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Fault-Zone Reactivation: Kinematics and Mechanisms

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Fault-zone reactivation: kinematics and mechanisms

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[Plate 1]

The kinematics and mechanisms of fault-zone reactivation are reviewed. Reactivation is dependent upon fault-zone orientation and the existence of weak mylonites along these zones. The recognition of reactivation within mylonite zones, and the softening processes that first concentrate deformation into these and secondly provide a weak medium for reactivation are discussed. Attention is given to the Darling Mobile Zone, Western Australia and the Redbank Zone (see Obee & White, this symposium) as examples of reactivated zones.

1. Introduction

The repeated tectonic reactivation of major fault zones in the continental crust was discussed by Hills some forty years ago (Hills 1946, 1956*a*, *b*, 1961). His hypothesis that major fault lineaments were planes of mechanical weakness was principally based upon the morphotectonics of continental Australia. At Melbourne University, Hills made a relief model of the country with a vertical exaggeration of *ca.* 40 times. He noted that many present-day prominent morphological features corresponded to major faults that criss-crossed Australia (figure 1) and that often they showed geological evidence for a long tectonic history of movement. He demonstrated emperically (Hills 1946, 1947) that most present basement tectonic trends are parallel to older trends and that some had long tectonic histories. A good example is the Muckleford Fault in Victoria, which was active from Ordovician to late Tertiary times. Hills (1956*b*) expressed the notion of structural rejuvenation of faults by using the term 'resurgent tectonics' to emphasize that movement along these faults was not continuous, but occurred after periods of quiesence or stability. So impressed was he by the extent of resurgent tectonics in Australia that he wondered what held the fractured and cracked continent together, but provided no answer. Such an answer will be attempted in this article.

Little attention was subsequently paid to resurgent tectonics until recently. Watterson (1975) recorded the tectonic persistence of major shear zones in Greenland and was the first to consider what feature might be responsible for making a major fault a plane of mechanical weakness in the crust. He advocated that this might be a reflection of the fine grain size, relative to the host country rock, of the mylonites which characterized these zones. White (1976) demonstrated how this might occur during a single deformation by invoking a change in deformation mechanism from dislocation creep to one in which grain-boundary sliding dominated and, consequently, the conditions for superplasticity pertained. A corollary of these two papers (Watterson 1975; White 1976) is that fine-grained mylonite zones could be planes of weakness for subsequent reactivation.

FIGURE 1. Fault-related morphotectonic features in Australia as recognized by Hills (1956*a*).

Slowly but surely a greater appreciation of the role of resurgent tectonics in the evolution of the continental crust has occurred (Watson 1980; O'Driscoll, this symposium). The notion of fault-zone reactivation has received a great boost from recent seismic reflection studies such as those across NW Scotland. The MOIST line (Brewer & Smythe 1984) revealed that thrusts have been reactivated as normal faults. A similar reactivation has been documented throughout the Cordilleran Metamorphic Core complexes of North America (Coney & Harms 1984) and in the Appalachians (Peterson *et al.* 1984). Thus the resurgent tectonics of major faults has a profound influence on the evolution of the continental crust and also on the siting of oceanic crust, as older faults may be the loci for later rifting. The hypothesis of resurgent tectonics is now becoming accepted, some forty years after being formulated by Hills.

In this paper it will be assumed that faults reactivate and we will concentrate upon the reasons for reactivation and the recognition of reactivation. However, first we will discuss the rocks that form along faults because these hold the key to reactivation; if they did not, reactivation would not take place and a new fault would propagate instead.

2. Fault rocks

Fault rocks can be subdivided into three groups on textural grounds (Sibson 1977; White 1982). These are the incohesive fault gouges, the cohesive cataclasites and mylonites. The distinction between the last two is that mylonites are foliated or banded and cataclasites are not. The terms used to describe different mylonites (i.e. proto-, blasto-, and ultra-), will be the same as those used by Sibson (1977) and White (1982). It is emphasized that the above distinction between mylonites and cataclasites does not restrict cataclasis to cataclasites. As noted by White (1982), a cohesive foliated fault rock (that is, a mylonite by the above definition), can arise mainly through cataclasis with subsequent deformation by crystal–plastic processes and is termed a c-mylonite. White (1982) uses the term r-mylonite for one arising by dominant crystal–plastic processes.

It is often assumed that the three fault-rock types represent a simple progressive depth sequence (Sibson 1977, 1982). The gouges, which may be foliated or non-foliated depending upon the original phyllosilicate content of the parent rock, are associated with shallow depths where temperatures are not high enough for sufficient clay mineral growth and precipitation of secondary minerals to bind up the rock after fragmentation. Binding occurs in cataclasites and mylonites which form at a greater depth. An additional assumption is that cataclasites are separated from mylonites by the brittle–ductile transition (used in this paper in the same sense as outlined by White (1984), namely the change from cataclastic to crystal–plastic deformation mechanisms) – the former fault-rock type occurring above the transition and the latter below it (Sibson 1977; Meissner & Wever, this symposium). However, this assumption is based upon a constant mineralogy between fault rock and parent, which is seldom exhibited: for, as will be shown later, a phyllosilicate-rich mylonite (phyllonite) is commonly derived from a granite or granitic gneiss and this has a different brittle–ductile transition from that of the granite. That is, one brittle–ductile transition applies for the establishment of the fault (that of the granite) and another (that of the phyllonite) for subsequent movement in the fault. A change from brittle to ductile behaviour and then back to brittle, or even cyclic brittle–ductile behaviour, can occur, as a fault zone with a reverse-movement component uplifts its hanging wall. The Alpine Fault in New Zealand is a good example where it has been shown that mylonites (c-mylonites) develop after initial cataclasis (Sibson *et al.* 1979; White & White 1983; White 1984). The brittle–ductile transition in the continental crust is unlikely to be sharp; it is more likely to be spaced over a temperature zone the width of which is dependent upon the geothermal gradient for a given rock type. The width and complexity of the transitional zone will be even greater of lithological variations occur. During reactivation, it is the brittle–ductile transition of the fault rocks that should be considered; these are likely to be different from those of the country rock and add further to the above complexities.

In this contribution, we shall concentrate mainly on deep mylonite zones in which ductile (crystal–plastic) deformation processes are dominant. We shall extend our comments to cataclasites and gouges where appropriate. We begin with an outline of features which may be used to recognize reactivation and then look at the reasons for the reactivation of a particular fault.

3. Recognition of reactivation

Reactivation is often easily recognized from stratigraphic criteria. It was these that Hills used to recognize reactivation of the Muckleford Fault. However, often it is not possible to get a clear picture of the movement history of a fault from stratigraphy alone. This is especially so in granite and gneiss terrains. In this section, the criteria which can be used to deduce movement directions within the fault rocks will be outlined, and in the following section, an example of the application of these to the Darling Fault in Western Australia will be given.

The key feature in deducing the movement direction in mylonites is the stretching lineation (Carreras *et al.* 1977). It gives the direction but not the sense of movement, which can be determined from other features adjacent to, and within, the mylonite zone (White 1982; Simpson & Schmid 1983; Evans 1984). These are (see figure 2):

(1) rotation of a pre-existing or generated foliation;
(2) rotation of deformed markers;
(3) asymmetry of intrafolial folds;
(4) microshears or C-bands;
(5) shear bands or S-bands;
(6) sheared porphyroclasts;
(7) rotation of fragments owing to shear fractures;
(8) rotation of fragments owing to tensile fractures;
(9) asymmetry of trails growing around rotating clasts;
(10) asymmetry of trails growing around non-rotating clasts;
(11) asymmetry of elongate recrystallized quartz grains;
(12) asymmetry of dragged-out mica porphyroclasts;
(13) asymmetry of quartz c-axis fabrics.

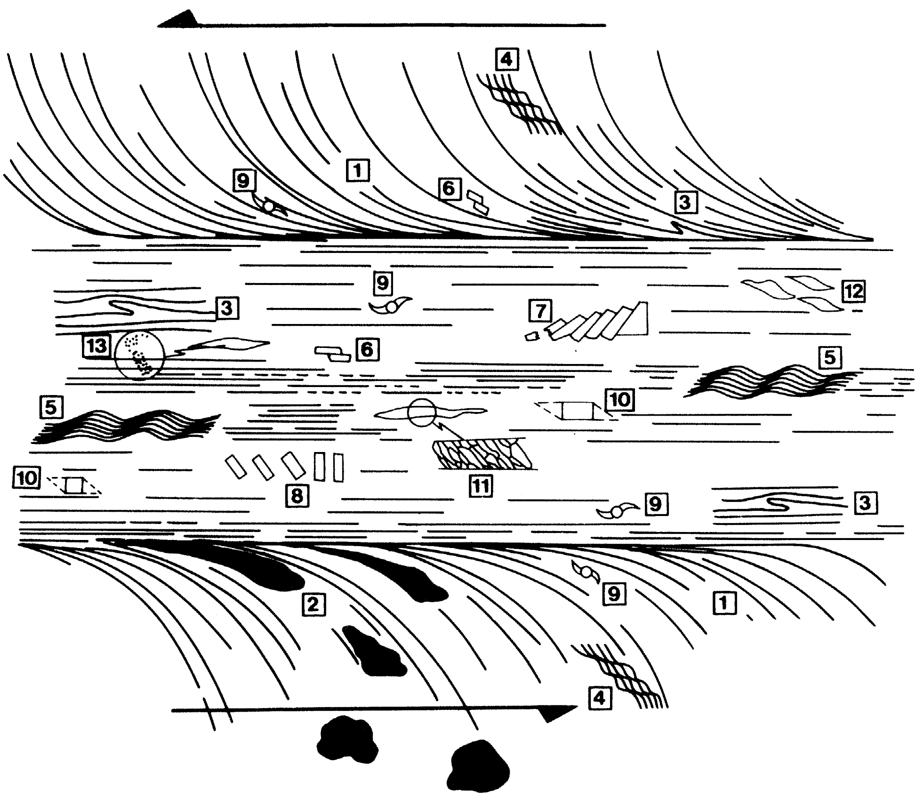

FIGURE 2. Kinematic indicators of the movement sense in mylonite zones. The features (1)–(13) are listed in the text.

The best features are those at the edge of the zones such as the rotation of pre-existing foliations or accompanying flattening foliations or the flattening and rotation of markers. Also at the edge of the zones, especially in granites and granitic gneisses, are microshears (C-bands, in the terminology of Berthe *et al.* 1979). Both are clear movement-sense indicators. Often it is difficult to find the edges of the zones, so movement indicators must be sought within the

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mylonites themselves. Shear bands (White 1979; White *et al.* 1980), also called extensional crenulation cleavages (Platt & Vissers 1980), are probably the most reliable of the criteria. They are best developed within phyllosilicate-rich (phyllonite) zones, although they can occur in monomineralic bands if these have a strong crystallographic preferred orientation (Gapais & White 1982). They occur after initial mylonite development. During movement, they are active shear planes and rotate backwards towards the primary mylonite foliation in a similar manner to slip planes in a crystal. Later shear bands may cut these earlier ones. The relation between S- and C-bands is clearly shown in figure 2 and no confusion should occur between them. The asymmetry of intrafolial folds are also a reliable indicator. This is particularly so for sheath folds (Carreras *et al.* 1977). Good examples of the use of these to deduce the movement direction can be seen in Carreras *et al.* (1977) and Evans & White (1984).

It is stressed that no single criterion is infallible and that as many as possible should be used. The likelihood of inconsistent movement-senses increases as the scale of the features used decreases and the effect of local heterogeneites becomes more apparent. This is particularly so with quartz *c*-axis fabrics, which can be affected by the presence of feldspar augens (Lister & Price 1978). Large augens can even affect the asymmetry of intrafolial folds. Particular care should be taken with shear fractures and rotated, pulled-apart, fibre-loaded particles. The shear fractures, especially in feldspars, are extensional features and, like shear bands, rotate in the opposite sense to the sense of shearing. The angle between the shear fracture and the primary mylonite foliation decreases with strain. As shown in figure 2, these may rotate into parallelism with the mylonite foliation and then undergo tensile fracturing. The pulled-apart fracture can be distinguished from shear fractures by the formation of wider infills, often with the central infill of a clast being the most pronounced. The fragments themselves rotate into parallelism with the mylonite foliation and then undergo a further phase of pulling apart. Care must also be taken with pressure shadows; rotated clasts have fringes which have the opposite sense of asymmetry to those arising from non-rotating clasts (cf. Simpson & Schmid 1983). Because inconsistent results can result from local heterogeneities, it is stressed again that many independent indicators from as many zones as possible must be used.

Movement directions can be deduced from cataclasites and gouges, although with greater difficulty. The main drawback is the lack of a lineation. Slickensides may be present, but normally reflect local movements of sliding blocks rather than the overall movement of the fault zone. The main features that can be used are the presence of microshears within the gouges and cataclasites and are illustrated in figure 3. These have the terminologies commonly used

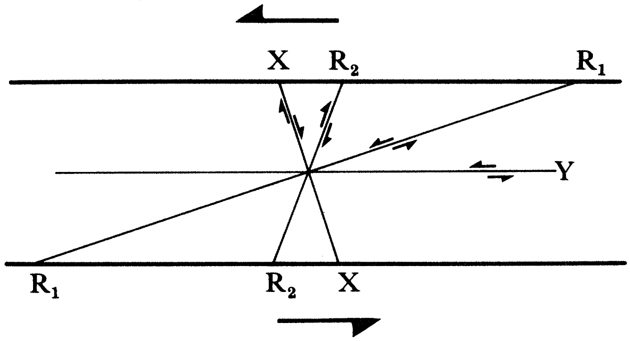

FIGURE 3. Microshear zone arrangements in gouge and cataclastic zones that can be used for deducing the sense of shear. They are marked by zones of more intense communication and also in gouge zones by aligned, shredded, and new phyllosilicates. The terms are from Logan *et al.* (1979).

in the gouge literature (Logan *et al.* 1979, 1981). The equivalents in the mylonitic literature are as follows: R_1-shears = S-bands; X-shears = complementary S-bands; Y-shears = C-bands; R_2-shears = kinks. The main shears encountered (Y and R_1) can be used to deduce the sense of movement in the same way as C- and S-bands. The direction of movement can be deduced from the intersection of R_1- and Y-shears or the intersection of each with rotated country foliations, which tend to be smeared out in the gouge or cataclastic zone. The movement is perpendicular to the intersection. Elongate clasts are also encountered, but give ambiguous results as they can originate, through shearing boundinage and pull-apart fracturing, perpendicular to the movement direction (figure 4). They are not equivalent to the elongate clasts which mark the extension lineation in mylonites.

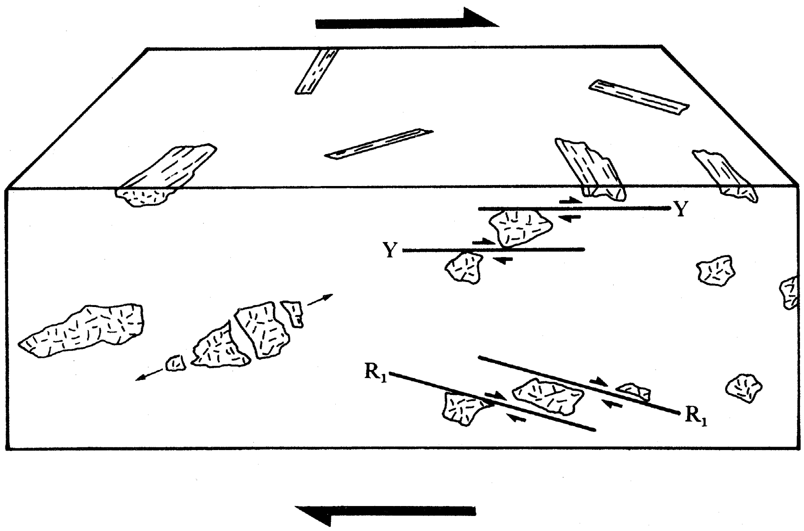

Figure 4. A sketch illustrating how elongate clasts may develop at high strains from hard particles in a gouge zone by stretching and shearing. These initially form perpendicular to the movement direction and may rotate into parallelism with this. They are not equivalent to elongate clasts in mylonites, which often mark the stretching lineation.

4. Reactivation of the Darling Fault, Western Australia

One of the examples of repetitive fault movement noted by Hills (1956*b*) is the Darling Fault in Western Australia. He noted that reactivation has occurred since the Devonian and raised the possibility that there may have been earlier movements.

The Darling Fault (Saint-Smith 1912) is a major continental fault situated along the western margin of the Australian continent (figure 5). The fault separates the Archaean Yilgarn Block from the Mesozoic Perth Basin. The last main period of movement along the fault occurred during early Cretaceous times and was dominated by steep normal displacements to the west. However, detailed field studies (Wilde & Low 1978, 1980; Wilde 1980; Wilde & Walker 1981, 1982) have shown that the fault was localized close to a major Precambrian zone of shearing (see Blight *et al.* 1981), which has been reactivated at several times during the Precambrian (Prider 1952; Wilson 1958). This zone of shearing is transitional to the Darling Mobile Zone (see Mathur & Shaw 1982), now buried beneath the vast accumulations of sediments which

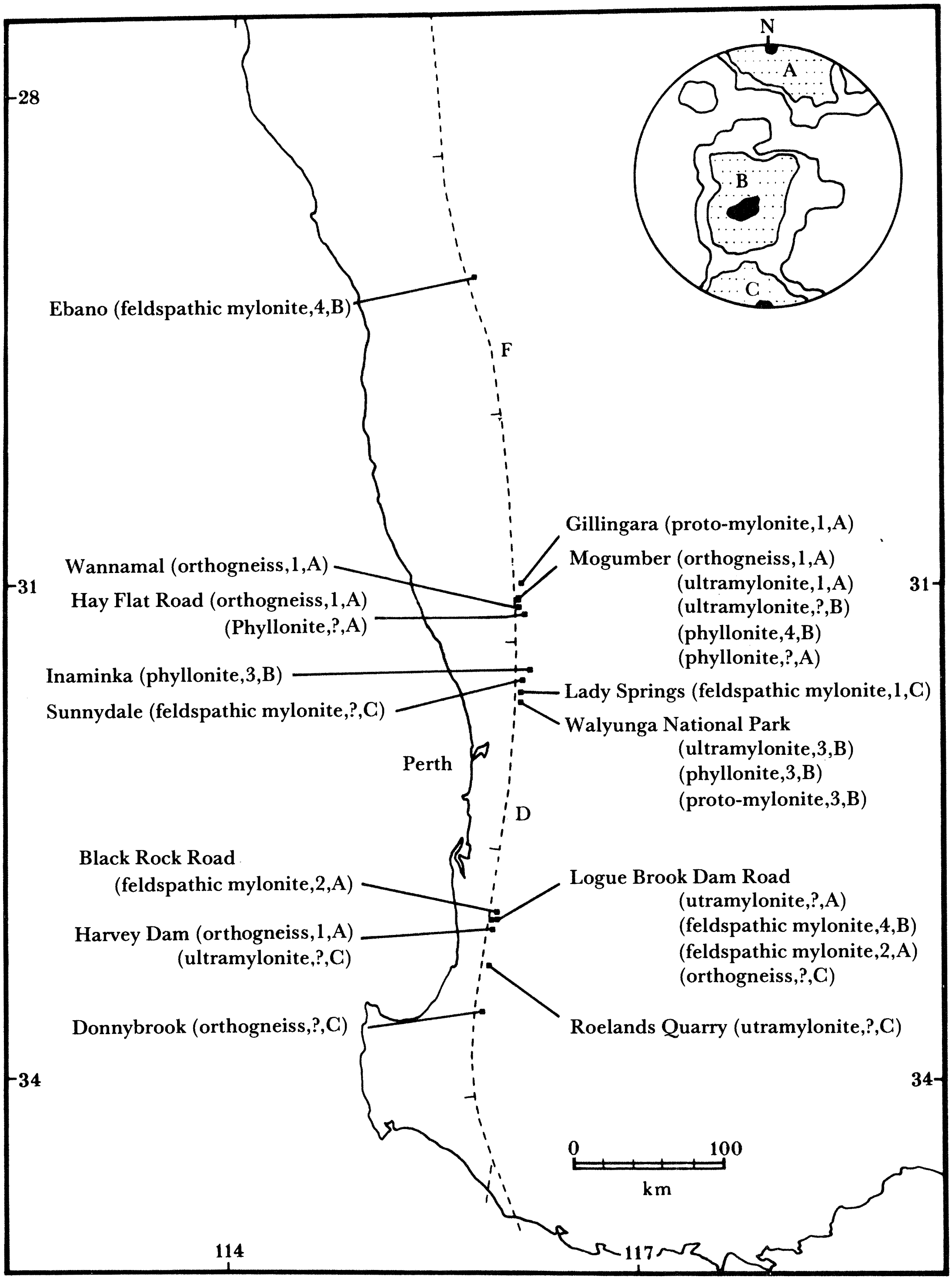

FIGURE 5. Summary of the structural data from mylonites found along the Darling Mobile Zone (DF). The lineation trends are given in the Lambert equal-area projection (top right). They are grouped into three main trends: A, sub-horizontal plunging to the NNE; B, steeply plunging to the W, SW and SE; and C, sub-horizontal plunging to the SSW. Also shown is the sense of shear (1, sinistral strike-slip; 2, dextral strike-slip; 3, normal displacement to the west; and 4, steep reverse movement to the east), for each mylonite zone studied.

comprise the Perth Basin (Bretan 1985). Mylonite zones within this transitional zone are spread over 600 km and are restricted to within 30 km of the Darling Fault (Wilde 1980). The mylonites occur as zones, possess well-developed stretching lineations and contain textural features similar to those illustrated in figure 2 which have been used to establish the sense of movement. The main features used were shear bands, which are particularly well developed within augen-rich mylonites (figure 6), rotated feldspar augens and asymmetric quartz crystallographic fabrics. Two broad groups of mylonites can be identified from these analyses (figure 5). The first group comprises ultramylonites, augen mylonites and some phyllonites, all of which possess sub-horizontal lineations plunging to the NNE and SSW. Except for the phyllonites (which are of low greenschist grade) all these mylonites are of amphibolite–granulite facies. The second group of mylonites comprises mid to low greenschist grade ultramylonites and phyllonites with lineations plunging steeply towards W and SW.

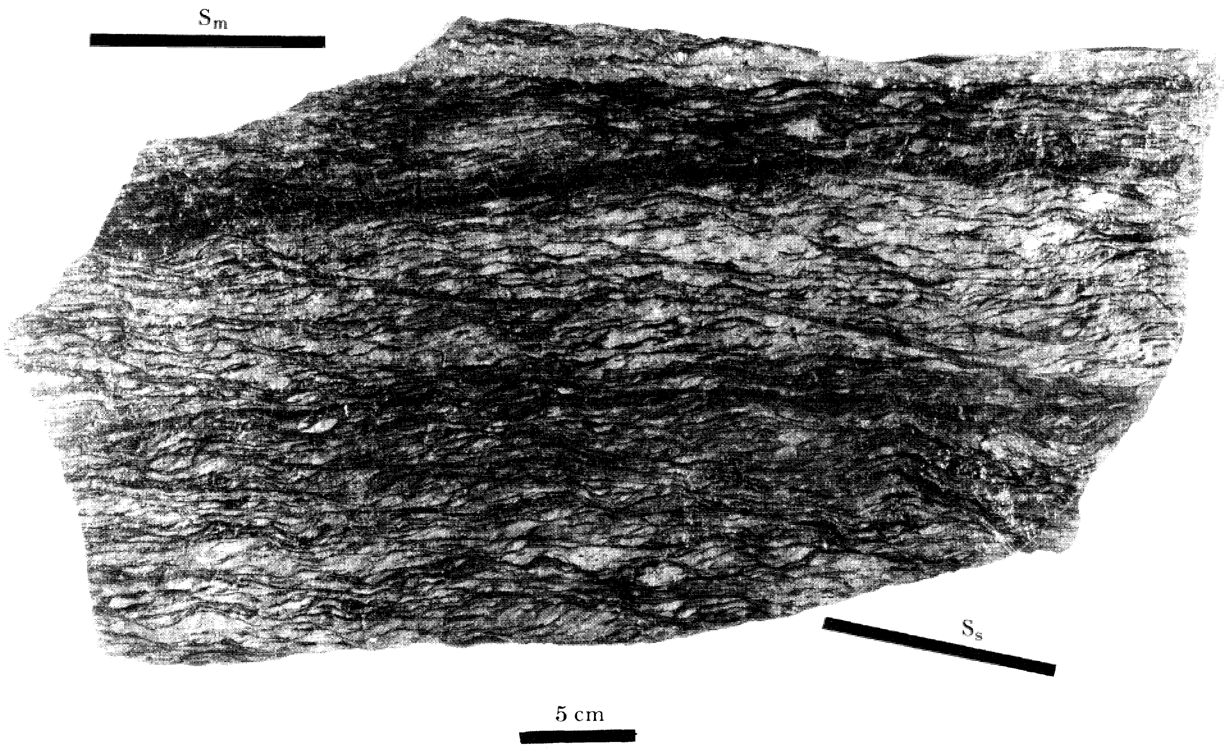

FIGURE 6. Shear bands are common in the augen-rich mylonites along the Darling Mobile Zone and were used extensively to deduce the movement direction. S_m and S_s mark the mylonitic and shear band foliations respectively.

Detailed studies of the movement-sense indicators have revealed that the first group of mylonites are dominated by sinistral strike-slip movements, with evidence in only a few zones for dextral movements. Kinematic indicators in mylonites in the second group reveal steep normal (to the west) and reverse (to the east) dip-slip movements.

Rare cross-cutting relations and geochronological data (see, for example, Blight *et al.* 1981) indicate that the high-grade mylonites with sub-horizontal lineations predate the relatively lower grade mylonites with steeply plunging lineations. The rotation of structure within the early

Proterozoic Albany Province (see Fletcher *et al.* 1983), adjacent to this zone of shearing, indicates that regional sinistral strike-slip movements occurred during the early to mid Proterozoic. The steep displacements may either be associated with the *ca.* 1000 Ma (Wilson *et al.* 1960) or with the 750 to 550 Ma thermal events (Wilson *et al.* 1960; Libby & de Laeter 1979) identified along the western edge of the Yilgarn Block. The timing and regional significance of these events will be discussed in a subsequent paper. So far as this paper is concerned, the important point is that kinematic analyses of the mylonites indicates four phases of movement and that Hills's (1956*b*) surmise that the Darling Fault has a long resurgent tectonic activity is correct.

5. Reactivation

The main reason for reactivation is the stressing of the continental crust. The faults which reactivate will depend upon their orientation with respect to the imposed stress field and their ability to accommodate the resultant strains. Of those correctly orientated, the weakest are expected to be preferred for reactivation.

The work of Hills (*op. cit.*) and O'Driscoll (this symposium) in Australia indicates that the continental crust, especially Precambrian basement, is so fractured that the orientation considerations are easily met (see figure 9 of O'Driscoll). That is, irrespective of the direction of stressing, there is always a fundamental fault, or group of faults, favourably orientated for reactivation. The other geometric consideration is the reactivation of one fault type as another. The geometric similarity between thrusts and normal fault systems suggests that one can as easily reactivate as the other, and as stated earlier, there is now abundant evidence for this. Similarly, White & Green (1985) have shown that the Alpine Fault in New Zealand, which initially had a dominant strike-slip movement has located in an earlier strike-slip zone and then changed into an oblique reverse-slip fault without creating a new fault trace. But can a strike-slip zone, which is usually vertical, reactivate as an inclined normal or a reverse fault? This obviously can occur because it is seen in the Darling Mobile Zone (see previous section and Bretan 1985). The answer lies in the internal structure of fault zones. Few consist of a single strand. Most consist of several anastomosing strands which give an effective width to a zone of tens of kilometres. Good examples are the Proterozoic mobile zones. The Darling Mobile Zone is at least 30 km in width, while the Halls Creek Mobile Zone in NW Australia is *ca.* 50 km in width (Dow & Gemuts (1967). Proterozoic Mobile Zones in Southern Africa are of similar widths (see Hunter 1981). The modern San Andreas Zone in California and parts of the Alpine Fault Zone in New Zealand are also *ca.* 50 km in width (Scholz 1977). The reactivation of selected strands within a given zone will result in an inclined fault during reactivation. Consequently, we see no geometric reason to prevent a reverse or normal fault developing from a strike-slip zone and vice versa.

6. Softening processes associated with the development of a mylonite zone

An indication of the potential weakness of a fault zone can be gained from a consideration of the softening processes that lead to the concentration of deformation in a zone during a single deformation associated with fault-zone initiation and development. Several softening processes

(White *et al.* 1980; Poirier 1980) have been postulated as being important for concentrating deformation in mylonite zones. These are:

(*a*) change in deformation mechanism because of grain refinement;
(*b*) reaction softening;
(*c*) geometrical or fabric softening;
(*d*) continual recrystallization;
(*e*) chemical softening;
(*f*) pore fluid effects;
(*g*) strain heating.

The first four are the most important. The degree to which deformation is concentrated is dependent upon the softening relative to the country rock and this will now be discussed.

(*a*) *Softening associated with a change in deformation mechanism owing to grain-size reduction*

As stated above, Watterson (1975) was the first to associate the mechanical weakness of mylonite zones with a fine grain size. But, as noted by White (1976), this is not universally true because the opposite can occur if the Hall–Petch relation applies:

$$\sigma = \sigma_i + kd^{-0.5}, \tag{1}$$

where σ is the flow stress, σ_i the internal frictional stress, k a constant and d is the grain size.

A change in deformation mechanism to one in which the strain rate is inversely dependent on grain size and which is dominated by grain-boundary sliding is required. There is good experimental evidence for the degree of softening that can be associated with a change in grain size of limestone (Schmid *et al.* 1977). In their experiments, the flow law changed from dislocation creep, obeying

$$\lg \dot{\epsilon} = -1.33 - \frac{71\,100}{RT\,2.3} + 4.7 \lg \sigma, \tag{2}$$

to grain-size-dependent creep, obeying

$$\lg \dot{\epsilon} = 4.98 - 3 \lg d - \frac{59\,000}{RT\,2.3} + 1.66 \lg \sigma, \tag{3}$$

where $\dot{\epsilon}$ is the strain rate (in reciprocal seconds), σ the differential stress (bars†) and d is the grain size (micrometres).

The degree of softening is summarized in figure 7, which is based upon the flow laws (2) and (3). A similar curve has been derived for quartz from theoretical considerations (White 1976) but no experimental data are available in spite of attempts to provide it by Kronenberg & Tullis (1984). The implication of the above marble experiments is that grain-size reduction could, in circumstances favouring grain-boundary sliding, produce a marked softening effect.

(*b*) *Reaction softening*

Most mylonites in major fault zones are characterized by a retrogressed mineral assemblage. An extreme, but common, case is the change of a granite or granitic gneiss into a phyllonite containing mostly white mica and chlorite with variable amounts of quartz. The stages in this process have been illustrated for the Lewisian gneiss associated with the Moine Thrust in NW

† 1 bar = 10^5 Pa.

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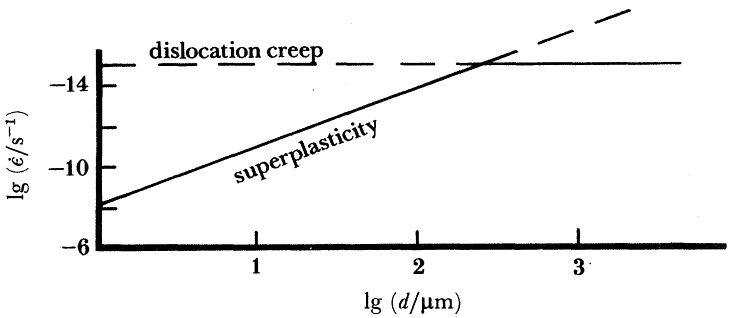

FIGURE 7. The strain softening, as evidenced by an increasing strain rate, accompanies the change from dislocation creep to superplastic behaviour as grain size decreases in limestones. The figure is based on data from Schmid *et al.* (1977). The plots assume a stress of 10 mPa at 400 °C.

Scotland (White *et al.* 1982). The first and last stages are illustrated in figure 8, plate 1. An indication of the strength changes as a result of phyllonitization should be obtained by comparing the strength of granite with that of slate, which is mineralogically and microstructurally similar to a phyllonite. However, while there are good rheological data for granite, there are few for slate- or mica-rich aggregates. There is no flow law available and consequently it is not possible directly to compare relative strengths under crustal conditions where phyllonites are the deforming medium. However, it is possible to compare strengths at specific conditions. Results for mica aggregates (Etheridge *et al.* 1974) and granite (Carter *et al.* 1981) indicate that the change from granite to phyllonite should lead to a permanent strength decrease by a factor of *ca.* 10.

This represents a gross effect of retrogression. There are more subtle effects, which include hydrolitic weakening of quartz, grain-boundary hydration, and the transient production of fine-grained products at the initial stages of reaction.

The relative weakening between quartz in the country rock and in mylonites is difficult to judge. Only trace amounts of water are required for hydrolitic weakening and there is unlikely to be a differential effect in a wet crust. This is more likely to occur in the deep crust if it is dry and if fluids are limited to fault zones. Weakening due to grain-boundary hydration is also likely to occur in the deeper crust. The result is to form a thin, disordered film in quartz or a retrogressed layer in anhydrous minerals such as plagioclase (White & White 1981), effectively widening the grain boundary, thereby aiding grain-boundary diffusion, and mechanically weakening the boundary. The strain rate for Coble creep and superplastic flow is proportional to grain-boundary width and the mechanical weakening of this has been shown in experiments and can be appreciated from figure 9 (from Kronenberg & Tullis 1984).

The production of fine-grained materials at the start of a reaction also causes appreciable weakening during the prograde reaction of serpentine to olivine (Rutter & Brodie 1985). A similar effect is likely during retrogression but the effect will be lost once coarsening occurs.

(*c*) *Fabric softening and continual recrystallization*

These two softening processes are considered together because when continual recrystallization occurs a preferred crystallographic orientation of the grains normally develops. The effect of the preferred orientation can be calculated through the Taylor factor (see Reid 1973), but in mylonites the situation is much more complex because of the accompanying recrystallization

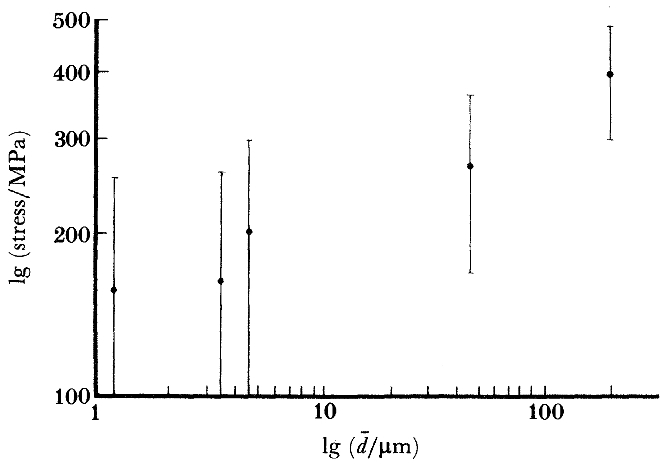

FIGURE 9. Softening associated with hydration effects along grain boundaries in quartzites (based on fig. 11 of Kronenberg & Tullis 1984). Strength decreases at the volume fraction of the hydrated grain boundary increases. Experimental conditions are listed in Kronenberg & Tullis.

and grain-size reduction. A further complication arises because the fabrics, especially crystallographic preferred orientations, are usually domainal (Celma 1982) and can vary throughout the mylonite and also during the course of deformation. Again there is need for experimental data but none are available for crystallographic preferred orientation development. A geometric softening effect will develop as the foliation is rotated into parallelism with the fault-zone margin in accordance with Schmid's law (Dieter 1961). Within mylonites an additional effect can result from compositional banding especially if bands of phyllosilicates form.

An idea of the degree of softening likely to be associated with the combined effects of crystallographic orientation development, grain-size reduction and continual recrystallization can be obtained from analogue experiments.

Figure 10 shows the results of using polycrystalline magnesium as an analogue for quartzite. The samples were deformed so that a shear zone developed, along which formed fine-grained recrystallized magnesium with the grains aligned for easy basal slip. The microstructure of the magnesium in the shear zones is identical to that of a quartz mylonite. The experiments are described in detail in the literature (Ion *et al.* 1982; Drury *et al.* 1985; White *et al.* 1985) and only the relevant results will be discussed here. There is a strength drop after shear-zone development, but it is temperature dependent (figure 10). The higher the temperature, the less marked the strength drop, until at temperatures above 0.65 T_m no shear zone developed and no marked strength drop was registered. An interesting point to note in the above is the slight effect that temperature has on the strength of the samples at large strains after shear-zone development. Also associated with fabric softening in mylonites is the preferential development of phyllo-silicate grains which have their boundaries parallel to the mylonitic foliation and are oriented for easy grain-boundary sliding. Oriented grain boundaries may also occur between quartz grains.

From the above, it is evident that mylonites can be softer than the country rock from which they were derived. Just how much weaker they are is not known quantitatively because they have not been the subject of intensive experimentation. Initial results of such a programme

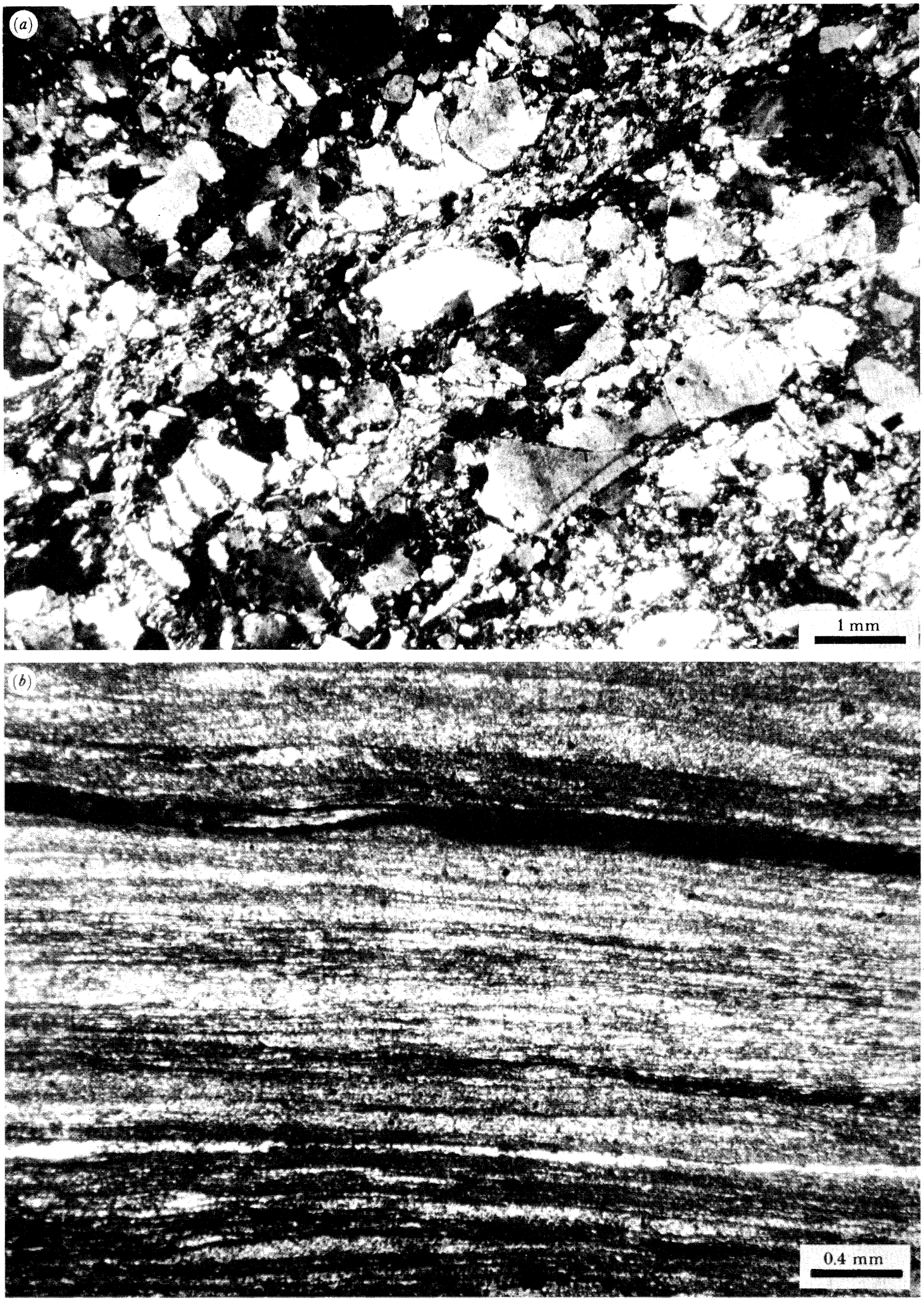

FIGURE 8. Microstructural changes associated with mylonite development in the Lewisian Gneiss along the Moine Thrust Zone, NW Scotland. Part A is the initial gneiss and part B is the quartz phyllosilicate mylonite (phyllonite) arising from A at higher strains. (Bar scales: A = 1 mm, B = 0.4 mm.)

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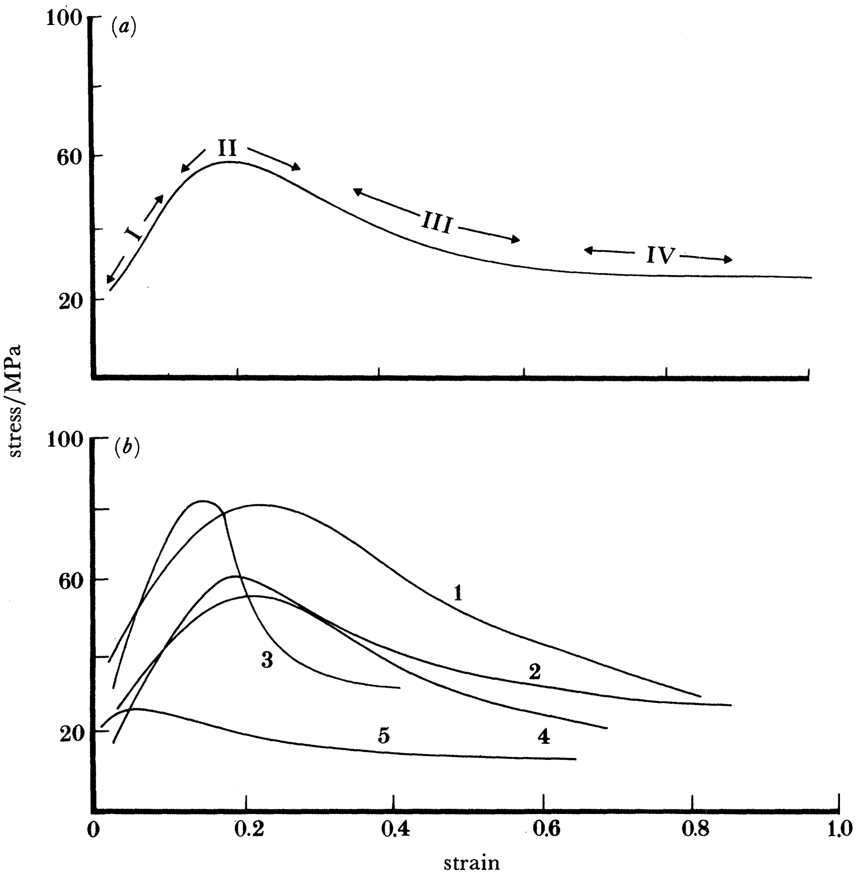

FIGURE 10. Softening associated with shear-zone development in polycrystalline magnesium which has been used as a quartz analogue (based on figures 1 and 2 in White *et al.* 1985). (*a*) A representative stress–strain curve showing the relations between the softening and shear-zone development: I, initial hardening; II, recrystallization dominant; III, shear-zone development; IV, steady-state microstructure and constant shear-zone width, with grains in the shear zone having a strong crystallographic preferred orientation. (*b*) Stress–strain curves for large strain compression tests of polycrystalline magnesium. Solid lines deformed at (1) 270 °C and $\dot{\epsilon} = 10^{-4}\ s^{-1}$, grain size 270 μm; and (2) at 400 °C. These show the Hall–Petch relation for peak stress. Other lines, with strain rate $10^{-5}\ s^{-1}$ and grain size 400 μm; (3) 150 °C–0.46 T_m; (4) 260 °C–0.58 T_m; and (5) 370 °C–0.7 T_m. No shear zone formed in (5) where softening was associated with pervasive recrystallization.

are encouraging. Experiments on ultrafine-grained quartz mylonites from the Redbank Deformed Zone (see Obee & White, this symposium) show a marked softening at slow strain rates approaching those that are likely to be representative of fault movements (figure 11). This is similar to those formed in gouges and is also thought to represent a change in deformation mechanism (Rutter & White 1979).

The softening effects to be found in cataclasites and gouges are a more controversial issue. Experimental data indicate that they are capable of maintaining high shear stresses (Logan *et al.* 1981), especially if free of montmorillonite (Morrow *et al.* 1982). There appears to be no reason for faults to concentrate in gouge zones other than the high pore-field pressure (Logan *et al.* 1981) or siting controlled by fault behaviour at greater depths.

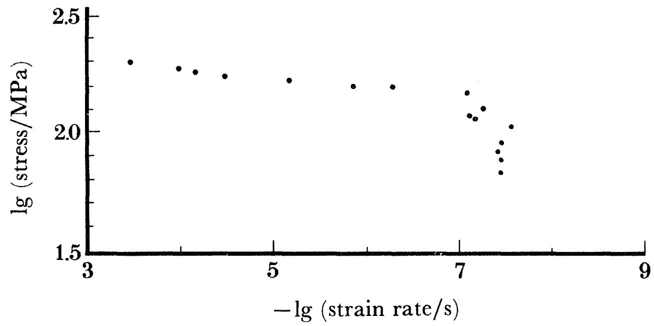

FIGURE 11. Softening associated with a change in deformation mechanism in a fine-grained quartz mylonite at slow strain rates. Stress relaxation test at 700 °C, stress = 200 MPa and P_{H_2O} = 50 MPa. The marked softening at strains less than 10^{-7} s^{-1} is consistent with a change in deformation mechanism to one that is diffusion or grain-size dependent or both (see Rutter & White 1979).

7. Softening and reactivation

The above processes strictly apply only after a fault has reactivated and are dynamic softening processes. Static softening will affect reactivation and consequently not all of the above will be important. However, the fact that faults do reactivate means that some aspects of softening accompanying earlier movements must influence the siting of reactivation. Two obvious ones appear from the work of Obee & White (this symposium); the first being a well-developed foliation and the second a marked retrogression. The former is an example of geometric softening and the second of reaction softening. Not all aspects of geometric softening during a former movement episode will provide weakness during the initiation of reactivation. If a change of slip vector is involved, all linear elements within the zone must be reoriented. Studies of crystallographic preferred orientation in quartz mylonites have revealed that the dominant slip plane and slip direction are oriented to accommodate slip in the foliation plane and in the approximate direction of the stretching lineation. Unless the new and old slip vectors are coincident, these grains are misoriented for easy slip at the initiation of reactivation and may cause local hardening and stress notches until they are. Johnson & White (1983) concluded that this occurred during a change in slip vector across the Alpine Fault in New Zealand.

Grain-size softening is a dynamic effect and is unlikely to contribute, other than locally, to the initiation of reactivation. The grain size for the advent of superplasticity can be seen, from (3), to be fixed in strain rate, temperature and stress space, and unless they are identical during the initiation of reactivation to those at the time of the mylonite development, it is likely that the Hall–Petch relation will apply. The result, in this case and also for zones of incorrectly oriented grains, is for the fault to seek out the weaker strands during reactivation, leaving the harder zones protected from reactivation. The preservation of older, fine-grained mylonite zones is found along the Darling Mobile Zone and within the Redbank Deformed Zone (see Obee & White, this symposium).

8. Conclusions

The continental crust is criss-crossed by fundamental fault zones which contain along them fault rocks that may constitute planes of weakness. These rocks, especially mylonites, are the 'glue' that holds the cracked and fractured continental crust together. When the crust is restressed, those faults that are correctly oriented to accommodate the imposed strains and which contain the weakest mylonites act as preferential loci for reactivation. Not all mylonites are necessarily weak, and these can be preserved in a reactivated zone and give an insight into earlier movement events.

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PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A

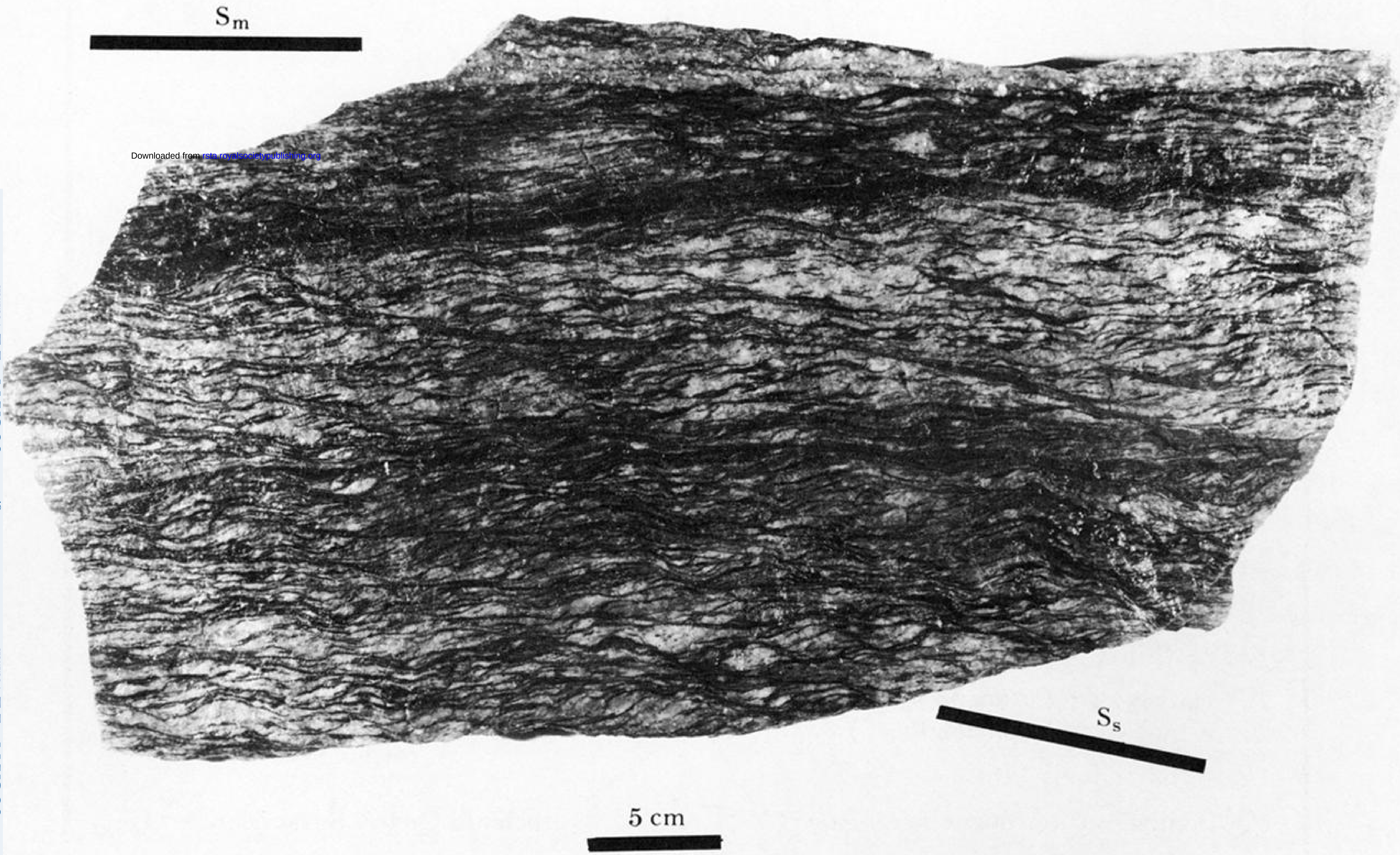

FIGURE 6. Shear bands are common in the augen-rich mylonites along the Darling Mobile Zone and were used extensively to deduce the movement direction. S_m and S_s mark the mylonitic and shear band foliations respectively.

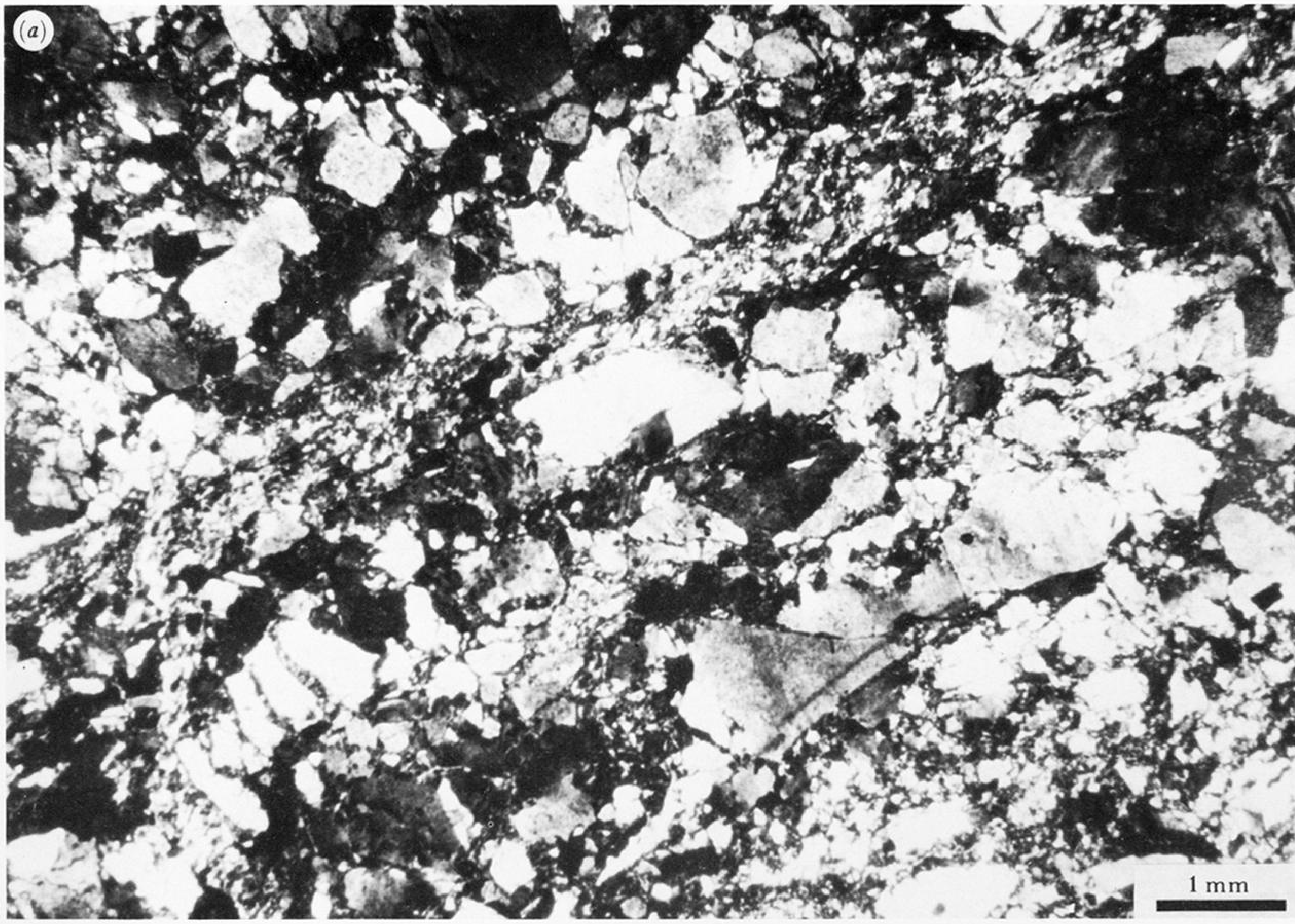

FIGURE 8. Microstructural changes associated with mylonite development in the Lewisian Gneiss along the Moine Thrust Zone, NW Scotland. Part A is the initial gneiss and part B is the quartz phyllosilicate mylonite (phyllonite) arising from A at higher strains. (Bar scales: A = 1 mm, B = 0.4 mm.)

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