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# Deformation mechanisms — recognition from natural tectonites

R. J. Knipe

Department of Earth Sciences, The University, Leeds LS2 9JT, U.K.

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**Abstract**—Deformation mechanisms which can operate in rocks include diffusive mass transfer, crystal plasticity, grain-boundary sliding as well as fracture and cataclasis. Each mechanism produces a range of characteristic deformation microstructures which help to constrain the conditions, dynamics and kinematics of tectonic events. Cyclic deformation events are particularly important to the development and preservation of microstructures. The review identifies four areas where future studies need to concentrate. The first is the identification of fracture mechanisms. The second is the detailed assessment of the characteristics and effects of cyclic deformation events or flow instabilities on deformation mechanism paths. The third is the need to quantify the microstructural stability of characteristic microstructures and the fourth is the need to incorporate analysis of *PTt* paths into the analysis of deformation mechanism paths.

## INTRODUCTION

A FUNDAMENTAL aim of structural analysis is to establish the kinematics, dynamics and rheology of rocks during tectonic events. During the last 10 years, microstructural investigation of deformation processes has become recognized as an important element in such analysis. The detailed observation of natural and experimental deformation products, combined with field mapping and theoretical modelling of deformation processes, has led to the recognition of microstructures characteristic of the different deformation mechanisms which operate during tectonic events (Schmid 1982, Borradaile *et al.* 1983, Zwart *et al.* 1987, Barber & Meredith in press). In addition, the individual mechanisms which operate in specific minerals or rock types and the estimation of flow laws associated with them (Hanks & Raleigh 1980, Poirier 1985, Hobbs & Heard 1986), have been identified. Microstructures and textures which develop at different stress levels and in different kinematic frameworks, (Simpson & Schmid 1983, Passchier & Simpson 1986, Twiss 1986, Cobbold *et al.* 1987) have aided interpretation of the dynamics and movement history in fault zones and the important interactions between deformation, metamorphic and fluid processes (Rubie 1983, in press, Etheridge *et al.* 1984, Brodie & Rutter 1985, Kerrich 1986, McCaig 1987) have been recognized.

This paper aims to review the developments made in the last 10 years which have aided the interpretation and assessment of natural deformation processes. The paper outlines the problems which mark the present frontiers to this research, reviews new applications and techniques which will help move those frontiers forward and highlights some of the likely directions which this research might take in the next 10 years.

The first part of the paper reviews how each deformation mechanism operates, how it is recognized, what information can be extracted from the preserved microstructures and what aspects of the deformation mechanism remain a problem for future studies. The second part of the paper discusses the progress being made towards understanding the deformation mechanism paths involved in creating the final rock textures and fabrics preserved in rocks. The likely effects of cyclic deformation processes and the stability of microstructures characteristic of different deformation mechanisms are emphasized. This second section to the paper also speculates on the future trends in the analysis of deformation mechanisms and emphasizes the need to integrate such studies with other related fields of geological research.

## DEFORMATION MECHANISMS AND MATERIAL PROCESSES

The response of a rock to deformation is a function of a large number of environmental and lithological variables. Figure 1 lists these variables and illustrates the ways in which these factors combine to activate a dominant set of material processes. The exact set of material processes which operate in the material under particular

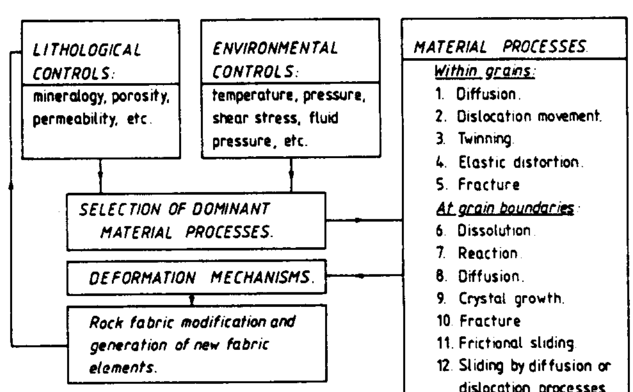


Fig. 1. Flow diagram illustrating the inter-relationships between lithological and environmental controls with material processes during rock deformation.

environmental conditions will control the deformation mechanisms which operate and the microstructures which evolve.

The basic prerequisites for the identification of deformation mechanisms from the preserved rock textures are that the microstructures generated by different deformation mechanisms are known and the likelihood that these microstructures are stable, and will survive, can be assessed. This second aspect, associated with the stability of microstructures, is critically dependent upon the exact history of the environmental conditions (e.g. temperature and stress conditions) which post-date the deformation event generating the microstructures. That is, some of the material processes listed in Fig. 1 will continue to operate after the main deformation event and can modify earlier microstructures. A later section of this paper argues that future quantification of the deformation processes and conditions depends upon careful assessment of microstructural stability.

Each of the deformation mechanisms possible in rocks is reviewed separately below, before a discussion of how these different processes interact during tectonic events. The mechanisms of deformation which can operate in rocks can be divided into the following headings.

(a) *Diffusive mass transfer* — Strain associated with the redistribution of material during deformation by diffusion processes.

(b) *Crystal plasticity* — Deformation by the movement of dislocations or by twinning in crystals.

(c) *Frictional sliding, fracture processes and cataclastic flow* — Deformation involving the creation of new surfaces, loss of cohesion by fracturing and frictional sliding along grain boundaries and surfaces.

## DIFFUSIVE MASS TRANSFER

Diffusive mass transfer (DMT) induces deformation by the transfer of material away from zones of relatively high intergranular normal stress to interfaces with low normal stresses (Kerrich 1978, Rutter 1983). The removal of material can lead to volume loss and to strain accommodation by compaction (Fig. 2). The driving force for the DMT depends upon the variation in chemical potential in the rock aggregate induced by stress variations in the aggregate (Wheeler 1987), fluid pressure gradients (Etheridge *et al.* 1984) or variations in the internal strain energy of grains (Wintsch 1985, Wintsch & Dunning 1985, Bell *et al.* 1986). Diffusive mass transfer is most likely to dominate the deformation in fine-grained material where the diffusion path length is low and the differential stress levels are low enough to inhibit crystal–plastic deformation mechanisms. The process of DMT can be considered a three stage process.

(i) *Source mechanisms* are associated with how the material enters a diffusion path and include the controls which influence activation of diffusion through the crystal structure and along grain boundaries and surfaces as well as those which dictate corrosion and reaction processes.

(ii) *Migration or diffusion mechanisms* involved in the transport of material along a range of mass transfer paths. The nature of these paths is critical to the DMT process and range from: (a) the use of the crystal structure as a diffusion medium, producing Nabarro–Herring Creep (see Poirier 1985); (b) diffusion along the distorted and disordered crystal structures of solid–solid grain boundaries, producing Coble creep (see Poirier 1985); (c) diffusion along a thin fluid film along grain boundaries, producing 'pressure solution' (Rutter 1983); and (d) transport in a bulk fluid which may itself be experiencing flow, producing infiltration of material (Etheridge *et al.* 1984, Gratier & Guiguet 1986, Spiers & Schutjens in press). The above list represents the end-member paths possible during DMT and one of the challenges to future studies is the characterization of the detailed properties of each path and the identification of which path(s) dominate DMT in different situations under different conditions.

(iii) *Sink processes* where material is precipitated or deposited in sites of crystal growth.

The microstructures associated with DMT may also be grouped according to this three-stage classification. Group (i) are microstructures indicative of the removal or redistribution of material, such as the truncation of fossils (Fig. 2), stylolites, pitted pebbles or mineral differentiation during crenulation cleavage development where quartz or carbonate migrates to hinge areas (McClay 1977, Rutter 1983, Houseknecht 1986). Group (ii) are microstructures which preserve evidence of the mass transfer path used, such as the preservation of reaction products along selected grain boundaries used for localized fluid flow (see McCaig 1987). Group (iii) microstructures are indicative of precipitation at sinks; for example overgrowths, pressure shadows and veins. Because of the number of processes which may create and transfer the material which ends up in sinks, it is difficult to infer source and transfer path details from the sink microstructures alone. This is highlighted by the recognition that newly precipitated phases can be different from the consumed phases. Beach (1979) termed this process "incongruent pressure solution". The growth of new phases in reaction sites where material is removed also emphasizes the importance of metamorphic reactions in DMT processes (see Knipe 1981). The complex interactions between straining and chemical process which can lead to softening or hardening of the material are reviewed by Brodie & Rutter (1985).

A number of flow laws have been derived to describe the relationships between strain rate and the environmental and material factors which influence DMT (see Table 1 and Spiers & Schutjens in press, for review). The range of equations reflects the range of diffusion paths and processes possible, especially those related to grain boundaries or interfaces. Understanding the structure and properties of grain boundaries and interfaces in tectonics remains an important goal in clarifying DMT processes. Recent investigations of grain boundaries by White & White (1981) have emphasized the potential role of networks of linked tubes and voids in the diffusion

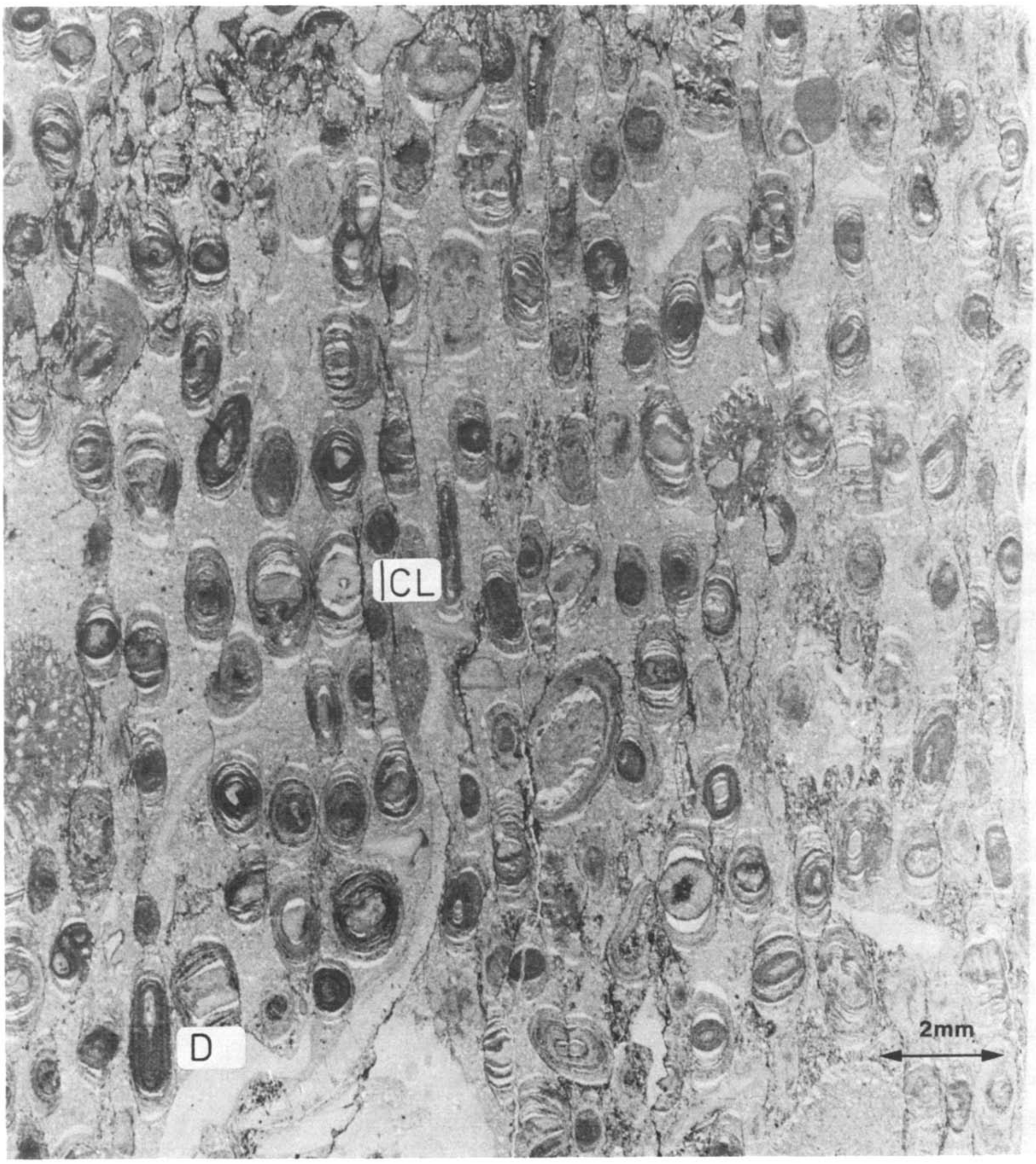


Fig. 2. Micrograph illustrating a weak cleavage (CL) developed by diffusive mass transfer (DMT) in an oolitic limestone. The internal oolite structure acts as a marker which is truncated during the removal of material on faces parallel to the cleavage. Note however that localized DMT is not the only deformation mechanism which has influenced the development of the final rock fabric. There is evidence of ductile flow producing a shape change in some ooliths (e.g. D) and almost every oolith exhibits evidence of sub-vertical extension by fracturing and crystal growth along the sub-horizontal sections of the internal concentric structure.

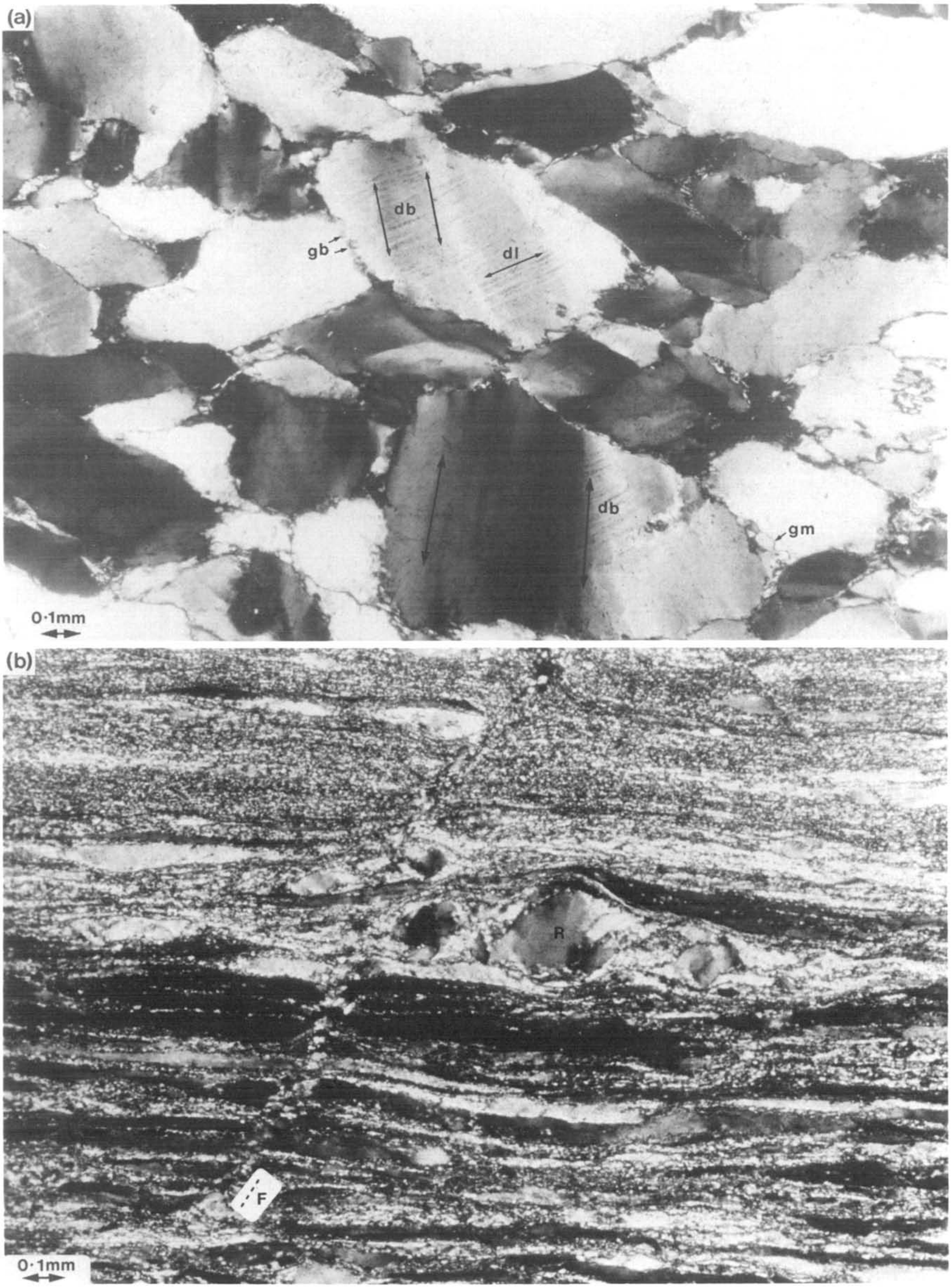


Fig. 3.

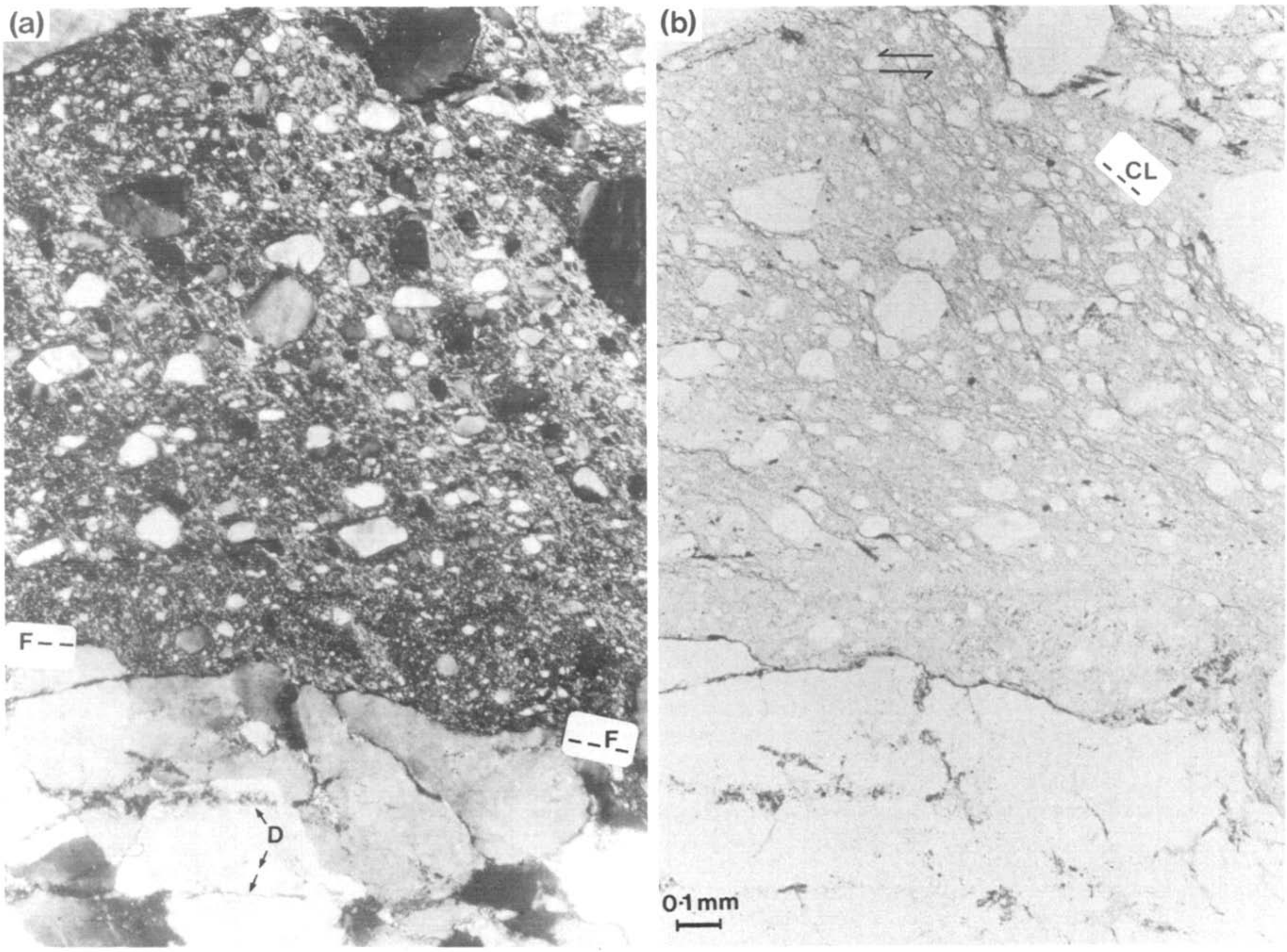


Fig. 4. Optical micrographs in crossed (a) and plane (b) polarized light of a cataclasite generated by brecciation of the same quartzite as shown in Fig. 3, (a) illustrates a microbreccia composed of sub-angular fragments set in a fine-grained matrix. Note the reduction in the number and size of fragments in a narrow zone adjacent to the host rock and the damage zones (D) which are sub-parallel to the main fault (F). (b) illustrates the presence of a weak oblique cleavage (CL) developed in the cataclasite. The sense of shear indicated by the cleavage is the same as the displacement which generated the cataclasite and suggests the deformation mechanism path for this specimen involved a change from accommodation of shearing by fracturing and grain-boundary sliding to diffusive mass transfer processes during the final stages of displacement on the fault when the displacement rate was decaying.

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Fig. 3. Optical microstructures developed by crystal–plastic processes operating in a quartzite. (a) Deformation lamellae (dl) and deformation bands (db) present in detrital quartz grains which have experienced only a small amount of strain (<10%). The plate also illustrates the localization of deformation features such as sub-grains and new recrystallized grains at grain boundaries (gb) and dentate grain boundaries characteristic of grain-boundary migration (gm). (b) Deformation features present in a mylonite where large strains have accumulated by crystal–plastic processes. In the top half of the micrograph approximately 90% of the rock has recrystallized from the texture similar to that shown in (a) to a fine-grained aggregate of new grains. The lower half of the micrograph is characterized by ribbon grains of quartz separated by narrow zones of recrystallized material. The ribbon grains may have arisen from either the selective grain-boundary migration from a fine-grained recrystallized aggregate or may represent elongated original detrital grains. Relict grains such as (R) contain well developed sub-grain structures. Note also the evidence for a late stage fracture which truncates the mylonitic foliation.

Table 1. Constutive equations for diffusion mass transfer processes and dislocation creep. $A_1$–$A_5$ are numerical constants. $\Omega$ is the molar volume of the solid. $D_v$ is the coefficient of volume diffusion. $D_b$ is the coefficient of grain-boundary (dry) diffusion. $D_f$ is the effective diffusivity of the solid in the grain-boundary fluid. $\sigma$ is the differential stress. $k$ is Boltzmann's constant. $T$ is the temperature and $d$ is the grain size. $S$ is the effective grain-boundary width for diffusion. C is the solubility (mole fraction of the solid in the grain-boundary fluid). $I$ is the velocity of dissolution or growth of the solid (the slowest). $G$ is the shear modulus. $b$ is the Burgers vector and $n$ is the dimensionless stress exponent

(1) Diffusional mass transfer

(a) Lattice diffusion (Nabarro–Herring creep):

$$\dot{e} = (A_1\ \Omega D_v \sigma)\ /\ (kTd^2)$$

(b) Grain-boundary diffusion — Dry (Coble creep):

$$\dot{e} = (A_2\ \Omega D_b \pi S \sigma)\ /\ (kTd^3)$$

(c) Grain-boundary diffusion — Wet (diffusion controlled):

$$\dot{e} = (A_3\ D_f\ C\Omega S \sigma)\ /\ (kTd^3)$$

(d) Dissolution/Precipitation control:

$$\dot{e} = (A_4\ I\Omega\sigma)\ /\ kTd)$$

(2) Dislocation creep

$$\dot{e} = (A_5 G D_v b/kT)\ (\Omega/G)^n$$

process. In addition, White & White (1981) suggested that many of the grain boundaries in tectonites are hydrated and that the grain-boundary width available for diffusion is larger than that proposed for metals (~5 × Burger Vector); the effect of this is to increase the diffusion flux possible along boundaries and thus enhance the DMT processes.

## CRYSTAL PLASTICITY

Crystal plasticity involves the accumulation of strain by intracrystalline processes such as the movement of dislocations (linear lattice defects) and twinning (Barber 1985).

The motion of dislocations through the lattice is controlled by the crystal structure and by the number of impurities and other lattice defects present (see Mitchell 1975, Poirier 1985, White 1985 for reviews). At low temperatures (usually <0.5 melting temperature at laboratory strain rates) deformation by this mechanism is dominated by dislocation glide where dislocation motion is confined to slip planes. This mode of deformation, termed low temperature plasticity, generally leads to dislocation tangles which restricts further dislocation motion and induces work-hardening (characterized by an increasing resistance to straining during deformation). Depending on the strain rate, the temperature and the amount of deformation possible before failure, the dislocation tangles may rearrange into a crude cell structure where the cell walls are defined by dense dislocation tangles. At higher temperatures (usually >0.5 melting temperature) thermally activated recovery processes such as dislocation climb (movement of dislocations out of their slip planes by point defect capture or emission) and cross-slip (alteration of the slip plane used by screw dislocations) help to reduce the work-hardening processes and increase the ductility of the material. The high temperature flow of a material where these recovery processes can counteract the hardening processes is termed dislocation creep. These processes can generate a set of characteristic microstructures (White 1976, Schmid *et al.* 1980, Vernon 1981, Olesen 1987), which include: (a) a stable dislocation density, (b) a well-defined sub-grain structure within grains, where low-angle grain boundaries, generated by the reorganization of dislocations, separate areas of the crystal with slightly different (<5°) lattice orientations, and (c) new grains enclosed by high-angle grain boundaries which develop by dynamic recrystallization processes associated with either grain-boundary migration or sub-grain rotation (see review by Urai *et al.* 1986). Cyclic changes in these microstructures during mylonite evolution involving dynamic recrystallization have also been recognized (White *et al.* 1980, Means 1981, Lister & Snoke 1984, Knipe & Law 1987). The flow law which has been derived for steady state dislocation creep is listed in Table 1. Examples of the microstructures generated by dislocation movement are shown in Fig. 3.

A key element in the assessment of crystal plasticity is the identification of the exact crystallographic orientation of slip and twinning systems possible under different conditions. One notable success during the last 20 years has been the identification of the slip systems operating during the flow of minerals (Nicolas & Poirier 1976, Hobbs & Heard 1986, Barber & Meredith in press). This success has contributed directly to the sophisticated interpretation of crystal–plastic processes in minerals which is now possible. There are three areas where the analysis of crystal–plastic processes have been particularly important to the understanding of deformation processes operating during tectonic events.

First, is the recognition that dislocation creep and dynamic recrystallization processes are fundamental to the softening mechanisms associated with the localization of deformation into shear zones (White *et al.* 1980, Schmid 1982). The major softening processes identified are geometrical softening, arising from the alignment of easy glide systems parallel to orientations of high shear stress and the operation of additional deformation mechanisms (e.g. DMT and grain-boundary sliding), possible in the fine-grained products of dynamic recrystallization. (Note that the grain-boundary sliding processes operating here differ from the frictional sliding described in the next section and are associated with solid state processes such as diffusion or the motion grain-boundary defects.)

Secondly, the refined understanding of the evolution of crystallographic lattice preferred orientations (CPOs) can now be related to the operating dislocation slip systems, the finite strain and the strain path followed (Lister & Hobbs 1980, Wenk 1985, Schmid & Casey 1986). The analysis of CPOs is now recognized as an important and powerful tool in tectonic analysis and detailed studies have aided the recognition of flow patterns in thrust sheets (Law *et al.* 1984, 1986, Platt &

Behrmann 1986, Schmid *et al.* 1987) and extensional terrains (Lee *et al.* 1987).

Thirdly an estimation of the stress levels operating in ductile deformation zones is now possible because the microstructures developed during dislocation creep and twinning are related to the flow stress level (Groshong 1972, Jamison & Spang, 1976, White 1979a, Edward *et al.* 1982, Ord & Christie 1984, Twiss 1986). Twinning can thus be used to assess the stress magnitude and orientation and the twin density related to the strain magnitude. In the case of dislocation creep, the dislocation density, sub-grain size and size of recrystallized grains can all be related to the flow stress level. Despite recognition of some refinements needed to this technique (White 1979b, Etheridge & Wilkie 1981, Ranalli, 1984, Twiss, 1986), the microstructures have led to important estimates of palaeostress levels.

Recent research into crystal plasticity has allowed the exact structures and energies of dislocations and grain boundaries in minerals to be modelled (White 1985, McLaren 1986), and the interaction of dislocations with point defects and impurities assessed (Hobbs 1981, Blacic & Christie 1984, Jaoul *et al.* 1984, Kirby 1984, Schock 1985, Ord & Hobbs in press). Such studies are proving crucial to the understanding of dislocation mobility and to the reactivity of minerals (Wintsch 1985, Bell *et al.* 1986, Wintsch & Andrews 1988). In addition, the analysis of grain-boundary migration and segregation process (Joesten 1983, Olgaard & Evans 1986) are providing important information on the processes which typically accompany crystal–plastic processes.

Amongst the new techniques which can dramatically help future studies on crystal–plastic processes (and aid other microstructural investigations), the recent advances in back-scattered electron imaging in the SEM are particularly significant. The ability to collect complete crystallographic orientation data from electron channelling patterns and to identify regions with different crystallographic orientations from orientation contrast images from large areas adds an important new dimension to fabric analysis (Lloyd 1987, Lloyd *et al.* 1987).

## FRICTIONAL SLIDING, FRACTURE PROCESSES AND CATACLASIS

Although all these processes are linked, they are separated here into (i) frictional grain-boundary sliding where fracture does not dominate the deformation and (ii) fracture processes.

### *Frictional grain-boundary sliding without fracture*

Deformation by frictional grain-boundary sliding involves the sliding of grains past each other. Individual grains are essentially undeformed and behave as rigid bodies. Borradaile (1981) termed this mode of deformation "independent particulate flow". Sliding starts when the cohesion and friction between grains is overcome, and is distinguished from the grain-boundary sliding possible at higher temperatures where diffusive or defect movement along grain boundaries controls the deformation and no loss of cohesion is involved.

Frictional grain-boundary sliding is a pressure sensitive mode of deformation and is enhanced by low confining pressures and high fluid pressures (low effective pressures). The mechanism is thus common in partially or unlithified sediments (Maltman 1984, Owen 1987) and in fault zones containing incohesive gouges (Wang 1986). The initiation of sliding is critically dependent upon the amount and strength of cement bridges between the grains.

Complex volume changes are possible during this style of deformation. High fluid pressures and fluid influxes may promote significant dilation and even fluidization and liquefaction of the material (e.g. sand volcanoes, slumps and fluidization breccias). Owen (1987) has recently reviewed the deformation processes associated with liquidization through fluidization and liquefaction. On the other hand, the deformation may involve a much lower level of dilation ($\leqslant 10\%$) on the hand-specimen scale, but still be associated with the creation and destruction of important dilation sites on a grain scale. In both cases the late stage loss of fluids may eliminate the transient dilation. Where clays and muds experience this deformation there is also the possibility of large volume losses and preferred alignment of grains during the porosity collapse of the grain framework as fluid is expelled. Such processes appear to be fundamental in the evolution of deformation features found in the partially lithified sediments recovered from D.S.D.P. cores including faults (Knipe 1986a), veins (Knipe 1986b) and scaly fabrics (Moore *et al.* 1986).

Knipe (1986b) has recently suggested that the interaction of fluid migration and deformation by frictional sliding in partially lithified sediments (caused by fluid overpressuring or by the regional stresses) can induce a migrating wave of deformation where a migrating sequence of dilation + fluid influx → disaggregation + displacement → collapse + grain alignment moves through the sediment. This process involving the migration of fluid 'packets' and pressure waves may be important in the evolution of bedding fabrics during compaction and the alignment of clay particles in fault and shear zones.

Analysis of the complex deformation sequences which may accompany frictional grain-boundary sliding in sediments has been aided by the recent application of concepts developed for Soil Mechanics (Jones & Addis 1986). The concept of 'Critical State Analysis' which identifies the conditions in effective stress/volume change space at which constant volume shear takes over from compaction or consolidation during increasing load, has proved particularly useful (see Jones & Addis 1986, Jones & Preston 1987). An additional way of gaining qualitative insight into the deformation histories associated with 'soft sediment' deformation is to consider the changes in terms of the important controlling variables on a 3D diagram. Knipe (1986c) chose lithification, strain rate and fluid pressure as axes in a first attempt at assessing the range of deformation histories which may be involved in the development of fabrics in soft sediments. This analysis highlighted the range of possible paths (i.e. families of surfaces in the 3D diagram along which the deformation path may lie) which could account for the fabric evolution. The advantage of attempting the construction of such diagrams (whatever axes are chosen) is that they help to identify the range of possible deformation mechanism paths and emphasize the post-deformation changes which contribute to the final fabric preserved in the sediment.

The recognition of frictional grain-boundary sliding in naturally deformed rocks is particularly difficult in that grain shapes, sizes and internal structure may all be unaffected by the deformation. Where the deformation was in poorly consolidated sands, only the offsets of bedding or the disruption of other sedimentary features indicates the presence of a deformation zone. However the lack of variations in the grain size, shape and internal structure in and out of such deformation zones confines the mechanism to one of frictional grain-boundary sliding. In some cases it is possible to detect changes in the packing of the grains within the deformation zones by assessing the separation distance and orientation of vectors linking grain centres using the centre-to-centre method described by Fry (1979) and Ramsay & Huber (1983). In aggregates containing elongate or platy grains, deformation may induce a change in the orientation pattern or a shape fabric either by alignment during fluid streaming and liquidization or by rotation during particulate flow (Freeman 1985). The quantification of the amount of dilation involved in particulate flow is particularly difficult in naturally deformed rocks because of possible modification of the flow fabric during the late stage collapse and fluid expulsion. Thus the final fabric preserved does not necessarily represent the fabric present during displacement. An additional problem arises where frictional grain-boundary sliding is only one of a number of deformation mechanisms which operate together. In these cases if the bulk strain and the strain from the other mechanisms can be estimated, then the contribution of grain-boundary sliding can be determined from the difference.

### *Fracture processes*

Fracture processes involve the nucleation, propagation and displacement along new surfaces created during deformation (Kranz 1983, Atkinson 1987). The fragmentation of material, together with the rotation and associated grain-boundary sliding and dilation, constitute cataclasis which dominates faulting at high crustal levels and generates gouges and breccias (Engelder 1974, Aydin & Johnson 1983, Wise *et al.* 1984, Sibson 1986a). Figure 4 shows an example of the products of cataclasis. Where rapid seismic slip takes place under dry conditions frictional melts or pseudotachylites can develop (Sibson 1975, Passchier 1982, see also Hobbs *et al.* 1986). At the other extreme, where repeated microfracturing events take place, fibrous veins or elongate grain fabrics can develop (Knipe & White 1979b, Ramsay 1980).

A number of recent approaches have helped towards the understanding of the mechanisms, dynamics and kinematics involved in the evolution of cataclastic faults. The integration of the fracture mechanics approach which defines the conditions and processes associated with single crack or fault propagation has helped enormously to interpret the failure modes and conditions in rocks (see Kozak & Waniek 1986, Atkinson 1987). The understanding of frictional sliding behaviour in faults has been advanced significantly by experiments and the development of constitutive laws which describe the weakening and strengthening behaviour patterns which take place under different slip velocity conditions (Dieterich 1981, Rice 1983, Blanpied & Tullis 1986, Rice & Tse 1986, Shimamoto & Logan 1986, Tullis & Weeks 1986). Detailed microstructural studies of experimental and natural gouges have allowed recognition of the criteria useful for the identification of the kinematics of fault zones (House & Gray, 1982, Blenkinsop & Rutter 1986, Rutter *et al.* 1986, Chester & Logan 1987) and in addition have produced important data on the rheological behaviour of gouges (Logan & Rauenzahn 1987). The role of extensional microcracks induced by the tensile stresses developed at point contacts between grains and fragments during the evolution of fault gouges has been emphasized by Gallagher *et al.* (1974) and Blenkinsop & Rutter (1986). Kranz & Scholz (1977) and Costin (1983) have proposed that a critical crack density may be required to initiate cataclastic failure. The amalgamation of data on earthquake repeat times and location with breccia characteristics has allowed Sibson (1986a) to relate brecciation processes to: (i) the progressive frictional wear along slip surfaces (attrition brecciation); (ii) impeded slip in the fault zone (distributed crush brecciation); and (iii) to rapid dilation events (implosion brecciation). The comparison of fracture array geometry and particle size (block) analysis on a range of scales has led to the recognition of self-similarity of the particle size distributions and of fractal geometries produced during comminution (Sammis *et al.* 1986, Turcotte 1986).

While each of the above approaches has helped refine the understanding of fracture processes in rocks, identification of the micromechanisms associated with fracture propagation, the separation of these from the mechanisms involved in displacement along the fracture and the exact role of fluids in all these events remains an important goal for future studies. Two features complicate the interpretation of natural fracture mechanisms and the achievement of these goals. Firstly there are a large number of fracture mechanisms and identification of the characteristic microstructures associated with each remains elusive. Secondly the modification (or even obliteration) of microstructures developed early in the fracture history is likely. Despite these problems, the understanding of fracture processes is moving forward rapidly. The range of fracture mechanisms and processes are reviewed here before a discussion of the modification of microstructures is given in a separate section on microstructural stability.

*Classification of fracture mechanisms*

A simple classification of fracture mechanisms is shown in Fig. 5, which emphasizes the range of pre-failure processes which influence or control the propagation of fractures. These processes include the following.

*Elastic strain accumulation,* where the elastic strain energy associated with a stress concentration at a crack tip controls propagation. Quantification of this stress intensity factor forms the basis of a number of theories which predict the conditions needed for crack extension at tips with different geometries under different loads (see Atkinson 1987, Ingraffea 1987). Fractures developed tend to follow weaknesses in the material and may be transgranular and follow cleavage orientations or may be intergranular and exploit grain boundaries. The frequency, orientation, shape and distribution of pre-existing surface cracks, flaws and grain-boundary voids and pores can all influence the amount of strain accumulated before failure by catastrophic fracture propagation.

*Crystal–plastic processes* can contribute to the fracture processes when dense dislocation tangles or high twin densities develop to restrict further deformation by crystal plasticity and induce rapid work-hardening leading to fracture (Atkinson & Meredith 1987). In addition, in polycrystalline aggregates variations in the ease of deformation between adjacent grains with different crystal structure orientations and thus different operating slip systems give rise to strain compatibility problems which can induce void formation along grain boundaries especially at triple points. The steady rise in the number of such features can eventually lead to failure. Both the work-hardening aspects of failure and the creation of voids will be suppressed at high temperature by the increased rate of recovery processes and the operation of additional slip systems.

*Diffusion processes* can lead to the development of voids at tensile grain boundaries or triple points by the concentration of point defects or vacancies. In addition the diffusion of impurities to grain boundaries can also lead to the 'embrittlement' of grain boundaries and fracture (Darot & Gueguen 1986, Atkinson & Meredith 1987).

*Phase transformations and reactions* by creating products with a different volume to the reactants can induce stress concentrations in an aggregate (Brodie & Rutter 1985) which may induce void formation leading to failure.

*Fluid processes* have a fundamental role in fracture processes. The mechanical effects arise where fluid pressure causes hydraulic fracture, and the effective stress exceeds the strength of the material (Fyfe *et al.* 1978, Sibson 1989). The chemical processes involve the control of fracture propagation by the corrosion and reactions taking place at the crack tip leading to sub-critical crack growth (Atkinson 1984, Darot & Gueguen 1986, Kerrich 1986, Atkinson & Meredith 1987).

Although the elastic strain energy accumulation is the main process associated with fast fracture propagation (brittle failure) all the above processes may be involved in sub-critical crack growth at lower propagation velocities below the critical stress level needed for catastrophic failure (Fig. 5). It is also common for slow crack growth (sub-critical) processes to be the precursors to large failures by fast crack growth.

Figure 5(b) emphasizes that the slow crack growth processes associated with stress corrosion can lead to failure under a wide range of conditions and may operate in conjunction with other mechanisms. For example the

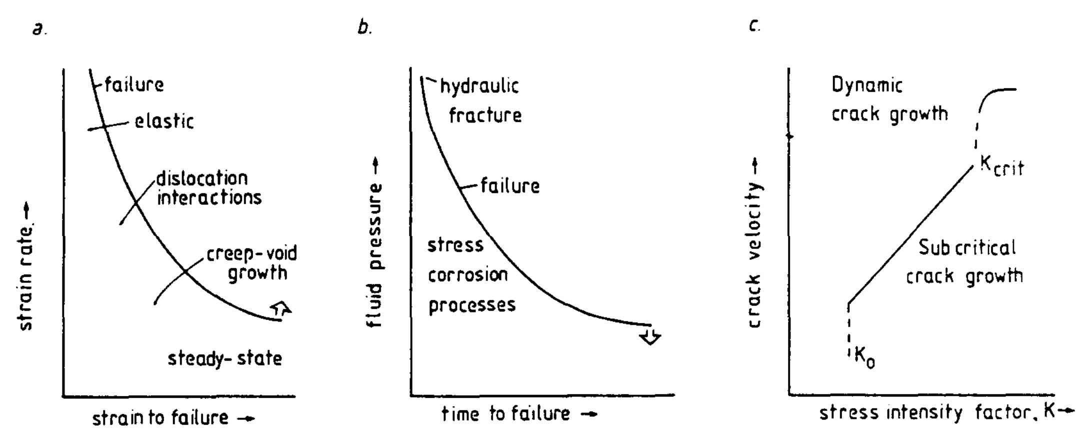


Fig. 5. Review of fracture processes. (a) The range of pre-failure processes which may operate during fluid-absent deformation under different strain rate conditions. The arrow represents the change in the failure line position associated with an increase in temperature or confining pressure. (b) illustrates the possibility of additional slow crack growth processes such as stress corrosion in a fluid present situation. The arrow indicates the movement direction of the time to failure line during an increase in the stress intensity factor (see c) and/or the alteration of the fluid chemistry to a more corrosive composition. (c) is the fracture mechanisms associated with different crack velocities and stress intensity factors. The stress intensity factor ($K$) can be considered to be the driving force for crack propagation. It describes the stress system around a crack tip and is dependent upon the loading conditions and the material properties. The values of $K$ range from $K_0$, below which sub-critical crack growth ceases, to $K_{crit}$ when propagation accelerates to approach the velocity of sound in the material.

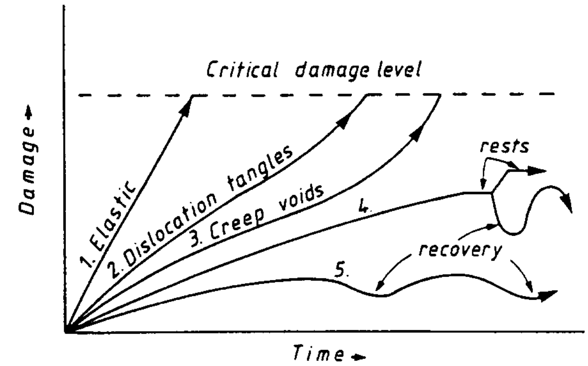


Fig. 6. Diagram reviewing the different 'damage' processes which can contribute to failure at a critical damage level. The different paths shown (1–5) illustrate different mechanisms of pre-failure damage accumulation which range from elastic strain energy accumulation (path 1), through dislocation tangles (path 2), to creep void development (path 3). Paths 4 and 5 illustrate deformation paths where the damage build up is interrupted by rest periods or recovery periods when damage is reduced.

growth of voids by crystal–plastic or diffusion processes may allow infiltration of fluid leading to stress corrosion, or the increased internal strain energy arising from dislocation creep may increase the rate of reactions and corrosion at crack tips. It is clear that the separation of fluid-absent and fluid-present situations in Fig. 5 is somewhat artificial and a range of mechanisms may be involved in the evolution of even a single fracture, thus necessitating a consideration of both 'end-member' situations.

An alternative way of representing the range of processes involved in fracturing is shown in Fig. 6. In this case, failure is considered to take place when the 'damage' in the material reaches a critical level. This damage can arise from an increase in the elastic strain energy, the plastic strain energy, the number, shape and orientation of creep voids (related to either crystal plasticity or DMT) or the microfracture geometry and density created by slow crack growth (see Costin 1987). Another important aspect of the damage processes associated with slow crack growth processes is that they can be considered to be 'sleeper' or background processes — which may produce only a very minor contribution to the bulk strain until the critical damage level is reached when they rapidly induce a change of mechanical behaviour and dramatic failure. Establishing the possible rates of damage build-up for the different processes, and identification of some of the processes which can reduce the damage is of crucial importance to fracture studies. Failure can be considered in terms of the competition between the rate of damage build-up and the rate of damage removal. The recovery of dislocation sub-structures and the infilling of creep voids by crystal–plastic, diffusion or precipitation processes, together with the blunting of crack tips by crystal plasticity, diffusion or reactions, may all contribute to the removal of deformation induced damage and thus prolong the pre-failure time period. A particularly important control of the rate of damage build up is the strain rate and stress history experienced by the material. Periods when the stress or the strain rate drops will be valuable rest periods which promote damage removal and thus the amplitude, wavelength and frequency of any stress cyclicity become fundamental to the failure or fatigue behaviour.

The brief review of fracturing processes given above emphasizes that a range of mechanisms can be involved in cataclasis. It is also clear that despite the progress made in recent years, understanding of fracture processes in rocks is still an infant science. More of the recent integrated studies in which fracture mechanics and microstructural studies are combined (Lawn 1983, Ferguson *et al.* 1987) and more fatigue studies (see Costin 1987) are required.

## DEFORMATION MECHANISM MAPS

Deformation mechanism maps developed by Weertman (1968) and Ashby (1972) were first applied to geological materials by Stocker & Ashby (1973), Rutter (1976), White (1976) and Atkinson (1977). The main purpose of these maps is to identify regions in stress–temperature or stress–grain-size space where different deformation mechanisms dominate the rheology of materials. The diagrams are based upon the calculation of strain-rate values using flow laws derived from steady-state deformation, and allow the rapid assessment of the probable deformation processes at different conditions (Langdon 1985). In addition, owing to the different microstructures associated with different mechanisms, they can be used to estimate deformation conditions involved in natural tectonic events. However the application of these maps to natural deformation is still limited by the amount of relevant experimental data on real rocks, the assumptions that steady state deformation operates and that a constant microstructure is present. The difficulty of identifying the relevant diffusion creep process, the problems of including fracture processes (especially slow crack growth) and the application of single crystal data to polyphase rocks impose additional limitations to the use of these diagrams. Freeman & Ferguson (1986) have addressed one of these limitations by including analysis of variable grain-size distributions.

More recently the idea of fracture mechanism maps developed by Ashby *et al.* (1979) and Gandhi & Ashby (1979) have been applied to geological material (Dennis & Atkinson 1981). In this case the maps show in stress–temperature space the different fracture mechanisms which produce failure in the shortest time. An alternative and preferable fracture mechanism map introduced by Fields & Fuller (1981) uses stress intensity and temperature as axes and contains contours of crack velocity. However, as Atkinson & Meredith (1987) note, such detailed maps cannot be constructed as yet for any geological material. Despite the present limitations there is no doubt that deformation mechanism maps are a valuable aid to interpretation of rock deformation and have provided a powerful first-base for an insight into the rheological behaviour of rocks during tectonic events. The maps are particularly useful for assessing the competition of deformation mechanisms and the probable distribution of deformation processes at different levels in the crust and mantle.

## DEFORMATION MECHANISM PATHS

The insight into deformation processes gained from the consideration of steady state flow laws and the ability to estimate strain rates and stress conditions using such flow laws has provided a framework for the quantification of deformation processes and construction of rheological models of the crust (Sibson, 1977, 1983, 1984, 1986b, Hobbs *et al.* 1986, Ord & Hobbs in press). By selecting an average strain rate and values for material variables it is possible to use the flow laws to construct profiles of shear resistance with depth through the crust (see Sibson 1984, 1986b). Such models separate an upper crustal frictional regime from a quasi-plastic regime and identify a peak stress region where the transition between these regimes occurs. (The change in behaviour with depth should not be termed the brittle–ductile transition — see Rutter 1986.) Sibson (1984, 1986b) considers the base of the seismogenic zone to be associated this frictional–quasi-plastic transition and that seismic events propagate down into the quasi-plastic regime. Hobbs *et al.* (1986) suggest that seismic events can also nucleate by plastic instabilities in the top of the quasi-plastic zone.

While shear resistance profiles with depth are extremely important for large-scale modelling of fault zones, consideration of the microstructural evolution of fault rocks also requires consideration of the stress and strain rate histories of faults at different depths. The assignment of a single characteristic stress level to deformation at different depths removes attention from the possible changes in strain rate and stress levels involved in deformation processes. This is especially true of high-level deformation in the frictional regime, where the average tectonic strain rate arises from the operation of intermittent high strain-rate events separated by periods of low or no deformation (Rice 1983, Sibson 1986b). It is also true of quasi-plastic deformation where high strain rate events or instabilities still characterize the unstable flow (Hobbs *et al.* 1986, Sibson 1986b). Only at high temperatures and deep levels where the deformation appears steady state are the average tectonic strain rates likely to reflect these experienced by the material.

The microstructural evolution and the mechanical behaviour are critically dependent upon the details of the cyclic events or instabilities, particularly the range of strain rates experienced and the time spent at each strain-rate level. This is because each strain-rate level experienced may be associated with a different deformation mechanism generating different characteristic microstructures. Thus, there is a need to characterize *deformation mechanism paths* (Knipe 1986c) which delineate the history of mechanisms and microstructural evolution involved in producing the final rock fabric. Future understanding of natural deformation processes will require characterization of the details of deformation mechanism paths associated with the transient events or instabilities which arise in material at different depths and in different tectonic settings, as well as an assessment of the changes in characteristic microstructures during subsequent history. Three aspects of deformation mechanism path recognition which are crucial to future studies are discussed separately below.

(i) The characterization of causes and effects of cyclic deformation events and flow instabilities.

(ii) The stability of microstructures.

(iii) The integration of *PTt* path data or burial-uplift histories into analysis of deformation mechanism paths.

### *Cyclic deformation and flow instabilities*

Any deformation event will involve an increase and then decrease in the strain rate and can be considered in terms of a cycle. On the large scale, such events can be considered to include an orogenic event or an extensional basin evolution where the strain rate builds up and then decays over a long time period (>20 Ma). These events or cycles can be thought of including a set of smaller-scale cycles: for example, the deformation cycle on one fault, which becomes active only during part of the larger cycle and is then abandoned as other faults take over the strain accommodation. On an even smaller scale, the movement on individual faults can be considered to be made up of a large number of repeated cyclic events. The multiple events associated with seismicity in the frictional regime and which involve cataclasis are well known (see Sibson 1986, 1989) but not fully understood in terms of mechanisms. Repeated fracturing events during the development of veins (Knipe & White 1979a,b, Ramsay 1980) and during the evolution of mylonites (Sibson 1980, White & White 1983, Knipe & Wintsch 1985, Hobbs *et al.* 1986) have also been noted. Oscillation in the strain rate in localized domains associated with the development of transiently soft or hard zones during plastic deformation was also considered important to the development of slaty cleavage by Knipe (1981) and to mylonite evolution by (Knipe & Wintsch 1985, Hobbs *et al.* 1986, Knipe & Law 1987). Hobbs *et al.* (1986) have discussed the potential of the instabilities arising during unstable plastic flow in mylonites for generating seismic events and have estimated the conditions required for instability development.

Cyclic deformation is also possible in material undergoing bulk steady state deformation. The results of an experimental deformation of polycrystalline para-dichlorobenzene conducted by the author is shown in Fig. 7 and illustrates the cyclic redistribution of strain accommodation between domains during bulk simple shear. The details of the experimental method used here are given by Means (1989). The diagrams illustrate the pattern of strain rate changes in a mylonite zone undergoing simple shear at constant temperature and bulk strain rate. At any one time, domains with increasing, decreasing or constant strain rate can be identified. A

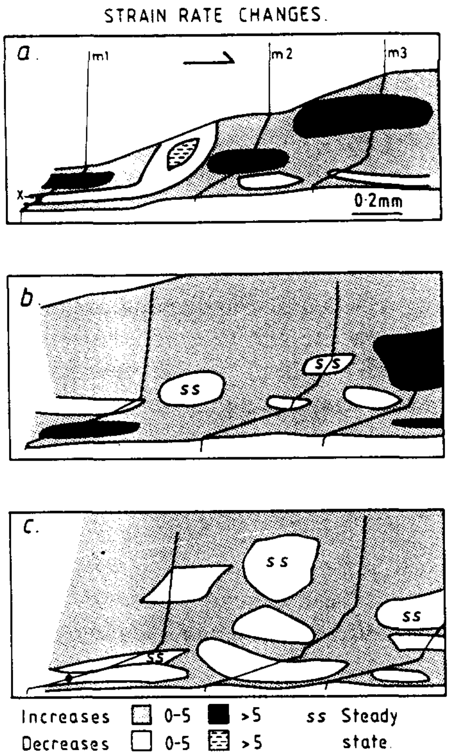


Fig. 7. Patterns of strain rate changes during progressive experimental dextral shearing of para-dichlorobenzene. The positions of three marker lines (m1–m3) during the deformation from (a)–(c) are shown. The increases and decreases in the strain rate of points in the shear zone during the deformation are shown by shading. Note the cyclic decrease–increase–decrease experienced by point X and the small areas which approximate to steady state deformation where changes in the strain rate could not be detected. The average strain rate during the experiment was $3 \times 10^{-4}\ s^{-1}$.

particularly interesting feature of the experiment shown was the oscillation of the strain rate in adjacent domains which are sub-parallel to the shear-zone margin.

The above discussion highlights the fundamental importance of cyclic deformation to tectonic events. Cyclic events or instabilities appear to be ubiquitous in deformation but the 'size' and effects of the different cyclic events vary. The detailed characteristics of the cyclic events will vary in rocks of different composition deforming under different conditions in different tectonic settings. The characteristic elements of deformation cycles which control the deformation mechanism paths followed and are important to the rheological behaviour (and the microstructural evolution) can be discussed in terms of the wave-like variation in the magnitude of the strain rate or stress with time (Fig. 8). The main characteristics of cyclic events therefore include, the range of strain rates involved (*amplitude*), the time period of the event (*wavelength*), the exact form of the strain-rate variation with time (*waveform*), the repeat time between events (*frequency*), the rate of migration of the peak strain rate associated with the event, if the event can be considered to be a migrating wave (*migration velocity*), the rate of change of the peak strain rate with time (*the growth or decay factor*) and the rock *volume affected* by the event (see Fig. 8). The detailed assessment of these characteristics is not possible at present, but certainly provides some important objectives for future studies. Despite these uncertainties it is possible to speculate on the possible strain-rate patterns expected at different depths along fault zones

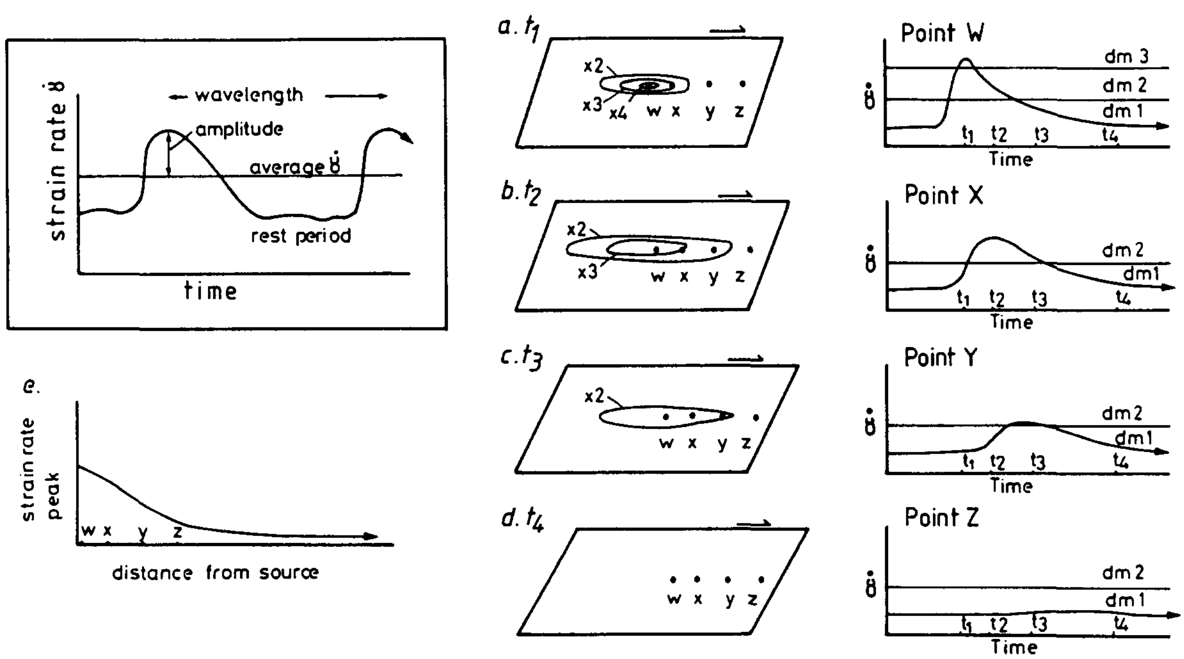


Fig. 8. Characteristic elements of cyclic deformation. The inset shows the main features of a strain-rate cycle. (a)–(d) illustrate the pattern of strain-rate contours associated with the development of a flow instability or cycle. The propagation of areas where the strain rate increases to values of up to four times (×4) the background strain rate and then decreases is shown in this example. The strain rate vs time histories of points w,x,y and z together with the time spent deforming in regimes where different deformation mechanisms (dm1–dm3) dominate are also shown. Note the different deformation mechanism paths of points located at different distances from the source of the instability. Note also that the interface between the fields dominated by different deformation mechanisms is diagrammatic: in the real situation the fields may expand or contract with time as microstructural changes take place, (e) shows the changes in the peak strain rate experienced by points at different distances from the source area of the instability.

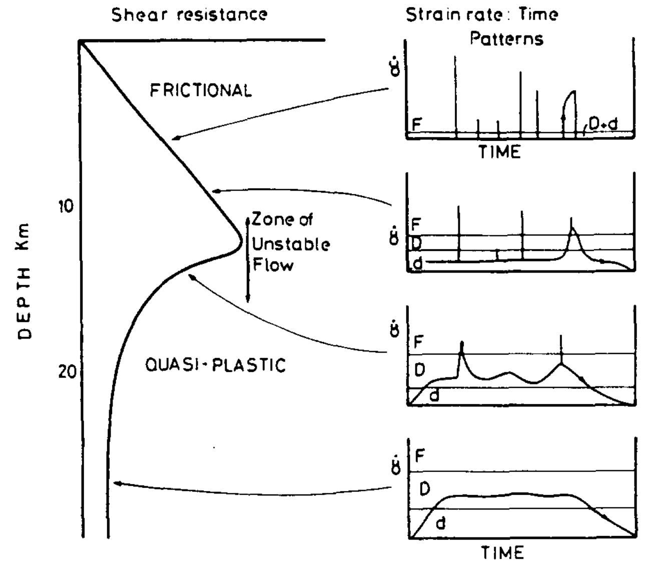


Fig. 9. Possible strain rate [$\dot{\gamma}$] vs time patterns for a strike-slip fault in the crust. The location of the different patterns on a shear–resistance–depth profile is shown together with the relative importance of different deformation mechanism fields: F=fracture, D=dislocation movement or d=diffusive mass transfer. See text for discussion.

and to use this speculation as a basis for a discussion on the material behaviour and the microstructural evolution. Figure 9 illustrates a strain rate–time pattern possible on a planar strike-slip fault and compares this with the familiar peak strength vs depth profile. The diagrams model the change from a pattern of intermittent high-amplitude, small-wavelength fracture events at shallow depths to one approaching possible steady state at depth. This leads to some important conclusions which are relevant to microstructural studies.

(a) Microstructures developing in rocks at different depths will be evolving *throughout* the history and the time spent in each strain-rate band associated with each deformation mechanism will influence the microstructures formed (see Figs. 8 and 9).

(b) The periods of lower strain rates between the transient high strain-rate events, and the decay period of the strain rate at the end of the longer deformation events, are important recovery periods when modifications to microstructures are possible.

(c) Where the deformation involves long periods at low strain rates separated by transient high strain-rate events (for example in the upper quasi-plastic regime and frictional regime), it is possible that strain accommodation is dominated by the short-wavelength, high-amplitude events, while microstructural evolution is dominated by the longer periods at lower strain rates.

(d) Steady-state deformation on a hand-specimen scale can still involve the cyclic relocation of strain rate between micro-domains and should perhaps be considered an end member situation where the cyclic deformation events have low amplitudes and affect small volumes.

(e) Fluid influxes may be associated with the transient high strain-rate events (Sibson 1981, 1989, Etheridge *et al.* 1984) which could induce additional microstructural modifications and reactions. Sealing of the fault zones (Stel 1985, 1986) and pulsed fluid migration may cause an alteration of open and closed systems which will influence reactions induced by the fluid (see Knipe & Wintsch 1985).

(f) The superposition of deformation events which behave as strain waves with different migration velocities also requires consideration when assessing the origin of instabilities in flow.

(g) Faults which move deforming material along a temperature path (e.g. thrusts and extensional faults) or deformation in areas where the heat flow is changing will generate more complex fault rocks as the microstructures characteristic of one set of conditions will be superimposed on another.

It is clear that the strain-rate patterns associated with different cyclic events and their causes are not fully understood at present. Hobbs *et al.* (1986) have assessed the role of work-hardening during crystal–plastic deformation as a source of 'damage' leading to instability generation. However, the large range of other processes which can lead to damage build-up noted in the earlier section on fracture processes and in Fig. 6 also need assessment. In addition, the different patterns of cyclic deformation associated with fault propagation (i.e. tip zone processes), fault jogs (Sibson 1985), fault bends (Knipe 1985) and the decay periods when fault activity either changes to a new location or dies at the end of large-scale tectonic events all require consideration. One consequence of the decay in the tectonic event (either as it migrates or as the driving force for the tectonic process itself decays) is that different points (i.e. specimens) which become involved in the deformation at different times or at different distances from the source area will experience strain waves or pulses with different characteristics. For example, Knipe (in press) argues that the ductile deformation found along high level, low temperature thrust faults at the base of the Moine Thrust zone represent the last, low-amplitude strain-rate events (i.e. the last thrust displacements) associated with the evolution of the Moine Thrust zone in N.W. Scotland. Knipe (in press) also notes that the ductile deformation features located at fault tips (e.g. folds at the tip zones of thrust faults) may not be characteristic of the fault propagation structures but represent 'death sequence' structures related to the lower strain rates associated with the end of the deformation cycle(s) located on that fault.

The brief assessment of cyclic deformation events presented also emphasizes the hierarchic nature of these events which can be considered at different levels of detail. That is, *large-scale tectonic cycles* (e.g. orogenic events) can be viewed as composed of an average strain rate cycle which provides the framework of more detailed analysis of deformation cycles associated with the strain accommodation patterns (e.g. faulting sequences and distribution of displacements) within the area effected by the large-scale tectonic cycle. Considered in even more detail, each tectonic cycle associated with the concentration of deformation into one fault zone is characterized by a range of *rheological cycles* characterized by changes in the strain-rate pattern and mode of failure induced by the mechanical behaviour.

The complications in microstructural evolution discussed above in no way reduce the value or potential of future microstructural studies of natural rocks. The additional questions which they raise are all aspects which future microstructural studies can help resolve. For example, what is the base level of strain rate between transient events on faults at different depths in different tectonic situations? How rapidly does the stress or strain-rate drop after a deformation event? How often do they occur and what volume of material is affected by the transient events? What is clear is that future studies require assessment of microstructural stability and analysis of the pressure–temperature–time (*PTt*) path experienced by the specimens studied, as discussed below.

### *Microstructural stability*

Recognition that strain-rate changes are likely during tectonic events, and that different microstructures will be generated during different parts of the event, necessitates consideration of how stable microstructures are during their subsequent history. Identification of microstructures which are sensitive enough to alter during deformation is important to the quantification of deformation conditions. On the other hand, we require the microstructures to be robust enough to survive for collection and analysis. This is particularly true of the microstructures generated during crystal–plastic processes such as the dislocation densities, the sub-grain size and the size of new recrystallized grains. It is generally assumed that such microstructures are preserved intact during their subsequent history. A more realistic view point is that they may be modified and then frozen in at some time, due to a decrease in the rate at which they change. Consideration of microstructural stability is therefore crucial to future studies in the field of palaeostress assessment and for assessing the details of the deformation cycles described above.

The susceptibility of deformation microstructures to modification arises primarily from the increased free energies stored in the form of elastic distortion of the crystal structure and the energy associated with defects such as dislocations grain boundaries and interfaces. The 'driving forces' inducing microstructural modifications relate to the magnitude of the changes in the chemical, strain and interfacial free energies which accompany the alterations. Although all of these three are linked, it is usually possible to identify one which dominates the particular process. The changes in the chemical free energy which are associated with chemical changes are often $\sim 10^3$ higher than those associated with strain or interfacial energy changes (Reitan 1977, Wintsch 1985, Wintsch & Dunning 1985, Wheeler 1987). Despite these differences there is no doubt that the latter driving forces can control microstructural changes.

There are a number of processes which can be involved in the modification or preservation of deformation microstructures. These include: (a) conversion of elastic distortion of the crystal to dislocation or twin arrays, (b) reorganization and reduction of dislocations by glide, climb and cross-slip to lower energy configurations (e.g. tangles become cells which become sub-grains; see Barber 1985, Poirier 1985, White 1985), and (c) migration of grain boundaries (low angle and high angle) which sweep away internal substructures in sub-grains or grains (Tullis & Yund 1982, Joesten 1983, Grest *et al.* 1985, Olgaard & Evans 1986).

The processes which will aid the preservation of microstructures include the segregation of solutes to form defect clusters, dislocation clouds or grain-boundary segregations or precipitates. Such segregations may arise directly from diffusion processes or may be related to the collection of impurities during dislocation movement or grain-boundary migration. Whatever their origin, these segregations will affect the mobility of dislocations and grain boundaries and often cause pinning (Kasen 1983, Grant *et al.* 1984) and thus preservation of these features. The rapid removal of the environmental conditions which induce microstructural changes (i.e. rapid stress or temperature drops) will also enhance preservation of microstructures. One of the critical factors in microstructural stability is the rate at which the preservation processes can take place. Two critical factors in the preservation of natural deformation microstructures are therefore the rate of stress drop at the end of the deformation event and the temperature history which post-dates the deformation. The second of these factors relates to the *PTt* path and is dealt with later in the paper. As an example of the role of the rate of change of the stress (stress rate), a simple estimate of the stability of the sub-grain structure is considered here.

Sub-grain size has been used extensively to estimate the stress levels operating in mylonite zones (see reviews in Ord & Christie 1984). The assumption is that sub-grain size adjusts to the deformation flow stress level and is not altered after the deformation event. One way of preserving the sub-grain size is if at the end of the deformation event the stress level drops fast enough so that the adjustment of the sub-grain to the 'equilibrium' size is not possible. By assuming that the sub-grain size adjustment is achieved by dislocation climb, it is possible to estimate the rate of sub-grain size change possible either between transient deformation events or at the end of a deformation event. Figure 10 presents the results of such an estimate using the equation for dislocation climb velocity given by Poirier (1985, p.62). The equation relating the flow stress and the sub-grain size used is that given by Ord & Christie (1984). Figure 10(a) illustrates the stress–time path needed to allow the sub-grain size to maintain its equilibrium size as the stress drops at different temperatures for quartz. One implication of the results of the estimate shown is that the preservation of sub-grain microstructures in mylonites deformed at 400°C and at 50 MPa requires that the stress drop is greater than 1 MPa per 2000 years while preservation from deformation at 200°C and 50 MPa

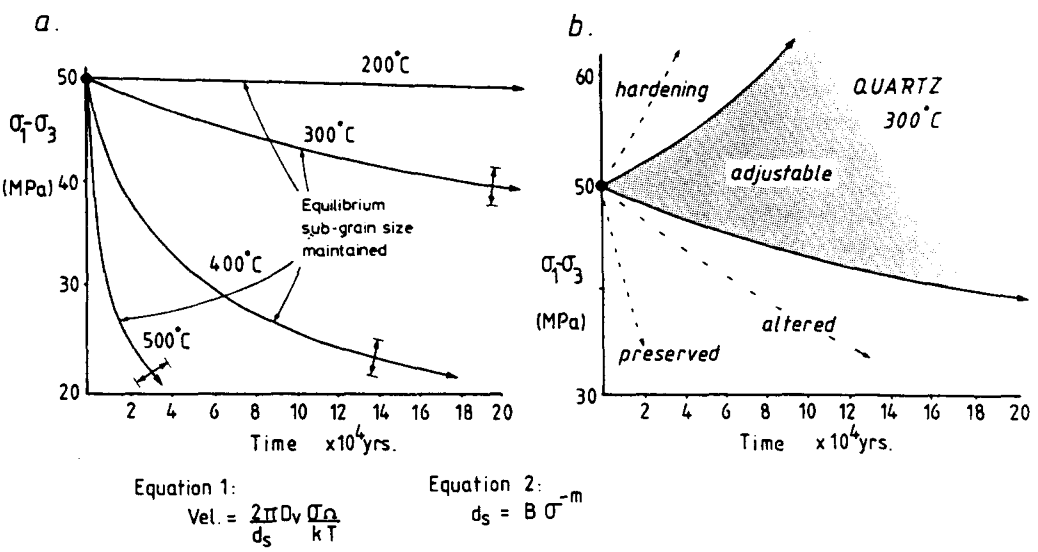


Fig. 10. (a) Differential stress vs time paths needed for sub-grain size to maintain an equilibrium size (given by equation 2) assuming the rate of sub-grain size change is controlled by the velocity of dislocation climb (equation 1). The rate of sub-grain boundary migration is lower at lower temperatures, thus necessitating a slower stress change to maintain the equilibrium size. (b) Influence of differential stress histories on the preservation or stability of sub-grain microstructures after deformation at 50 MPa and 300°C based on data used in (a). The stress path needed to maintain the sub-grain size in equilibrium with a stress change is shown for both an increase and a decrease in stress. If the stress increases too rapidly, hardening may occur, while if the stress drop is rapid, the size may be preserved unchanged. The values of the variables used in the construction of the graphs are the same as those listed by Rutter (1976).

requires a stress drop faster than 1 MPa per 12,000 years. Figure 10(b) illustrates in more detail the sub-grain size changes in quartz at 300°C. In this case the range of stress rates which allow adjustment, modification or preservation are shown. Note also that the diagram illustrates the conditions of stress increase where the stress build up is too fast for the equilibrium sub-grain size to develop. Such a situation may cause work-hardening and the generation of creep-damage leading to failure or the initiation of an instability. Although the estimates presented above are only approximate, they do provide an insight into the rate of change of microstructural features and identify the need for future experimental programmes and modelling to address this problem. If the accuracy of palaeostress estimates and the assessment of the effects of repeated deformation events on microstructural evolution and preservation are to be improved, then the stability of the dislocation density also requires assessment as does the kinetics of grain-boundary migration.

It is commonly assumed that a test of the validity of these palaeostress indicators is obtaining an identical stress estimate from each indicator (dislocation density, sub-grain size and recrystallized grain size) recorded from a single specimen. However this should only be expected if the rate of stress decrease is fast enough to prevent modification of all these indicators. Where the stress drop is slow then the freezing in of the different indicators will be at different stress levels and result in a range of stress levels being indicated from one sample. The range of values will be related to the rate of stress or strain rate decrease (Fig. 11). Therefore, if future studies can improve the understanding of the rates at which these microstructural changes take place, it may be possible to use the range of stress levels indicated by the different indicators to quantify to some extent the stress histories on ductile faults.

*Integration of deformation mechanisms and* PTt *path assessment*

The need to integrate detailed metamorphic studies and tectonic events has been gaining momentum over the last 10 years (Brodie & Rutter 1985, Platt & Lister 1985, Platt 1987, Knipe in press, Handy submitted). However, the likely changes in the patterns of cyclic deformation and instability characteristics at different temperatures, together with the different rates and modes of cooling during uplift, render the integration of *PTt* paths (Thompson & England 1986, Thompson & Ridley 1987) and deformation mechanism path analysis essential to future studies. The different rates of strain rate or stress changes during and after a deformation

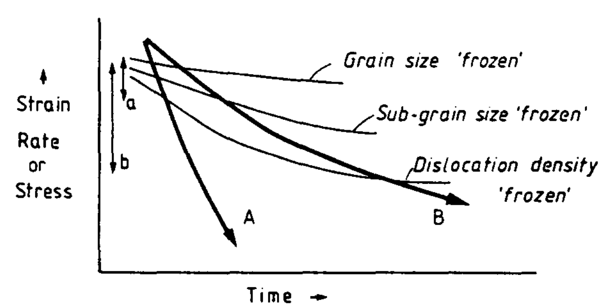


Fig. 11. Influence of the changes in the strain rate or stress at the end of a deformation event on the 'freezing' in of palaeostress indicators. Rapid decreases in the strain rate (path A) will promote the chances of each indicator reflecting very similar stress levels (a), while a slower decrease (path B) may result in a larger range of stress levels (b) being indicated from the different microstructures.

event and the temperature history will be important to both the mechanical behaviour of the material and the preservation of microstructures. For example, the time period, temperature and stress levels experienced between deformation events will influence the frequency of deformation events or instabilities developed; and the stress and temperature path at the end of the deformation will influence the preservation of microstructures. Some deformation microstructures induced during progade metamorphism may not be preserved at all.

Another reason for integrating these studies is that different deformation mechanisms are often associated with different parts of a tectonic history and combining analysis of the geometrical evolution of the area with microstructural and metamorphic aspect can allow detailed assessment of the tectonic evolution. For example, in thrust belts individual faults often accommodate deformation for a limited time period before displacement is transferred to a different fault (Boyer & Elliot 1982, Butler 1987). Each fault in the array therefore will record a different part of the thrust zone evolution. Where the new faults are developed in the footwall to the earlier faults, then the deeper level deformation products are uplifted by displacement on later faults. Thus fault rocks developed at depth experience a stress drop as displacement is transferred; followed by a slow decrease in temperature during uplift and exhumation. A recent analysis of this situation by Knipe (in press) of the microstructural evolution of fault rocks associated with the different thrusts in the Moine Thrust zone of N.W. Scotland has allowed the integration of deformation mechanism paths with likely *PTt* paths. The resulting pressure–temperature–strain-rate–strain–time (*PTėet*) paths are different for each thrust

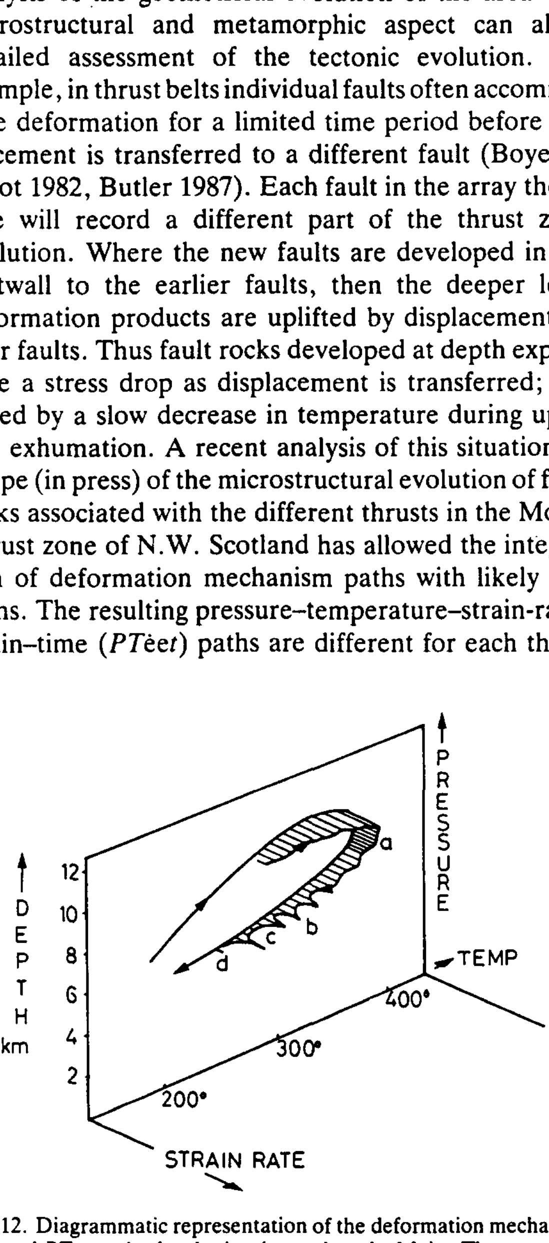


Fig. 12. Diagrammatic representation of the deformation mechanism path and *PTtė* path of mylonites located on the Moine Thrust zone in Scotland. The possible change in the type of strain rate pattern with time is shown. Deformation during the section *a* involves dislocation creep processes during shearing. Section *b* involves alternation of dislocation and fracture processes. The transition from *b* to *c* is associated with a change from shearing during overthrusting to thrust sheet thinning and can be related to the transfer of displacement to lower level thrusts. This transition is associated with a change in the dominant deformation mechanism to DMT. Section *c* to *d* is dominated by DMT and occasional fracture events, which may be related to movement on underlying thrusts.

sheet and reflect the range of conditions which characterize deformation in each. The *PTtė* path for one of these sheets, the mylonitic rocks located on the Moine Thrust itself, is reproduced in Fig. 12. The history involves a change from plastic shearing by dislocation creep to an alternation of fracturing and localization of shearing into selected domains. This history can be interpreted as indicating a transition from stable to unstable flow as the mylonite moves along a decreasing temperature path. In addition, there is a progressive transfer from a shearing deformation to a vertical shortening, the final stages of which are achieved by diffusive mass transfer. This change indicates that a decrease in the strain rate and a change from shearing to thrust sheet thinning, marks the transfer of displacement to the other underlying faults. This study also provided an example of the effect of temperature history on microstructural stability in cataclastic gouges. The gouges studied from the different thrust sheets experienced cooling from different initial temperatures and their microstructures reflect these different annealing histories. Those gouges developed at the highest temperatures are now composed of an aggregate of strain-free hexagonal grains while those cooled from lower temperatures preserve progressively more complex microstructures composed of crude dislocation cells and high dislocation tangles (see Knipe in press for details).

## CONCLUDING STATEMENT

The review presented above highlights four research directions which are crucial to the future understanding of microstructures preserved in tectonites. The first is the identification of fracture mechanisms in natural fault zones. The second is the detailed assessment of the characteristics and effects of cyclic deformation events or flow instabilities. The third is the need to quantify the stability of deformation microstructures and their susceptibility to modification during their subsequent history. The fourth is the need to integrate analysis of deformation mechanism paths with the geometrical evolution of structures and to incorporate these into *PTt* or burial–uplift histories.

The review has illustrated the important role which microstructural studies have in the understanding and quantification of tectonic processes. As Sorby (1858) wrote "... some geologists may perhaps be disposed to question the value of the facts I have described, and to think the objects so minute as to be quite beneath their notice, and quite inadmissible, ... however ... though the objects I have described are minute, the conclusions to be derived from the facts are great".

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