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On mylonites in ductile shear zones

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Abstract—In this contribution, shear zones are treated as zones of inhomogeneous deformation in which strain softening has occurred. The mylonites which form in ductile shear zones are the softened medium. The development of mylonite microstructures and fabrics are discussed from this point of view. Seven possible softening processes are discussed. They are the advent of superplasticity, geometrical softening, continual recrystallization, reaction softening, chemical softening, pore fluid effects and shear heating. Attention is also given to the brittle deformation of hard minerals in a soft ductile matrix. It is concluded that these fracture because of localized stress concentrations and that their microstructures give no information on deformation in the matrix. An oblique shear band foliation may develop in mylonites as shear strains continue to increase and may destroy the earlier-formed mylonite foliation. It is thought that the foliation may indicate hardening of the soft mylonite.

INTRODUCTION

DUCTILE shear zones offer a unique opportunity to study the progressive development of mylonite microstructures and fabrics with increasing strain. They also provide, through these features, an insight into the deformation processes associated with large strain deformation and the softening mechanisms which must occur if deformation is to be concentrated into a shear zone.

We shall summarize recent work on microstructures and fabrics in mylonites, especially in quartzofeldspathic rocks and it will provide a background when discussing softening mechanisms. Reference will be made to recent studies on shear zones in metals. The article will conclude with a discussion of the microstructural changes that occur during the continuing deformation in a shear zone and during changes in deformation parameters. Although we will mainly deal with ductile deformation we will also discuss the evidence for localized brittle deformation of hard minerals within the ductilely deforming mylonite matrix. We will commence by outlining the sense in which the terms 'shear zone' and 'mylonite' will be used in this contribution.

SHEAR ZONES AND MYLONITES

Shear zones are commonly encountered during the deformation of materials. They occur on all scales from sub-microscopic shear bands and slip planes in metals to the wide zones of intense deformation, tens of kilometres wide, which occur in the Pre-Cambrian gneisses of Greenland (Bak *et al.* 1975, Watterson 1979). Irrespective of the scale, all are examples of inhomogeneous deformation although on a relatively larger scale the deformation may be judged to be homogeneous.

Ductile shear zones are those in which crystal plasticity is dominant. Shear zones, in general, are thought to form when the hardening capacity of the host material has been exceeded; with their development marking the onset of softening processes (Argon 1973, Palmer & Rice 1973, Cobbold 1977, Chan & Asaro 1979, Burrows *et al.* 1979). They mark areas that have undergone localized strain softening and are initially sited by localized stress concentrations or material heterogeneities. The development of shear zones in laboratory specimens is known to produce an overall reduction in stress in constant strain rate tests (Burrows *et al.* 1979, Chan & Asaro 1979) or an increase in strain rate in constant stress tests (Post 1977). This indicates that there has been a reduction firstly in any localized stress concentration and secondly in the bulk applied stress in these specimens. The question remains as to the extent and to the scale that these are reduced in nature. It is likely that localized stress concentrations giving rise to shear zones have been relaxed to the regional value, but the extent of relaxation of the regional stresses and the area over which it occurs is not known. It is probably valid to assume that the area relaxed can be related to the size of the shear zone or shear zone array. It is unlikely that a stress build up would occur if the rock could accommodate imposed regional strain rates. Consequently, throughout this contribution we will regard a shear zone as a 'planar zone of concentrated, dominantly simple shear, deformation and which by itself or with associated zones helps to accommodate, or wholly accommodates, an imposed regional, or local, strain rate which the country rock cannot accommodate by bulk deformation'. It follows that a shear zone is a zone of strain softening, the degree of which will depend on the mechanical properties of the rock type which develops within the shear zone. In ductile shear zones these are mylonites which we regard as 'the rock produced in a ductile shear, or fault, zone and which allows the zone to accommodate the imposed strain rate by dominantly ductile processes'. As will be discussed later, localized fracturing of resistant minerals can occur. The important point is the behaviour of the matrix; in a ductile shear zone it deforms by crystal plastic processes, in a brittle shear zone it deforms by friction sliding and rigid body rotation.

The accommodation of a given imposed strain rate will require the development of a critical volume of mylonite. Consequently, a shear zone will grow laterally or new ones will form until this is achieved; the mylonite formed before the critical volume is reached will not be able to accommodate the imposed rate and it will harden. During growth, a given zone should be continuous with the country rock whereas discontinuous zones (cf. Burg & Laurent 1979) are likely to form after the critical volume of mylonite has formed. The amount of mylonite required will depend upon deformation parameters and on the softening processes that operate. Evidence for the latter should therefore be contained in mylonite textures and fabrics and will be discussed in the following sections.

MYLONITE DEVELOPMENT IN DUCTILE SHEAR ZONES

From the outset we wish to re-emphasise that we regard a mylonite as the material instrumental in strain softening. The fact that mylonites are normally fine-grained indicates that a fine-grained material is, in most circumstances, softer than its coarser grained equivalent. There are circumstances in which coarse grains may be more efficient (discussed later in this section).

The term mylonite was first used by Lapworth (1885) to describe fine-grained rock in the Moine Thrust Zone at Eriboll. The term implies grain size reduction by brittle processes and this has remained the universal view until recently (see recent reviews by Christie 1963, Higgins 1971, Zeck 1974). It had been noted frequently that the 'milled' rock had a recrystallized texture which was thought to result from post tectonic grain growth of the fragments.

Extensive grain growth was invoked to explain the presence of large, apparently undeformed porphyroclasts. Consequently, a complex terminology based on the amount of milling and grain growth emerged (Higgins 1971). However, recent microstructural, petrofabric and transmission electron microscopy studies (Bell & Etheridge 1973, White 1973, Lister et al. 1977) have indicated that the recrystallization was syntectonic. This led to considerable simplifying of the mylonite terminology and has resulted in a clear division between cataclasites and mylonites (White 1976, Sibson 1977). In the former, grain refinement was by cataclasitic processes with subsequent deformation mainly by the sliding and rotation of fragments. In the case of mylonites, the grain refinement was by recrystallization or neomineralization with subsequent deformation being by ductile processes. There is the grey area between the two, namely when grain refinement has been by fracturing and subsequent deformation has been by water-assisted diffusive processes (pressure solution). Although this must exist, no microstructural criteria for its recognition have emerged. As will be discussed later, fragments with associated fibrous overgrowths need not give any information about flow in the surrounding matrix.

The above simple distinction between cataclasite and mylonite is for an ideal monominerallic rock. The distinction can be applied to polyminerallic rocks by considering the behaviour of the matrix—in a mylonite the matrix behaves ductilely, in a cataclasite it deforms mainly by rigid body rotation of the fragments. It is not uncommon to find that harder minerals in a mylonite may undergo grain refinement by fracturing, for example feldspar grains in quartz feldspar mylonites (Wakefield 1977, Mitra 1978) and enstatite in mylonitised ultrabasic rocks (Nicolas et al. 1973, Basu 1977). It is also known that hard phases in metals fracture during the ductile deformation of the matrix. Thus subsidiary brittle deformation may occur during mylonitisation. In fact if we re-examine a mylonite from Lapworth's type locality (Teall 1918) we find that it consists of a matrix of fine-grained, well aligned micas and chlorites with angular grains of fractured quartz forming porphyroclasts (Fig. 1). There is no evidence for any brittle deformation in the matrix, grain refinement occurred by neomineralisation. Thus the above sense in which the term 'mylonite' is used does encompass Lapworth's type specimen.

The development of a mylonite microstructure with increasing shear strain has been the subject of many papers (see papers in Lister et al. 1977 and papers by Brodie 1980, Brown et al. 1980 and Hudleston 1980). The percentage of recrystallized grains increases until at shear strains (γ) between 2 and 5 a 'steady state' microstructure consisting of recrystallized grains and a few porphyroclasts forms. The stages are summarized for a quartz mylonite in Fig. 2. As strain increases, the coarse-grained quartzite is converted into a protomylonite, then into a mylonite and finally into an ultramylonite (see Sibson 1977 for a discussion of the terms) as the percentage of recrystallisation increases. In a monomineralic mylonite the porphyroclasts that resist recrystallization are those that are most unfavourably orientated for slip or those that are perfectly orientated (Bouchez 1977, Carreras et al. 1977). The former tend to form globular porphyroclasts (Tullis et al. 1973, Buiskool Toxopeus 1976, Bouchez 1977) which eventually recrystallize, whereas the latter tend to form tabular or ribbon grains parallel to the foliation (Bouchez 1977, Carreras et al. 1977) and remain to high shear strains. We will return to ribbon porphyroclasts later.

Recent work has indicated that the ultimate size of the recrystallized grains in a monomineralic mylonite is stress, strain rate, and to a lesser extent temperature dependent (Mercier et al. 1977, Twiss 1977, White 1979 a, b). The role of temperature is controversial with some regarding it as having no significant effect upon grains size during dynamic recrystallization (Twiss 1977). However, field observations suggest that temperature is

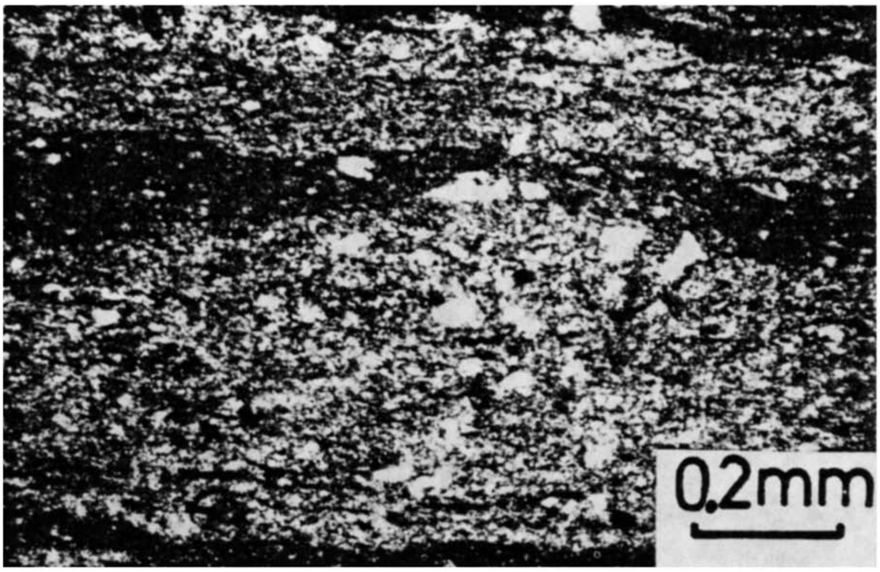

Fig. 1. A mylonite from Lapworth's type area, Eriboll N. W. Scotland. The mylonite consists of quartz fragments in a mica and chlorite matrix (plane polarized light).

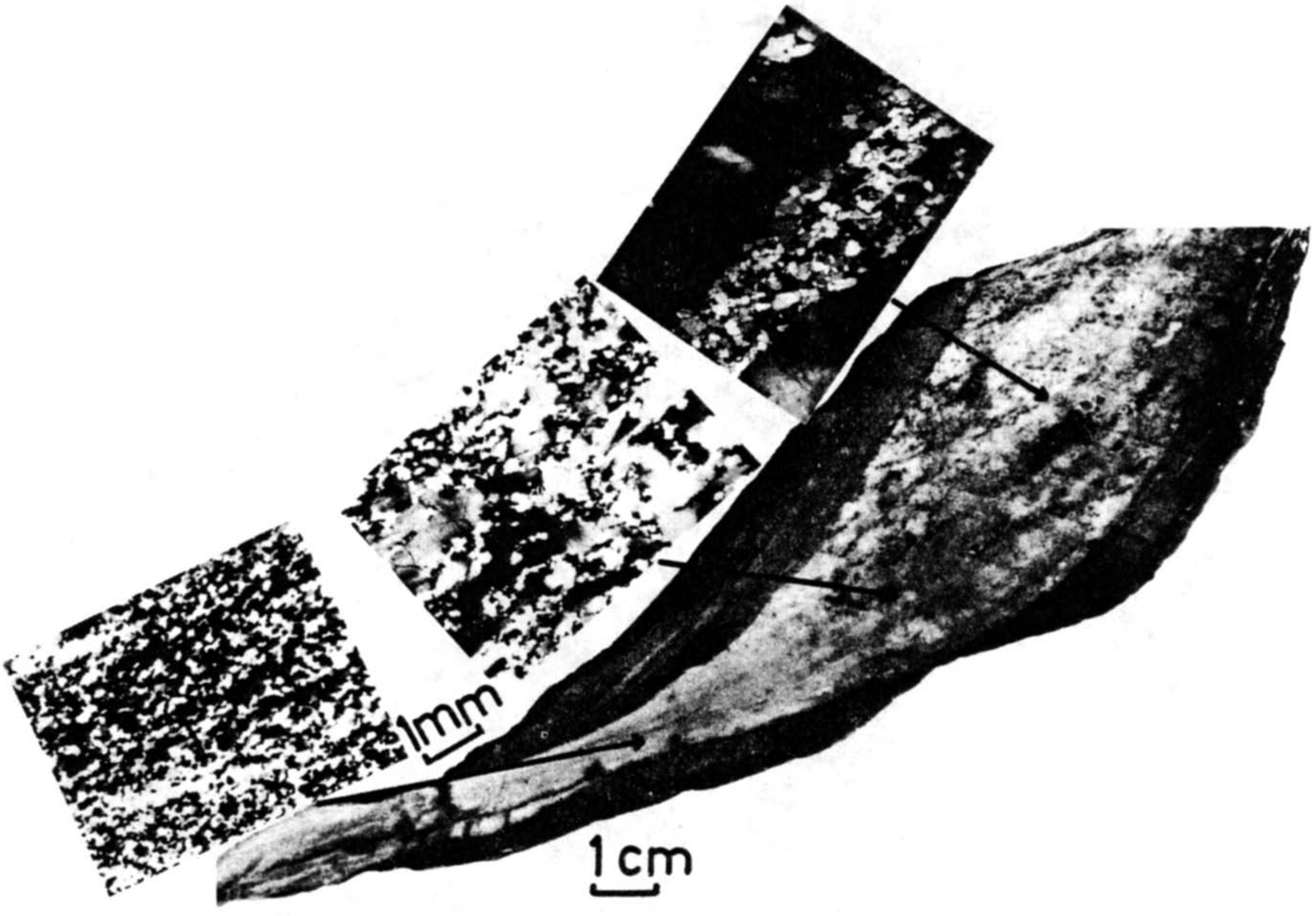

Fig. 2. The progressive development of a mylonite microstructure with increasing shear strain, quartz band, Cap de Creus, N. E. Spain. The stages from a protomylonite to a mylonite and to an ultramylonite are shown. Note that, in this case, the ultramylonite is relatively coarse-grained.

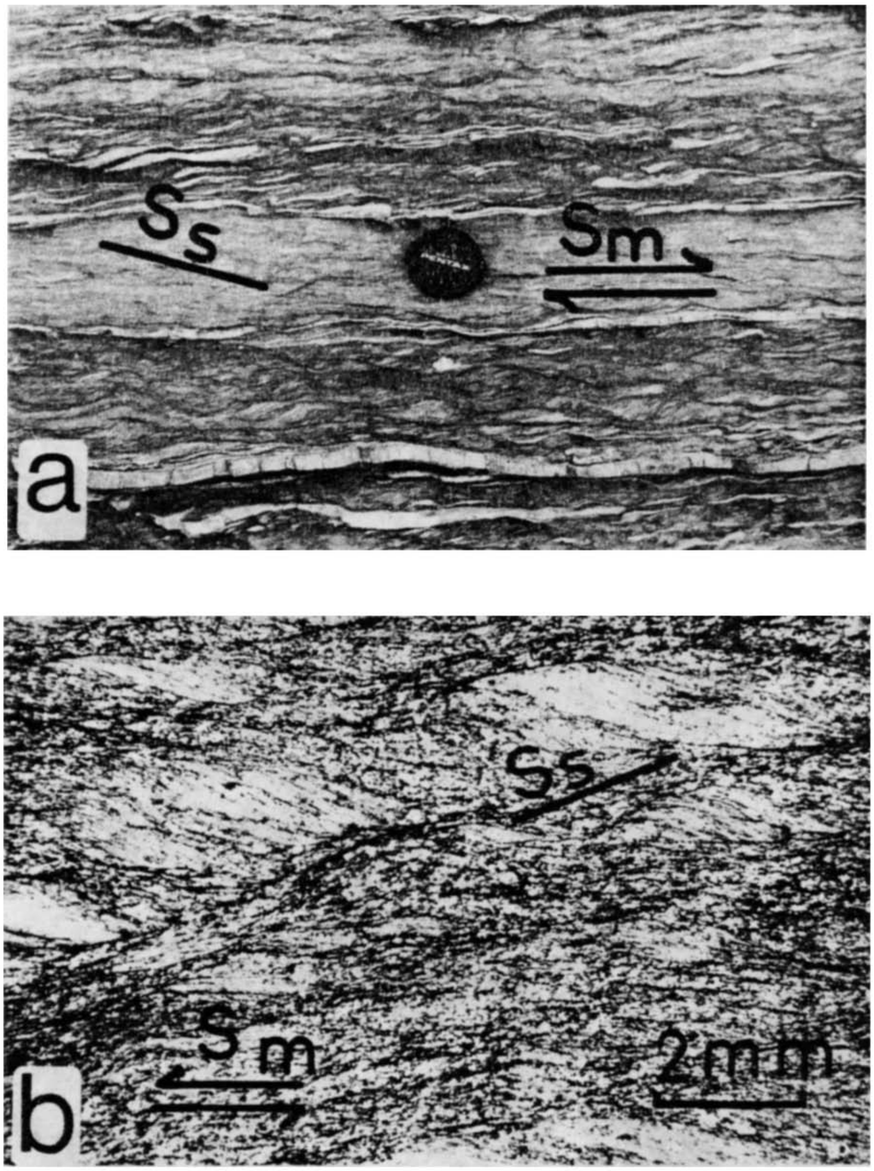

Fig. 6. Shear band foliation in the Cap de Creus phyllonites. The direction of shearing is marked on both the field photograph (a) and on the micrograph, (b) of a thin section from the phyllonites (a) showing the details of the foliation. S_m and S_s mark the mylonite and shear band foliations respectively. The acute angle between S_m and S_s always points in the shear direction.

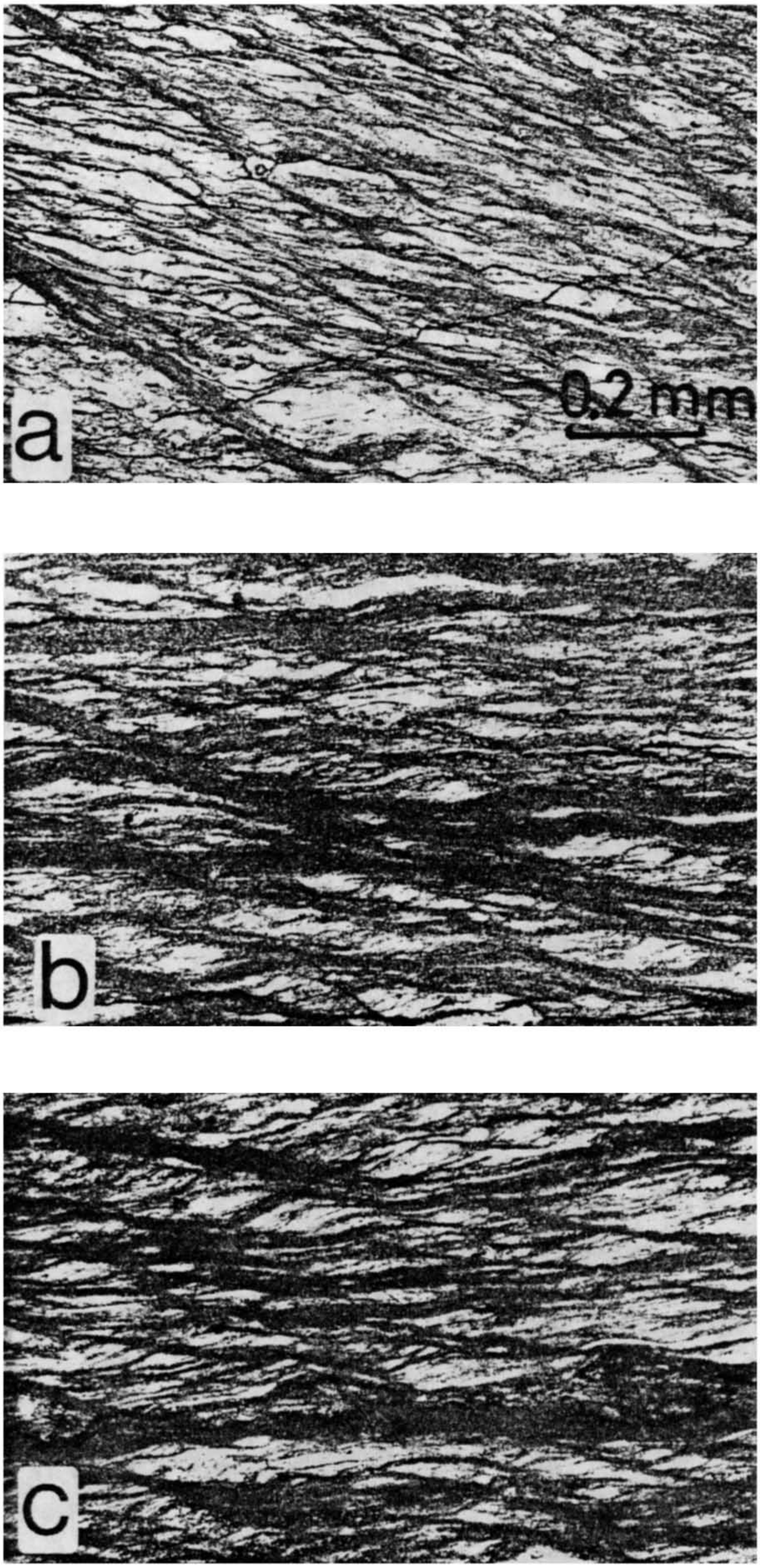

Fig. 7. Shear bands progressively destroying the mylonitic foliation in a calcite ribbon mylonite with increasing shear strain (a–c). Samples from the Tutt Head Shear Zone, Gower, Pembrokeshire (plane polarized light — all micrographs have the same magnification).

an important factor, mylonites formed at lower temperatures tend to be finer grained than those from high grade environments. Watterson (1979) has recently pointed out that in many high temperature shear zones, the rocks become heavily deformed but do not show a marked grain refinement.

The development of, a mylonitic microstructure in polyminerallic rocks is more complicated. Minerals harder than the matrix tend to remain as porphyroclasts, e.g. feldspar grains in quartzo-feldspathic mylonites and orthopyroxenes in peridotite mylonites. Feldspar grains normally exhibit limited ductility in low temperature (greenschist facies) mylonites, thought to reflect high resistance to slip (Peierls stress) and limited climb (White 1975). As a result of the latter, dislocation densities build up rapidly and the grains recrystallize or fracture before exhibiting any marked elongation. Marked intracrystalline plasticity of plagioclases, other than albite, and especially of alkali feldspar is only seen in high grade (amphibolite–granulite facies) mylonites (Goode 1978). The fracture behaviour of hard minerals in a mylonite depends greatly upon the presence or absence of prior ductility. Shear fractures are common in minerals such as plagioclase or amphibole (see Allison & Latour 1978) which have undergone limited prior ductility. However some minerals, such as tourmaline, which do not show any evidence of prior plastic deformation, have extensional fractures, often with subsequent pulling apart of fractured segments. The ultimate size of these fragments is likely to be stress controlled. It has been established that the size of similar fragments in ductilely deformed metals is dependent upon the flow stress in the matrix. Two explanations have been offered, based on the fibre loading mechanism and the dislocation pile up mechanism (Kelly 1968, Gurland 1972, Broek 1973, Goods & Brown 1979). In the former it is assumed that the tangential stress exerted on the particle surface by the flowing matrix is balanced by the tension in the non-deforming particle. The maximum stress forms at the centre of the particle and the fractures form perpendicular to the extensional axes. Thus elongated or fibrous particles break at the centre into halves, the halves into quarters and so on until an equilibrium length to width ratio, $\frac{l}{w}$, forms which is related to the flow stress σ in the ductile matrix and the fracture strength of the brittle matrix, σ_f, by the following equation

$$\frac{l}{w} = \frac{\sigma_f}{\sigma}.$$

The second theory is based on dislocation pile ups. These are thought to occur at points along the particle–matrix interface and the stress associated with the pile up exceeds the fracture strength of the particle, again producing extensional fractures. It has been found that the flow stress must exceed σ_c

$$\sigma_c = \left(\frac{d}{x+d}\right)^{\frac{1}{2}} \left(\frac{2\gamma\mu}{\pi(1-\nu)}\right)^{\frac{1}{2}}$$

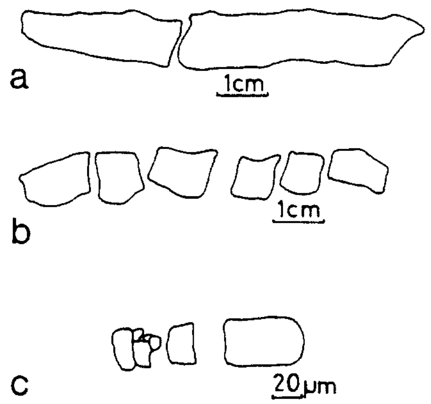

Fig. 3. Sketches of fractured brittle minerals in mylonites. (a) and (b) Fractured tourmalines, sketched from a colour transparency, in a mylonitized pegmatite, Cap de Creus, N. E. Spain, (a) is from the edge of the zone, (b) from within. (c) Fractured apatite sketched from Allison & Latour (1977). Note the wider spacing of the central fractures, indicating fibre loading.

where d is the width of the particle, x the length of the pile up, γ the crack surface energy, μ the shear modulus of the matrix and ν is Poissons ratio of the matrix. It is seen that σ_c is independent of the length of the particle and cracks can occur anywhere along the length but will preferentially form in large grains. This enables the one process to be distinguished from the other. In the fibre-loading mechanism the particle fractures in the centre and then pulls apart, resulting in a maximum spacing between pulled apart particles at the centre. Evidence for the fibre loading mechanism can be seen in tourmaline in mylonitised pegmatites at Cap de Creus and also in apatites in some Grenville Front mylonites (Allison & Latour 1977). These are sketched in Fig. 3.

GRAIN REFINEMENT PROCESSES

The recrystallization and also neomineralization that produces most mylonites, preferentially occurs around the edges of the original grains, especially in the case of quartz, carbonate, olivine and plagioclase and is a reflection of the core and mantle structure (White 1976) common in many deformed minerals. It is due to the marked difference in behaviour between grain centres and grain edges during the deformation of a polycrystalline material and it is particularly evident in minerals as they seldom have the five independent slip systems required for homogeneous deformation (von Mises 1928). As pointed out by White (1976, 1977) for quartz, the grain interiors behave essentially as single crystals; the accommodation of constraints imposed by the internal and external boundary conditions occurring in the mantle. It follows that the deformation in the mantle must vary both spatially and temporarily, as the constraints around the grain continuously alter during flow, and is the means by which cohesion across adjacent grain boundaries is maintained. Consequently, slip within the mantle will be largely governed by geometrical considerations. The difference in subgrain shapes (White 1976, White *et al.* 1979) between the core and mantle of quartz grains plus recent Burgers Vector determinations (White et al. 1979) indicate the operation of harder slip systems in the mantle. The smaller subgrains in the mantle suggest a stress concentration adjacent to grain boundaries and this, in turn, would activate harder slip systems.

The concentration of deformation within the mantles produces sharp and variable lattice misorientations. Consequently recrystallization is concentrated at grain margins with new grains developing by strain-induced boundary migration and sub-grain rotation (White 1973, Nicolas & Poirier 1976). Walls of geometrically necessary dislocations, which must undergo a continuous and erratic movement in the mantle as constraints alter, are likely to be important sources of new grains. The *c*-axis fabrics of the new grains formed by recrystalization processes are similar to those for the old grains, with the orientation of newly formed grains being contained within the range of misorientations in the mantle, which may be appreciable (White et al. 1979).

Neomineralization processes such as the formation of chlorite from biotite as seen in some of the Cap de Creus mylonites and phyllonites and of hornblende from pyroxene as occurs in the shear zones in the Lewisian Scourie dykes of NW Scotland (see Beach 1980) will also occur preferentially in grain mantles. This is due to the enhancement of reaction rates by the presence of high dislocation densities as described by White (1975).

The preservation of small grains in mylonites indicates that little growth has occurred after recrystallization. In some instances this may be due to grain growth inhibition by grain boundary inclusions and bubbles (White 1979c). In others there is no clear reason for the fine grain size but it is thought to be due to the rapid equilibration of dislocation densities from grain to grain after recrystallization (McQueen & Jonas 1975, White 1976, Twiss 1977, Twiss & Sellers 1978). It follows that porphyroclasts present in monomineralic mylonites should be partially recrystallized old grains. This is certainly true for ribbon-shaped porphyroclasts in low temperature mylonites as they show evidence of deformation. However, in high temperature mylonites, although orientated for easy slip (Boullier & Bouchez 1978), they often show less evidence for intragranular deformation than do the adjacent smaller matrix grains. An explanation for this difference may be found in recent experiments on the deformation and recrystallization of camphor (Urai et al. 1980). Ribbon grains formed in the recrystallized matrix during high temperature (>0.5Tm) deformation. The camphor grains that grew were those orientated for easy slip. The authors speculated that this was because elastic strain energy would easily be relieved by slip in such grains, whereas it would be stored in grains with other orientations; that is, the difference in stored elastic energies would become the driving force for growth. Strain incompatibility at the ends of the growing grain can be accommodated by local diffusion at elevated temperatures whereas at lower temperatures it would tend to lead to a build up in the elastic strain energy which would remove the driving force for further growth. Thus ribbons due to preferential grain growth are favoured by high temperatures whereas 'stretched' ribbons are therefore likely to form at lower temperatures where diffusion is less rapid.

FABRIC DEVELOPMENT

Although numerous fabric studies have been reported for a variety of mylonites over the past fifty or so years, it is only recently that attention has turned to the progressive development of fabrics with increasing shear strain (Hara et al. 1973, Bouchez 1977, Carreras et al. 1977, Burg & Laurent 1978, Berthe et al. 1979a).

These latter studies have shown that a fabric progressively develops at low shear strains with a well developed fabric pattern having formed by $\gamma \simeq 5$. The intensity of the maxima is dependent upon the amount of other minerals present, e.g. feldspar in a quartz feldspar mylonite (Price 1978, Lister & Price 1978, Starkey & Cutforth 1978) and possibly temperature (Hopwood 1976). In most quartz and olivine mylonites, the fabric remains more or less steady as deformation proceeds to higher strains, but in some fine-grained mylonites it may tend to become more random (Berthe et al. 1979 a, b). The latter is thought to reflect a change in deformation mechanism after grain refinement to one in which grain boundary sliding dominates. The effect of this on fabrics is shown in the experimental work on calcite by Schmid et al. (1977). The former behaviour is due to a maintenance of a dislocation deformation mechanism which is likely to be of a warm or hot working type. However, differences in fabric have been reported in given quartz and olivine mylonite bands and may result from local differences in deformation path due to the presence of sheath folds (Carreras et al. 1977) or to the wrapping of the ductile matrix around hard porphyroclasts (Buiskool Toxopeus 1976, Lister & Price 1978). The susceptibility of a fabric to deformation path is well illustrated in the computer simulations of Lister et al. (1978). These simulations (cf. Lister & Price 1978) and the published data above emphasize that the fabric is related to the kinematic framework in which the important elements are the shear plane, direction and sense of shear and orientation of the stretching axis—local variations in any of these will produce changes in elements of the steady state fabric.

There appear to be three main types of *c*-axis fabrics in quartz mylonites, type I and type II girdles (defined by Lister 1977) and a point maxima perpendicular to foliation and lineation. They are illustrated in Fig. 4 and are thought to primarily reflect the operation of different slip systems (Tullis et al. 1973, Wilson 1975, Lister et al. 1978) along with different amounts of climb (Ball & White 1978). For example, fabrics a and b and c in Fig. 4 have been related to dominant prism plus rhomb and basal slip respectively, in the case of type I girdles this has been confirmed by Burgers vector analyses (Carreras et al. 1977).

The girdles, especially the type I girdles, and the point

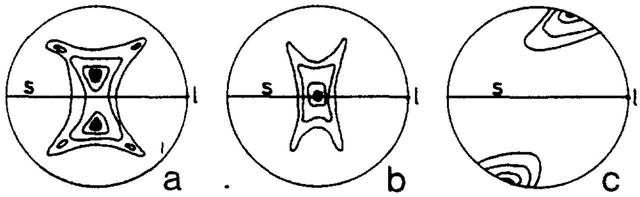

Fig. 4. Sketches of three typical c–axis fabric patterns for quartz mylonites. (a) Type I girdle (after Carreras *et al.* 1977). (b) Type II girdle (after Bouchez 1977). (c) c-axis point maxima (after Wilson 1975). *s.* and *l* are the foliation plane and lineation direction respectively.

maximum fabric as shown in Fig. 4(c) are often asymmetric, the sense of which has been related to the flow direction in the shear zone. However this is, at present, a source of controversy (see Simpson 1980) as in some mylonites it is in the same sense as the sense of shearing (Lunardi & Baker 1975, Bouchez 1977, Burg & Laurent 1978); in others it is in the opposite sense (Lister & Price 1978) and some are symmetrical (Hara *et al.* 1973). Carreras *et al.* (1977) show all three variants existing within a single band of Cap de Creus mylonite. This again emphasizes the effects that local variations in the kinematic framework can have on fabrics.

As stated earlier the conclusions drawn recently by several authors from fabric studies is that the matrix grains in both quartz and olivine mylonites and the elongated porphyroclasts remaining at high shear strains are orientated for easy slip (Wilson 1975, Buiskool Toxopeus 1976, Bouchez 1977, Carreras *et al.* 1977, Burg & Laurent 1978, Lister & Price 1978). Recent *a*-axis fabrics for quartz mylonites with type II *c*-axis girdles show that there is a preferential alignment of the *a*-axes parallel to the stretching lineation in the mylonite (Bouchez 1978), again consistent with the grains being orientated for easy slip. The same effect has been demonstrated for mylonites formed in experimentally formed shear zones in magnesium metal (Burrows 1979, Burrows *et al.* 1979). These observations indicate that the slip planes rotate into parallelism with the shear plane during a simple shear deformation and not the flattening plane as is the case in an axi-symmetric deformation.

STRAIN SOFTENING

Little attention has been paid to the softening processes that must accompany the localisation of deformation in shear zones. If marked softening did not occur, the shear zone would widen rapidly and would become a self-arresting instability akin to a Luders band in a metal (Reid 1973). The softening, in mechanical terms, can be expressed as a reduction in stress at constant strain rate or an increase in strain rate at constant stress. It is likely to be a combination of both in shear zones. It has been argued that shear zones may initially be localized by stress heterogeneities which are relaxed by the formation of the mylonites, alternatively localized softening such as a metamorphic reaction (see below) could lead to an increase in strain rate in the reaction products (Cobbold 1977, White 1979c).

Strain softening can occur by a number of processes (see also Poirier 1980) which will be discussed below. These include:

1. a change in deformation mechanism;
2. geometric, or fabric, softening;
3. continual recrystallization;
4. reaction softening;
5. chemical softening;
6. pore fluid effects;
7. shear heating.

1. *Change in deformation mechanism*

As the grain size is reduced during recrystallization, it is likely that the deformation mechanism will change from warm or hot working to a mechanism in which grain boundary sliding dominates, resulting in superplastic behaviour (White 1976, Schmid *et al.* 1977, Twiss 1977). On the other hand, if water is present, a form of pressure solution will be favoured. In both cases, strain rate becomes inversely proportional to the second or third power of the grain size; the former if sliding is accommodated by dislocation processes (Schmid *et al.* 1977) and the latter if accommodation is by diffusive processes (White 1976) or by pressure solution (Rutter 1976). All processes are capable of producing a marked softening (Rutter 1976, White 1976, 1977, Schmid *et al.* 1977).

The above changes in mechanism can be recognised microstructurally as discussed by White (1976, 1977, 1979c). For superplasticity to occur in a monomінerallic mylonite, the matrix grains should be smaller than the sub-grains in remaining porphyroclasts, they will tend to have square or rectangular shapes and will display no marked crystallographic fabrics (see also Schmid *et al.* 1977). The general requirement of a fine grain size means that this softening process is favoured by high initiating stresses, low homologous temperatures and the presence of a dispersed second phase which inhibits grain growth. Superplasticity is most likely to occur at low temperatures in fine-grained ultramylonites. However, the rarity of weak or random mylonite fabrics suggests that superplasticity may not be as common as first thought (White 1976).

2. *Geometrical softening*

Geometrical softening has been discussed at length in the recent metallurgical literature (Dillamore *et al.* 1979) but has received little attention from geologists. This is in spite of well developed fabrics in most mylonites which can be related to the re-orientation of the grains for easy slip (see the previous section). Geometrical softening can be best envisaged as the result of the increased resolved shear stress for intragranular slip on a given plane as it rotates into parallelism with the edges of the shear zone and as the slip direction rotates into parallelism with stretching lineation. The flow stress for deformation by dislocation glide processes is then at a minimum. Geometric softening is most pronounced in materials with limited slip systems (Reid 1973). The extent of softening will depend on the starting fabric and its orientation to the shear plane and shear direction. There is no information available for the softening of polycrystalline minerals as a fabric is developed but it can be calculated from the Taylor (1938) and Bishop-Hill (1951) theories. However, this is outside the scope of the present paper. During experiments, geometric softening is normally accompanied by continual recrystallization. Some idea of the degree of softening produced by their combined effects can be obtained from the magnesium data of Burrows *et al.* (1979) which is reproduced in Fig. 5. It should be noted that the flow stress after softening is almost independent of temperature but that the extent of softening is strongly temperature dependent and reflects a more intense fabric at the lower temperatures, a result of limited available slip systems. The softening occurred over the period in which the shear zones initiated and grew. No marked softening occurred in the samples in which a shear zone did not develop.

3. *Continual recrystallization*

This leads to a softening by ensuring that, in a deforming mylonite, there are always some newly formed strain-free grains. The extent of the softening will depend upon the volume fraction of these at any given instance. As continual recrystallization is normally associated with geometric softening, it is difficult to establish its effectiveness as a softening mechanism. However, metallurgical studies (Stuwe & Ortner 1974, McQueen & Jonas 1975) indicate that it only leads to a slight softening.

4. *Reaction softening*

Enhanced ductility due to a metamorphic reaction has been discussed briefly by White & Knipe (1978) but, again, has been rather neglected by geologists. White & Knipe (1978) concluded that a metamorphic reaction will aid softening by producing small grains that are likely to promote superplasticity and also by producing soft, strain-free grains. Furthermore, hard phases such as feldspars may be converted to soft assemblages such as quartz and sericite. There is the additional weakening effect due to 'transformational superplasticity'. This is the softening that occurs when deformation is concurrent with a phase change or mineral reaction (see Nicolas & Poirier 1976, White & Knipe 1978) and results from stresses induced by volume changes and by increased diffusion rates during the transformation or reaction.

5. *Chemical softening*

Chemical softening results from changes in the trace element content of a mineral. The best example is the water weakening of quartz. The introduction of water into a shear zone may lead to retrogression (see Beach 1980) but in a quartz mylonite, the quartz itself may take up small amounts of water. The weakening effects of lattice-bound 'water' are dramatic, for example Carter (1976) lists a drop in flow stress of 50% for 'wet' quartz single crystals. There are no data available for the weakening of quartzites. Water weakening has also been recorded for olivine (Post 1977) and enstatite (Ross & Nielsen 1978). It is possible that other trace elements such as the alkali ions in quartz also lead to weakening but to a lesser extent (White 1975, Knipe in press).

6. *Pore fluid effects*

The presence of pore fluids can significantly reduce the strength of a rock by a mechanical and by a physico-chemical effect (Fyfe *et al.* 1978). The first is a lowering of the fracture strength of a rock by lowering the effective stress and therefore leading to cataclasis, and is outside the scope of this paper. The second effect is the increase in ductility (Rutter 1972) due to the Rebinder effect (Westwood *et al.* 1967). Rutter's (1972) experiments on calcite rocks suggest that pore fluid effects would produce a significant softening in those shear zones which act as fluid channelways.

7. *Shear heating*

Shear heating is the increase in temperature that can accompany deformation, especially if it is concentrated into narrow zones. This topic is reviewed by Brun & Cobbold (1980). Significant shear heating can result from high stresses and fast strain rates. However, it is unlikely that high stresses are maintained after softening has occurred. Shear heating is not regarded as a significant process during shearing (see Brun & Cobbold, 1980) and is unlikely to be a major softening process in most shear zones.

From the above, it can be seen that there are a number of processes which, singularly or in combination, can lead to strain softening. The extent of the softening with respect to the adjacent country rock and consequently the width of the shear zone or shear zone system required to accommodate a given imposed strain rate at a given temperature will be dependent upon the operative processes. These in turn will reflect the mineralogy and petrology of the rocks in which the shearing has occurred. However, for a given rock type, the softening may be more effective at lower rather than at higher temperatures (see Fig. 5). Both superplasticity and fluid effects, the latter possibly leading to reaction and chemical softening, are favoured by low temperatures, that is by conditions favouring fine-grained ultramylonites. Geometric softening, as indicated by well developed crystallographic fabrics, will be important in coarser grained ultramylonites, mylonites and protomylonites. However, it will be less effective as temperature increases because of the observed weakening of both mineral and metal fabrics at higher temperatures (Hopwood 1976, Burrows 1979). Thus it is expected that the difference between the strength of a mylonite and its parent rock will decrease as temperature increases and may be one explanation, along with a general weakening of all rocks with increasing temperature, for the increasing width of a ductile fault zone with depth (Sibson 1977).

THE EFFECT OF VARYING DEFORMATION PARAMETERS ON MICROSTRUCTURE, FABRICS AND SOFTENING PROCESSES

A central theme of this contribution has been that a mylonite microstructure and fabric is the net product of the softening processes that localised the deformation into the mylonite zone. Both have developed specifically to meet set conditions of strain rate, stress and temperature with each mineral assemblage responding in whatever way it can to produce a soft product. Most mylonites are *L–S* tectonites and therefore form to accommodate shear on a set plane in a set direction. It is interesting to consider what happens if the flow direction, strain rate or temperature alters. Again this is illustrated for a quartz mylonite.

An incrêase only in the imposed strain rate should have no major effect on either the microstructure, fabric or softening processes, but should lead to the production of more mylonite, i.e. lateral growth of existing shear zones or the initiation of new ones. However, a change in flow direction in a mylonite zone will have a great effect upon geometric softening processes but little effect on other processes. If we consider a fabric such as that in Fig. 4(a) then the *c*-axes girdle is perpendicular to the lineation; that is, to the movement direction. If this alters due to a reorientation of the external stress field, then the majority of grains in the girdle are no longer orientated for easy slip. Either a reorientation of the fabric must occur or a new mylonite form. Until this happens, there will be an effective hardening and an associated increase in stress within the shear zone, i.e. the zone will temporarily lock up. Changes in temperature are likely to be even more significant as besides

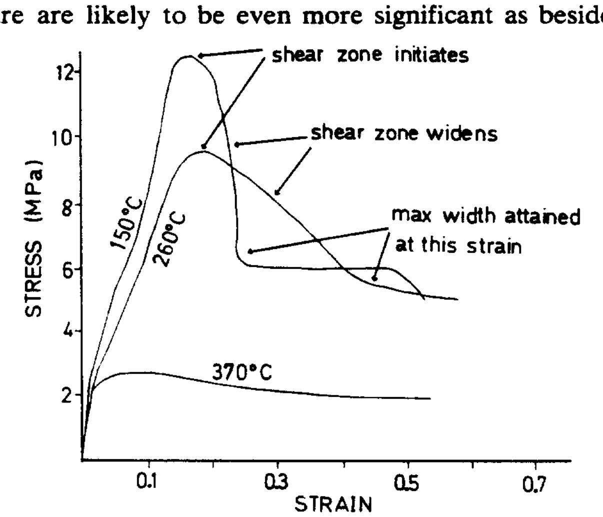

Fig. 5. Stress strain curves for polycrystalline magnesium showing the stress decrease associated with the development of shear zones. Shear zones developed at 150°C and at 260°C. No shear zone formed at 370°C. Note the more marked softening at lower temperatures (based on Burrows *et al.* 1979).

affecting strain rate, it also affects diffusion rates, slip systems, grain size and possibly softening processes. We will concentrate on the effects of a temperature decrease in an active zone as this gives some insight into the effect of uplift. Again, a quartz mylonite zone will be used for illustration. If activity has ceased, the mylonite zone, because of its finer grain size, will be stronger than the country rock because of the Hall Petch relationship (Honeycombe 1967) described by

$$\sigma = \sigma_0 + kd^{-2}$$

where σ is the flow stress, σ_0 is the internal or friction stress, k is a constant and d is the grain size. Consequently the mylonite zone will be shielded by the softer country rock and will not be affected by uplift. This will not be the case for an active zone.

In most high temperature quartz mylonite zones, i.e. above mid greenschist facies, the softening should be by continual recrystallization and geometric softening, with the prism plane being the favoured slip plane in the grains. As the temperature decreases the slip system changes to the basal plane (Tullis *et al.* 1973) and the grains must reorientate, if geometric softening is to be maintained. The effect is similar to that described above for a change in flow direction, viz. the grains must reorientate. However, continual recrystallization is expected to continue during any temperature decrease and will produce smaller grains. Eventually this could result in a change in deformation mechanism to a superplastic process which, as stated previously, appears to be a most effective softening process. If it is a more effective softener than the combined geometric and continual recrystallization process it replaces, it will cause a decrease in the width of the active shear zone and produce a bordering inactive zone with a fabric and microstructure reflecting the prior deformation. A continuing decrease in temperature will lead to a slowing down in the grain boundary sliding rate which is thermally activated. Although any ingress of water can initiate softening by chemical and/or pore fluid processes, ultimately the imposed strain rate will not be accommodated, and the stress within the shear zone will build up and result in fracturing. However, the fine-grained mylonite will be much stronger than either the coarse grained mylonite or the country rock (see above equation) and cataclasis should be concentrated in them. The result is a laminated shear zone of gouge, cataclasite, ultramylonite and mylonite such as that occurring along the Alpine Fault in New Zealand (Sibson *et al.* 1979).

HIGH STRAIN DEFORMATION IN A MYLONITE ZONE

Up until this point, we have been mainly concerned with the microstructures, fabrics and softening mechanisms that lead to the establishment of a mature shear zone. Little attention has been given to deformation after the formation of the mylonite. Sheath folds frequently form at high strains (see Cobbold 1980).

Although they do not appear to lead to any obvious change in deformation mechanism, they may lead to local variations in fabric and microstructure (Carreras *et al.* 1977). However, the most obvious feature that appears at high strains is the development of a late penetrative foliation at a low angle (<45° and typically at circa 35°) to the mylonite foliation. It is common in the pelitic mylonites (phyllonites) in the Cap de Creus area (Fig. 6) and has been reported in other phyllonites (Sibson 1977, Bell 1978, Platt 1979) and in quartz and quartzo-feldspathic mylonites (Berthe *et al.* 1979b, Gapais 1979). Occasionally two sets may develop both at ≃ 35° to the mylonite but the main one is always approximately parallel to the complementary shear plane to the active shear zone. If only one is formed it can be used to deduce the shear direction as shown in Fig. 6. Morphologically, the foliation resembles shear bands which form during the high strain deformation of metals particularly during rolling (Fargette & Whitham 1976, Dillamore *et al.* 1979, Malin & Hatherly 1979). They mark, in metals, the breakdown of homogeneous flow and are thought to be a consequence of the anisotropy that developed during the earlier stages of the deformation (cf. Dillamore *et al.* 1979). The shear bands break up this anisotropy and may at even larger strains bring about the re-establishment of homogeneous deformation; the breaking up of the foliation in a carbonate ribbon mylonite by shear bands is clearly visible in Fig. 7. In phyllonites, the anisotropy is due to the orientation of the phyllosilicates—in quartz and calcite it is due to the crystallographic fabric. The geologically significant feature of the second foliation is that it developed during the same deformation that produced the mylonite but at a later stage. It is probable that in many instances the shear bands may represent the final phase of ductile deformation in a shear zone as the temperature drops. Also it is one way in which an existing fabric or microstructure can be destroyed to be replaced by another which is softer.

SUMMARY

Mylonites in ductile shear zones are the soft materials produced by crystal plastic deformation mechanisms. They are responsible for the concentration of deformation within the shear zones. Continuing deformation is again by crystal plastic processes which may occur only in the matrix grains. Hard porphyroclasts may deform by brittle processes.

The mechanical weakness that is envisaged may be due to a number of softening processes. However, the existence of well defined crystallographic fabrics in most mylonites suggests that geometric (fabric) softening and continual recrystallization are the most important of these. Superplasticity which tends to be characterized by a weak or random fabric may not be as prominent as previously thought. The volume of mylonite generated in a given individual shear zone or shear zone array should be governed by its strength relative to that of the country rock and by the imposed strain rate to be accommodated. Mylonite generation will continue until accommodation can be achieved. A continuous shear zone is likely prior to this, a discontinuous one once it has happened. An oblique shear band foliation may develop in mylonites at high shear strains or as uplift commences, if the shear zone is still active.

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