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Semi-brittle deformation within the Alpine fault zone, New Zealand

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Abstract—This paper reports the results of an attempt to identify the deformation micromechanisms in the brittle–ductile transition zone of the Alpine fault, New Zealand. Characterization of sequentially developed microstructures was carried out using optical microscopy, HVEM and STEM analysis. It was found that the transition zone corresponds to a broad zone of retrogression produced initially by microcracking and fluid infiltration. The non-catastropic nature of the fracturing indicates that fluid-enhanced sub-critical cracking is a significant crustal deformation mechanism. After the initial phase of retrogression, both ductile and brittle processes coexist and cyclically alternate in response to stress cycling. The relative importance of macroscopic brittle processes may decrease with time as more ductile minerals, especially phyllosilicates, develop during continued retrogression.

INTRODUCTION

MODES of deformation within both active and ancient fault zones are widely variable. Sibson (1977) has rationalized these observed variations in deformation and their attendant fault-rock types in terms of the variation in deformation mechanism with crustal level. Brittle deformation is attributed to seismic release in the upper portions of the crust, while non-brittle deformation (hereafter called ductile) occurs at depth due to thermal activation of aseismic flow mechanisms.

Between these end-member rock types, a transitional fault rock is expected to occur which is neither uniquely brittle nor ductile (Sibson 1977). A lack of detailed studies of semi-brittle behaviour in fault zones leaves several aspects of this transition zone to be clarified. These include: (1) identification of the micromechanisms which control the bulk rock deformation; (2) delineation of the relative importance of these micromechanisms at different crustal levels and (3) determination of the degree of interdependence of the active deformation mechanisms. Experimental contributions towards resolving these questions have been made by Heard (1963), Heard & Carter (1968), Tullis & Yund (1977, 1980), Carter & Kirby (1978) and Carter *et al.* (1981).

The above problem can be approached by examining microstructures at different crustal levels within a fault zone on the premise that these will be indicative of the micromechanisms active at that crustal level. This method has been used in a study of the Alpine fault zone (AFZ), New Zealand. Microstructural characterization of fault zone rocks from the transition zone in the AFZ was carried out using optical microscopy, high voltage transmission electron microscopy (HVEM) and scanning transmission electron microscopy (STEM) with X-ray analytical capabilities.

REGIONAL GEOLOGY AND DEFORMATION HISTORY

The Alpine fault is a major transcurrent fault exposed in South Island, New Zealand, separating continental crustal blocks of the Pacific and Indian plates. The basic distribution and structure of fault rocks along the Alpine fault have been described by Sibson *et al.* (1979). Deformation along the fault is concentrated in a 1 km wide zone bounded by quartzo-feldspathic, amphibolite facies schists (Alpine schists) to the east with granitoid basement rock and early Palaeozoic sediments to the west. The eastern boundary of this zone is gradational, with pre-existing structures in the schists progressively transposed into sub-parallelism with the principal mylonitic foliation. The western boundary is defined by the Alpine fault lineament. Changes in plate motion at around 5 Ma altered the then predominantly dextral strike-slip motion, which had caused cumulative displacements on the order of 350 km, to one of oblique slip with a component of thrusting from the east. Continued thrusting of the Pacific plate toward the Indian plate is indicated by the creation of the Southern Alps mountain range for which active uplift rates of 20 mm year^{-1} have been measured (Walcott 1979). Structural analysis (Sibson *et al.*, 1979) within the fault zone indicates that all surface-exposed fault rock fabrics were impressed dur-

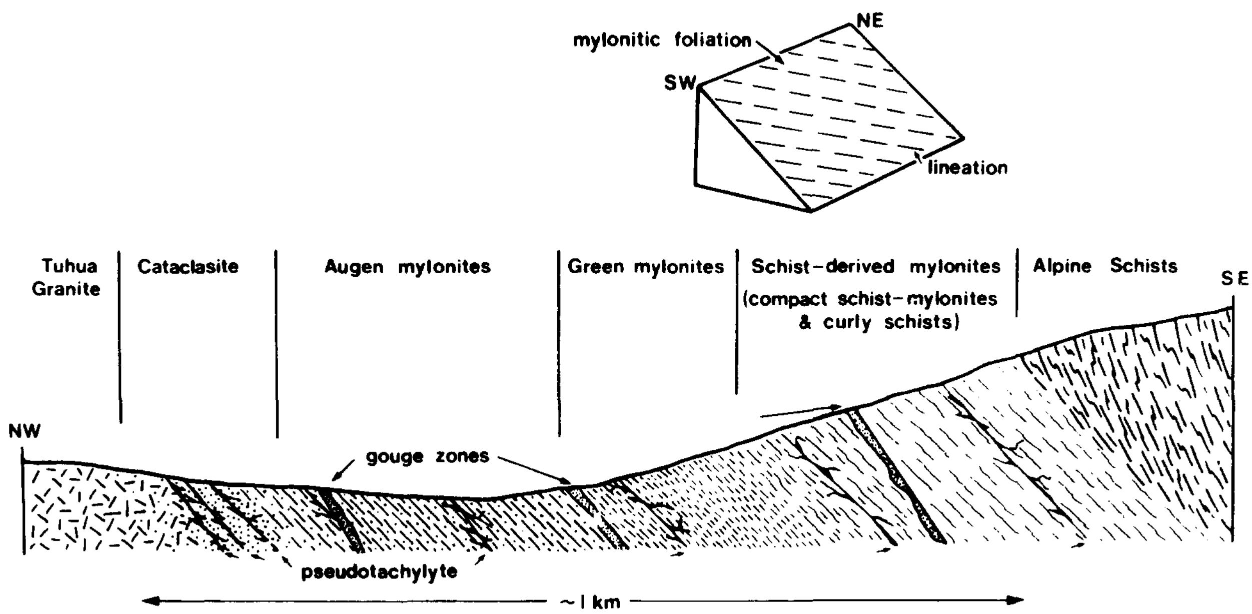

Fig. 1. Cross-section through the Alpine fault zone (after Sibson *et al.* 1979).

ing this still-active phase of oblique reverse faulting along a fault plane dipping approximately 50° southeast.

The wide range of fault rock types exposed within the AFZ make it particularly amenable to an examination of microstructural variations (Fig. 1). The across-strike variation from mylonite to cataclasite is believed to represent a crustal section through the fault zone which preserves deformation features representing different crustal levels (Sibson *et al.* 1979). The AFZ zonation can then be viewed as a reflection of different P–T conditions during deformation. The Alpine schists and schist-derived mylonites represent conditions at the deepest level of the shear zone while cataclasites and gouges along the western margin of the AFZ lineament mark the highest level of deformation. Between the schist-derived mylonite and cataclasite zones bounding the AFZ is a zone, up to 500 m wide, of mixed brittle and ductile deformation, hereafter referred to as the transition zone. Transition zone rocks were classified in the field as either augen or green mylonites following Reed (1964), based on the respective presence or absence of macroscopic feldspar phenocrysts. The augen mylonites are thought to have been derived from granites, and the green mylonites from schists.

METHOD

Successive changes in microstructure will be described sequentially through the transition zone from schist-derived mylonite to cataclasite. Observations are based on samples collected along the Wanganui River and Harold Creek traverses of Sibson *et al.* (1979). These traverses were chosen because they provide a good collection of transition zone rocks, particularly at the schist-derived mylonite/transition zone boundary.

After the completion of optical work, selected areas of the standard petrographic thin sections were chosen for examination by TEM. These areas were thinned by standard ion-beam milling procedures and examined in an AEI EM7 1000 kV electron microscope. Additional observations and identification of mineral phases were carried out using a JEOL 120-CX STEM with an attached energy dispersive X-ray analytical system.

OPTICAL MICROSTRUCTURES

General statement

Studies of regional deformation show a progressive increase in strain and strain rate as one approaches the AFZ from the Alpine schists (Walcott 1979, Sibson *et al.* 1979). Macroscopically this is indicated by the transposition of pre-existing structures in the schists into subparallelism with the mylonite foliation within the AFZ (also see Wood 1978, fig. 5). Microscopically there is a well-defined tectonic reduction in grain size by dynamic recrystallization concurrent with the progression from Alpine schist to schist-derived mylonite (Atkinson *et al.* 1979). Near the transition-zone contact the schist-derived mylonites are characterized by shear bands (White *et al.* 1980) re-orienting an earlier mylonitic fabric. Between the unequivocally ductile schist-derived mylonite and brittle cataclasite assemblages, three aspects of the microstructural development are considered especially important:

(1) continued reduction in grain size within the transition zone rocks as the cataclasites are approached;

(2) large-scale changes in bulk mineralogy from that of the schist-derived mylonites and

(3) the persistence of multiple foliations with consistent geometries.

Grain size reduction

Minerals observed in the schist-derived mylonite, particularly quartz, plagioclase, biotite and muscovite continue to show progressive grain size reduction within the transition zone. Quartz-rich layers exhibit core and mantle structures with recrystallized grains around the edges of the largest grains, while discrete fragments of quartz are also formed by intergranular cracking (Fig. 2a). Microcracks are generally restricted to grain boundaries of newly recrystallized grains. Intergranular cracking is ubiquitously associated with infiltration of secondary minerals along the microcracks. Common infiltration phases are chlorite, calcite and quartz; these secondary minerals exhibit ductile deformation in the form of twins, deformation bands and sub-grains (quartz and calcite) and can themselves be cross-cut by later cracks and infiltration minerals with similar features.

Biotite and muscovite do not indicate as clear a grain-size reduction path. Muscovite grains tend to resist size reduction to a greater degree than biotite, but they represent only a small fraction of the total primary phyllosilicates in the schist-derived mylonite. Biotite grain sizes continue to decrease across the transition zone down to the resolution of the light microscope at which time only very fine-grained masses of foliated phyllosilicates are observed. Grain size reduction in biotite occurs variously by kinking, polygonization and microfracturing, but as will be described, the primary mechanism involves mixed chemico-mechanical recrystallization during hydration reactions (Johnston & White 1983).

Although the complete grain size reduction sequence is not generally preserved, several modes of brittle deformation have been observed at what is considered to be a critical point in the developmental sequence of plagioclase microstructures. Discrete shear on cleavage planes occurs where planes are preferentially oriented in the foliation so as to incur a high resolved shear stress (Fig. 2b). Porphyroclasts (Figs. 2b–d), are also observed with cracks perpendicular to the principal foliation. Crack orientation in these cases is independent of feldspar crystallography; both twins and cleavage are intersected (Fig. 2c). Cracks initiate at the edge of the porphyroclasts and propagate towards the centres of the grains. The degree of penetration by the cracks varies, suggesting a series of relatively slow propagating events as opposed to catastrophic brecciation. Narrow alteration zones of epidote, chlorite and calcite occur along these microfractures (Fig. 2d). Ultimately, by coalescence of several cracks or the preferential propagation of a single fracture, tensile cracks perpendicular to the foliation produce a gross fragmentation of the porphyroclasts. Large grains of secondary minerals, particularly epidote and chlorite develop in the ensuing void space.

Mineralogical changes

The absence of metamorphic retrogression within the schist-derived mylonite has been discussed by several authors (Scholtz *et al.* 1979, Sibson *et al.* 1979, Johnston & White 1983). This presumed isochemical behaviour does not hold for the transition zone rocks and hence, for much of the AFZ. Concurrent with the progressive decrease in grain size across the transitional zone is a marked change in the bulk mineralogy. Lack of continuous outcrop limits this discussion to qualitative observations. The appearance of new minerals is directly linked to the initiation of microcracks in the AFZ. As the intensity of microcracking increases toward the cataclasite zone, so does the modal fraction of secondary chlorite, epidote and calcite. These new phases develop as:

(1) infiltration precipitates along intergranular microcracks (e.g. calcite along quartz grain boundaries);

(2) alteration zones along intracrystalline microcracks (e.g. epidote, calcite and chlorite in plagioclase) and

(3) products of hydration reactions, especially biotite reacting to chlorite.

All secondary minerals in these rocks are compatible with retrogression caused by the introduction of fluids. Secondary mineral generation in concert with continued grain size reduction creates an ultra-fine-grained, foliated chlorite–calcite–epidote–quartz rock; that is, complete mineralogical alteration occurs effectively at the same time as grain size reduction.

The primary consequence of an increase in the modal fraction of phyllosilicates within the transition zone is the creation of a well-defined, domainal foliation parallel to the principal foliation of the AFZ (Fig. 3). In the finest-grained rocks, individual foliation domains are 10–20 μm wide and are defined by alternating phyllosilicate-rich and phyllosilicate-poor layers. Oblique foliations defined by phyllosilicates and discrete shear planes occur antithetically to each other at angles of 5–15° to the main foliation.

TEM MICROSTRUCTURES

Preliminary TEM microstructure observations on schist-derived mylonites have been reported by White in Atkinson *et al.* (1979). Within these rocks, intracrystalline defect structures and dynamic recrystallization are contemporaneous with the most recent phase of mylonite fabric development. Microstructures in the transition zone will be described in terms of the individual minerals in which they occur.

As in the schist-derived mylonites, quartz can show a well-developed sub-grain structure defined by quasi-stable dislocation arrays (Fig. 4a). However, there is now the introduction of a more blocky sub-grain structure with sub-grain walls consisting of hedges of dense dislocation tangles as opposed to more geometrically regular picket-fence type dislocation networks (Fig. 4b). The blocky sub-grain structure commonly shows several orders of sub-grain size, suggesting a de-stabilization of the pre-existing dislocation microstructure. Free dislocation densities within individual quartz grains vary over several orders of magnitude (10^7–10^{10} cm^{-2}) in a very inconsistent manner, although there is a qualitative increase in the frequency of grains with locally high densities as the cataclasites are approached.

Quartz frequently occurs as dislocation-free or slightly dislocated grains with well-delineated crystal faces, both at mica–quartz and quartz–quartz interfaces (Fig. 4c). These features suggest that the grains are not syntectonic recrystallization products, but rather solution-precipitated grains forming in void spaces within the rock.

A striking feature of quartz grains within the transition-zone rocks is the large number of healed intragranular fractures that are not observed in the schist-derived mylonites. These fractures often act as generation sites for dislocation loops (Fig. 4d). Other healed cracks exhibit numerous voids. These are frequently marked by narrow zones of quartz with fewer and straighter dislocations than the rest of the grain linking the healed fractures. Similar dislocation arrays are known in quartz overgrowths (Grant & White 1978). This suggests that the crack was originally open and sealed by quartz infiltration, that is, the crack-seal process (Knipe & White 1979, Ramsay 1980). Preferential etching during sample preparation inhibited detailed examination of intercrystalline fractures. However, in calcite-filled grain boundary cracks, calcite was heavily dislocated and twinned, indicating that the secondary fracture-filling mineral phases had been subjected to later intracrystalline deformation.

Feldspar microstructures were sampled principally from the plagioclase (oligoclase) porphyroclasts described in the previous section, although small fragments and grains from the rock matrix were also examined. The most characteristic feature of the feldspars is cracking on a range of scales. Again, preferential etching of cracks during sample preparation made examination of these fractures by TEM particularly difficult. On this scale fractures are rarely discrete cracks, but rather curvilinear features, often in quasi-parallel arrays (Fig. 5a) or zones of irregular comminution. Alteration or formation of secondary mineral phases along these fractures is common. Intracrystalline defect structures are developed in feldspars throughout the transition zone but are particularly well-developed in the western part of the zone near the cataclasites. Very small cellular structures are developed, some possibily defined by single dislocation loops, while very small recrystallized grains (Fig. 5b) were also observed in these rocks.

Primary phyllosilicates, particularly biotite, show grain size reduction by syntectonic recrystallization and/or cracking. As previously indicated, there is a bulk change in mineralogy across the transition zone, with a progressive increase in the amount of chlorite and epidote. Phyllosilicate reactions are common and generally related to voids within the rock. In Fig. 6(a) biotite grains are overlain by a mass of fine, new chlorite laths in an area with noticeable opening between individual grains. Two foliations (Fig. 6b) can still be observed at high magnifications and can be related to foliation orientations at the optical microscope scale. This emphasizes the pervasiveness of multiple foliations in these semi-brittle rocks.

DISCUSSION

General statement

Fault rocks within the AFZ characterize many of the problems attendant with both descriptive and genetic classification of deformation in large shear zones. Macroscopically, there is a maintenance of material continuity across the schist-derived mylonites and transition-zone rocks until evidence for brecciation can be observed in the cataclasites. However, optical and TEM observations indicate that deformation within the transition zone can be both brittle and ductile. Such mixed-mode deformation is reminiscent of transient creep deformation described by Carter *et al.* (1981). A general feature of the AFZ is the continual trend toward smaller grain sizes, either by dynamic recrystallization or comminution. This trend is independent of the active deformation mechanisms which simply supply a particular path in each faulting environment.

Deformation mechanisms

Microstructures within the transition zone indicate the occurrence of dislocation deformation processes, brittle failure in the form of microcracking and fluid-assisted diffusion processes such as quartz dissolution–precipitation and crack alteration. Active deformation mechanisms are alternating and cyclic as shown in the disruption of intracrystalline creep structures by fractures which are later healed and act as dislocation generation sites.

Dislocation creep features characteristic of deep crustal shear zones have been described from the schist-derived mylonites. These features, particularly the development in quartz of equant sub-grains and dynamically recrystallized grains continue to develop well into the transition zone. A trend toward de-stabilization of these quasi-steady-state creep features is reflected in the development of cell-structures defined by dislocation tangles. Because dislocation climb is thermally activated, a decrease in its effectiveness would lead to such work hardening features and is consistent with decreasing temperature as the transition zone is entered from the east.

Dislocation glide and climb features are not as well-developed in the feldspars until near the cataclasite zone. The very fine cell structures generated in this region could indicate an increase in shear stress in the upper portions of the shear zone according to the relationship $\sigma = \mu(b/d)$, where μ is the shear modulus, b is the Burgers vector and d is the sub-grain diameter (White 1979). This is consistent with White's interpretation (Atkinson *et al.* 1979) of stress magnitudes based on sub-grains and dislocation densities. Before a stress gradient can be verified, temperature and pressure

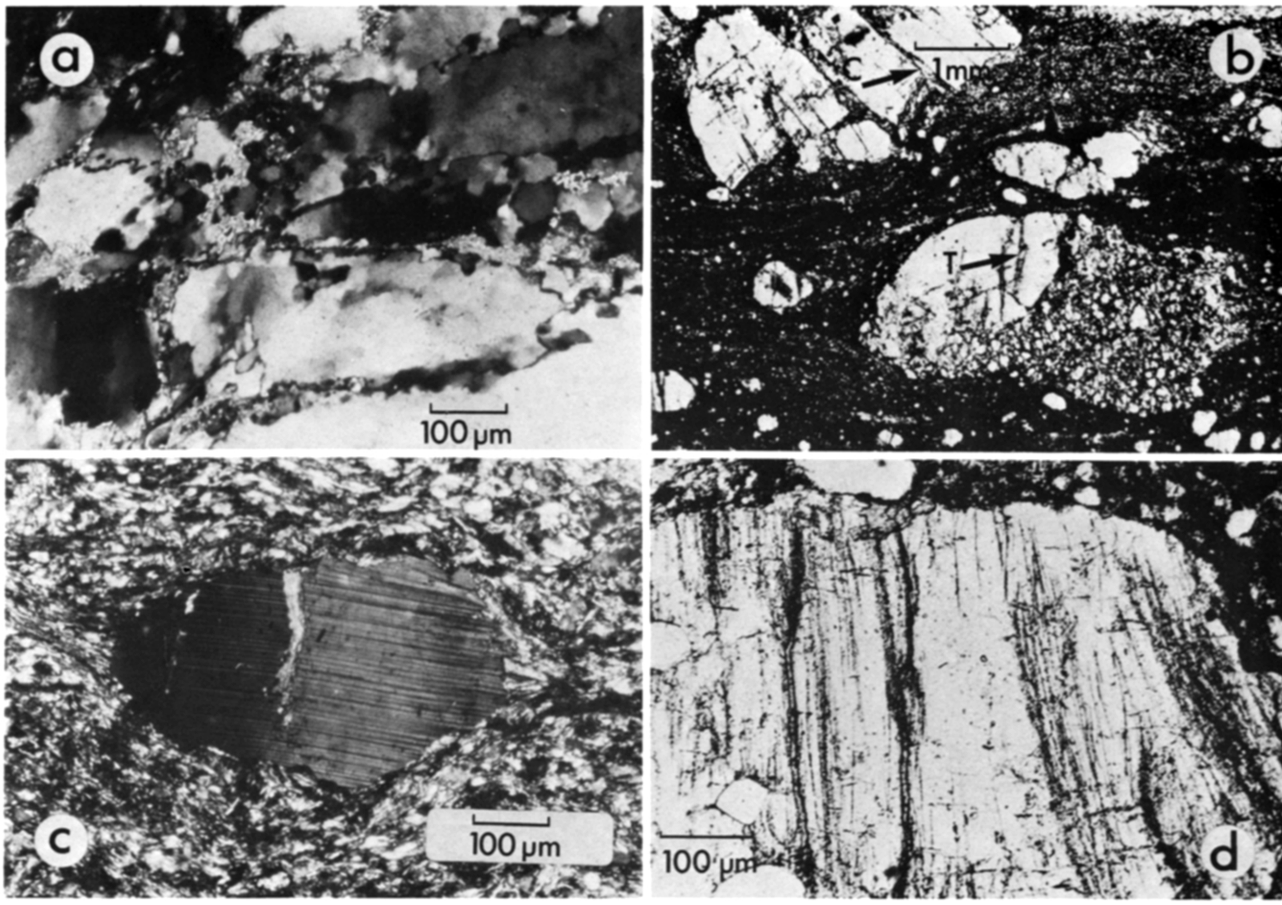

Fig. 2. Optical microstructures in quartz and plagioclase. (a) Grain boundary cracks filled with calcite in a dynamically recrystallized quartz-rich layer. (b) Brittle deformation of oligoclase porphyroclasts by shear along cleavage planes (C) by deformation of foliation—normal cracks (T) and general comminution. (c) Tensile cracks filled with calcite and epidote transecting cleavage in an oligoclase porphyroclast. (d) Foliation—normal cracks in oligoclase with alteration selvages of calcite, chlorite and epidote.

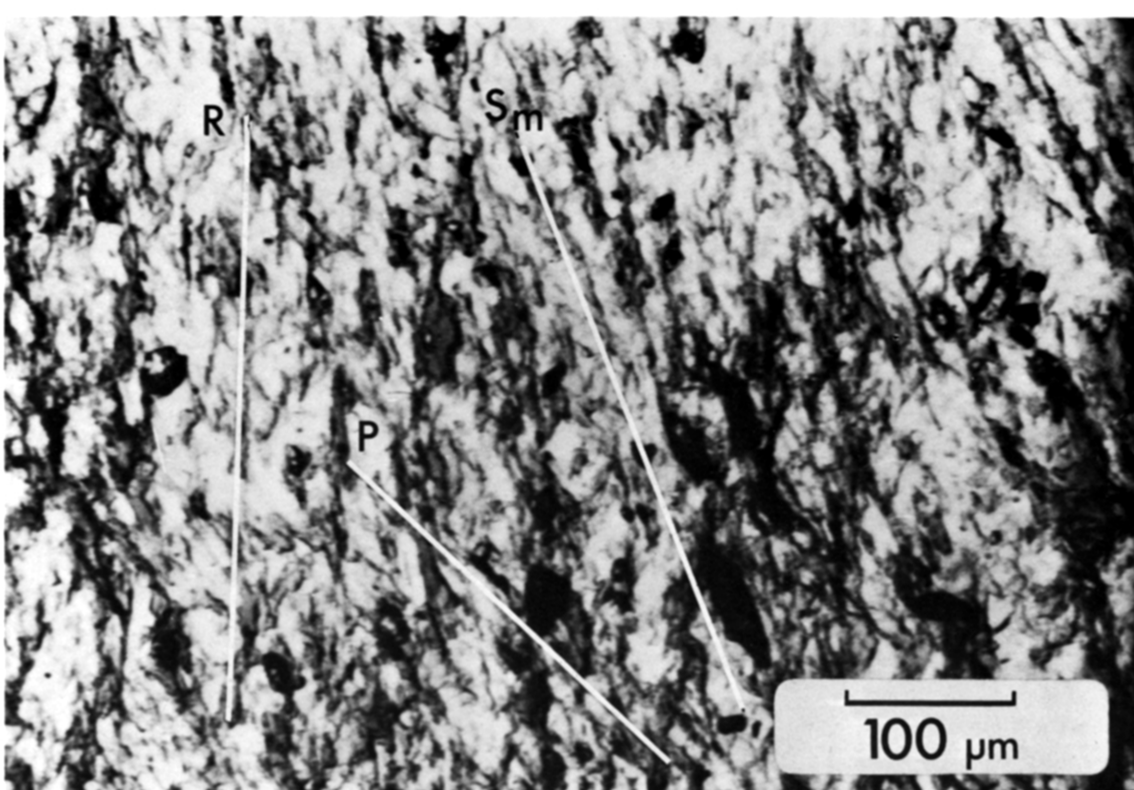

Fig. 3. Multiple foliations in green mylonite. S_m is the principal foliation, R is defined by shear planes and P is a dimensional preferred orientation produced by rotation of S_m during the slip on R.

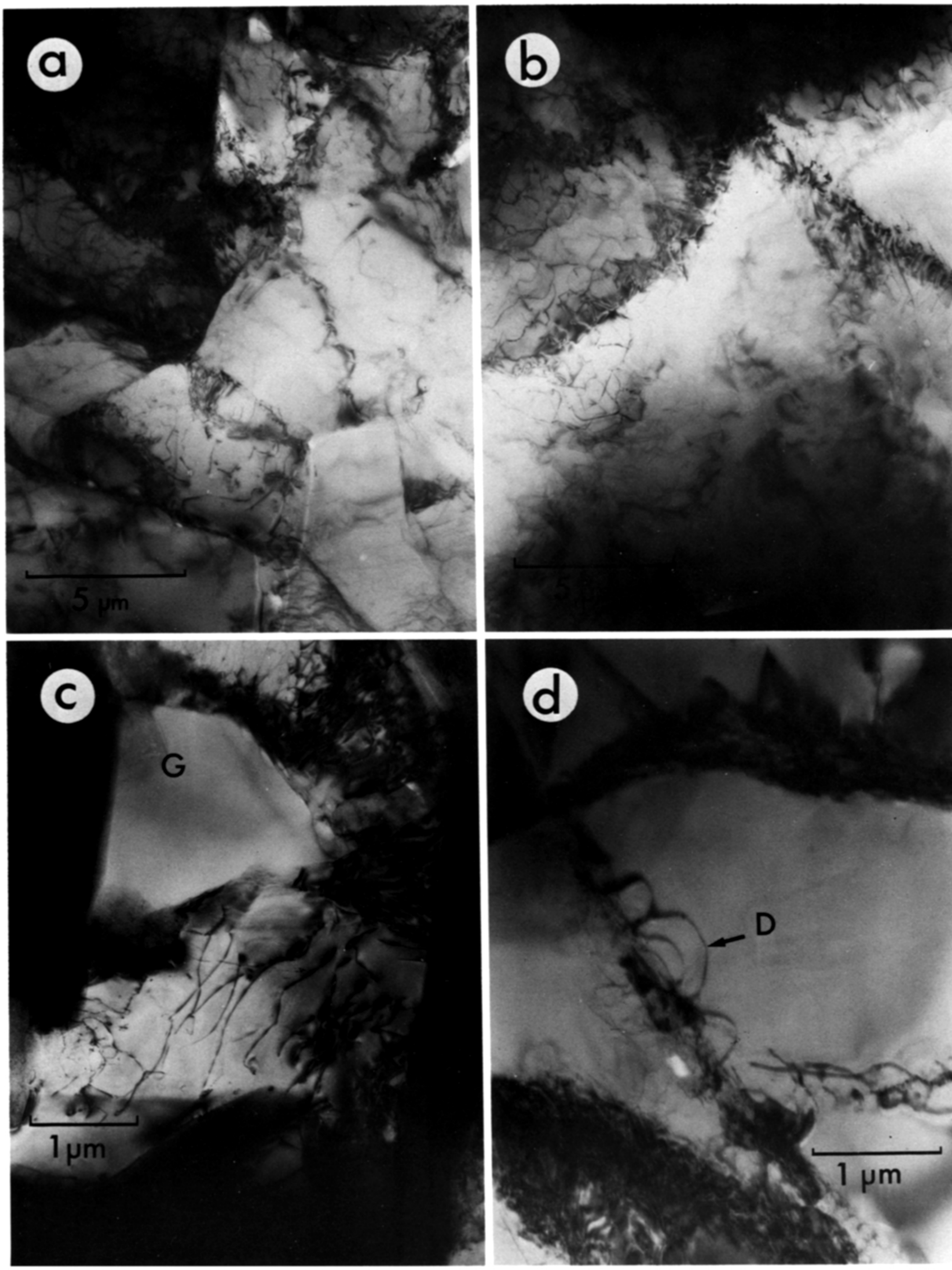

Fig. 4. Electron micrographs of quartz microstructure. (a) Equant sub-grains in quartz from the transition zone near the boundary with schist–mylonite. (b) Blocky sub-grain structure defined by dislocation hedges. (c) Dislocation-free quartz grain (G) with a euhedral outline, probably formed by solution precipitation. (d) Generation of dislocation loops (D) along a healed crack in quartz.

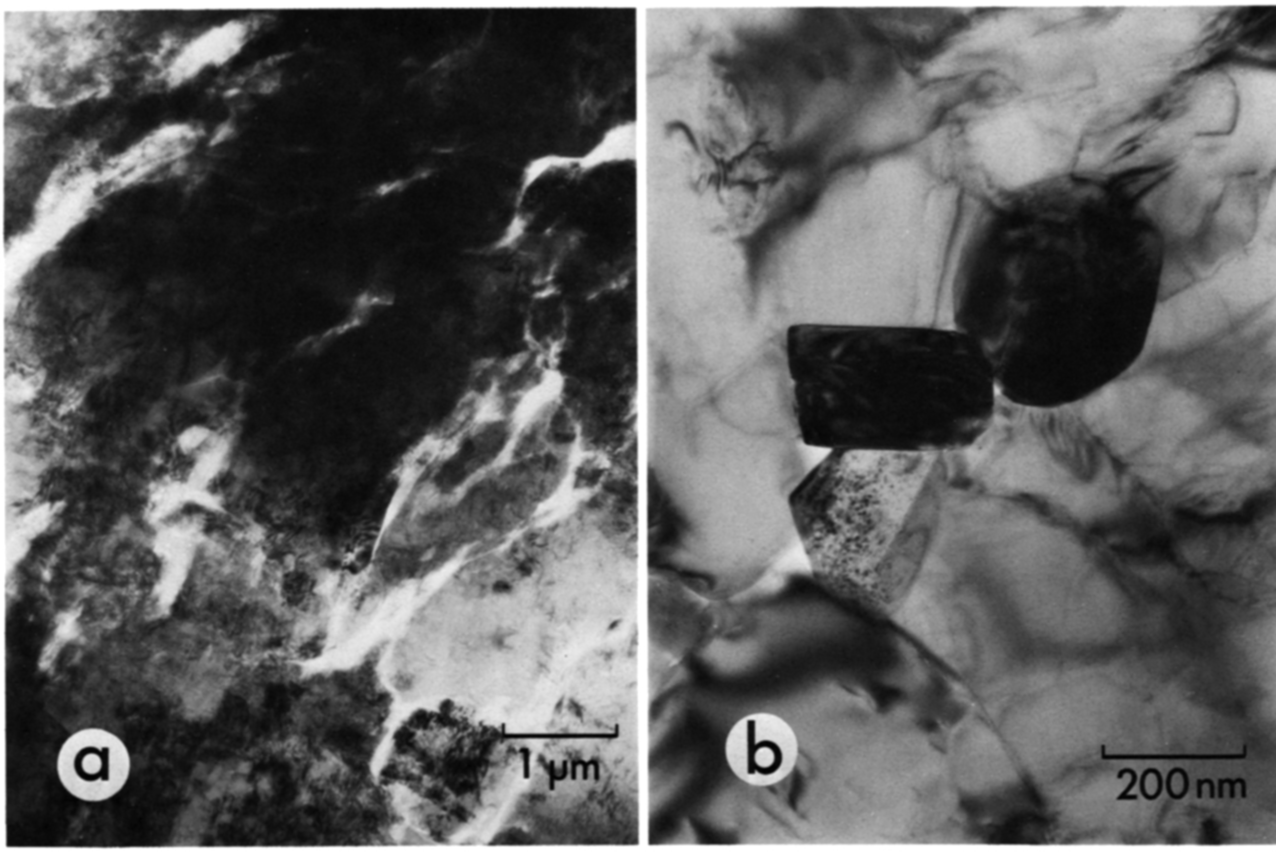

Fig. 5. Electron micrographs of plagioclase. (a) En échelon microcracks in an oligoclase grain. (b) Recrystallized oligoclase grains.

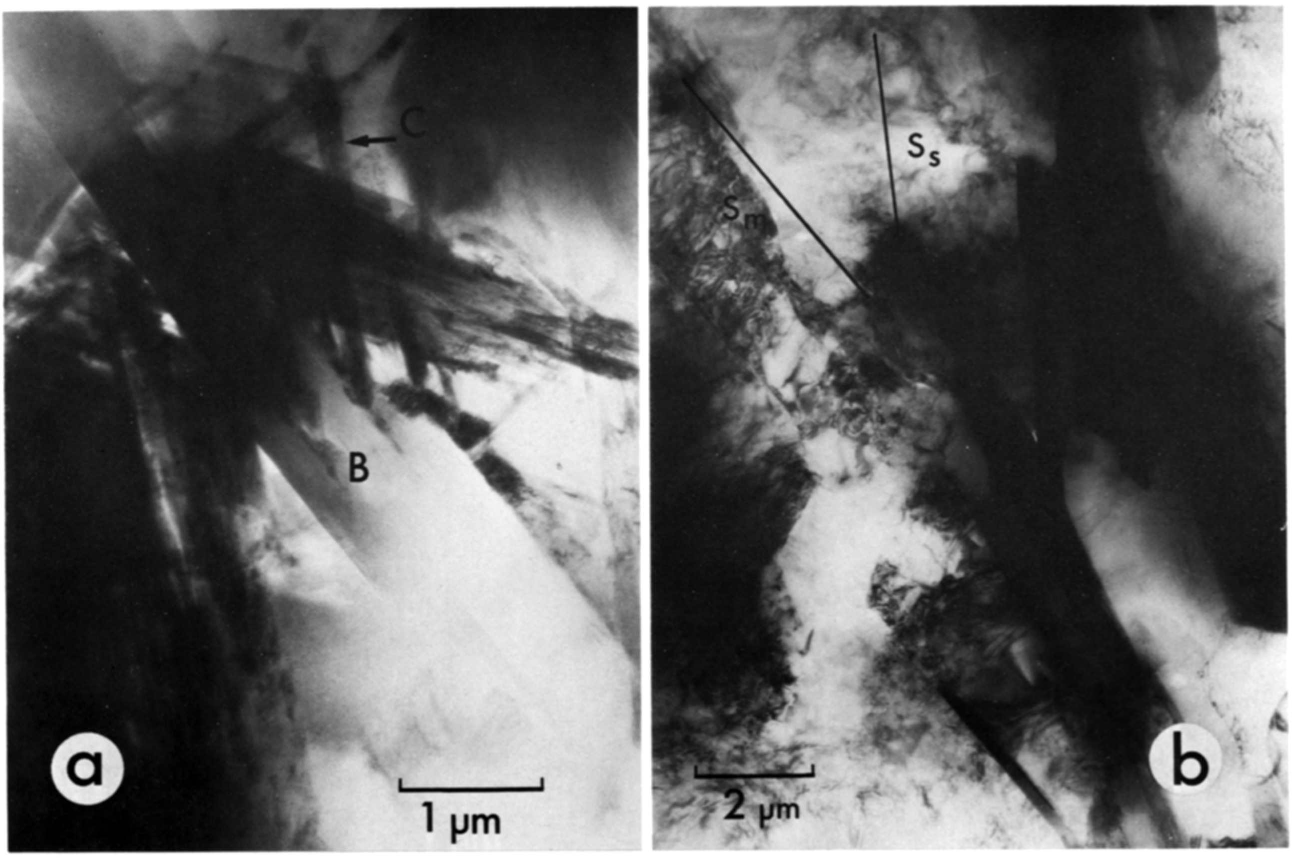

Fig. 6. Electron micrographs of phyllosilicates. (a) Large biotite grain (B) hydrating to form chlorite (C). (b) Primary mylonite foliation (S_m) and secondary foliation (S_s) formed by alignment of chlorite grains.

effects on microstructure must be considered, as these must be significant over a fault zone deformation depth range of 20–30 km.

Brittle deformation features within the transition zone are interpreted in the following ways.

(1) Microcracks are initiated in the schist-derived mylonites at their boundary with transition zone rocks.

(2) The principal mechanism generating microcracks is time-dependent, sub-critical crack propagation as indicated by quasi-stable crack arrays and an absence of obvious catastrophic brecciation.

(3) The association of solution precipitates and hydration reactions with microcracks emphasizes the abrupt increase in fluid ingress at the base of the transition zone and suggests that brittle failure occurred by fluid-assisted sub-critical cracking, that is, stress corrosion.

The mechanism of fracture will now be discussed in detail.

Brittle failure mechanisms in the transition zone

Sub-critical crack growth describes the non-catastrophic propagation of cracks at crack-tip stress concentrations below the critical stress concentration required for dynamic propagation. Scholz (1972) and Atkinson (1979) indicate that the most likely rate controlling mechanism for sub-critical cracking in rocks is chemical interaction of minerals with the fluid environment. Hence, although the association of hydration reactions and solution precipitates with microcracking is only circumstantial evidence for fluid-assisted cracking, stress corrosion is a reasonable expectation. In addition to the possible secondary mineral-microcrack relationship, the concentration of microcracks along grain boundaries in the quartz aggregates is a general characteristic of reaction-controlled sub-critical crack growth for non-geological materials (Anderson & Grew 1977).

Grain boundaries in rocks can consist of relatively wide zones (*c.* 10–100 nm) of disordered crystal structure along which occur fluid-filled voids and/or amorphous hydrated layers (White & White 1981). As such, they are analogous to arrays of pre-existing cracks (elliptical voids) filled with potentially corrosive agents and they suggest that stress corrosion can be expected to concentrate along grain boundaries. Grain boundary sliding which is an integral part of creep will generate larger voids which can coalesce to form a crack which during any stress increase associated with work hardening will propagate rapidly and further aid stress corrosion.

Crack arrays in the large plagioclase fragments are extremely straight relative to those in quartz aggregates and have propagated inwards for different distances towards the grain centres. These plagioclase fractures must have developed early in the history of the transition zone rocks (i.e. near the base of the zone). This follows because the oldest cracks, in terms of sequential development, are expected in the largest grains; once commenced, fracturing and alteration very quickly produce small comminuted feldspar fragments. Geometrically, these fractures resemble load-parallel cracks that develop perpendicular to the foliation; that is, locally, the principal compressive stress must be perpendicular to the foliation at the clast–matrix interface. The continuity of these cracks reflects the absence of grain boundaries in the plagioclase clasts that can act as preferential cracking sites. Holcombe & Stevens (1980) show that material inhomogeneity is the most likely cause for the development of tensile crack-generating stresses in a macroscopic compressive stress system. If an inclusion has a larger shear modulus (i.e. is stiffer) than the matrix, then tensile stresses will be generated within the inclusion. It is argued that this is the case in these plagioclase grains.

There are certain generalized properties of sub-critical cracking, in addition to the above observation, that show crack growth is temperature dependent (Atkinson 1979). Also, Scholz & Koczynski (1979) have shown experimentally that, at high mean stress, stress corrosion dominates over other time-dependent cracking mechanisms, such as crack working during stress cycling. Unlike the classical view of fast fracture, increasing temperature and pressure both promote slow crack growth. Within the limits of the seismogenic layer in the crust, stress cycling is expected (Sibson 1980). Relaxation of stresses and subsequent reduction in crack-tip stress concentrations are expected to have a significant effect on fast fracture, but a more limited effect on the maintenance of slow cracking (Anderson & Grew 1977). Fast fracture, at least locally, would simply be superimposed on a longer-term slow fracturing process. Because of this relative independence from large stress concentrations related to peaks in the stress cycle, stress corrosion could exist well into the general regime of crustal creep. Considering all these factors, one expects stress corrosion, relative to fast fracture to occur at deeper, higher-temperature, higher-pressure levels of the crust. Extrapolating this to the AFZ, it is argued that the transition from schist-derived mylonite to green mylonite (Sibson *et al.* 1979) occurs by stress corrosion cracking and associated mineral retrogression.

Implications for fault zone deformation

Deformation mechanisms leading to the development of transition zone rocks by transformation of schist-derived mylonites to green mylonites (Sibson *et al.* 1979) during upward movement through the crust are summarized in Fig. 7. Of prime importance are the temporal cyclicity and spatial coexistence of brittle and ductile deformation. Such behaviour is most probably related to tectonic stress cycling along the fault zone. Although stress corrosion does not equate with classical pore fluid effects, the damage by corrosion could produce the dilatancy that is ultimately responsible for fluid migration and pore pressure changes observed during seismic events (Nur 1972, Scholz 1972, Scholz *et al.* 1973, Rice 1975).

Dilatancy in the AFZ must extend to a depth equivalent to the schist-derived mylonite/green mylonite transition. If the maximum width at the base of the AFZ is

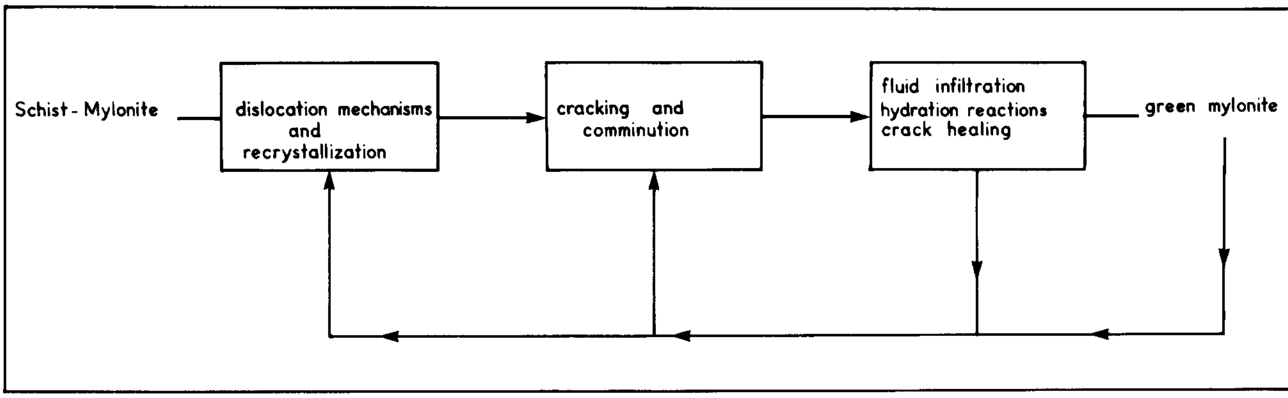

Fig. 7. Summary of mechanisms involved in the generation of transitional fault rock in the Alpine fault zone.

assumed to be 1 km and the transition begins when the active fault zone is 500 m wide (see Sibson *et al.* 1979), then the base of the transition was initially at one-half the depth to the base of the 1 km wide shear zone. This assumes a linear decrease in active zone width from 1 km at the base of the shear zone to a discrete plane at the surface. Schist-derived mylonite fabrics are coeval with amphibolite facies metamorphic assemblages developed at about 500–550°C (Johnston & White 1983). If a 20°C km^{-1} gradient is assumed at the time oblique reverse faulting initiated (Allis *et al.* 1979), then the schist-derived mylonites necessarily formed at a depth of about 25–30 km and the transition to green mylonites would occur at about 12–15 km depth.

The transition temperature will be 250–300°C and is independent of the thermal gradient chosen. This corresponds well with other estimates of the brittle–ductile transition temperature at this scale of observation (Tullis & Yund 1977, 1980, Carter *et al.* 1981). Thermal activation of intracrystalline creep in quartz at 300°C was used by Sibson (1977) in defining the upper limit of his quasi-plastic zone in quartzo-feldspathic crust. A recently published fracture mechanism map for quartz (Atkinson 1982) shows that the fields defining tensile fractures, intergranular creep fractures and intragranular fractures meet at a triple point which corresponds to a temperature in the range 300–400°C and a tensile stress of about 10 MPa. In our model we postulate that the microstructural features indicative of each mechanism coexist partly because of stress cycling; hence, the stress indicated by their coexistence is only a rough estimate. The temperature range is likely to be a more accurate estimate.

As well as a depth sequence, these rocks reflect a progressive time-deformation sequence. This can be described by considering there to be a static P–T reference frame through which rocks are cycled from deep to shallow levels. However, if an initial equilibrium thermal distribution is assumed (whether defined by a normal geothermal gradient or generated by shear heating) thrusting will perturb the isotherms upward. If isotherms cannot re-equilibrate faster than the rate of thrusting, then the P–T reference frame will not remain static. The observed geothermal gradient will be time-dependent and only pressure–depth relationships can be assumed effectively constant. Therefore, it is necessary to remember that observed microstructures reflect both a continuous and continuing deformation cycle.

Associated with dilatancy are the hydration reactions that produce lower-temperature mineral phases. Generation of these secondary minerals ultimately leads to a complete transformation in the bulk mineralogy of the transition zone rocks. If the hydrating fluids are related to stress corrosion processes, then deformation of the fault zone cannot be isolated from retrograde metamorphism. As schist-derived mylonites are moved upward into colder deformation environments, the penetration of cracking down to the base of the transition zone allows an influx of water, giving, effectively, a hot rock–water interaction. The green mylonites can then be simply described as a broad alteration zone.

A fundamental problem with fault zone deformation is strength recovery following dilatancy. Unless this can occur, progressively larger stresses are required to generate seismic release in the upper crust, which is inconsistent with most observations. Rice (1978) has suggested healing of cracks as a strength recovery mechanism. Two types of crack healing were observed in the AFZ transition zone rocks. First, fluid-assisted transport and cementation processes are observed in terms of 'pressure solution' quartz and calcite infillings. This porosity and permeability reduction provides the potential for recovering elastic strength in the seismogenic layer during the relaxation phase of the stress cycle. In addition to the above, more 'perfect' intragranular healing was observed in quartz due to simple closing.

The combination of continual grain size reduction, hydration reactions producing phyllosilicates and increased access to fluids enhances fluid-assisted diffusion processes. These processes aid in maintaining macroscopic ductility within the transition zone and in this sense brittle deformation is self-defeating. This grossly ductile behaviour is best seen in the pervasive development of multiple foliations which are geometrically analogous to those found in experimental and natural gouges (Logan *et al.* 1979, Hall & Rutter 1981). Microscopically, the most extreme transition zone rocks near the top of the transition zone cannot be differentiated from natural gouges although the latter are not as materially coherent.

CONCLUSIONS

(1) Within the Alpine fault zone, there is a broad zone of alteration produced by microcracking and fluid infiltration. The base of this alteration zone corresponds to the schist-derived mylonite/green mylonite contact and broadly defines the maximum depth of dilatancy in the fault zone.

(2) Stress corrosion cracking, not classical cataclasis and comminution, is the predominant brittle deformation mechanism. Deformation micromechanisms are not strictly independent and occur cyclicly. Ductile and brittle behaviour at the grain scale alternate in response to stress cycling in the fault zone.

(3) At a macroscopic scale, the brittle deformation mechanisms are self-defeating in that increased permeability generated by microcracks leads to increased hydration reactions that form more ductile minerals, particularly phyllosilicates. Hence, brittle deformation is suppressed as the relative proportion of ductile minerals increases.

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