Geoscience Canada

Flow Mechanisms in Rocks:
Microscopic and mesoscopic structures, and their relation to physical conditions of deformation in the crust

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Volume 5, Number 3, September 1978

URI: https://id.erudit.org/iderudit/geocan5_3art01

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Publisher(s)

The Geological Association of Canada

ISSN

0315-0941 (print)
1911-4850 (digital)

Explore this journal

Cite this article

Kerrich, R. & Allison, I. (1978). Flow Mechanisms in Rocks:: Microscopic and mesoscopic structures, and their relation to physical conditions of deformation in the crust. *Geoscience Canada*, *5*(3), 109–118.

Article abstract

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Brittle deformation with grain fracture leading to a reduction of particle diameter is characteristic of cataclastic flow.

Pressure solution produces grain shape fabrics by intercrystalline diffusion assisted by the presence of water. Grains may change shape at constant mass, or decrease in mass (and therefore in size) by long range diffusion: mass is then not locally conserved. Reduction of grain diameter leads to increased rates of deformation (strain softening).

Distinctive spaced cleavage zones form by pressure solution in which mineral species are redistributed due to different rates of deformation: the displacement field is discontinuous and deformation non-isochemical. Tectonic veins associated with pressure solution structures probably form by local mass transport; thus brittle and ductile mechanical behaviour coexist.

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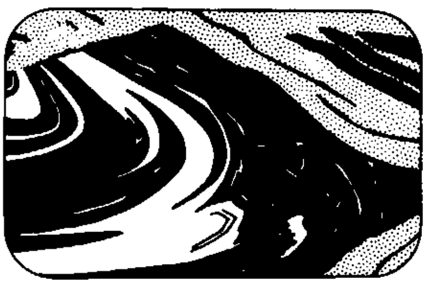
Articles



Flow Mechanisms in Rocks:

Microscopic and mesoscopic structures, and their relation to physical conditions of deformation in the crust

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Summary
Deformation of the crust is believed to occur dominantly by cataclasis at low temperatures and/or effective confining pressure, by pressure solution at intermediate temperatures, and by dislocation creep at high temperatures. Each flow mechanism gives rise to distinctive microscopic and small scale structures.

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Dislocation creep produces grain shape fabrics by intracrystalline deformation, and may cause grain size reduction by subgrain formation and recrystallization. Preferred crystallographic orientations can arise from dislocation glide. Mass is conserved and deformation is believed to be essentially isochemical. Small scale structures formed by dislocation creep are ductile, with a continuous displacement field.

Introduction

Recent advances in materials science have led to a better understanding of the solid state flow mechanisms in crystalline materials. These concepts have been applied to deformation of rocks, with particular reference to flow in the mantle (Ashby and Verrall, 1977; Weertman, 1978).

The objective of this paper is to present a brief synthesis of the flow mechanisms believed to be important in crustal deformation. Microscopic and small scale structures typical of each mechanism are illustrated, and placed in the context of the physical conditions prevailing during deformation. A comprehensive treatment of flow mechanisms is not attempted: the appropriate works are referred to. We emphasise, and present observational evidence for, the relation between the physical environment of crustal deformation and the nature of the operative flow mechanism.

Flow Mechanisms

Crystalline aggregates may deform by a number of independent mechanisms. These are: 1) *Cataclasis*, involving repeated grain fracture and frictional sliding of particles; 2) *Dislocation creep*, in which strain is largely produced by the gliding motion of dislocations; 3) *Boundary diffusion creep*, involving mass transport at grain boundaries (generally known as Coble creep); 4) *Volume diffusion creep*, which takes place by means of the diffusion of point defects through grains (generally termed Nabarro-Herring creep). These flow mechanisms and the factors which influence them are reviewed by Ashby, (1972); Stocker and Ashby (1973); and Ashby and Verrall (1977). Other mechanisms of crystal plasticity such as kinking and twinning (see Gilman, 1969) do not in general contribute large strains.

Pressure solution (Sorby, 1863) is a term used by geologists to describe a process of diffusive mass transport by which rocks deform at low temperature, probably assisted by the presence of fluid at grain boundaries; and which therefore has similarities to mechanism (3) above. Nabarro-Herring creep has been identified in the deformation of fine grained metals at high values of homologous temperature ($T/T_{melting}$). This creep mechanism is not thought to be significant in deformation of the crust or mantle, and is not considered further here.

Each mechanism has a constitutive flow law which relates strain rate to rheological properties (such as defect diffusivities), and the state variables temperature and differential stress. The ensuing discussion is concerned principally with the deformation of quartz and calcite because the mechanical behaviour of these species is better understood than that of other minerals.

Cataclasis

Cataclastic flow involves repeated grain fracture and frictional sliding of rock particles: both of these processes are dependent on the magnitude of normal stress (Stearns, 1968; Hamil and Sriruang, 1976).

At greater crustal depths progressively higher values of confining pressure inhibit cataclasis by increasing the normal stres across potential fracture and sliding surfaces, thereby favouring mechanisms of crystal plasticity.

The effect of pore fluids is to reduce the total confining pressure by an amount equal to the fluid pressure, yielding an 'effective' confining pressure ($P_{effective} = P_{total} - P_{fluid}$). Under conditions of normal fluid pressure ($P_{fluid} \sim \frac{1}{3} P_{lithostatic}$) this transition from brittle to ductile behaviour takes place at a depth of a few kilometres in the earth's crust. In the presence of abnormal fluid pressure where P fluid may approach or exceed the confining pressure brittle mechanical behaviour may extend to significantly greater crustal depths. (Price, 1970; Carter, 1976).

In proximity to the terrestial surface, where conditions of confining pressure and temperature are both low, deformation of soft sediments is accommodated by frictional sliding of rock particles. This is illustrated for folded sediments and buckled anisotropic tuff in Figures 1A and 1B respectively. The microscopic structure of these rocks exhibits a low degree of grain fracture. Soft sediment deformation is discussed by Hobbs *et al.* (1976) and Woodcock (1976).

At higher values of confining pressure, and in consolidated rocks, deformation occurs by means of repeated grain fracture, coupled with frictional sliding and rotation of rock particles. These processes are best developed along fault zones, in which repeated fracturing leads to grain size reduction and generation of a 'fault gouge' (Engelder, 1974; Fig. 1C).

A special case of brittle deformation occurs in 'dry' consolidated rocks at high confining pressure, subjected to displacement at fast strain rates. Under these conditions fracture and frictional sliding at high velocity lead to temperature rise and generation of a melt phase known as pseudotachylyte (Fig. 1D). This feature has been discussed by Sibson (1975), and is believed to represent a locus of palaeo-seismic faulting. The influence of strain rate on deformation is discussed by Price (1975) and White (1975).

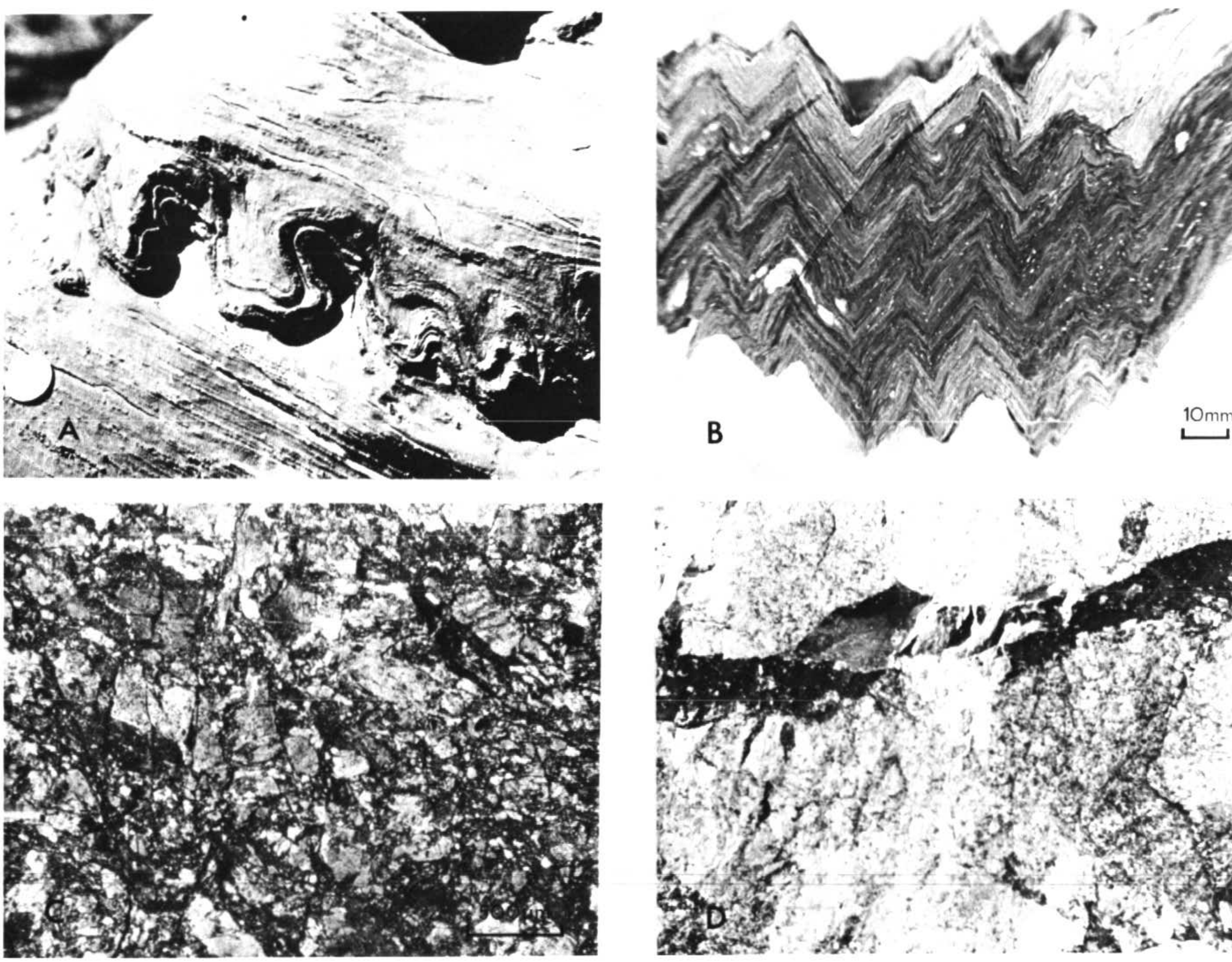

Figure 1

A: Folded layers in sea floor gravity slump. B: Crenulation folds formed by buckling of anisotropic sediment in a sea floor slump. C: Grain size reduction by fracture, along a fault. D: Pseudotachylyte vein in granite: rapid frictional sliding along a shear plane leads to melting.

Pressure Solution

Pressure solution is a deformation mechanism involving intercrystalline diffusive mass transport in response to applied stress. Mass transport is directed from grain boundaries at high normal stress to grain boundaries at low normal stress, permitting the shape change of grains. For the deformation of polycrystalline materials mass transport may occur by means of the surface diffusion of point defects (Coble creep, see Elliott, 1972), by diffusion through an adsorbed fluid layer (Weyl, 1959; Rutter, 1976), or by diffusion through a static pore solution. Rutter (1976) has emphasised that the presence of water at grain boundaries probably yields effective diffusivities in the aqueous phase at low temperatures (< 400°C) of the order of solid-state diffusivities at much higher temperatures, where boundary diffusion creep is observed to operate in the deformation of 'dry' metals and ceramics. This assumption is reasonable inasmuch as pressure solution is encountered principally in sediments which have undergone diagenesis, and in rocks deformed under conditions of low grade metamorphism. In these low temperature environments rocks generally have a relatively high porosity and high content of hydrated mineral phases. The subject of pressure solution has been reviewed by Kerrich (1978).

Grains change shape by removal of material along boundaries exposed to high normal stress, with diffusive transport to boundaries at lower normal stress and epitaxial overgrowth. In this state the mass of individual grains is approximately conserved (Fig. 2A). Alternatively, groups of grains under high normal stress may lose mass to groups of grains at lower normal stress, by long-range diffusion; or precipitate in structures such as pressure shadows (Fig. 2B). In this situation mass is not locally conserved.

It is possible that a preferred alignment of optic axes may develop by pressure solution due to differences in the crystallographically controlled rates of growth and solution. An attempt to detect this effect in a rock with textural evidence for dominant pressure solution met with a negative result (Allison and Kerrich, unpublished data).

Penetration of one grain into another by removal of mass along grain boundaries is illustrated by the deformed orthoquartzite in Figure 2C. The absence of intracrystalline deformation is indicated by the presence of rectilinear arrays of secondary fluid inclusions within grains. An example of mass removal at pebble contacts by pressure solution in a deformed conglomerate is given in Figure 2D.

The presence of layer silicates appears to enhance the rate of deformation of quartz and calcite relative to monomineralic rocks. Thus features of pressure solution are best developed in impure lithologies.

There is a large variation in the rate at which different minerals deform by pressure solution. For instance quartz and calcite deform faster under given conditions than phyllosilicates or metal oxides. Because of this differential response to pressure solution one mineral may be selectively removed from domains in a polymineralic rock. This results in an increase in the relative proportion of less mobile minerals, and permits immobile platy minerals to rotate into a parallel alignment. Deformation is then locally non-isochemical.

This effect typically develops along a series of subplanar zones which form a spaced pressure solution cleavage, or along tectonic stylolites in rocks with a high abundance of carbonate minerals. Structures formed by this process are illustrated in Figures 2E, 2F and 3A. Planar markers oriented obliquely with respect to the zones of pressure solution appear to be displaced because of volume loss (Fig. 2F), and this feature has been cited as evidence for the new contra-indicated shear hypothesis of cleavage formation.

Separation of minerals with an initially homogeneous distribution by pressure solution is a typical feature of buckled mineral fabrics. The separation process involves diffusion of a mobile phase from limbs into hinges in response to stress gradient, with concomitant enrichment in the limbs of layer silicates and other immobile phases (Williams, 1972; Stephansson, 1974). This is the typical crenulation cleavage structure, examples of which are given in Figures 3B and 3C. The displacement field in these structures is not continuous. Under certain conditions which are not well understood both quartz and micas are transported at approximately the same rate and may form quartz-mica "beards" at grain boundaries under low normal stress (see Fig. 3D).

Pressure solution zones are initiated at the boundaries of rigid particles which act as stress raisers in the adjacent matrix. Examples of these are pyrite crystals, fold hinges, pebbles and fossils in a fine grained matrix. The pressure solution zones are believed to propagate normal to the inferred direction of maximum principal stress.

Zones of pressure solution are frequently associated with tectonic vein arrays, and exhibit systematic geometrical relations to the veins. Tensile fractures initiate in a direction parallel to the maximum principal stress; pressure solution zones propagate normal to this direction. Such veins are generally composed of minerals that reflect the composition of the host rock (Figs. 3E, 3F). This observation, coupled with the structural relations, has led to the conclusion that the vein minerals are derived largely by diffusive transport from the adjacent rock (Durney and Ramsay, 1974). Control of vein composition by the host rock is shown in Figure 3F where quartz is precipitated in a fracture adjacent to the felsic part of a graded unit, and chlorite adjacent to pelite.

Tectonic veins are generally abundant in rocks deformed under conditions of low grade metamorphism where pressure solution is operative. The mechanism of mass transport into veins by pressure solution avoids difficulties inherent in the alternative hypothesis of precipitation from hydrothermal systems where the water/rock ratio is constrained to be high by mineral solubilities.

Large crustal volumes may undergo non-isochemical deformation by pressure solution if the mobile phase diffuses into sinks such as veins (Kerrich *et al.*, 1978).

The constitutive flow law for pressure solution given by Elliott (1972) and Rutter (1976) assumes diffusion to be rate controlling and is essentially similar to that for Coble Creep (Ashby, 1972), with a linear dependence of strain rate on differential stress and an inverse cube dependence on grain diameter. This accounts for the fact that pressure solution is favoured in fine grained lithologies; and also predicts that grain size reduction by mass loss leads to increased rates of deformation (strain softening).

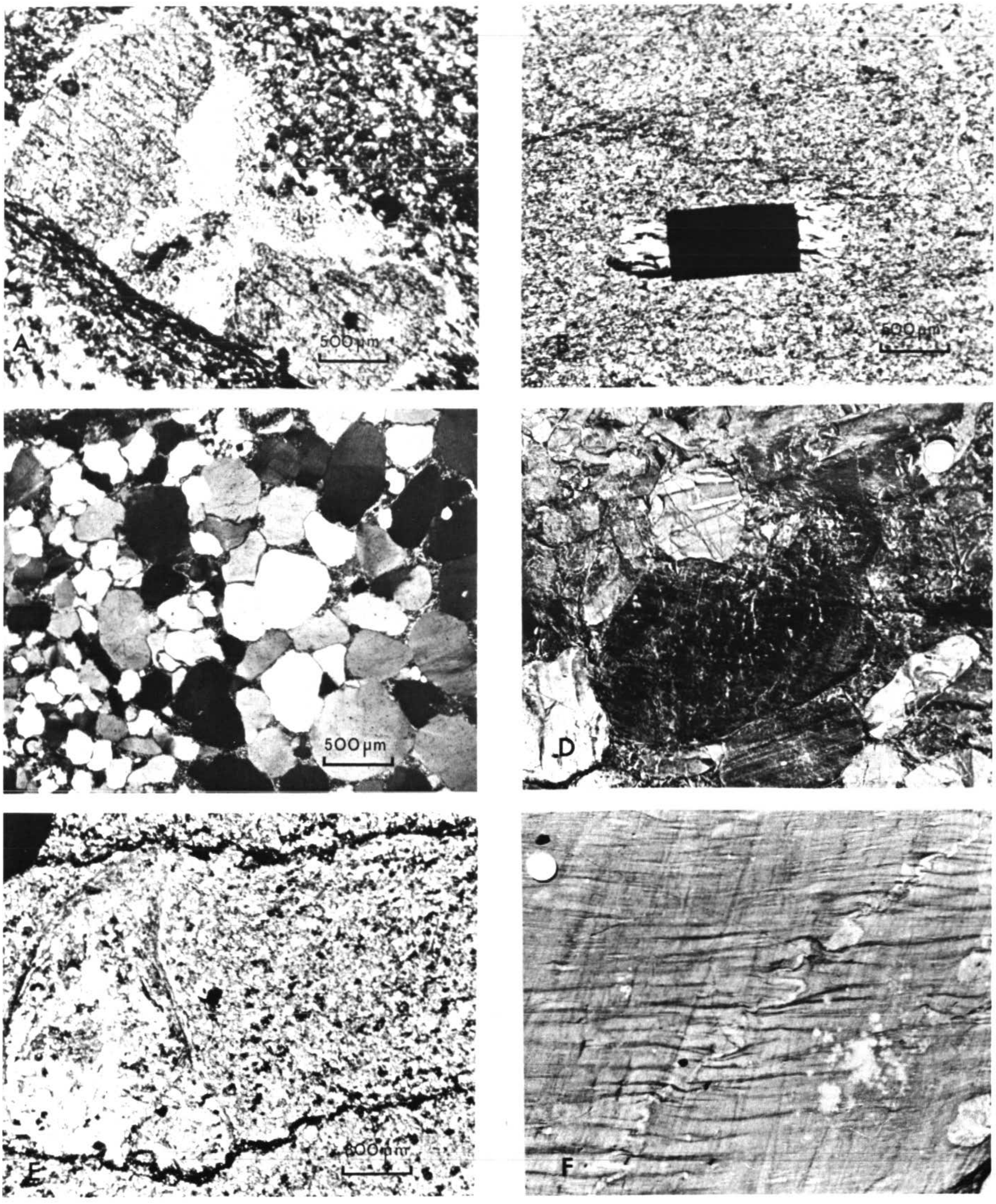

Figure 2
A: Mass loss of calcite in a tectonic stylolite with epitaxial overgrowth of pure calcite in pressure shadows. B: Redistribution of quartz around a rigid particle from boundaries exposed to high normal stress to boundaries at lower normal stress (pressure shadows). C: Mass loss at indented grain boundaries in a deformed orthoquartzite. D: Mass loss at indented boundaries in a deformed conglomerate. E: Tectonic stylolites transect a fossil and define spaced cleavage. F: Spaced pressure solution cleavage zones propagating from a buckled quartz vein. Volume change indicated by displacement of vein across cleavage zones.

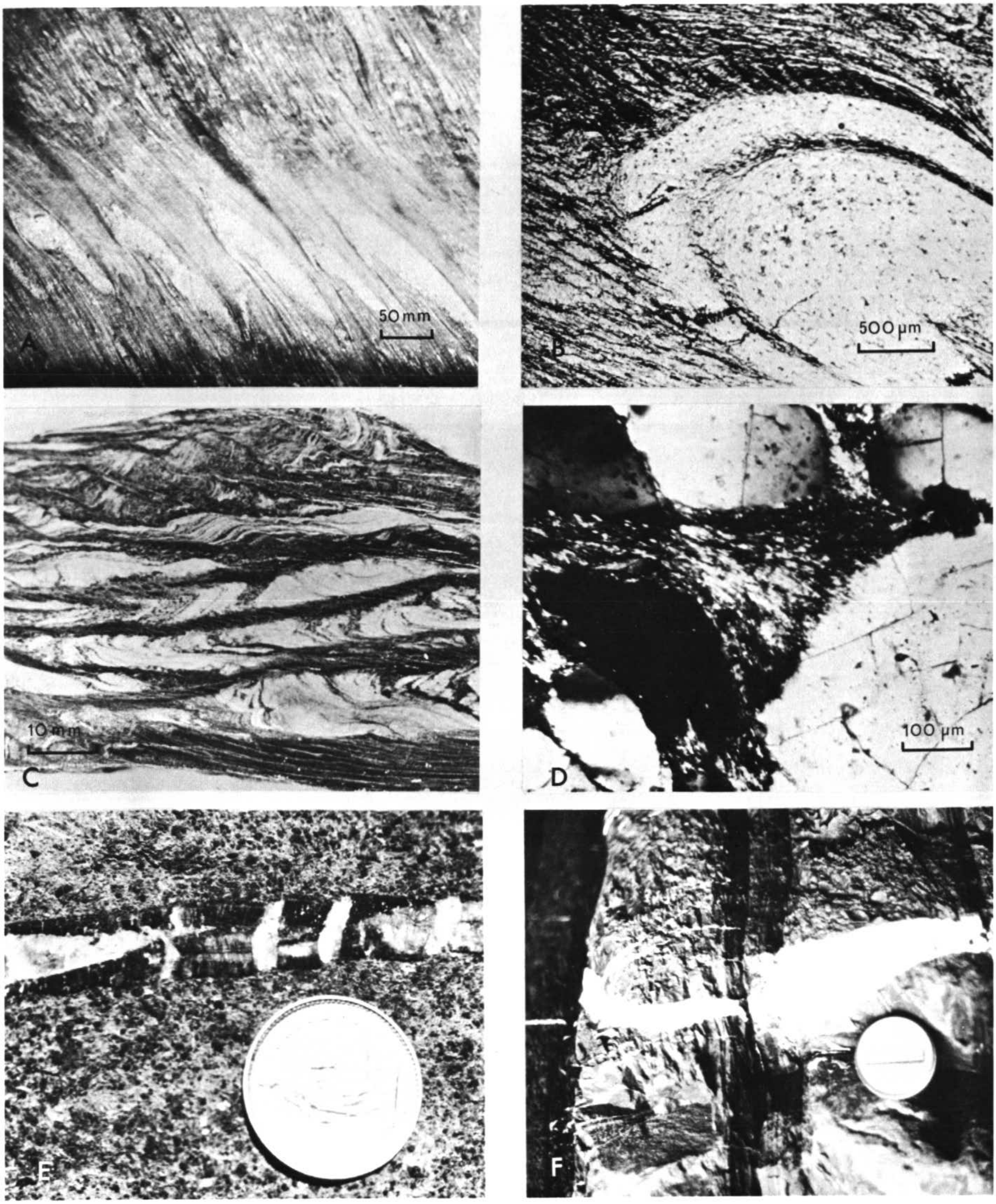

Figure 3
A: Folded bedding transposed into tectonic striping by pressure solution with removal of inverted limbs. B: Spaced pressure solution cleavage in the buckled matrix around a microfold. C: Crenulation cleavage formed by redistribution of quartz and micas by pressure solution in a buckled phyllite multilayer. D: Quartz and mica overgrowth in a domain of low stress. E: Boundinaged belemnite with migration of calcite from deformed matrix into pressure shadows. F: Deformed greywacke displaying local redistribution of quartz in grit layers and chlorite in shale.

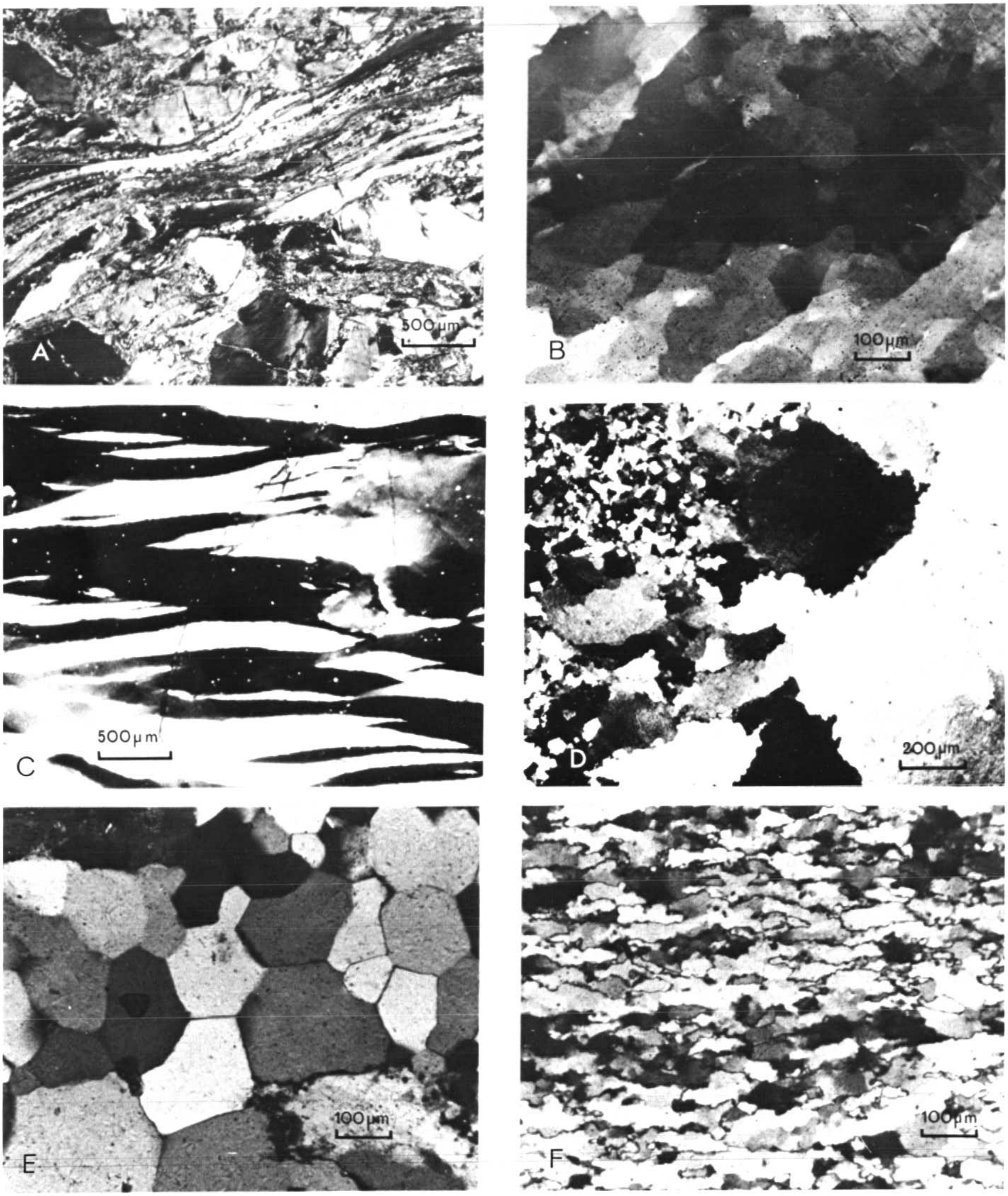

Figure 4.

A: Ribbon quartz produced by dislocation glide. Slight recrystallisation is evident on some quartz grain boundaries. B: Subgrains in quartz: the slight angular misorientation between subgrains is revealed by differences in interference colour. C: Deformation bands in quartz. D: Tectonic grain size reduction of mm size quartz grain into a mosaic of subgrains. Note the serrated grain boundaries. E: Equant polygonal quartz grains with equilibrium triple junctions (foam microstructure). Stable end point of annealing recrystallisation. F: Quartz mylonite. Irregular elongate grains with subgrains and undulatory extinction, pressured to have formed by dynamic recrystallisation.

Superplastic Flow. Superplastic flow is a term used to denote the phenomenon by which polycrystalline materials attain very high ductile strains. Several independent mechanisms have been implicated in superplastic flow, notably grain boundary sliding accommodated by diffusive transport of dislocation creep. Superplastic behaviour in metals and ceramics has been reviewed by Edington *et al.* (1976). Ashby and Verrall (1973) presented a model of superplastic flow in which essentially frictionless grain boundary sliding is accommodated by grain boundary diffusion.

The extent to which superplastic flow contributes to deformation of the crust is not known because of the difficulty of identifying switching of neighbouring grains in naturally deformed rocks. Boullier and Gueguen (1975) presented textural evidence for superplastic flow in some mylonites. This mechanism is favoured in materials having small equant grains, deformed at temperatures above 0.5 $T_{melting}$

Dislocation Creep

Dislocation creep is the most common mechanism of high temperature plastic flow in crystalline materials. Microstructures indicative of dislocation creep are abundant in rocks deformed under conditions of upper greenschist and higher grades of metamorphism. This mechanism involves a combination of glide and climb of dislocations. Under an applied stress dislocations are continuously produced (multiply) at Frank-Read sources and glide through the crystal producing a change of shape (see Nicholas and Poirier, 1976, for a review of dislocation creep).

At the lower temperature range of intracrystalline plasticity gliding dislocations may become pinned by obstacles such as impurities, inclusions and dislocation tangles. Grain boundaries also serve to inhibit the motion of dislocations (grain boundary hardening). Dislocation multiplication coupled with pinning causes the density of dislocations to rise. This has the effect of increasing the stored strain energy leading to work hardening, in which the applied stress must be continuously increased to produce a constant strain rate.

At higher temperatures ($T/T_{melting} >$ 0.5) dislocations may cross-slip or climb past obstacles from one slip plane to another, and may annihilate one another, thus reducing the dislocation density and strain energy. These processes by which dislocations are arranged into a lower energy configuration are termed *recovery*. Steady state dislocation creep is a balance between work hardening and recovery, permitting high ductile strains to be attained. Although the strain is due to the gliding motion of dislocations strain rate is dependent on the climb of dislocations by diffus on. During recovery dislocations are arranged into dislocation walls which separate slightly misoriented sub-grains containing fewer dislocations: this is termed polygonisation.

Experimental work has implicated structural 'water' as an important factor weakening quartz and other silicates relative to the strength of 'dry' crystals (Griggs, 1974; Blacic, 1975; Jones, 1975). The mechanism is not well understood, and the various hypotheses are reviewed by Nicholas and Poirier (1976).

Constitutive flow laws proposed for dislocation creep have the following general form (taken from White, 1976a):

$$\dot{e} = \frac{AD_v Gb}{kT} \left(\frac{\sigma}{T}\right)^n$$

where $\dot{e}$ is strain rate, A is a constant, D_v is the volume diffusivity, G is the shear modulus, b is the Burgers vector, k is Boltzmann's constant, T is temperature (°K), σ is differential stress, and n a constant.

The following features are characteristic of dislocation creep (as reviewed by Carter, 1976; Nicholas and Poirier, 1976): 1) Strain rate varies as a power (n) of the applied stress where n, determined empirically, varies between 3 and 6; 2) Dislocation creep processes result in a strain rate that is independent of grain size. However in polycrystalline aggregates the stress required to overcome grain boundary hardening is inversely proportional to the square root of grain size and thus fine grained aggregates require a higher flow stress than coarser ones. This is the Hall-Petch

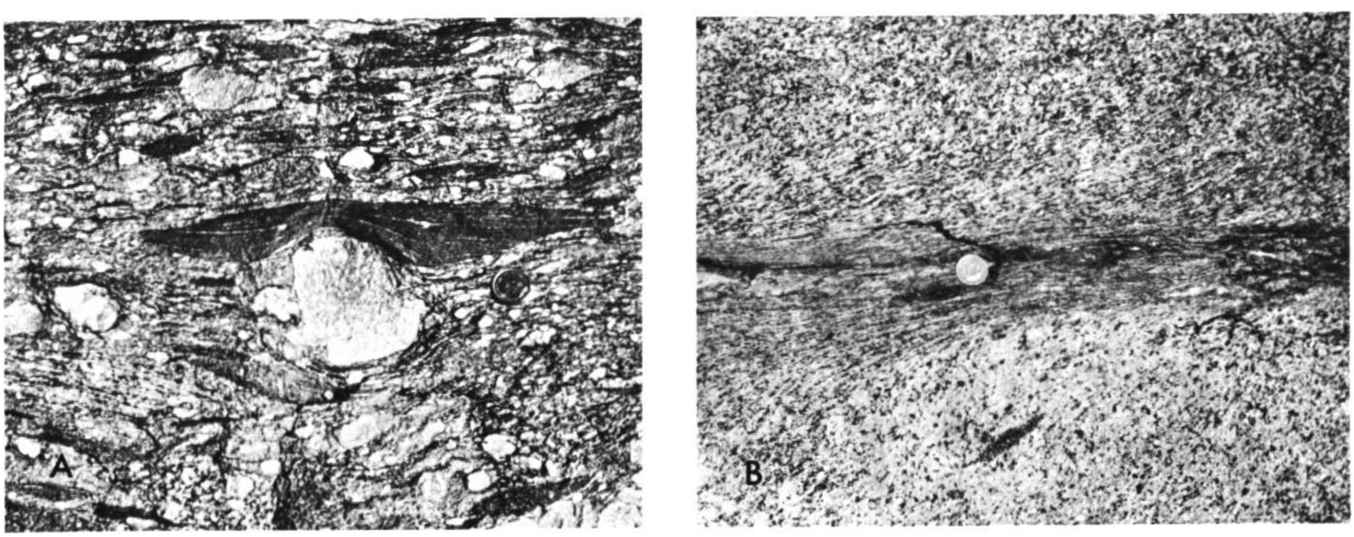

Figure 5.
Ductile shear zone: progressive development of schistosity at increasing states of strain from an initially undeformed granite. Schistosity defined by grain shape. B: Internal deformation of pebbles in conglomerate.

law (Nicholas and Poirier, 1976, p. 129); 3) There exists a substructure of polygonal cells whose size is inversely proportional to stress.

Microstructures. The high ductile strains which may be attained by dislocation motion are illustrated by the quartz ribbons in Figure 4A. Undulatory extinction is a feature generally present in deformed quartz (Figs 4A, 4D, 4F), arising from binding of the crystal structure, accommodated by dislocation motion. Subgrains formed by polygonisation during recovery are shown in Figure 4B. During recovery dislocations may climb into planes of high dislocation density which separate domains of lower density. These kink bands or deformation bands are distinguished optically by their bonded extinction (Fig. 4C). Optical features arising from dislocation creep and their relation to dislocation structures are described and discussed by White (1973, 1976b).

Dislocation creep may be complicated by recrystallisation, a more severe process than recovery for reducing the internal strain energy of crystals. The presence of small, equant, strain-free grains is indicative of recrystallisation (Fig. 4D). Recrystallisation generally occurs first at grain boundaries of kink bands within crystals. Prior to the formation of discrete new grains the boundaries become serrated (Fig. 4D).

It is possible to distinguish between annealing and dynamic recrystallisation. The former occurs at elevated temperature in the absence of differential stress, following deformation, and leads to the formation of equant polygonal stain free grains (foam microstructure, Fig. 4E). Dynamic recrystallisation occurs during deformation at high temperatures. The newly recrystallised grains are themselves deformed and are generally elongate. The microstructures typical of quartz mylonites which exhibit grain size reduction (Fig. 4F), are the products of dynamic recrystallisation.

Although recrystallisation is not itself a deformation mechanism it does alter the microstructure and change grain size, and hence may influence the relative importance of competing flow mechanisms. White (1977) presents a comprehensive account of recovery and recrystallisation in quartz.

The presence of undulatory extinction alone implies that the state of strain is low and it is unlikely that steady state creep has been attained. Conversely, a well developed subgrain structure is indicative of steady state creep (White, 1976b). Preferred crystallographic orientations are believed to be the product of dislocation gliding.

Small Scale Structures. Grain shapes which arise from intracrystalline deformation may contribute to a grain shape fabric which defines mesoscopic L and/or S fabric elements (Fig. 5A). Zones of high deformation such as mylonites typically exhibit extreme grain size reduction as a result of dynamic recrystallisation; not as a product of grain fracture.

Ductile deformation produced when the strain is accommodated by dislocation creep leads to small scale structures that have a continuous displacement field, and in which the mass of original grains is conserved. Deformation by dislocation creep is believed to be essentially isochemical in the absence of fluid-rock interaction (Kerrich *et al.*, 1978). Rock structures reflect the nature of the dominant deformation mechanism which in turn has an overall correlation with the tectonic environment. Grain size reduction by fracture in a submetamorphic environment (Fig. 1C) may be distinguished from the fine grained microstructure of a mylonite produced by dynamic recrystallisation under amphibolite conditions (Fig. 4F). The crenulation folds illustrated in Figure 1B which formed on the seafloor have neither the penetrative schistosity that might be anticipated for high temperature deformation (c.f. Fig. 5A), nor pressure solution cleavage characteristic of deformation at intermediate temperatures (Fig. 3C).

Deformation of a quartzite by mass transport at grain boundaries, with no internal strain, under sub-greenschist facies conditions, is illustrated in Figure 2D. This contrasts with the high ductile stain of quartz grains deformed at amphibolite grade (Fig. 4A), where dislocation creep predominates. The difference between strain accommodated by mass removal at boundaries and by internal deformation is further illustrated by the mechanical behaviour of conglomerates deformed at low temperature (Fig. 2C) and high temperature (Fig. 5B) respectively.

Deformation Regimes

A large body of field evidence has led to the conclusion that deformation of the crust is dominated at low temperatures by cataclasis, at intermediate temperatures by pressure solution, and at high temperatures by dislocation creep. The basis of this generalization is the observation that pressure solution is abundant in rocks deformed under conditions of low grade metamorphism; with microstructures indicative of dislocation creep becoming increasingly important from conditions of upper greenschist to higher grades of metamorphism. Cataclastic flow appears to be dominant under sub-metamorphic conditions.

Because of the rheological complexity of rocks, and the large number of independent variables which influence each flow mechanism it is difficult to place quantitative limits on these three deformation regimes. A wide and variable degree of overlap is present.

The transition between the pressure solution and dislocation creep regimes for quartzites of 1000 μm, 100μm and 10 μm grain diameter has been estimated at 550°C, 450°C and 300°C respectively. Corresponding figures for calcite rocks of the grain sizes quoted are 360°C, 300°C and 200°C. This transition does not appear to be sensitive to total confining pressure (Kerrich *et al.*, 1977).

A temperature dependence of the dominant deformation mechanism is further supported by the observation that rocks retaining some record of high grade metamorphic mineral assemblages deformed at lower temperature may exhibit features of pressure solution. Examples of these features are not abundant, contrasting with the ubiquitous effects of pressure solution during low temperature prograde metamorphism. This may be attributed to two factors. Firstly, high grade metamorphic rocks are relatively coarse grained, thus inhibiting pressure solution. secondly, during retrograde metamorphism hydration reactions absorb water, in contrast to the excess water present during prograde metamorphism.

A transition in the creep behaviour of quartzites from dominant pressure solution to dislocation creep with increasing strain rate has been established by Mitra (1976).

In a number of highly deformed rocks grain size reduction by dislocation creep appears to cause a switch to pressure solution, which permits faster flow rates at smaller grain diameters (White, 1976; Kerrich *et al.*, 1978). These observations support the conclusion that transition between the flow regimes is dependent on temperature, grain size, and strain rate.

Deformation mechanism maps. The overall rate of deformation of a polycrystalline solid depends on which independent mechanism contributes the fastest strain rate: this in turn is influenced by the rheological properties of the material (i.e., defect diffusivities) and the state variables temperature and differential stress (Stocker and Ashby, 1973). The inter-relationship of the different creep mechanisms as a function of the state variables, for a material of specified grain diameter, may be depicted on Ashby deformation mechanism maps (Ashby, 1972). These maps are constructed by solving the constitutive flow laws simultaneously, and display the stress-temperature space in which mechanisms are dominant (i.e., contribute the fastest strain rate).

Deformation-mechanism maps have been constructed for several geological materials including olivine (Stocker and Ashby, 1973), quartz (Rutter, 1976; White, 1976a), calcite (Rutter, 1976), and some ore minerals (Atkinson, 1977). These maps provide a theoretical basis for the observations presented above that pressure solution is dominant over dislocation creep for fine grained materials, at low values of temperature and strain rate.

Conclusions

It is believed that natural deformation of rocks is dominated by dislocation creep at high values of differential stress and temperature, and that pressure solution is an important mechanism in fine grained lithologies under conditions of low temperature and low strain rate. At very low temperatures and/or low values of effective confining stress cataclasis probably constitutes the dominant mechanism of crustal deformation.

Microstructures of rocks may therefore be used to infer the broad environment of deformation. It is important for instance to distinguish between rocks deformed at low temperatures and subsequently subjected to static metamorphism at higher temperature, from rocks deformed exclusively under higher temperature conditions.

For example, a lack of awareness of this factor may have given rise to the epigenetic replacement hypothesis for certain deformed strataform massive sulfide ore deposits, whereas many such deposits contain evidence of seafloor slumping during or after emplacement. A later overprint of deformation during regional metamorphism may lead to features that could be interpreted in terms of replacement and interpretation of all the structures as the product of high temperature tectonics.

Acknowledgements

The authors have benefitted from extensive discussion of rock deformation with B. K. Atkinson. E. H. Rutter and S. White. We are grateful to B. E. Gorman and J. White for critical reading of the manuscript. I. A. acknowledges National Research Council of Canada grant A3555.

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MS received March 30, 1978.