Journal of Structural Geology, Vol. 6, No. 1/2, pp. 51 to 61, 1984.
Printed in Great Britain

0191–8141/84 $03.00+0.00

Brittle to ductile transition due to large strains along the White Rock thrust, Wind River mountains, Wyoming

GAUTAM MITRA

Department of Geological Sciences, University of Rochester, Rochester, NY 14627, U.S.A.

(*Accepted in revised form* 28 *June* 1983)

Abstract—The deformed foreland of the Rocky Mountain Cordilleras is characterized by major reverse faults that bring Precambrian basement rocks to the surface. The White Rock thrust in the Wind River mountains of Wyoming is one such fault that developed at fairly shallow levels in the crust. Detailed study of grain-size variations and textures within the fault zone provide a framework for discussion of the mechanics of movement along the fault. Textural criteria are suggested for distinguishing between dominantly brittle and dominantly ductile deformation. A transition from brittle fracturing to ductile deformation is seen with increasing strain and resultant decrease in grain-size along the fault zone. Diffusional creep in the final stages of deformation leads to strain softening and may allow large strains to take place along narrow zones at low deviatoric stresses. This may be one way of obtaining large displacements along major faults at shallow levels in the crust.

INTRODUCTION

A TRANSITION from brittle to ductile mode of deformation in rocks with changing conditions of temperature and pressure has long been recognized (Byerlee 1968). Experimental work on granites (Tullis & Yund 1977, 1980) has helped to define the brittle–ductile transition in these rocks as a function of pressure, temperature and fluid content. Such work is useful in understanding the mechanical behaviour of rocks in the upper crust, which are largely granitic in composition.

In naturally deformed granitic basement rocks, the brittle–ductile transition is more difficult to recognize, since the macroscopic behaviour of the rock is not observable at the time of deformation. Microscopic criteria suggested here may be used to recognize the brittle–ductile transition.

The brittle–ductile transition signifies a change in the dominant deformation mechanisms in a rock from fracturing to cataclastic flow or to a diffusion or dislocation controlled mechanism. In this sense, it is similar to any other change in deformation mechanisms (e.g. dislocation glide to dislocation creep) observed in rocks undergoing deformation under varying physical conditions. Such a change in deformation mechanisms is controlled by numerous factors such as pressure, temperature, fluid content, rock composition, as well as strain rate, strain, and grain size. Here, I will discuss a transition from brittle to ductile deformation along a fault, caused by large strains and the resultant reduction in grain size. In many ways, such a transition in deformation mechanisms is similar to changes in deformation mechanisms due to grain size reduction seen along other faults, such as the Glarus thrust (Schmid 1975, 1976). Although the actual deformation mechanisms involved are different, due to different pressure–temperature conditions in the two cases, both show work softening along the fault due to change to a new deformation mechanism resulting from a reduction in grain size.

REGIONAL GEOLOGY

The Wind River mountains of Wyoming are part of the deformed foreland of the Rocky Mountain Cordilleras (Fig. 1). The deformed foreland is characterized by large blocks of Precambrian crystalline basement that were brought to the surface along large reverse faults during the late Cretaceous to early Tertiary Laramide orogeny.

The major Laramide structure exposed in the Wind River mountains is a large, doubly plunging anticline that plunges to the north at the northern end and toward the southeast at the southern end (Fig. 1). The fold involves Palaeozoic and Mesozoic beds, and is cored by Precambrian crystalline basement made up of granitic gneisses and intrusive granitoids which are all older than 2700 Ma. Along the western flank of the anticline the Precambrian basement rocks are thrust up against Palaeozoic–Mesozoic and early Tertiary sediments of the Green River basin, as seen on the down-plunge projection of the northern end of the anticline (Fig. 2a). This major fault, the Wind River thrust, has a minimum displacement of 21 km at the southern end of the structure, and can be traced to depths of over 20 km on seismic sections (Zawislak & Smithson 1981). The fault has a somewhat smaller minimum displacement (~15 km) at the northern end of the anticline; the movement here is taken up along some smaller faults, such as the White Rock thrust, which are possible imbricates of the Wind River thrust (Fig. 2a).

The Precambrian basement rocks show evidence for repeated deformation from Precambrian through early Cenozoic times (Mitra & Frost 1981), with networks of several generations of ductile and brittle deformation zones. Laramide deformation zones and faults were the last-formed tectonic structures in these rocks, and they formed concurrently with the Wind River anticline and the Wind River thrust during late Cretaceous to early Tertiary times. These deformation zones and small faults

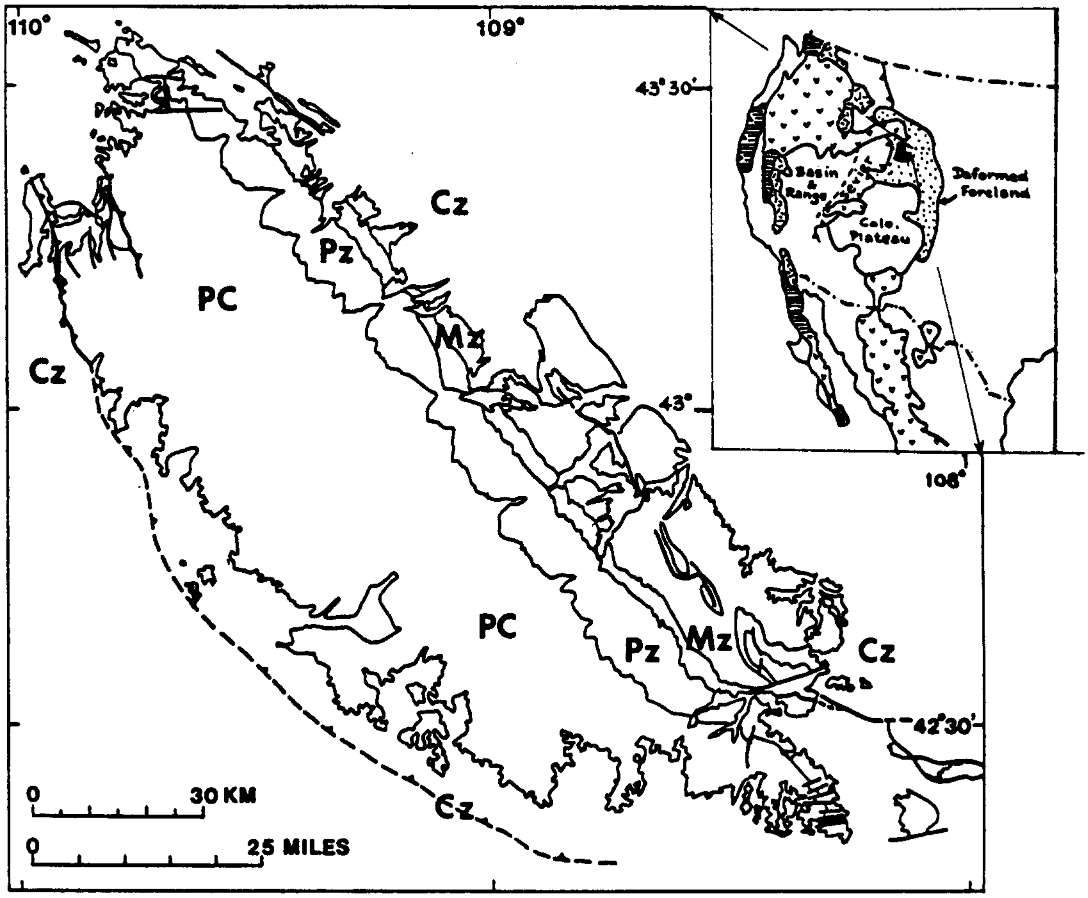

Fig. 1. Generalized geologic map of the Wind River mountains showing the folded Palaeozoic–Mesozoic succession, and the major faults. Inset map shows the location of the Wind River mountains within the deformed foreland of the Rocky Mountain Cordilleras. Symbols used are PC, Precambrian crystalline basement rocks; Pz, Palaeozoic sedimentary rocks; Mz, Mesozoic sedimentary rocks; Cz, Cenozoic sedimentary rocks.

form a pervasive and intricate network through the basement rocks, and must represent basement shortening corresponding to shortening represented by folding in the cover sediments. The pressure–temperature conditions of formation of these structures can be estimated by unfolding and unfaulting the Wind River anticline to determine the original depths of present-day surface rocks at the start of Laramide deformation (Fig. 2b).

CHARACTERISTICS OF BRITTLE AND DUCTILE DEFORMATION

The brittle–ductile transition may often be recognized in experiments by the macroscopic behaviour of the specimen being studied, such as a sudden drop in stress due to failure along a brittle fracture. This is usually not possible in naturally deformed rocks, since the deformation took place some time in the geologic past. It is, therefore, useful to recognize textures in thin-sections of the deformed rocks that will allow one to distinguish between brittle and ductile deformation. In using such distinguishing criteria one must keep in mind that individual minerals in a rock have different physical properties and may behave differently from one another under the same physical conditions. Hence, an overall ductile deformation of a rock may be accompanied by brittle deformation of an individual mineral (Mitra 1978, Boullier 1980). A careful study of fracturing in thin-sections shows two main recognizable types of fractures, making it possible to distinguish between dominantly brittle and dominantly ductile deformation.

Brittle deformation is characterized by unstable fractures (Lawn & Wilshaw 1975, Paterson 1978) that propagate spontaneously across grains and grain boundaries. They are observed as continuous fractures cutting across grains of different compositions and orientations (Fig. 3a). The grains appear to be otherwise undeformed since there is little or no plastic deformation within the grains prior to brittle fracturing. Under certain conditions, plastic deformation may occur in the immediate vicinity of the fracture tip as the fracture propagates; this may give rise to a narrow zone of plastic deformation immediately adjacent to the fracture. Individual fractures may originate as shear fractures or extensional fractures, and may show evidence for shearing and/or extensional movement on them after they have formed. Certain narrow zones of intense deformation are characterized by a series of closely spaced, unstable fractures. I will refer to these zones as Brittle Deformation Zones (BDZs) (Mitra & Frost 1981, Mitra 1980, in preparation); such a general term is preferred since the zones may have components of both shearing and extensional movement on them.

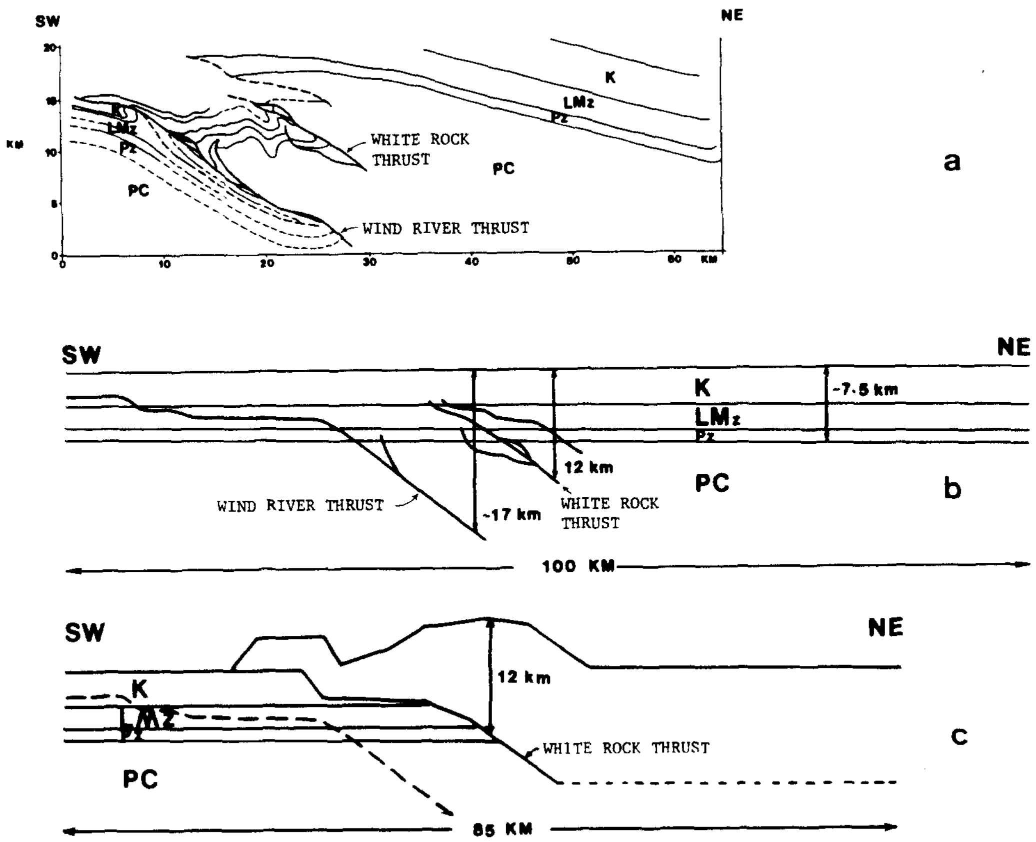

Fig. 2. (a) Down-plunge projection of structures at the northern end of the Wind River mountains, showing relationships between major folds and thrusts. (b) Restored cross-section showing original depths of present-day surface rocks along the Wind River thrust and the White Rock thrust, and on the east limb of the Wind River anticline. (c) Structural model of the Wind River mountains immediately after movement on the White Rock thrust, drawn using a simple kink-fold geometry. If thrusting took place significantly faster than erosion, present day surface rocks would have been overlain by a 10–12 km thick thrust sheet throughout the thrusting episode. Symbols used are: PC, Precambrian crystalline basement rocks; Pz, Palaeozoic sedimentary rocks; LMz, Lower Mesozoic sedimentary rocks; K, Cretaceous sedimentary rocks.

Ductile deformation, close to the brittle–ductile transition regime, is characterized by stable fracturing of grains (Lawn & Wilshaw 1975, Paterson 1978). Stable fractures die out within individual grains, or are stopped at grain-boundaries where the stresses at the fracture tip are dissipated by plastic deformation in adjacent grains (Fig. 3b). Thus, no continuous fractures are seen in the rock. Many of the more ductile grains in the rock may show evidence for diffusion or dislocation controlled deformation, such as pressure solution, undulose extinction, deformation twinning and deformation bands. Stable fracturing leads to grain size reduction in zones of intense deformation giving rise to mylonitic rocks, with a strong foliation produced by diffusion or dislocation controlled processes in the fine-grained material; this process is only one of the ways in which mylonitic rocks may be produced, and is characteristic of deformation at lower greenschist-grade conditions close to the brittle–ductile transition. I will refer to such zones of intense deformation as Ductile Deformation Zones (DDZs) (Mitra 1978): once again, such a general term is preferred since the overall motion on such a zone may have components of shearing parallel to the walls of the zone as well as extension or shortening perpendicular to the walls.

Massive quartzo-feldspathic rocks making up crystalline basement generally deform very inhomogeneously, with much of the deformation being concentrated along narrow zones of intense deformation. These zones may be BDZs or DDZs depending on the conditions of deformation. Transition from the brittle to the ductile regime may be controlled by various factors such as pressure, temperature, strain rate, the mineralogy of the rock, and the presence or absence of fluids and how these fluids change the mineralogy during deformation. The transition may also be controlled by the grain-size of the rock and by the total strain. Depending on the dominant deformation mechanisms that are active during deformation, increase in strain may be accompanied by changes in microstructure even at large strains (Schmid 1982). One of the most obvious microstructural changes is a reduction in grain-size in zones of mylonites (ductile) and cataclasites (brittle), which may lead to a change in the dominant deformation mechanisms.

THE WHITE ROCK THRUST

The White Rock thrust, a possible imbricate of the Wind River thrust, is a major reverse fault at the northern end of the Wind River mountains (Figs. 1, 2a and 4). The fault zone appears to have had a fairly long and complicated history, as suggested by the presence of fragments of older mylonites in the fault gouge along parts of the fault. I will concentrate here on the deformation features produced only during the last phase of deformation, namely the Laramide orogeny. Based on the offset of Palaeozoic and Mesozoic beds, the fault has about 8 km of displacement, and places Precambrian granitic basement over overturned Palaeozoic and Mesozoic sediments. The restored cross-section (Fig. 2b) suggests that the basement rocks exposed along the fault had an overburden of 10–12 km during Laramide deformation. Motion along the fault and later erosion of the sedimentary cover have exposed these rocks at the surface. Thus, at the time of deformation, the pressures were about 4 kbar and temperatures were less than 250°C assuming a geothermal gradient of 20°C/km. Hence the rocks were at essentially 'brittle conditions' at the start of deformation, although they were close to brittle-ductile transition conditions as defined by Sibson (1977). During thrusting, if the rate of thrusting is much faster than the rate of erosion, then a 12 km-thick sheet of material is emplaced along the fault; this would maintain pressures and temperatures at approximately the same levels (4 kbar, 250°C) during the entire period of movement on the fault (Fig. 2c). If, on the other hand, erosion kept up with deformation, both pressures and temperatures would decrease during progressive deformation.

Initial Laramide uplift in the Wind River mountains took place in Campanian times, about 72 Ma ago (Dorr *et al.* 1977). The Precambrian crystalline rocks were breached about 52 Ma ago, as evident from deposits in the early Eocene Pass Peak Formation of the Hoback Basin (Dorr *et al.* 1977). Assuming that the movement on the major faults (Wind River thrust and White Rock thrust) took place successively, the average movement rates on the faults were approximately 1 mm/year. If most of the deformation on the White Rock thrust took place in a 10 m thick zone, the average shear strain rate within this zone of concentrated strain would be approximately 3×10^{-12}/sec, that is a principal strain rate of 1.5×10^{-12}/sec. This is an average value over the entire period of movement, and actual strain rates at individual instants of time during deformation could have been significantly higher or lower.

The fault shows a 50 m thick zone of strong deformation (which is usually highly weathered), although the most intense deformation is concentrated in a zone less than 10 m thick. On the basis of detailed studies on other major BDZs in the area (Mitra 1980, in preparation) it is reasonable to assume that the fault zone (which may be treated as a major BDZ) grew in thickness with time and increasing deformation. At the same time, the middle of the zone accumulated progressively larger strains. Hence, the progression of structures from the edges to the middle of the fault zone provides a time sequence of development of structures within the zone. A detailed microscopic study of the samples taken in transects across the fault zone allows the dominant deformation mechanisms responsible for the growth of the fault zone, and any changes in these deformation mechanisms with progressive deformation to be determined.

As indicated by samples from the edges of the fault zone, deformation starts with unstable fracturing in both feldspars and quartz in coarse-grained granitic basement (Fig. 5a). Such fracturing can be described in terms of four possible micromechanisms of crack propagation, following Ashby and others (Ashby *et al.* 1979, Gandhi & Ashby 1980) and Atkinson (1982). Briefly, the four possible mechanisms are as follows.

Cleavage 1. Fracture is controlled by pre-existing cracks, with no plastic deformation, except locally at crack tips.

Cleavage 2. Fracture is controlled by cracks generated by small amounts of plastic strain within grains. Bulk plastic strain is less than 1%. Pile-ups of dislocations at obstacles or grain-boundaries give rise to stress concentrations that cause the cracks to nucleate.

Cleavage 3. Fracture occurs at higher temperatures or in material with smaller grain-size. Up to 10% plastic strain, by crystal plasticity or grain-boundary sliding, precedes fracture. Cracks may grow in a stable manner until their increased length, and higher stresses due to work hardening, cause them to propagate unstably.

Intergranular creep fracture. Cracks nucleated by dislocation creep grow by local diffusion. Linking of voids and cracks eventually leads to fracture after long periods of creep.

The fractures at the edge of the fault zone originate as Cleavage 1 or Cleavage 2 type fractures (Gandhi & Ashby 1980) after little or no plastic deformation within the grains, and may grow by coalescing with one another before propagating unstably under critical stresses or as subcritical cracks under the influence of stress corrosion (Atkinson 1982).

With increasing deformation, a series of closely spaced fractures develop. These fractures usually show small amounts of shearing movement (Fig. 5b). The fractures may anastomose with one another along their lengths producing lenticular fragments (Fig. 5c). Such fragmentation, together with the production of wear fragments as a result of frictional sliding on fractures, results in an overall reduction in grain size (Fig. 5c). Cleavage 1 (Griffith cracking) is a crack-size dependent process, and is, therefore, ultimately controlled by grain-size (Brace 1964); Cleavage 2 is also grain-size dependent (Atkinson 1982). Grain-size reduction results in strain hardening, with higher stresses being required for fracturing to continue by these processes.

If the remote applied stress remains approximately constant, deformation moves into the Cleavage 3 field where large plastic deformation (up to 10% strain) precedes fracturing (Fig. 6). Relatively brittle, coarse grains of quartz and feldspar still form a stress-supporting framework surrounded by a matrix of finer grains which behave in a more ductile manner. The coarse grains may continue to fracture, but these fractures are often stopped at grain-boundaries, due to ductile deformation in the matrix. This type of stable fracturing

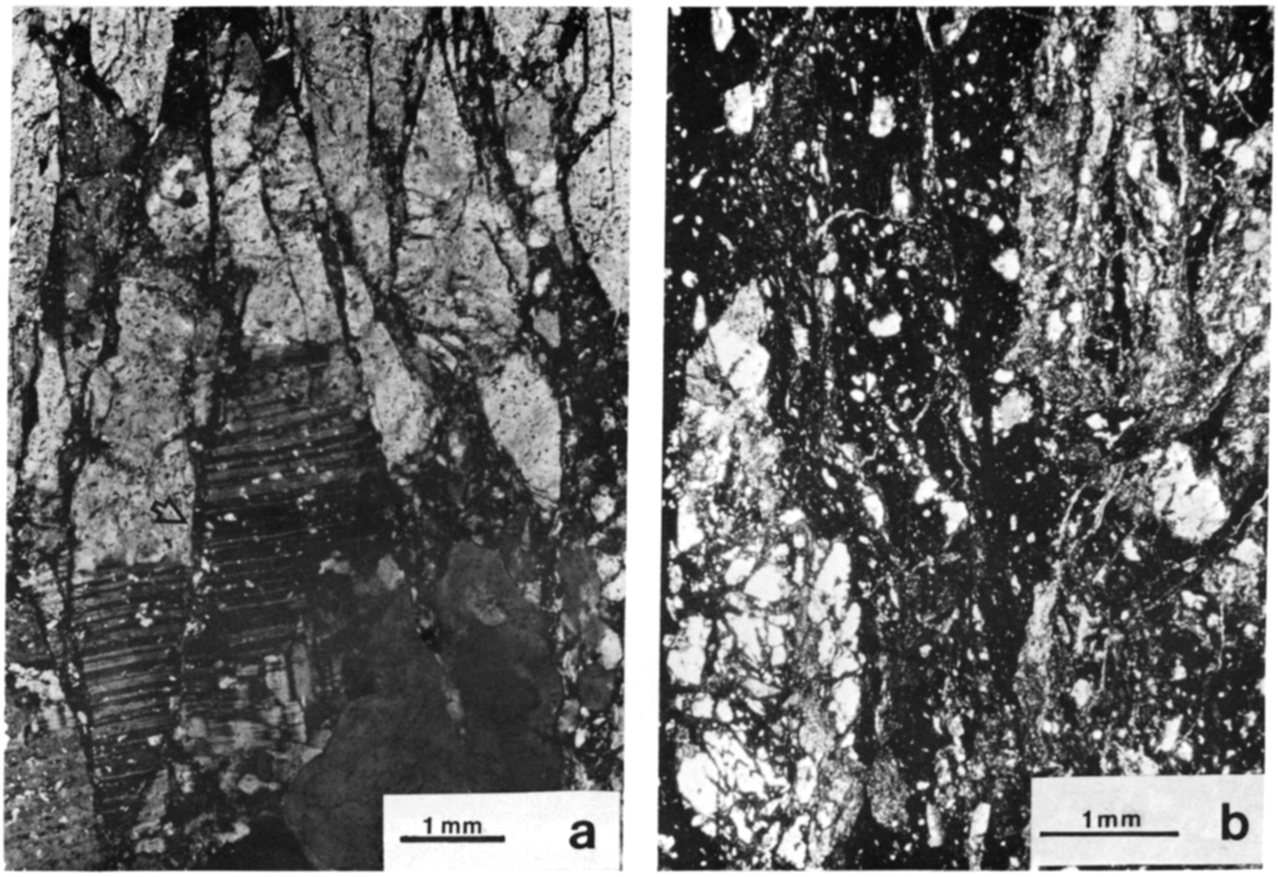

Fig. 3. (a) Brittle deformation zone defined by closely spaced unstable fractures which cut across a polyphase aggregate of randomly oriented grains. Some granulation is present along individual fractures. Offset of a plagioclase grain along one fracture (arrow in lower-centre part of photograph) indicates typical displacement along fractures. (b) Ductile deformation zone characterized by grain-size reduction due to stable fracturing of grains. Individual fractures die out at the grain-boundaries of coarse grains. Chemical alteration of minerals may take place in the fine-grained matrix which forms weak zones in the rock. Some calcite-filled fractures, formed during late stages of deformation, are seen cutting across both matrix and framework grains.

Fig. 4. The White Rock thrust (arrow) at the northern end of the Wind River mountains. The fault places Precambrian crystalline basement rocks over an overturned sequence of Palaeozoic and Mesozoic sediments.

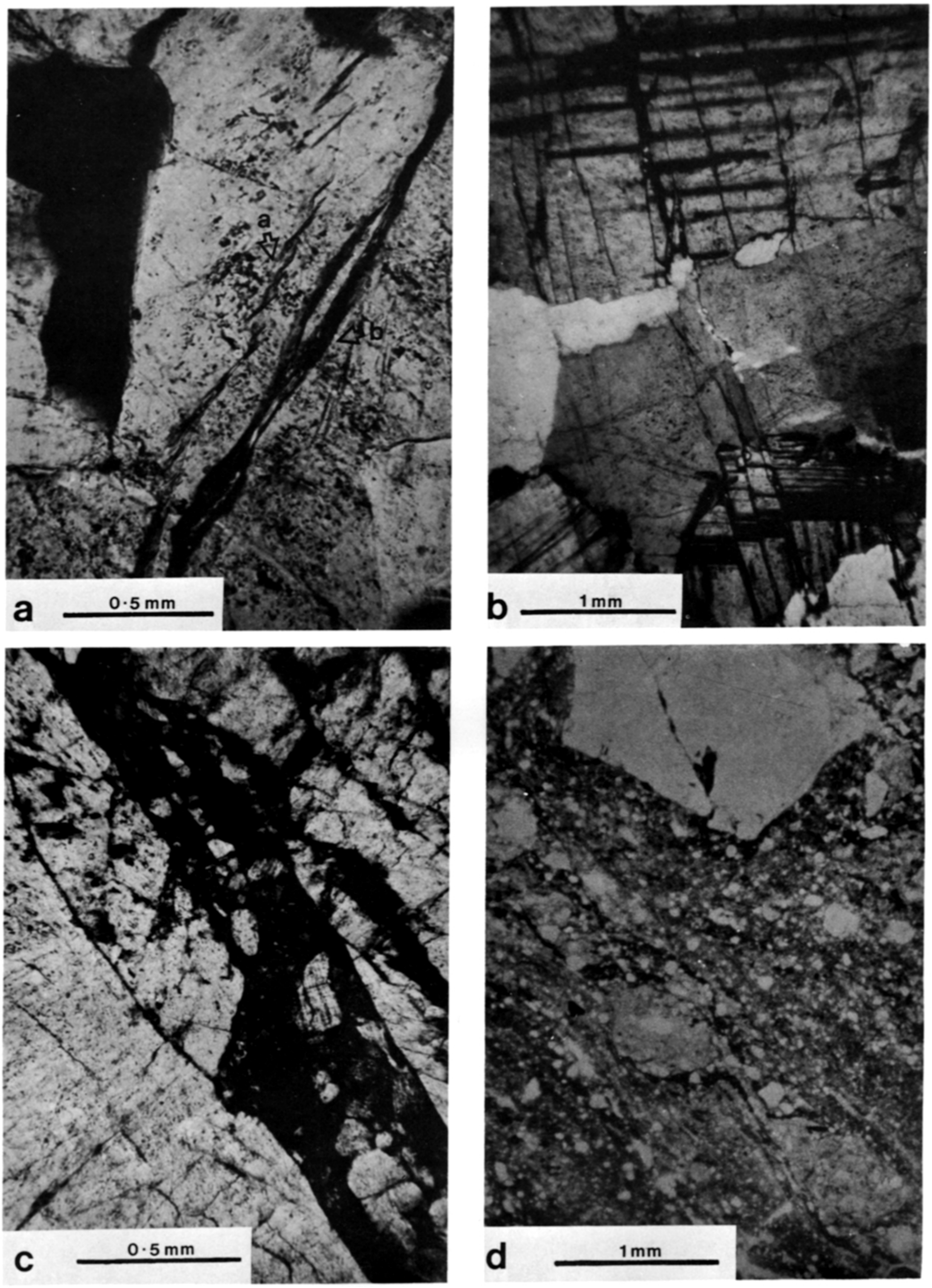

Fig. 5. (a) Coarse feldspar grain showing initiation of Cleavage 1/Cleavage 2 type cracks. Cracks start as en échelon extension fractures (a) parallel to cleavage in the feldspar grain. The small cracks may coalesce to form a larger crack (b) that propagates spontaneously across grain boundaries. (b) Closely spaced unstable fractures cutting across plagioclase and quartz grains. Each fracture has a small shear displacement on it, as indicated by offset albite twin planes in plagioclase grain in lower-centre part of photograph. (c) Anastomosing unstable fractures giving rise to lenticular fragments from a single large K-feldspar grain. Earlier fractures grow into zones of gouge with progressive deformation as wear fragments form due to frictional wear as a result of sliding along the fractures. (d) Narrow zone of intense deformation in the middle of the fault zone. Coarse crystals of quartz and feldspar continue to fracture stably, resulting in grain-size reduction. Pressure solution of finer grained material in the matrix results in concentration of opaque minerals along certain zones giving rise to a crude foliation.

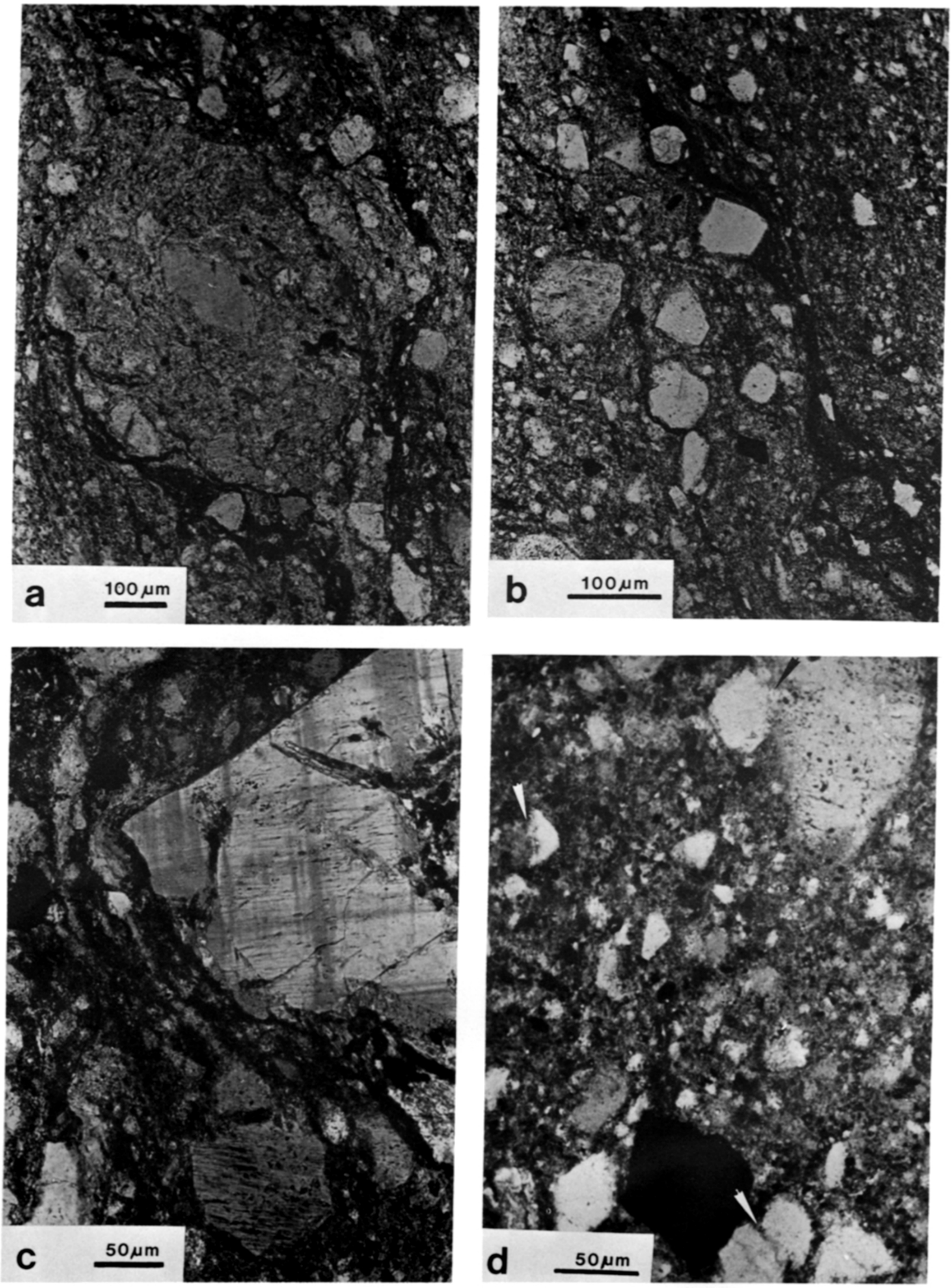

Fig. 9. (a) Fragment of coarse-grained gouge material (mean grain-size >10 μm) in a finer-grained matrix (mean grain size ~5 μm). (b) Extremely fine-grained matrix (~5 μm mean grain-size) made up of quartz and opaque minerals. Pressure solution of quartz results in opaque minerals being concentrated into zones defining a foliation. The matrix shows ductile deformation textures, flowing around more brittle, coarser fragments of quartz and feldspar. (c) Fine-grained ductile matrix 'flowing' around a coarse grain of microcline. Small 'tails' of quartz are precipitated in the pressure shadows of small quartz and feldspar fragments. (d) Small quartz grains (~10 μm in size) show evidence for pressure solution, forming indentations into one another along common grain-boundaries (arrows in upper left, upper centre and lower right of photograph).

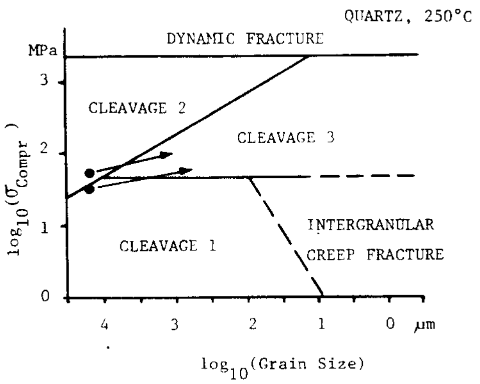

Fig. 6. Generalized stress vs grain-size fracture mechanism map for quartz at 250°C, based on data presented in Atkinson (1982, fig. 1). Approximate positions of various fracture fields are shown. Grain-size reduction during deformation would result in a shift to the Cleavage 3 field from the Cleavage 1 and Cleavage 2 fields, if the stresses remain approximately constant during progressive deformation.

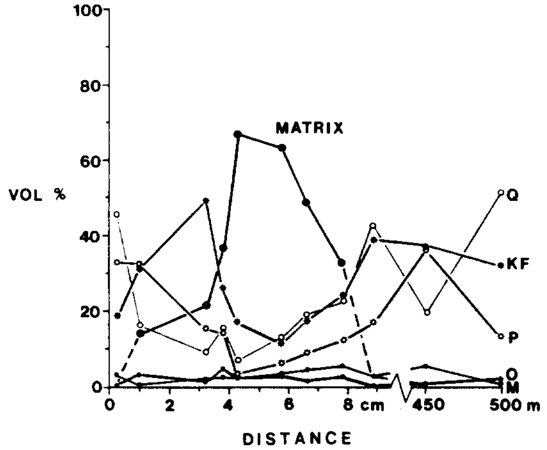

Fig. 7. Variation in volume percentages of various minerals across a typical narrow zone of intense deformation within the White Rock thrust zone. Such narrow zones surround lenticular fragments of less deformed material, allowing these lenses to slide past one another. Volume percentages from two samples outside the fault zone are shown for comparison. 'Ductile matrix' is defined as all material below 10 μm in grain-size. Symbols used are: Q, Quartz; KF, K-feldspar; P, Plagioclase; O, Opaque minerals; M, Micas.

continues to reduce the overall grain-size of the rock, until the large, brittle grains are 'floating' in a ductile matrix. When this situation is reached the behaviour of the matrix begins to control the overall mechanical behaviour of the rock, and this may be achieved with matrix fractions as small as 10% by volume of the rock (Mitra 1978).

The middle of the fault zone may be either a narrow zone of intense deformation, or, more likely, anastomosing narrow zones (less than 10 cm thick) of intense deformation surrounding lenses of somewhat less deformed material. Typical textures in one such narrow zone are shown in Fig. 5(d). Measurements of relative proportions of different minerals across a typical zone of intense deformation indicate that the volume fraction of ductile matrix (for the purposes of this study, this is defined as material made up of grains less than 10 μm in size) increases substantially within such a zone (Fig. 7). Size reduction of coarser fragments may continue by stable fracturing if stress concentrations occur at their boundaries where glide bands in the ductile matrix are blocked (Mitra 1978). These coarser fragments may be either coarse, single crystals of brittle materials (Fig. 5d) or fragments of gouge material that are coarser grained than the matrix (Fig. 9a). Such grain-size reduction is evident from grain-size measurements across such a zone (Fig. 8).

The matrix is made up mainly of quartz and iron oxides, with small amounts of feldspars and micaceous minerals. The grain-size in the matrix is close to the limits of optical resolution. The opaque grains, with their high optical contrast with the surrounding lighter grains, provide the closest measure of this grain-size; the grain-size of the opaque grains varies from < 2 to 15 μm, with a mean grain-size of less than 6 μm, in the most intensely deformed rocks (Fig. 8). The matrix appears to deform in a ductile manner, flowing around brittle fragments (Fig. 9b), and forming 'tails' in the pressure shadows behind large brittle grains (Fig. 9c). Its dominant deformation mechanism is probably some form of diffusional creep accompanied by grain-boundary sliding. Diffusional creep is suggested by evidence for pressure solution along the boundaries of small quartz grains where the grains are indented into one another (Fig. 9d). Grains as large as 20 μm in diameter show evidence for pressure solution indicating that fluids introduced during the late stages of deformation extend the process to larger grain-sizes. Many of the other fine grains in the matrix retain their overall equant shapes even in zones of intense deformation, suggesting considerable accommodation by grain-boundary sliding (Figs. 9c & d).

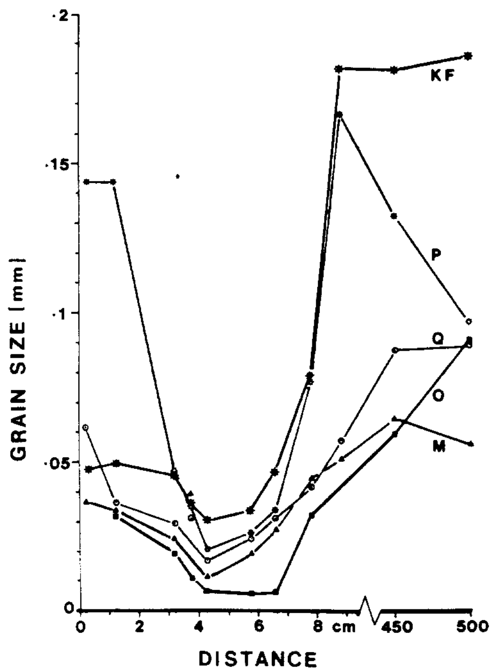

Fig. 8. Variation in mean grain-size of various minerals across a typical narrow zone of intense deformation which forms part of a network surrounding lenses of less-deformed material within the White Rock thrust zone. Grain-sizes from two samples outside the fault zone are shown for comparison. For meanings of symbols refer to Fig. 7.

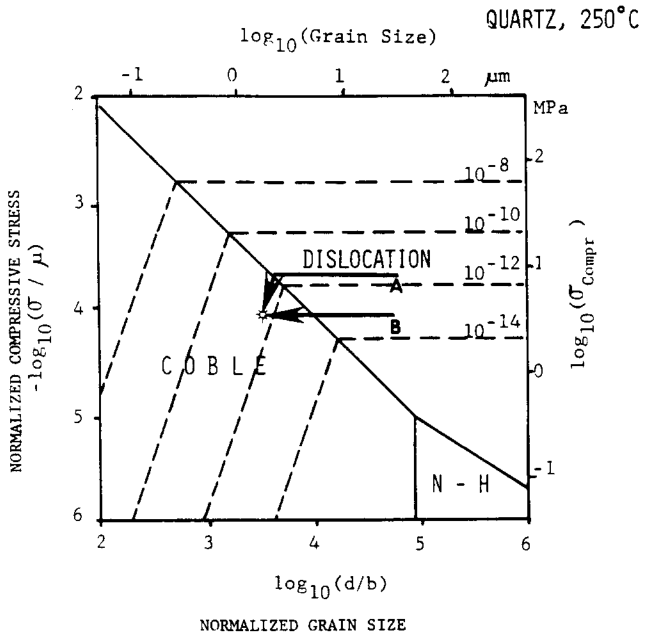

Fig. 10. Stress vs grain-size deformation mechanism map for quartz at 250°C, showing dominant deformation mechanism fields and strain-rate contours. For the smallest, optically measurable grains in the matrix (~2 μm grain-size) deformation takes place in the Coble creep field at average strain rates of 10^{-12}/ s. Two end case deformation paths are shown: (A) constant strain rate; (B) constant stress. Since geologic deformation takes place under compression, compressive stress has been used as one of the axes on the map. The calculations are made with the assumptions that compressive strength is usually 10–20 times larger than the tensile strength.

A deformation mechanism map plotting stress against grain-size (Mohamed & Langdon 1974) allows the deformation history to be followed with decreasing grain-size (Fig. 10). Since quartz makes up the major component of the matrix, and the matrix determines the overall behaviour of the rock during the late stages of deformation, a stress vs grain-size deformation mechanism map was constructed for quartz at 250°C under dry conditions, using data mainly from White (1976), and also from Rutter (1976). Even at these near surface conditions the Coble creep field is very large if the grain-size is sufficiently small. At average strain-rates of 10^{-12}/s along the fault, for a matrix grain-size of 2 μm, deformation would take place by Coble creep at flow stresses of about 3 MPa; at the same strain-rate, grains that are 5 μm in size would deform by dislocation creep at flow stresses around 7 MPa. If the matrix grain-size in the deforming fault-rock was small enough, deformation in the matrix would take place close to the transitional regime between dislocation creep and Coble creep, with the finest-grained material being deformed by diffusion controlled processes even under dry conditions. The actual strain-rate during deformation would also play a very important part in determining the dominant deformation mechanism; at typical matrix grain-sizes (2–10 μm), a shift in the strain-rates by an order of magnitude to slower or faster rates would change the dominant deformation mechanism from the transition regime to either diffusion or dislocation controlled processes.

Introduction of even small amounts of water along the fault during the late stages of deformation would significantly enhance grain-boundary diffusion, resulting in the Coble creep field on the deformation mechanism map (Fig. 10) being replaced by a considerably larger pressure solution field (Rutter 1976). This would allow fairly large quartz grains (~20 μm) to undergo pressure solution during the late stages of deformation (Fig. 9d). Since a deformation mechanism map for quartz has been used to represent a polymineralic aggregate, it only approximately represents the actual behaviour of the rock. The exact positions of the boundaries between the fields is, therefore, not critically important. The map (Fig. 10), however, is instructive because it points out that below a 'critical' grain-size the dominant deformation mechanism along a fault zone may suddenly switch from a grain-size independent process to a grain-size dependent process.

The late stages of the deformation history can be approximated by working backwards with the deformation mechanism map using two end-case assumptions. If the deformation took place at a constant strain-rate of 10^{-12}/s, then there was probably a significant stress drop when the dominant deformation mechanism changed from dislocation creep to Coble creep as a result of reduction in grain-size (path A on Fig. 10). If the deformation took place at a constant flow stress of about 3 MPa, then a significant increase in the strain-rate would occur when the dominant deformation mechanism changed to Coble creep (path B in Fig. 10). In either case, this type of strain softening would localize the deformation to a narrow zone, causing very large strains to accumulate along these zones.

CONCLUSIONS

A detailed study of grain-size and textures of rocks along the White Rock thrust in the Wind River mountains, suggests a history of development of major faults at shallow crustal conditions. Initially, brittle fracturing leads to strain hardening, causing the fault zone to grow wider with progressive deformation as new fractures form in adjacent 'fresh' rocks. With progressive grain-size reduction along the fault zone, there is a transition from dominantly brittle to dominantly ductile deformation mechanisms. In the final stages of deformation, the grain-size is small enough that the dominant deformation mechanism is diffusional creep; this may be either Coble creep in the absence of large amounts of fluids in the rock, or pressure solution (Rutter 1976) if large amounts of fluids are present. Since fault zones are weak zones in the crust, one may expect large amounts of fluids to be channeled through them at some stage during the deformation, and this would favour pressure solution in the rocks. In the case of the White Rock thrust, there appear to have been very little if any fluid present during much of the deformation since most of the feldspars remain almost totally unaltered through the deformation. Small amounts of fluid may have been introduced late in the deformation, resulting in enhancement of pressure solution in quartz grains. Large amounts of fluid were introduced along the fault at very late stages of deformation, giving rise to thick calcite veins and breccias with calcite in the matrix. However, these coarse calcite grains remain almost entirely undeformed suggesting that they were introduced after the end stages of the main deformation along the fault.

Diffusional creep is usually accompanied by strain softening, concentrating the deformation into narrow zones, and allowing very large displacements to take place along these zones at very low deviatoric stresses. Many major faults at shallow levels in the earth's crust show extremely fine-grained gouge, often along very narrow zones. The model above suggests that large amounts of movement along these faults may have taken place by ductile flow, with or without fluids being present along the faults.

Acknowledgements—I am grateful to J. Hull and A. Yonkee for help in the field, and for their helpful discussions and comments on the original manuscript. The critical comments of two anonymous reviewers were very useful and helped in improving the paper. This study was supported by NSF Grants EAR-7927161 and EAR-8207395.

REFERENCES

Ashby, M. F., Gandhi, C. & Taplin, D. M. R. 1979. Fracture-mechanism maps and their construction for F.C.C. metals and alloys. *Acta metall.* **27**, 699–729.

Atkinson, B. K. 1982. Subcritical crack propagation in rocks: theory, experimental results and applications. *J. Struct. Geol.* **4**, 41–56.

Boullier, A. M. 1980. A preliminary study of the behaviour of brittle minerals in a ductile matrix: an example of zircons and feldspars. *J. Struct. Geol.* **2**, 211–217.

Brace, W. F. 1964. Brittle fracture of rocks. In: *State of Stress in the Earth's Crust.* (edited by Judd, W. R.). Elsevier, New York, 111–174.

Byerlee, J. D. 1968. Brittle–ductile transition in rocks. *J. geophys. Res.* **73**, 4741–4750.

Dorr, J. A. Jr., Spearing, D. R. & Steidtmann, J. R. 1977. Deformation and deposition between a foreland uplift and an impinging thrust belt: Hoback basin, Wyoming. *Spec. Pap. geol. Soc. Am.* **177**, 1–82.

Gandhi, C. & Ashby, M. F. 1980. Fracture mechanism maps for materials which cleave: F. C. C., B. C. C. and H. C. P. metals and ceramics. *Acta Metall.* **27**, 1565–1602.

Lawn, B. R. & Wilshaw, T. R. 1975. *Fracture of Brittle Solids.* Cambridge University Press, Cambridge.

Mitra, G. 1978. Ductile deformation zones and mylonites: The mechanical processes involved in the deformation of crystalline basement rocks. *Am. J. Sci.* **278**, 1057–1084.

Mitra, G. 1980. Brittle and ductile deformation zones in granitic basement rocks of the Wind River mountains, Wyoming: a look at the brittle–ductile transition. *Geol. Soc. Am. Progr. with Abs* **12**, 485.

Mitra, G. & Frost, B. R. 1981. Mechanisms of deformation within Laramide and Precambrian deformation zones in basement rocks of the Wind River mountains. *University of Wyoming Cont. Geo.* **19**, 161–173.

Mohamed, F. A. & Langdon, T. G. 1974. Deformation mechanism maps based on grain-size. *Trans. Metall. Soc.* **5**, 2339–2345.

Paterson, M. S. 1978. *Experimental Rock Deformation — The Brittle Field.* Springer, New York.

Rutter, E. H. 1976. The kinetics of rock deformation by pressure solution. *Phil. Trans. R. Soc.* **A283**, 203–219.

Schmid, S. M. 1975. The Glarus overthrust: field evidence and mechanical model. *Eclog. geol. Helv.* **68**, 247–280.

Schmid, S. M. 1976. Rheological evidence for changes in the deformation mechanisms of Solnhofen limestones towards low stresses. *Tectonophysics* **31**, T21–T28.

Schmid, S. M. 1982. A model for strain induced change in deformation mechanism leading to work softening. Unpublished manuscript.

Sibson, R. H. 1977. Fault rocks and fault mechanisms. *J. geol. Soc. Lond.* **133**, 191–213.

Tullis, J. & Yund, R. A. 1977. Experimental deformation of dry Westerly granite. *J. geophys. Res.* **82**, 5705–5718.

Tullis, J. & Yund, R. A. 1980. Hydrolitic weakening of experimentally deformed Westerly granite and Hale albite rock. *J. Struct. Geol.* **2**, 439–451.

White, S. 1976. The effects of strain on the microstructures, fabrics, and deformation mechanisms in quartzites. *Phil. Trans. R. Soc.* **A283**, 69–86.

Zawislak, R. A. & Smithson, S. B. 1981. Problems and interpretation of COCORP deep seismic reflection data, Wind River range, Wyoming. *Geophys.* **46**, 1684–1701.