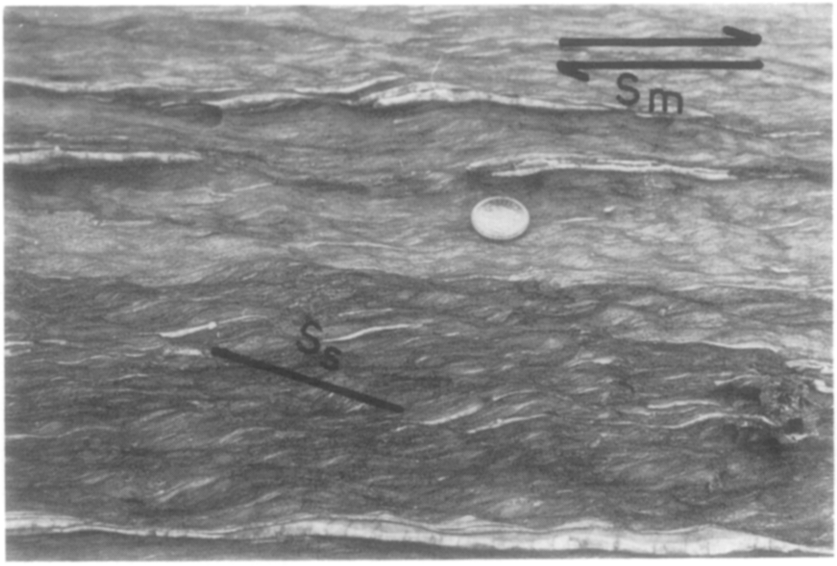

Fig. 1. A shear band foliation (S_s) in phyllonites in the Cap de Creus shear zones (discussed by White *et al.* in press). The mylonite foliation (S_m) and the shear directions are marked. Note the small angle between S_s and S_m.

(4) *Fe b.c.c.* (Mathur & Backhofen 1973, Willis & Hatherly 1978, Davidson & West 1979)

0–80%	pencil glide and cell or microband formation
>80–95%	shear bands, but not as frequent as in other metals

The exact nature of microbands and shear bands is not yet understood, but the shear bands observed in metals deformed at room temperature are remarkably similar to those observed within some mylonites. It is clear that substructures formed by cold deformation are not homogeneous, nor do they remain constant during the course of deformation and it is not therefore reliable to extrapolate low strain observations to higher strains.

A comparison of the deformation substructure in copper and 70/30 brass after cold rolling. A. A. Ridha and W. B. Hutchinson, Department of Metallurgy and Materials, University of Aston, Birmingham, England.

Despite numerous investigations the mechanisms of deformation of metals at high strains are not well understood. The present work was carried out on a metal of moderately high stacking fault energy (copper) and an alloy (70/30 brass) of low stacking fault energy. These were cold rolled with reductions in the range 20 – 98% (true strains 0.2 – 4) and examined by optical, transmission electron, and scanning electron microscopy. Scanning electron microscopy in the back-scattered mode is especially useful as it reveals the deformation substructure by diffraction contrast from the surface of large, solid specimens. Preferred orientations were also determined. From these observations we have attempted to identify the dominant mechanisms of deformation at various stages of reduction.

In copper, octahedral slip is believed to be the only deformation mode. However, at various levels of reduction this may be more or less homogeneously distributed. Between about 20 and 70% reduction much of the slip is concentrated in localised microbands that are usually parallel to the slip plane. At higher reductions, macroscopic shear bands are occasionally found although these are probably not the main deformation mechanism.

In brass, deformation occurs initially by octahedral slip but mechanical twinning intervenes in the range 40 – 60% reduction. Inhibition of normal slip by the numerous twin interfaces causes further deformation (60 – 95%) to become concentrated in shear bands. Subsequently, when the twinned structures are completely destroyed by the shear bands, deformation becomes again more homogeneous.

A conclusion of this work is that different mechanisms of deformation occur at different strain levels. Although localised shear banding is frequently a high strain phenomenon, it may in some cases be a transient mechanism which is subsequently replaced by homogeneous deformation.

The plasticity of polycrystalline materials at high temperatures. F.J. Humphreys, Department of Metallurgy and Materials Science, Imperial College, London, England.

A knowledge of the deformation behaviour of metals at high temperatures is important in understanding creep and hot working. Although the bulk mechanical properties of many metals are well known, the understanding of the microstructural mechanisms occurring at high strains is incomplete, and investigations have generally been confined to cubic metals.

When deformed in compression or torsion, cubic metals deform rather homogeneously, and, except at very high temperatures and low strain rates, the microscopic mechanisms are dislocation glide and climb, accompanied by the softening processes of dynamic recovery and dynamic (syntectonic) recrystallisation. The softening processes are competitive, and dynamic recrystallisation occurs only if recovery is slow, for example in metals of low stacking fault energy. The mechanism of dynamic recrystallization is thought to be strain induced grain boundary migration, somewhat similar to that occurring during static (post-tectonic) recrystallisation. The recrystallised grain size is a function of the applied stress, strain rate, and temperature of deformation (Roberts & Ahlblom 1978).

The behaviour of non-cubic metals is less well known, although a detailed study has recently been made of magnesium (close packed hexagonal structure) (Burrows *et al.* 1979). The behaviour of magnesium is determined to a large extent by the number of available slip systems. At low temperatures, slip is possible only on the basal plane, and the metal has low ductility, whereas at high temperatures, sufficient slip systems are available for homogeneous deformation. In the intermediate temperature range, 200–350°C, the microstructures after deformation show some similarity to those of minerals such as quartz, olivine and calcite.

The high stresses necessary for slip on other than the basal plane in magnesium results in strain incompatibilities at the grain boundaries. These are accommodated by lattice rotations in the mantles of the grains, and, at higher strains, new grains are formed in these regions by dynamic recrystallisation, resulting in a duplex grain structure similar to a quartz mylonite. The occurrence of dynamic recrystallisation by progressive lattice rotation in grain boundary regions has not been reported in cubic metals although there is evidence that it occurs in minerals (White 1976, Poirier & Guillope 1979).

In magnesium, the regions of dynamic recrystallisation extend across the specimen at higher strains and lead to the formation of shear zones in which subsequent deformation is concentrated.

Although high strain laboratory deformation of polycrystalline minerals is difficult, useful information on the development of microstructure with strain can be obtained by studying the deformation of low symmetry organic crystalline solids *in situ.*

In summary, it would appear that polycrystalline materials — metallic, mineral or organic, with limited numbers of slip systems show certain similarities in their development of microstructure during high temperature deformation, and that these similarities arise from the inhomogeneity of deformation in the grain boundary regions. A study of analogue materials may thus be of use in understanding the plasticity of minerals, but it is clear that caution must be exercised in applying data and theory for cubic metals to materials of other symmetries.

Superplasticity. C. Hammond, Department of Metallurgy. University of Leeds, England.

Superplasticity deformation occurs at homologous temperatures greater than 0.5 Tm and the flow properties are characterised by: (1) zero strain hardening, (2) a low value of the stress sensitivity index n (< 2) as defined in the equation $\dot{\varepsilon} = K\sigma^n$, and (3) randomisation of any prior deformation texture (fabric). It is the low n value which gives rise to slow stable notch growth and hence the high elongations which are observed in tension. The n values are determined by measuring the flow stress σ as a function of strain rate $\dot{\varepsilon}$, either from stress relaxation or, preferably, from strain rate cycling tests. In general it is found that n varies with $\dot{\varepsilon}$ and the n-value falls in three regimes: (1) at low $\dot{\varepsilon}$, $n>2$, (2) at intermediate $\dot{\varepsilon}$ (the superplasticity regime) $n < 2$, and (3) at higher $\dot{\varepsilon}$, $n > 2$.

The microstructural requirements for superplasticity are a fine grain or subgrain size (< 10 μm). Grain growth during a strain rate cycling test gives rise to an apparent strain hardening and loss of superplastic flow properties. Apart from the problem of grain growth, it is found that the microstructure remains unchanged irrespective of the amount of deformation, and hence microstructural features cannot be used to determine the extent of prior deformation.

Theories of superplastic deformation are based on (1) diffusion creep models and (2) grain boundary sliding models. The diffusion creep models of Herring & Nabarro and Coble are based on the assumption that grain (or subgrain) boundaries are perfect sources and sinks for vacancies which diffuse either through the lattice (Herring –Nabarro) or along the grain boundaries (Coble). Both predict Newtonian viscous flow ($n = 1$) but different stress dependencies on the subgrain size. These models may be modified to account for the possibility that the rate of emission or absorption of vacancies at grain boundaries is determined by the non-conservative movement of dislocations, and this gives rise to a predicted value of $n = 2$. Although diffusion creep is expected to give rise to grain elongation, the associated grain boundary migration and possible development of subgrain structures may lead to grain division and hence a retention of an overall equiaxial grain shape.

Models of grain boundary sliding essentially differ in terms of the strain rate controlling accommodation processes and the geometry of the translation processes. If accommodation is by non-conservative motion of dislocations a value of $n = 2$ is predicted, if accommodation is by diffusional creep processes a value of $n = 1$ is predicted.

It is emphasised that in the assessment of the applicability of the models to a given material, the most important factor is the indentification of the appropriate diffusivity (lattice or grain boundary) and the estimate of its value and temperature dependence.

Ductile shear zones: a guide to the large strain deformation of rocks. S. White, Department of Geology, Imperial College, London, England.

Shear zones are examples of inhomogeneous deformation arising because the country rock is incapable of accommodating an imposed strain rate by bulk homogeneous deformation. Large strains concentrate in these zones because of softening within them. The following are possible softening mechanisms: (1) a change in deformation mechanism; (2) geometric or fabric softening; (3) continuous recrystallization; (4) reaction enhanced ductility; (5) chemical softening; (6) pore fluid effects; and (7) shear heating. They have been discussed in detail by White *et al.* (in press) who concluded that a combination of (2) and (3) were the main processes in most quartz, olivine and carbonate mylonites, and a combination of (3) and (4) in mylonites produced by mineral reactions rather than simple recrystallization.

Once initiated, shear zones will grow, or others will form, until there is sufficient soft material to accommodate the imposed regional strain rate. Subsequent deformation in the mylonite at large strains is mainly affected by the anisotropy due to the developed foliation or grain fabrics. Any hardening in the shear zone due to, for example, an increase in imposed strain rate, a decrease in temperature or a change in shear direction, will lead to shear band formation. The bands form a crenulation at a small angle to the foliation and may eventually destroy the foliation (White *et al.* in press). Shear band cleavages are best developed in phyllosilicate-rich mylonites or phyllonites, but have also been observed in quartz and carbonate mylonites, and are usually related to the shear direction (Fig. 1).

It follows from the above that with increasing strain, deformation has changed from a bulk homogeneous deformation into an inhomogeneous deformation represented by shear zone formation and may eventually even become inhomogeneous within the shear zone. In addition a new rock, a mylonite, has developed during the straining and has a rheology very different from its parent.

Shear zones: problems with measuring large strains. M. P. Coward, Department of Earth Sciences, University of Leeds, England.

Major shear zones occur not only at plate margins but also within crustal plates, commonly over 1000 km from compressive margins. They indicate that crustal plates are not rigid but that they have soft margins. The shear zones may be steeply dipping with a strike-slip sense of displacement, gently dipping or even flat-lying. Many shear zones are corrugated, in that they change orientation from flat-lying to steeply dipping even though their displacement vectors remain constant. Movements of many tens of kilometres have been recorded from these shear zones, generally from offsets across the zones.

To understand the nature of plate margin deformation it is important to measure the strains and displacements and their variations along the shear zones. Ramsay (1967) and Ramsay Graham (1970) produced methods to estimate these parameters using variations in strain ratio and strain trajectory orientation across the zone, assuming simple shear alone. In certain small-scale shear zones, especially in sedimentary rocks, volume change must also be taken into account. However, many shear zones in metamorphic rocks show strains which cannot be explained by simple shear alone (Coward 1976). In many of these zones, volume change may be ruled out as (a) the volume change would be too great, (b) there are no changes in density of the rock or in volume of the deformed particles across the shear zone, and (c) there would have to be volume increase in one part of the zone and volume decrease in another part. The strains may be explained, however, by considering them as a combination of simple shear and shear-parallel extension or shortening. Assuming plane strain deformation there must be extension and shortening at the ends of shear zones due to differential displacement as the shear dies out. These extensions have also been found in the central portions of shear zones (Coward 1976). They may have formed by differential shear early in the history of the shear zone. As the shear zone propogated, simple shear strains would be superimposed upon these extensional strains. Alternatively, late differential displacement along a shear zone would cause the addition of layer-parallel extensions on simple shear strains.

In three dimensions, there may be differential displacement where a shear dies out normal to the movement vector. Such differential displacements are easy to recognise in thrust tectonics where a large area of the shear plane is available for study. They must also occur in steeply dipping zones. Combinations of these shear strains and extensional strains result in a wide range of ellipsoid shapes from oblate to prolate (Coward & Kim in press). Oblate ellipsoids do not necessarily mean that there has been extension of the shear zone parallel to its length.

To conclude, in large-scale shear zones, estimations of displacement from strain markers need to take into account shear-parallel extensions formed by differential movement of the shear zone; assumption that there was simple shear alone will give the wrong answer. Small-scale textural and fabric studies also need to take into account these strains; preferred crystallographic orientations will depend on the magnitude of the layer parallel extensions and also on the order of superimposition of extensional and shear strains. Many of the small-scale structures in shear zones, small-scale folds and shear bands, may be due to extensional strains formed by differential movement of the zone.

Strain history of a mylonite. M. J. Watts and G. D. Williams, Department of Geology, University College, Cardiff, Wales.

Large natural strains in mylonites arise from a steady state ductile flow which operates when strain hardening processes are balanced by strain softening processes. A granitoid mylonite from the North Armoricain Shear Zone in northwest France possesses microstructures formed in response to both these processes.

Early in the deformation history both quartz and feldspar behaved plastically; quartz showing undulatory extinction, deformation bands and ribbon structures (whilst dynamic recovery and dynamic recrystallisation also operated), and feldspar showing undulatory extinction, deformation twinning and kinking. Quartz *c*-axis fabric distributions show that both basal and prism slip have occurred. Feldspar exhibits slip along the twin plane. Textural relationships indicate an increase in strain heterogeneity as the mylonite developed. Some narrow bands comprising small, equidimensional, dynamically-recrystallised quartz grains, commonly with small dispersed micas, show a progressive randomizing of the quartz fabric as a result of grain boundary sliding.

As deformation proceeded, a ductility contrast between plastic quartz and brittle feldspar resulted in the fracturing of feldspars along cleavage planes, and the subsequent rotation of feldspar fragments towards the X direction. The distribution and orientation of cracks in feldspars favours a model involving their initiation by a fibre-loading mechanism as opposed to the generation of cracks through dislocation pile-up.

The component of strain in quartz due to basal slip has been estimated as $\gamma = 3.4$. However, that due to prism slip and other mechanisms such as grain boundary sliding remains unknown. In the feldspars, post-fracture longitudinal strain varies between $\varepsilon = 0.1$ and $\varepsilon = 0.6$ which represents only a small component of the strain of the mylonite. At a late stage, shear bands at an angle of about 35° to the mylonitic foliation were developed.

Sheath folds and large strains in rocks. P. R. Cobbold, C.A.E.S.S., University of Rennes, Rennes, France.

A large strain causes reorientation of linear and planar elements in a material. This kinematic effect has important geological consequences. For example, fold shapes and attitudes become strongly modified. This is especially obvious in three dimensions, where the cylindricity of folds may become more or less pronounced. In many situations, an intense plane strain produces sheath-like folds with long axes almost parallel to the principal direction of extension (Carreras *et al.* 1977). If folding is dominantly passive, sheath folds provide valuable information concerning the principal values and orientations of the bulk strain ellipsoid (Cobbold & Quinquis, in press).

Sheath folds have now been recognized from various geological environments, including large transcurrent shear zones (Berthé & Brun, in press), mylonites (Lister & Price 1978), thrust zones (Dalziel & Bailey 1968, Williams & Zwart 1977, Quinquis *et al.* 1978, Bell 1978) and even salt domes and salt glaciers (Talbot 1979).

Good examples also exist in the Pennine Zone of the Swiss Alps, including the area of flat-lying nappes and the so-called root-zone or Insubric zone (see Milnes 1974 for a review of the geological structure). In the nappes, the gneissic foliation is flat-lying and forms sheaths pointing NNW–SSE along a strong stretching lineation. In the Insubric zone, in contrast, the geneissic foliation is nearly vertical, but sheaths and lineation plunge very gently. This suggests that the Insubric zone is in fact a major transcurrent shear zone, dextral, about 10 km wide, with an internal shear strain of the order of $\gamma = 100$ and a corresponding displacement of up to 1000 km. In the absence of other markers, sheath folds are here useful indicators of the nature and intensity of deformation.

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