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Development of interlaced mylonites, cataclasites and breccias: example from the Towaliga fault, south central Appalachians

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Abstract—A variety of cataclastic rocks from crush breccias to cataclasites and silicified breccias are associated with retrograde mylonites along the Towaliga fault zone of south central Appalachians in Georgia. A zone of alternating breccias and quartz ultra-mylonites, bordered by quartz mylonites that are roughly laminated with mm-scale mica-rich bands of cataclasites, occur subparallel to the mylonitic foliation. Elsewhere along the fault zone, evidence for plastically deformed microfractures is found.

In discussing the mechanical evolution of these rocks two possibilities are considered: (a) the cataclasis is an overprint representing deformation in an entirely brittle regime; and (b) cataclasis and mylonitization occurred at $T > 300°C$ in a predominantly plastic regime. Within the framework of the latter model it is suggested that the cataclasite bands are either due to the relaxation of dilational stresses following downward propagation of seismic ruptures or represent strain-induced seismic instabilities during plastic shearing.

INTRODUCTION

EARLY models of the continental fault zones have a two-layer rheological structure (Brace & Kohlstedt 1980, Sibson 1980). In these models a pressure-sensitive, brittle upper layer and a ductile–plastic lower layer are separated by a rather narrow brittle–plastic transition zone that is usually shown to coincide with a strength maximum and the onset of quartz plasticity. Exhumed mylonites overprinted with brittle deformation tend to support this model (White *et al.* 1980). More recent models (Hobbs *et al.* 1986, Scholz 1988, Shimamoto 1989), however, suggest a rheological structure that provides for an extensive and more complex transitional behavior in an intermediate regime. Mylonites interlaced with cataclasites, pseudotachylytes and *S–C* mylonites, are interpreted as evidence of transitional behavior in the intermediate regime (Sibson 1980, Stel 1981, 1986, Passchier 1982, 1984, Wenk & Weiss 1982, Hobbs *et al.* 1986, Shimamoto 1989).

A key question is the possibility of microfracturing and cataclasis in an otherwise ductile–plastic regime. As Sibson (1989) noted, the end-products of faulting processes should be considered in a dynamic earthquake-related environment. For instance the analysis of aftershock activity of some continental earthquakes reveals that some seismic ruptures may extend downward along fault zones, well below the background seismogenic zone (Strehlau 1986). Results of experimental deformation indicate that strain-induced seismic instabilities may occur within the intermediate regime (Shimamoto & Logan 1986, Shimamoto 1989). Physical evidence is needed for confirmation of the occurrence of such events and as a constraint on estimates of their pressure–temperature conditions. This paper presents a study of some naturally-deformed microstructures from the Towaliga fault zone in the south central Appalachians where in some parts the products of cataclasis and plastic deformation are found side by side and show a close structural relationship. Traditionally, the cataclasis is considered to be a late event in sequence with the mylonites (e.g. White *et al.* 1980). The analysis attempts to contrast this model with more recent views that under certain dynamic conditions the two deformation mechanisms may occur concurrently (Hobbs *et al.* 1986, Strehlau 1986, Scholz 1988, Shimamoto 1989).

THE TOWALIGA FAULT ZONE

The Towaliga fault zone (TFZ) of the south central Appalachians in Georgia (Fig. 1) is an example of a

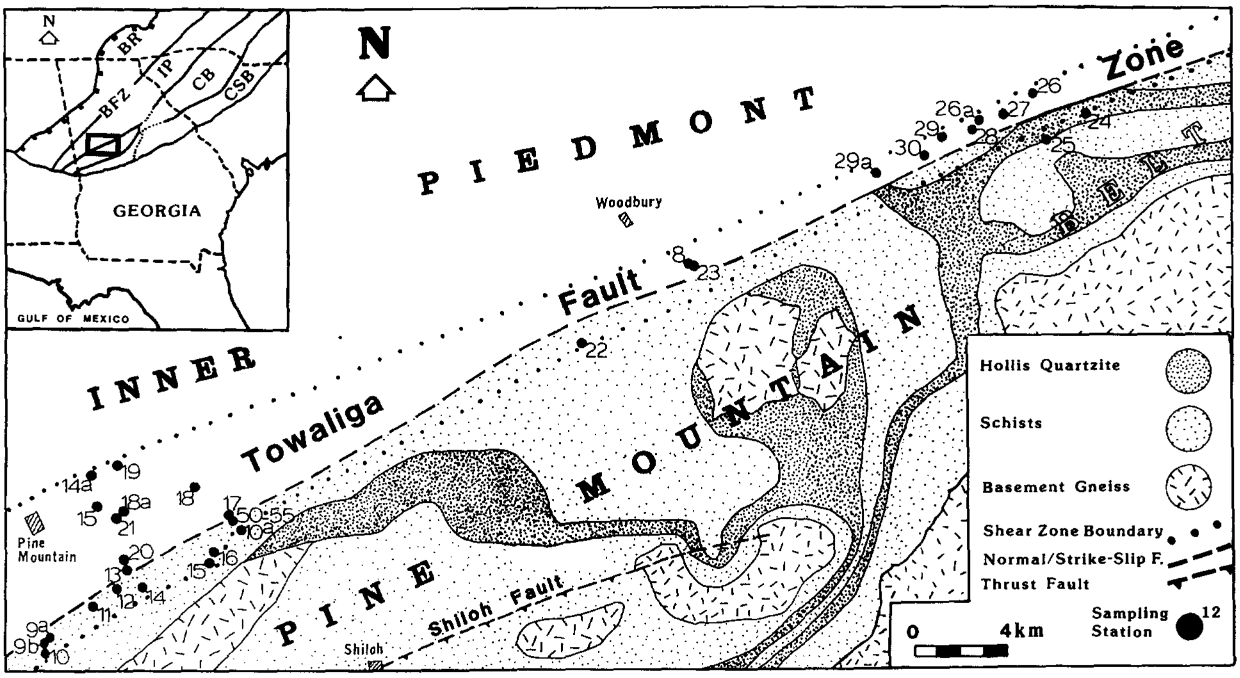


Fig. 1. Location map and sampling sites along the Towaliga fault zone, south central Appalachians, Georgia. Modified after Schamel *et al.* (1980). Regional index map (inset) after Kish *et al.* (1985). BFZ, Brevard fault zone, BR, Blue Ridge, CB, Charlotte Belt, CSB, Carolina Slate Belt, IP, Inner Piedmont.

deeply exhumed fault zone where a suite of sheared rocks are exposed. For the most part this fault is marked by retrograde mylonites that are believed to have developed under a middle to lower greenschist environment (Hooper & Hatcher 1988b). The fault zone borders the northwestern side of the Pine Mountain Belt (window) (Schamel *et al.* 1980, Sears & Cook 1984, Hooper & Hatcher 1988a). Kinematic indicators within the mylonites and some structural relationships suggest that the TFZ is a high-angle dextral strike-slip fault of late Paleozoic age (Grant 1968, Hooper & Hatcher 1988a, Steltenpohl 1988). Based on COCORP (Consortium for Continental Reflection Profiling) data, however, Nelson *et al.* (1987) contend that the TFZ is a NW-dipping normal fault with an inferred offset of some 9 km. In the study area the shear zones generally strike 070° and dip 60°NW as suggested by the mylonitic foliation. The dominant stretching lineation, usually defined by the elongated quartz grains, plunges at 52° ± 5° toward 025°. Previous descriptions of the brittle deformation along the fault zone are mainly restricted to bodies of silicified breccias that are believed to post-date the mylonites of the Towaliga fault (Kish *et al.* 1985, Hooper & Hatcher 1988b). Reinhardt *et al.* (1984) pointed to the similarity of microstructures in the silicified breccias of the TFZ and those along Cretaceous and Cenozoic faults elsewhere in the southeastern Appalachians. Some earlier work on the TFZ applied the general term "cataclastic rocks" to the mylonites (Higgins 1971, Davis 1980). Throughout this paper we adopt the fault rock classification of Sibson (1977).

FIELD AND MICROSTRUCTURAL OBSERVATIONS

Cataclastic deformation

The observed cataclastic deformation features throughout the study area can be divided into three groups: (1) mm-size bands of cataclasites subparallel to mylonitic foliation; (2) brecciated mylonites; and (3) the silicified breccias of Dixon Mountain (locations 8 and 23, Fig. 1). The latter group has been reported and described by others (Crickmay 1933, Davis 1980, Kish *et al.* 1985) and is not discussed here. Micro-faults and fractures also occur in the intact rock, commonly associated with the cataclasite bands.

Cataclasite bands in quartz mylonites

A unit of fine-grained (*ca* 60 μm grain-size) banded quartz mylonite is exposed in a roadcut at location 17 (Figs. 1 and 2a). In the outcrop, the banding is penetrative and defined by 0.5–2 mm thick, soft, oxide-stained material subparallel to laminations in the mylonite spaced at 1–3 cm and dipping 55–60°NW (Fig. 2b). The quartz mylonites easily split along the bands, revealing surfaces with strong slickenlines (Fig. 3a) that are subparallel to the stretching lineation. Sections cut parallel to and through the bands reveal the same linear fabric, in a less pronounced form, defined by cataclasite fragments (Fig. 3b). Under the microscope, evidence of brittle deformation within the bands includes rotated angular fragments of the host mylonite embedded in a fine-grained mica-rich matrix (Fig. 4a). Cataclastic rotation is indicated by a mismatch of crystallographic orientations in multigrain fragments with respect to those in grains outside the band. Some of the thicker bands also have internal microstructures (Fig. 4b) resembling *P*-type shears or the incipient shape fabric described by Bartlett *et al.* (1981) that is shown to evolve in brittle, hardening shear zones during and after strain localization (Logan *et al.* 1979, Shimamoto 1989). Most of the thinner bands (≤0.2 mm) are relatively short (10–15 cm) and wedge out in simple microfractures. This suggests that at least some of the bands were developed by propagation of microfractures at their peripheries. Highly fractured and rotated mica fish and feldspar grains are also common at the margins of the bands. A persistent set of microfractures cuts across at high angles and displaces the bands (Fig. 4c). The apparent displacement on microfaults so produced is typically 0.5–1.5 times the average band thickness.

The term shear band has been used to describe small-scale zones of both ductile (Simpson 1986) and brittle (Platt & Vissers 1980, Passchier 1984) shear within a larger non-coaxial shear zone. However, to avoid confusion with the kinematic implications of the term "shear band" as defined by Berthé *et al.* (1979), we use the term cataclasite band for the microstructures described here. Although it is difficult to ascertain the overall position of the bands with respect to a variable mylonitic foliation, locally the lineation associated with slip on the bands appears to be subparallel to the stretching lineation. Thus a distinction may be drawn between the bands described here and similar cataclastic features described by others as granulation seams (Pittman 1981), deformation bands (Aydin & Johnson 1983) or shear faults (Blenkinsop & Rutter, 1986), all of which were apparently unrelated to the host-rock fabrics.

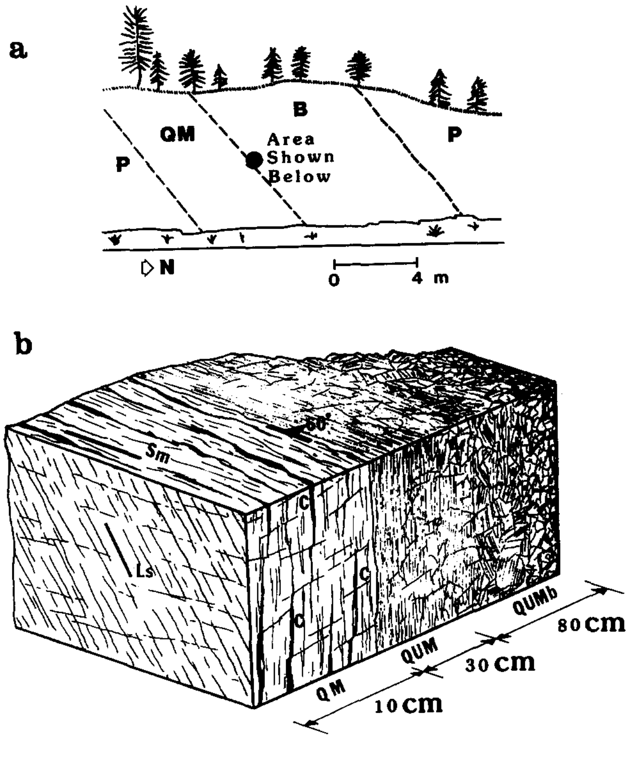


Fig. 2. Outcrop relationships. (a) Diagram of outcrop (traced from a field photograph) at location 17 (see Fig. 1), looking west. P, weathered phyllonites; QM, quartz mylonites containing cataclasite bands; B, zone of crush breccias and protocataclasites interlaced with quartz ultramylonites. (b) Schematic block diagram illustrating relationship between mica-rich quartz mylonites QM, containing cataclasite bands C, intact quartz ultramylonites QUM, and brecciated quartz ultramylonites QUMb. Scales show spacing of cataclasite bands and the typical distance of transition from intact ultramylonite to brecciated ultramylonite. Sm and Ls show orientation of mylonitic foliation and stretching lineation, respectively.

Brecciated mylonites

A breccia zone comprising quartz mylonites interlaced with quartz ultramylonites is in sharp contact (see Fig. 2a) with the banded quartz mylonites. The thickness of the zone normal to strike is about 7 m, and within it layers of breccia are separated by several irregularly-spaced thin lenses or wedges of intact quartz ultramylonite each about 30–40 cm thick. In outcrop, the brecciated mylonites occur as units clearly concordant with the local shear zone attitude (070°/60°NW) and the brecciation appears to affect ultramylonites alone. Within each breccia layer, transition from intact ultramylonite to crush breccias to protocataclasites takes place typically over a distance of less than 1 m, starting with a narrow zone of very densely fractured rock (Fig. 2b). Reduction in clast size is coincident with increased matrix fraction and rounding of the clasts. The extent of comminution and textures of these crush breccias and protocataclasites is shown in Figs. 5(a)–(d). Fresh fracture surfaces of the coarser breccias show isolated open microfissures and cavities 1–2 mm in diameter that contain only very minor secondary quartz growth. Separated fragments often bear strong slickensides with striations up to 3 cm long marked on thin films of indurated gouge separating the fragments. Very little veining and almost no secondary cementation was observed.

Optical microscope observations reveal some details of the process of progressive brecciation and cataclastic flow. Larger clasts of quartz-ribbon mylonite tend to fracture and shatter parallel to the length of the ribbons. Rotation of the loose ribbons and extension microfractures normal to their length indicate cataclastic flow within the breccia matrix (Fig. 6a). The ribbon fragments, however, maintain their plastic deformation grain substructures (subgrains, deformation bands and lamellae). The matrix texture shows random fabric and untidy grain boundaries probably due to presence of sub-micron intergranular gouge fragments, typical of cataclastic deformation textures (Fig. 6b). In plane polarized light (not shown) the matrix material is marked by trails of semi-opaque impurities that are mostly concentrated along the fragment grain boundaries.

Porosity variations across the breccia zone

Experimental studies of cataclastic deformation have shown that porosity plays an important role in the processes of cataclasis and textural evolution during cataclastic flow in rocks (Mogi 1965, Paterson 1978, Hirth & Tullis 1989, and others). If the original porosity of a cataclastic zone is not significantly modified by later events like silicification (as in the breccias described here), then variations and distribution of finite porosity throughout a zone may reflect the extent, spatial distribution and state of evolution of localized cataclastic deformation of the rock. Provided that the initial porosity of unfractured rock is known or can be estimated, this tectonic porosity may also give an estimate of magnitude of volume change (negative for high initial porosities) during cataclasis.

A number of porosity measurements were conducted using water-saturation for crush breccias and mercury porosimeter (Robertson 1982) methods for protocataclasite samples. The samples, selected to be texturally very similar to those shown in Figs. 5(b)–(d), had average total porosities of 1.57, 2.36 and 5.72%, respectively. Analysis of the mercury intrusion and withdrawal behavior for the protocataclasite samples (Fig. 5d) showed that the average pore diameter was $0.034 \pm 0.0087\ \mu m$.

Plastic deformation

Evidence of syntectonic plastic deformation is widespread along the TFZ particularly in southeastern parts of the zone (Fig. 1). Width of the fault zone varies from approximately 5 km in the southwest to under 3 km in the northeast (Fig. 1). Although both type I and type II *S–C* mylonites (Lister & Snoke 1984) are present, type I mylonites generally are restricted to the Piedmont (northwest) side of the fault. This is mainly because the TFZ separates the Inner Piedmont gneiss from the mica-rich quartzites and schists of the Pine Mountain Belt to the southeast. The interior of the fault zone is dominated by alternating quartz-ribbon mylonites and quartz ultramylonites (e.g. locations 10, 12, 17, 22, 28 and 29a on Fig. 1). Hooper & Hatcher (1988b) reported similar observations from the TFZ southeast of Jackson, about 40 km northeast of location 26 (Fig. 1).

Microstructural evidence indicates that episodes of microfracturing with subsequent healing and plastic shear may have occurred. Figure 7(a) shows healed microfractures that appear to be curved and bent opposite to the dextral sense of shear in the mylonite (inferred from the asymmetrical recrystallized grain-shape fabric). Closer examination of these curved microfractures (Fig. 7b) reveals a weak crystallographic preferred orientation and grains with subhedral shapes and smoothly rounded corners suggestive of dynamic recrystallization.

DISCUSSION

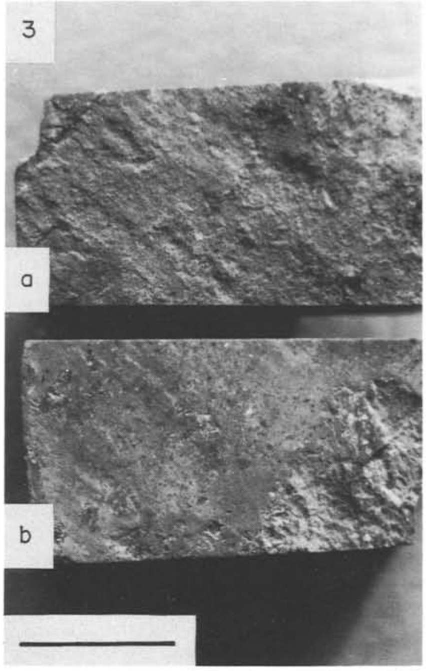
Three different cataclastic features are recognized in the study area: (1) cataclasite bands and related cross-cutting microfaults; (2) brecciated mylonites; and (3) silicified breccias. The first two of these features are related to mylonites and ultramylonites of the TFZ. We consider two possibilities: (a) that the cataclasis is an overprint that represents deformation in an entirely brittle regime; and (b) that cataclasis and mylonitization both occurred in a predominantly plastic regime; i.e. at $T > 300°C$ in a greenschist environment (Kohlstedt & Weathers 1980, Turner 1981). The two models differ in that in (a) cataclasis is in sequence with mylonitization and marks the cessation of plastic flow in quartz.

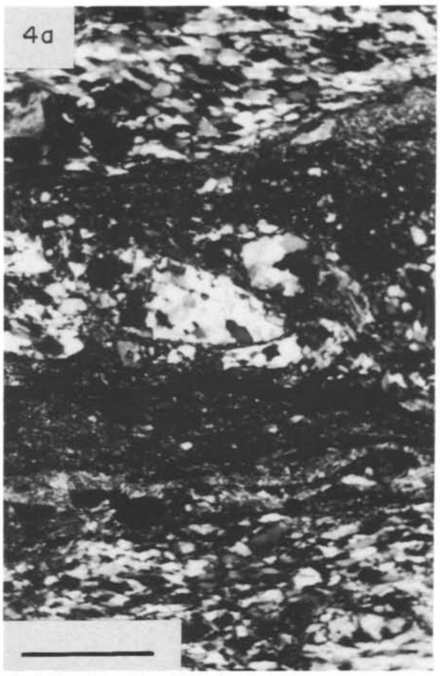
(a) *Brittle overprint model*

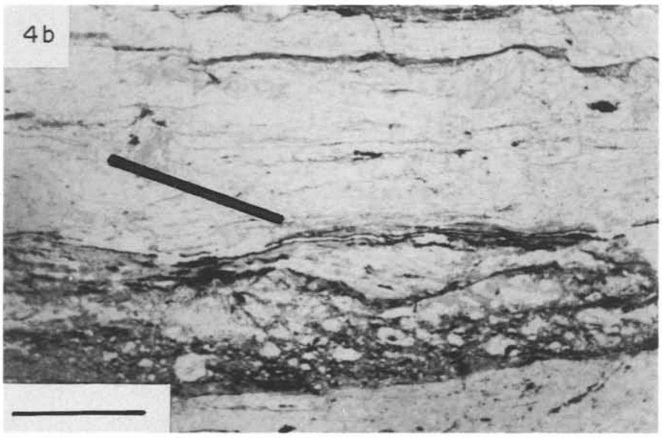
Gradual uplift of an active fault zone may result in a laminated shear zone of gouge, cataclasites, ultramylonites and mylonites (White *et al.* 1980, Lister & Snoke 1984). Layers of gouge should be particularly noticeable as they are diagnostic of persistent localized displacement along shallow brittle fault zones. Deformation features associated with abundance of water, such as vein-filled microfracture systems and evidence of pressure solution may also attend the uplift of an active fault zone. Although such features were not found in the TFZ, some of our observations support the progressive uplift model. There is a general grain-size reduction in quartz mylonites toward the interior of the fault zone where the cataclastic rocks occur, and brecciation almost exclusively affects the ultramylonites. Dynamic recrystallization at progressively lower temperatures is expected to produce the grain-size reduction effect and, as argued by White *et al.* (1980), cataclasis should preferably occur in fine-grained mylonites when an active fault zone is elevated into the brittle regime. The grain-size reduction effect, however, could also result from an increase in stress or strain rate (Mercier *et al.* 1977, Twiss 1977, White 1979). One difficulty with this model is that if the entire zone is considered, the mylonites with cataclasite bands and the plastically deformed microfractures remain unexplained.

(b) *Mixed brittle and ductile deformation in a predominantly plastic regime*

Indications of intermittent brittle deformation within an otherwise plastically deforming mylonite zone have been presented by a number of investigators (Sibson 1975, 1980, Passchier 1982, 1984, Casas 1986, Hobbs *et al.* 1986, Stel 1986, Bowler 1989) where the most convincing evidence for interchanged deformation mechanisms is often plastically deformed pseudotachylytes. Direct evidence for alternating cataclastic and plastic deformation in the TFZ includes plastically deformed healed microfractures and cataclasite bands. The conforming attitudes of brecciated mylonites, cataclasite bands and mylonitic foliation suggest the possibility that the brecciation and development of the cataclasite bands







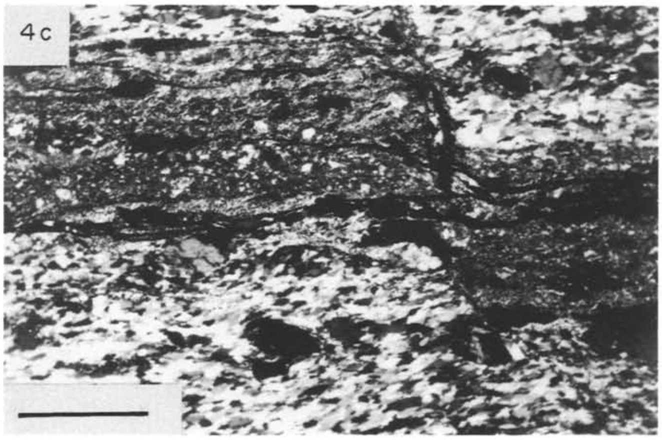
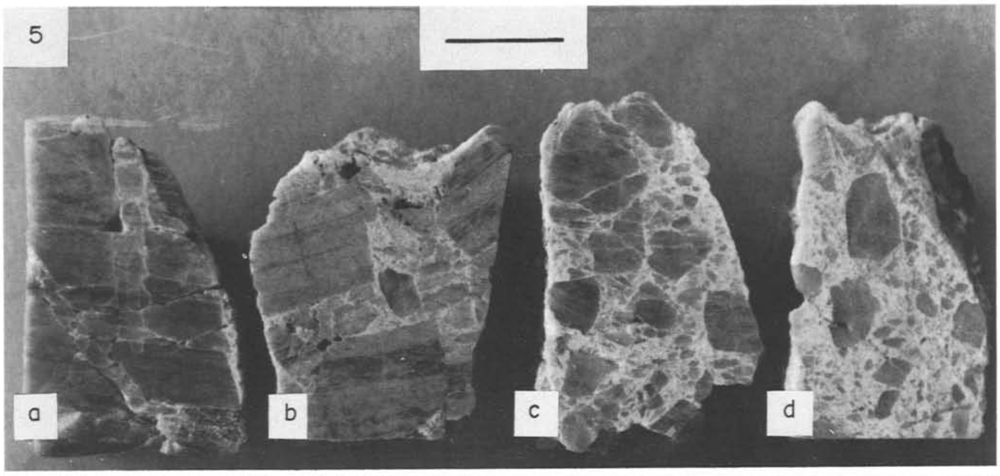
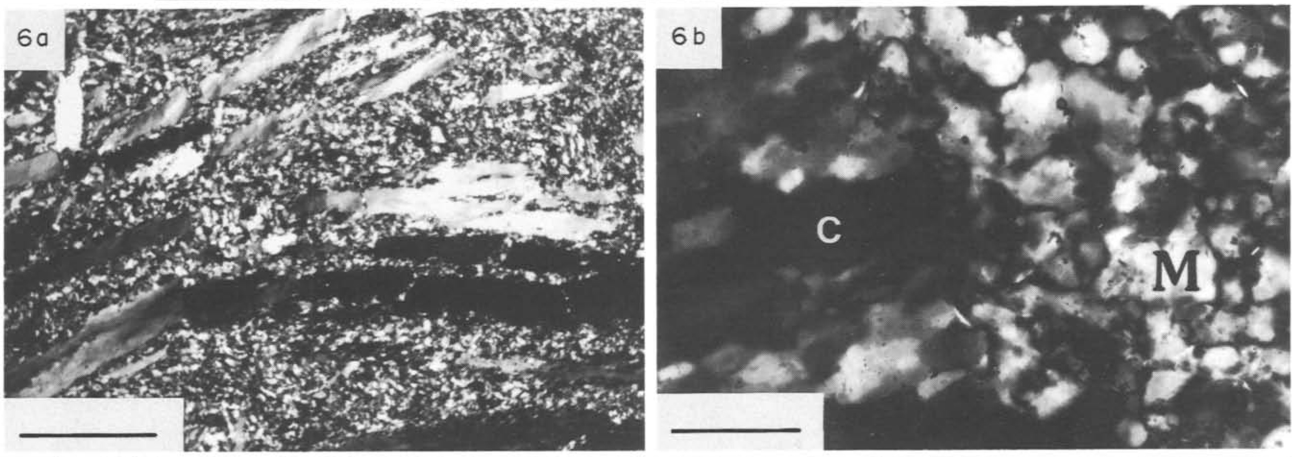


Fig. 3. Linear fabrics associated with cataclasite bands. (a) Slickenlines revealed by splitting the mylonite along bands. The parting surfaces often coincide with films of very fine-grained oxide-stained material and white mica that anastomose within the bands. (b) A surface cut through a band parallel to its shear plane shows a lineation defined by alignment of clasts as well as streaks of coarse and fine matrix in the cataclasites. Material plucked off the surface discloses the underlying mylonite. Scale bar = 1 cm.

Fig. 4. Typical microstructures of cataclasite bands and evidence of cataclasis. (a) Layered structure of a single band defined by fragment size. Note clustering of large composite fragments from host mylonite in middle of band and anastomosing streaks of extremely fine-grained material. Mylonitic fabric in fragments is rotated with respect to host-rock fabric. Crossed polars. (b) *P*-type shears developed within thickest part of band. Section is cut parallel to linear fabric shown in Fig. 3 and shear type identified with respect to asymmetric shape fabric in the mylonite (solid line). Plane polarized light. (c) A microfaulted cataclasite band. Note drag associated with fault. Crossed polars. Scale bars = 0.5 mm.





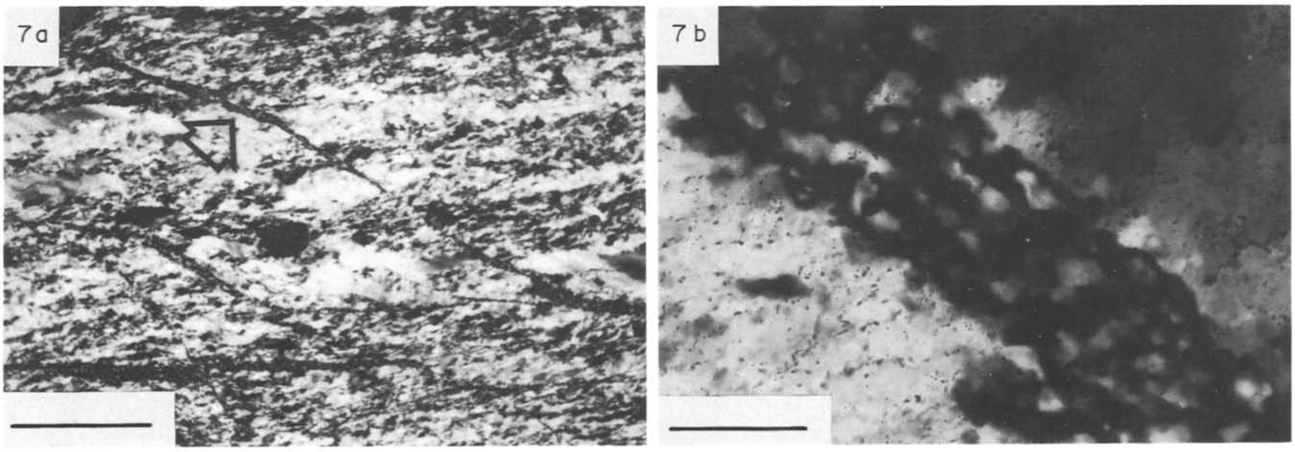


Fig. 5. Polished surfaces showing progressive cataclasis of Towaliga fault ultramylonites. (a) Densely fractured quartz ultramylonite. (b)–(d) Progressive comminution from crush breccia to protocataclasite. Outcrop distance between (a) and (d) is typically about 0.8 m. Scale bar = 2 cm.

Fig. 6. Typical microstructures of brecciated mylonite. (a) Ribbon fragments in a crush breccia matrix. Note extension fractures normal to the length of ribbons. (b) Close-up view of matrix texture M, in crush breccias and protocataclasites. Part of a clast, C, on left. (a) and (b) in crossed polars, scale bars for (a) 1 mm and (b) 50 μm.

Fig. 7. Plastically deformed microfractures. (a) Curved healed microfractures in fine-grained quartz mylonite. The asymmetric grain-shape fabric indicates dextral shear in mylonite. (b) Close-up view of healed microfracture, area arrowed in (a). Note fabric developed along site of original microfracture. (a) and (b) in crossed polars. Scale for (a) 1 mm and (b) 40 μm.

are closely related. However, it is also equally possible that the brecciated mylonites alone were produced in an entirely brittle regime, that is according to model (a) above.

Strehlau (1986) and Scholz (1988) suggested that an earthquake rupture that nucleates well above the mylonitic regime can propagate downward and impose a strain rate several orders of magnitude greater than the prevailing tectonic strain rates of about 10^{-12} s^{-1}. Such momentary events can produce zones of breccia and pseudotachylytes in deeper parts of an active fault zone, preferably along pre-existing zones of weakness with a lower coefficient of dynamic friction (Strehlau 1986). It is conceivable, however, that the seismically induced strain rates rapidly diminish and the affected volume begins to undergo a period of stress relaxation followed by a return to aseismic strain rates under which plastic deformation can resume (Sibson 1980). Because cataclasis in low porosity rocks is usually a dilatant process, relaxation of stresses immediately following the brittle event could be affected by the nearly instantaneous volume increase within or adjacent to the propagation front. Norrell *et al.* (1989) show that for a fixed shear zone width, the strain due to a volume increase in the zone is accommodated by extension parallel to the foliation plane. At the onset of a relaxation period, while strain rates are still high, deformation is mainly brittle and strain due to dilation will be partly accommodated in the adjacent mylonites by producing the cataclasite bands. At the same time it is possible that some of the pre-existing cataclasite bands would be reactivated and further extended parallel to their length. Assuming that the porosity across the breccia unit (location 17, Fig. 1) has not increased since the cessation of tectonic deformation, the measured 5–6% finite porosity of the protocataclasites may be a rough measure of the total dilation due to cataclasis that had to be partly accommodated within the adjacent mylonites in the form of cataclasite bands, microfaults and fractures.

Another possibility within model (b) is that the cataclasite bands are evidence of mechanical instabilities that were initiated within the mylonitic regime. Based on experimental deformation of halite, Shimamoto & Logan (1986) and Shimamoto (1989) suggested that plastic deformation in the deeper parts of the intermediate (semi-ductile) fault regime is potentially unstable and the zone is in part seismogenic. The internal structures of the cataclasite bands described here indicate localized frictional sliding processes and lend support to Shimamoto's (1989) suggestion of semi-ductile rheology. In this case the cataclasite bands would result from an instability within the plastically deforming mylonites after a critical shear strain value was reached, rather than from an imposed change in strain rate. The existence of cataclasite bands with various thickness and lateral extent indicates an ongoing process in which the thinner bands mark the latest events. The development of Riedel shears within the bands may further strengthen the interpretation that the bands are involved in the stick–slip processes (e.g. Moore *et al.* 1989). However, if it is assumed that the cataclasite bands are developed in a semi-ductile regime, it is difficult to show that the zone of brecciated mylonites is associated with the same processes and *P–T* environment.

CONCLUSIONS

The deformation features reported in this study indicate the co-operation of brittle and plastic mechanisms in quartz-rich rocks of a section of the TFZ and are best explained by a fault zone model that includes an intermediate regime in which both brittle and plastic deformation mechanisms transiently operate. It appears that none of the existing models alone can account for the variety of deformation in the TFZ; a problem which probably reflects the complex rheology of a crustal transition zone. Our observations suggest that further investigation of exhumed low-grade shear zones, along with the study of experimentally produced microstructures associated with the transition from fracture to crystal plasticity in rocks, is necessary in order to reach a better understanding of the partitioning of strain and rates of deformation and type of mechanical behavior within the intermediate regime.

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