Journal of Structural Geology, Vol 13, No 9, pp 1013 to 1023, 1991
Printed in Great Britain

0191-8141/91 $03.00 + 0.00
Pergamon Press plc

Significance of brittle and plastic fabrics within the Massawippi Lake fault zone, southern Canadian Appalachians*

ALAIN TREMBLAY and MICHEL MALO

INRS Georessources, 2700 rue Einstein, Ste Foy, Québec, Canada G1V 4C7

(Received 12 June 1990; accepted in revised form 27 February 1991)

Abstract—Brittle and plastic deformation structures are common features of many well described brittle–ductile fault zones. These fabrics coexist in the Massawippi Lake fault zone (MLFZ), a NW–SE oriented fault related to the La Guadeloupe fault in the Dunnage Zone of the Canadian Appalachians. Kinematic analysis of the MLFZ indicates that it is a NW directed ramp thrust underlain by tectonic slices of mafic–ultramafic and granitic rocks. The association of brittle and plastic deformation structures is well demonstrated within the granitic rocks. Brittle fabrics include cataclasites and quartz filled fractures and microfractures. Plastic deformation is superimposed on brittle textures. Microstructures and c axis fabrics suggest plastic and superplastic regimes of deformation. Analysis of coexisting brittle and plastic microstructures in terms of deformation mechanism associations reveals inconsistencies between brittle and ductile related conditions of deformation. Combined with the regional structural and metamorphic history, this suggests that brittle fabrics are relict structures preserved within a predominantly ductile shear zone.

INTRODUCTION

THE coexistence of cataclastic and mylonitic fabrics within a single fault zone is frequently described in the literature (Eisbacher 1970, Sibson 1977, Gibson & Gray 1985, Simpson 1986, and others). It is generally interpreted as indicative of a brittle-to-plastic transition shear zone (Rutter 1986) and attributed to a progressive phase of deformation within a single phase of faulting (Sibson 1977, Watts & Williams 1979, Mitra 1984, Gibson & Gray 1985, Simpson 1986, Stel 1986). More rarely, it is attributed to different tectonic events leading to relict fabrics within a reactivated fault zone (Flinn 1977, Obee & White 1986, Gaudemer & Tapponnier 1987). However, it is frequently difficult to distinguish between characteristic fabrics of progressive fault-related deformations and those associated with reactivated faults. In internal domains of orogens with a complex tectonic history, superposition of profoundly different types of fault rocks requires careful analysis before conclusions can be drawn on the presence of any brittle–plastic shear zone.

In this paper, we describe coexisting brittle and plastic textures and structures within the Massawippi Lake fault zone. Detailed observations on the nature of the microscopic and intracrystalline textures, along with the regional structural history, reveal an inconsistency between brittle related and plastic related deformation mechanisms, which is best explained by a two-stage history of the fault zone.

GEOLOGICAL SETTING

The Massawippi Lake fault zone (MLFZ) is located in the Dunnage Zone (Williams 1979) of the Canadian Appalachians (Fig. 1a). In southern Québec, the Dunnage Zone is made up of Cambro-Ordovician rocks lying between the Baie Verte–Brompton Line to the NW (Williams & St Julien 1982) and the La Guadeloupe fault to the SE (St Julien *et al.* 1983, Tremblay *et al.* 1989b). Deformation and metamorphism were Acadian in age (Tremblay & St-Julien 1990) although some localities of northern Vermont record the imprint of an older metamorphic phase (Laird *et al.* 1984). Acadian metamorphism in the area was at the biotite-grade greenschist facies (Laird *et al.* 1984, Sutter *et al.* 1985).

The MLFZ is a NW–SE striking fault viewed as a lateral ramp between the Rivière Magog fault (Tremblay in press) and the La Guadeloupe fault (Fig. 1a) (St-Julien & Slivitzky 1985). The fault zone is well exposed on the west shore of the Massawippi Lake (Fig. 1b). The MLFZ is a tectonic suture (Tremblay 1990) between the Ascot Complex and the Magog Group (Fig. 1). The former is the remnant of an Ordovician volcanic arc (St Julien & Hubert 1975, Tremblay *et al.* 1989a) and the latter is an arc related sedimentary basin lying along the northwestern side of the Ascot Complex volcanic belt (Cousineau 1988).

The MLFZ is underlain by tectonic slices of felsic and ultramafic igneous rocks (Fig. 1b). Between and along these rocks slices, there are felsic red brownish mylonitic schists and phyllonites. The protoliths are strongly foliated and folded albitic schists and fine grained volcaniclastic rocks which crop out along the northeastern side of the fault zone (Fig. 1b). Rocks on either side of the MLFZ are polydeformed. Regional folds are coeval with an axial planar S_2 foliation. Pre D_2 deformation is expressed by a S_1 schistosity locally associated with F_1 folds (Fig. 1b). A late NW dipping crenulation cleavage (S_3) overprints earlier deformation structures.

There is a sinistral rotation of F_2 folds and S_2 foliation along the MLFZ (Fig. 1b). The regional foliation is progressively rotated and reoriented parallel to the

*Published with the permission of the Ministère de l'Energie et des Ressources du Québec.

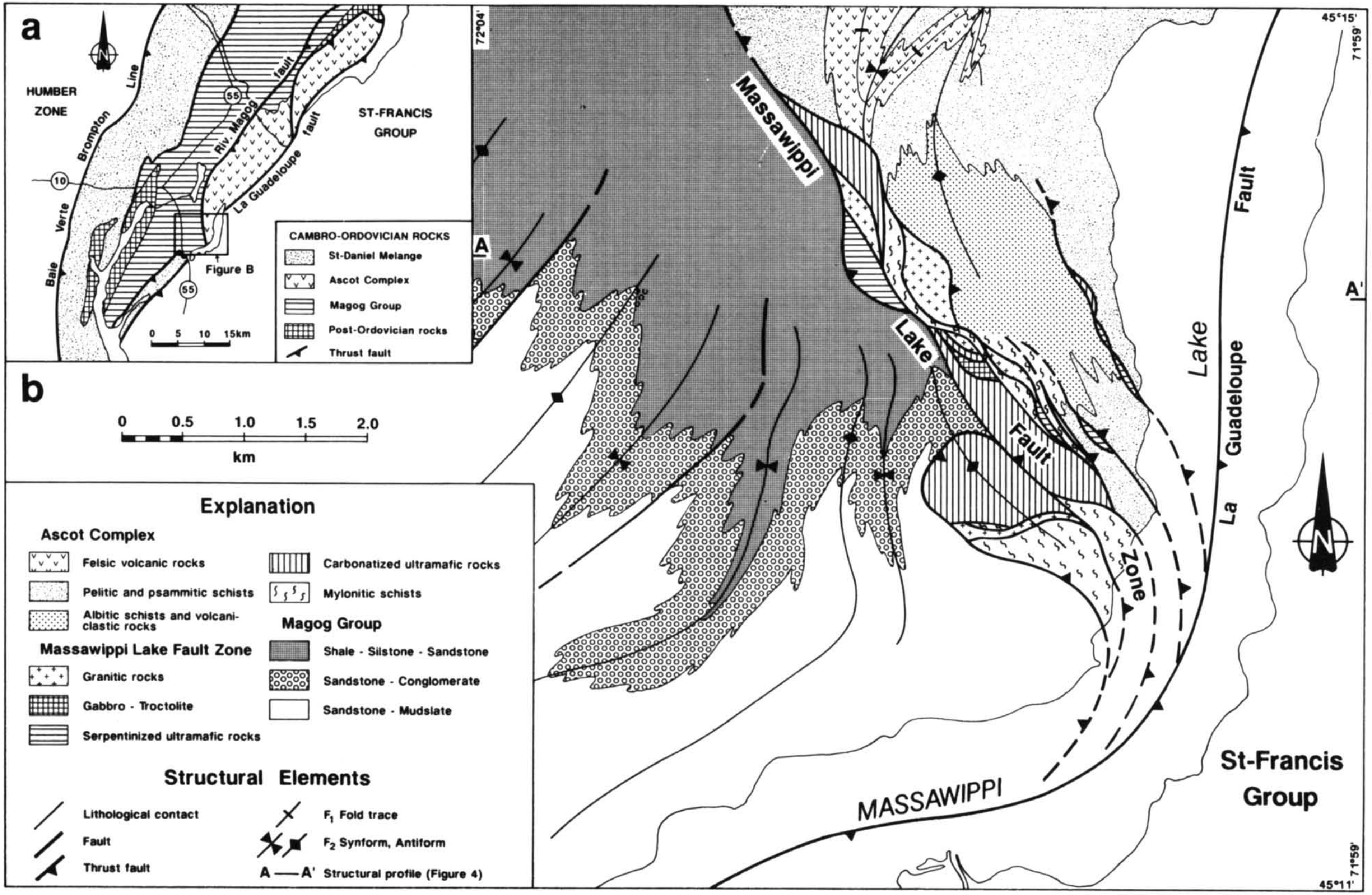

Fig. 1. (a) Geology of the Dunnage Zone in the southern Quebec Appalachians. (b) Geological map of the Massawippi Lake fault zone and adjacent units in the Massawippi Lake area.

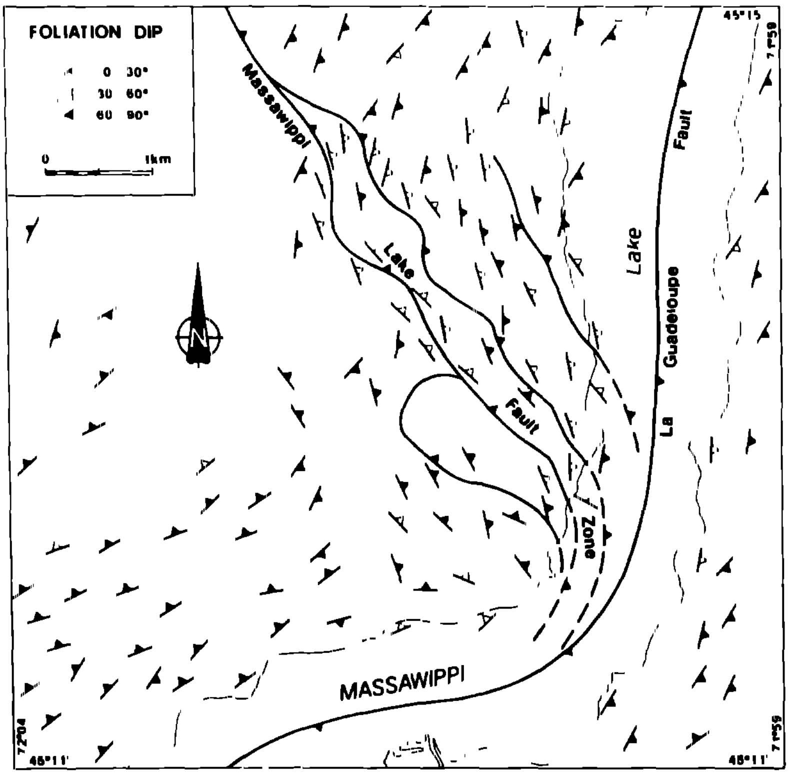

Fig 2 Structural map of regional foliation (S_2) within and around the MLFZ Same area as Fig 1(b). Note gradual sinistral rotation of foliation along the MLFZ

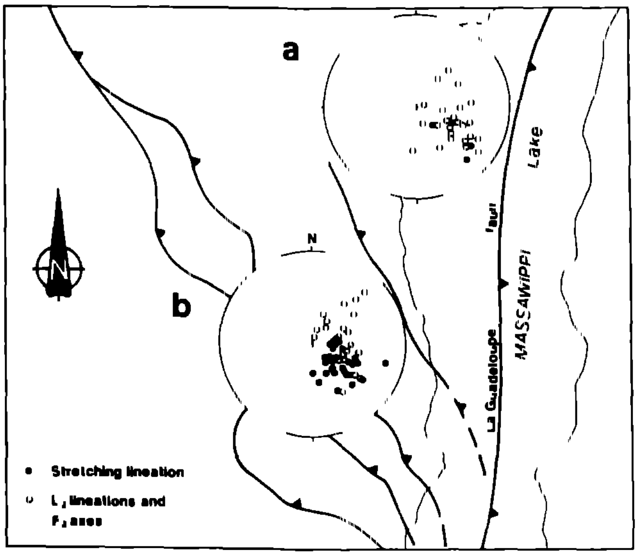

Fig 3 Lower hemisphere equal area stereographic plots of stretching lineations, F_2 axes and L_2 lineations along (a) the Ascot Complex and (b) the MLFZ See text for discussion

MLFZ in both Ascot and Magog rocks (Fig 2). The fault zone dips to the NE and contains SE plunging stretching lineations (Figs 3a & b), interpreted as parallel to the tectonic transport direction (e.g Sanderson 1973, Williams 1978, Ramsay 1980) Stretching lineations are parallel to those along the La Guadeloupe fault to the NE of the Massawippi lake (Labbé & St-Julien 1989, Tremblay *et al.* 1989b), which suggests that the MLFZ is a subsidiary fracture zone of the La Guadeloupe fault. In the studied area, F_2 axes are rotated into a unimodal clustering close to the mean stretching lineation (Figs 3a & b) and this is attributed to the development of sheath folds (Quinquis *et al* 1978, Cobbold & Quinquis 1980) Stretching lineations, structural relationships and kinematic analysis of the MLFZ (see below) indicate that it is a fault along which the Ascot Complex is thrust toward the NW over the Magog Group (Fig 4)

Although the actual geometry of the MLFZ shows a relatively common style of thrust related deformation, careful analysis of associated textures and structures suggests a complex history which could be easily overlooked

FAULT-RELATED FABRICS OF THE MLFZ

We use Sibson's (1977) nomenclature to classify the various types of fault rocks within the MLFZ The terms 'ductile' and 'plastic' are used in the sense of Rutter (1986) and Shimamoto (1989) and refer to the mechanical and microstructural aspects of the deformation.

Fault related rocks found within the LMFZ vary from brittle to ductile (Fig 5) However, most of the MLFZ is characterized by ductile deformation The mylonitic schist and phyllonite which form a sheared matrix between exotic slices show numerous *S-C* fabrics and porphyroclastic textures (Fig. 6a) at both the mesoscopic and microscopic scales. Sandstones of the Magog Group are also strongly sheared along the MLFZ and crystal-plastic deformation textures are well-developed (Fig. 6b). Serpentinite rocks belong to the 'incohesive serpentinite' type of Norrell *et al.* (1989) Meter to decameter scale blocks of undeformed serpentinite occur within scaly serpentinite zones. Serpentinite mylonites (Norrell *et al.* 1989) are locally present. Fault-related fabrics in granitic rocks of the MLFZ vary from cataclastic to mylonitic (Fig. 5), and are the focus of our study.

BRITTLE DEFORMATION FABRICS

Brittle deformation fabrics of the MLFZ are characterized by cataclastic structures. Granitic rocks range from fractured microgranite to cataclastic breccia (Fig. 5a). The protolith is an equigranular, fine grained (< 1 mm) granitic rock made up of quartz, plagioclase and microcline, plus muscovite, epidote and zircon as secondary minerals. Quartz filled microfractures are sparsely distributed (Fig. 5a).

Brittle deformation products of granitic rocks vary from protocataclasite to cataclasite (Figs. 6c & d). A fine grained matrix, containing grains of 50 μm and less in diameter, makes up to 50% of the rock volume. The matrix consists of sericite and crushed fragments of quartz and feldspar. A variable amount of ferruginous and insoluble material is also present. A crude foliation, interpreted as an incipient pressure solution cleavage, is sometimes visible in the matrix (Figs. 6d & e). Angular fragments of granitic composition range in size from 100 μm to more than 1 cm. Fragments are mono- or polymineralic and are often fractured. Most fractures are intragranular although some are also found in the matrix. Some fragments show an internal cataclastic texture.

Millimeter- to centimeter-wide quartz veinlets form a dense network of filled fractures in cataclastic samples. These veinlets cross cut the matrix and fragments (Fig. 5a), but are locally found in cataclastic particles (Fig. 6c). Induration of the cataclastic fabric prior to veining is inferred from the sharp contact between the host cataclasites and most of the quartz veinlets.

Extensive quartz veining is good evidence for hydrothermal activity along fault zones (Stel 1981). Vein filling in cataclasites can be derived by pressure solution, introduced during deformation by high fluid pressures, or associated with post kinematic hydraulic fracturing (Groshong 1988). The close relationship between quartz veinlets and the intensity of cataclasis suggests the presence of fluids during the brittle deformation. Occurrence of fragments of cataclasites and quartz veins in the cataclastic breccia of the MLFZ indicates multiple stages of fracturing, veining and gouge formation associated with the cyclic development of the brittle fault rocks.

PLASTIC DEFORMATION FABRICS

Intracrystalline plastic deformation is found in both the cataclastic and mylonitic rocks. All quartz crystals observed in fault rocks of the MLFZ are plastically deformed even in the cataclasites. The nature and intensity of plastic deformation textures of quartz are variable, but are all attributed to ductile faulting along the MLFZ.

Plastic deformation in cataclastic rocks

In cataclastic rocks, plastic deformation textures include ubiquitous undulose extinction, deformation bands and deformation lamellae. A well developed core-and mantle structure (White 1976) is visible in all quartz veinlets cross-cutting the cataclasites (Fig. 6f). Quartz crystals of 100–600 μm have core domains that vary from 80 to over 500 μm in diameter and are mantled by subgrains and new grains of up to 50 μm. Dynamic recovery and recrystallization are responsible for the development of the core-and-mantle structure (White 1976, 1977). Increasing strain leads to an increase of dynamic recrystallization over dynamic recovery (White 1977). In the quartz veinlets of the MLFZ rocks, new grains predominate over subgrains and indicate the

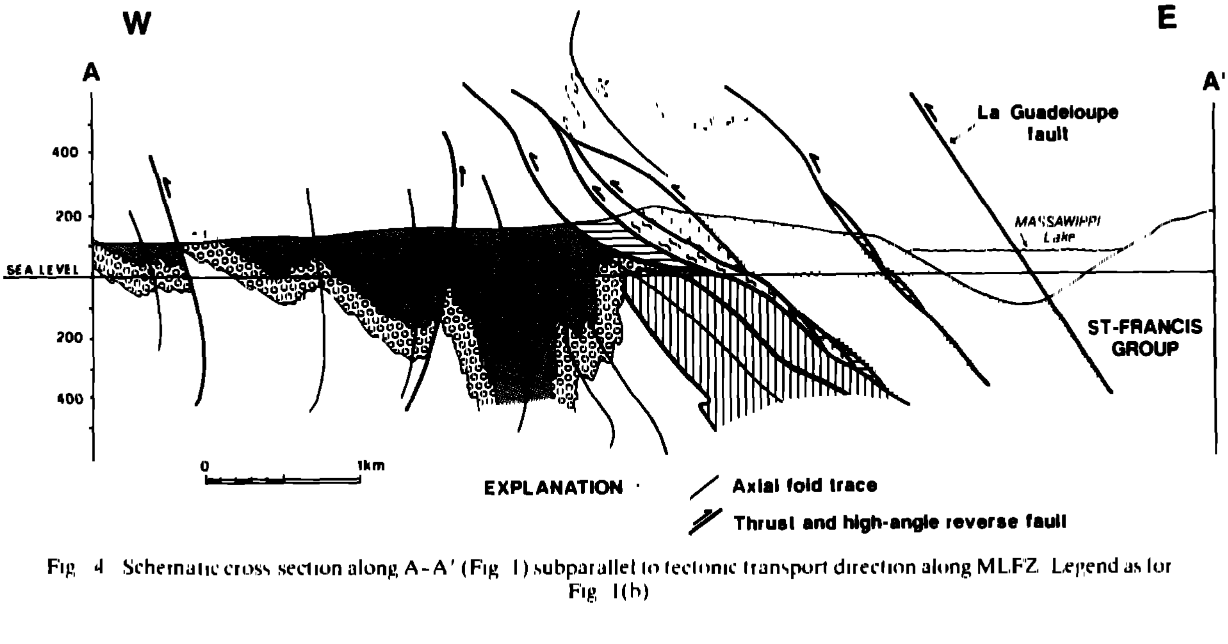

Fig. 4. Schematic cross section along A–A' (Fig. 1) subparallel to tectonic transport direction along MLFZ. Legend as for Fig. 1(b).

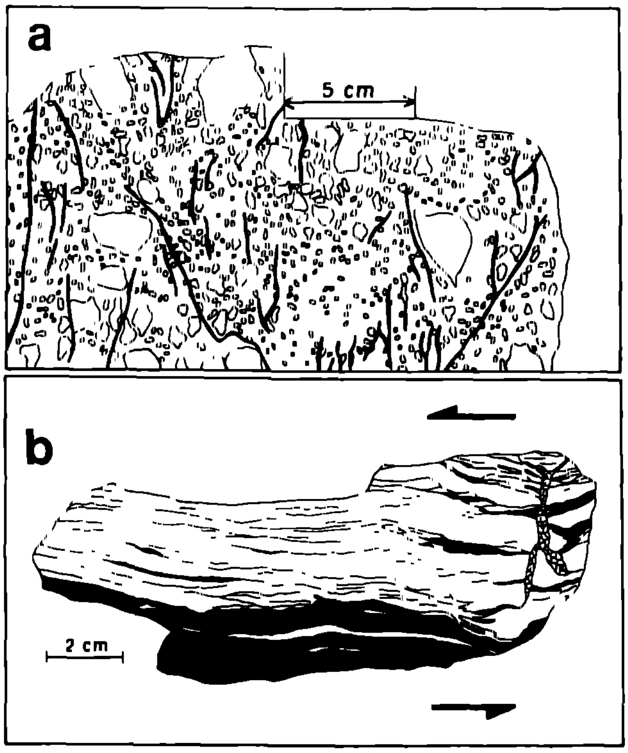

Fig. 5. Hand sketches of samples from two end member fault rocks of the MLFZ. (a) Cataclasite of granitic composition. Quartz filled veins are shown in black. (b) Granitic mylonite shown in section parallel to the stretching lineation. The mylonitic foliation is defined by alternating quartzo feldspathic (white) and phyllosilicate (black) layers. Shear sense is deduced from c axis measurements shown in Fig. 8(a).

predominance of dynamic recrystallization. In more deformed samples, cores are made up of $< 10\ \mu m$ sized subgrains (Fig. 6f). Quartz grains are elongate and their long axes define a foliation plane that cross cuts the veinlet margins, and that is also crudely defined by fractures and parallel sericite and muscovite flakes outside the veinlets. The contrasting plasticity of deformation within the quartz veinlets with that in the surrounding cataclastic rocks is attributed to a high proportion of more competent feldspar in the latter (e.g. Schmid 1982, Olsen & Kohlstedt 1985).

Plastic deformation in mylonitic rocks

The granitic mylonites are characterized by a ribbon structure defined by alternating phyllosilicate and quartzo feldspathic layers (Fig. 5b). Phyllosilicate domains of sericite and muscovite range from 100 to 300 μm wide, and quartzo feldspathic ribbons vary from less than 200 μm up to 1 mm in width. More than 80% of the ribboned felsic material consists of quartz. Feldspar is present as large porphyroclasts, rarely as minute crystals, that show evidence of kinking and lattice bending (Fig. 7a). Tullis & Yund (1987) attributed this type of feldspar deformation to the cataclastic flow regime which is effective in feldspars up to middle amphibolite P-T conditions. Simpson (1985) found similar fabrics in feldspars from upper greenschist facies shear zones.

Quartz grain size varies from 10 to 50 μm within the ribbons. Smaller quartz grains (less than 5 μm) are found in phyllosilicate rich ribbons where they often form an elongate mosaic of microcrystalline quartz. Two texturally distinct types of quartz ribbons are recognized, those in which quartz forms elongate grains and those in which it forms equant grains.

Ribbons of elongate quartz are made up of strongly deformed quartz crystals that range in size from 10 to 20 μm (Fig. 7b). They form large ribbons with occasional feldspar porphyroclasts. Quartz crystals show a well developed grain shape alignment (Fig. 7b). c axis measurements over this quartz population give a single girdle concentration of maxima (Fig. 8a), similar to a type I girdle distribution (Lister & Williams 1979), which obliquely cross cuts the foliation trace. The shear sense inferred from c axis measurements is consistent with other kinematic indicators (S-C fabrics, asymmetrical porphyroclasts) found along the MLFZ. The c axis fabric in Fig. 8(a) is defined by maxima and submaxima near the inferred Z and Y axes of the finite strain ellipsoid (where X is parallel to the stretching lineation of the sample, Y is perpendicular to X in the foliation plane, and Z is the normal to the XY plane). A weaker submaximum is developed near Y. This is interpreted as evidence for basal, rhomb and prismatic slip active within the quartz crystals (Bouchez & Pêcher 1981). Rhomb and prismatic submaxima are attributed to higher temperature conditions of deformation ($> 350°C$, Watts & Williams 1979, Bouchez & Pêcher 1981). This type of c axis fabric reveals a strong preferred crystallographic orientation in response to a plastic deformation regime (Schmid 1982).

Ribbons of equigranular quartz are made up of nearly rectangular quartz crystals of 20–50 μm grain size (Fig. 7c) which form composite ribbons separated by thin layers of phyllosilicates or opaque minerals. Grain boundaries show numerous triple point junctions, rarely at 120°. Quartz grains are strain free except for slight undulose extinction. Presence of 'pinning' and 'dragging' microstructures (Jessel 1987) point to the importance of grain boundary migration (GBM, Fig. 7d). Long axes of crystals are sub perpendicular to the ribbon margins. c axis orientations are different from those in ribbons of elongate quartz (Fig. 8b), a point maximum is orthogonal to a weak girdle of c axes.

In quartz rich mylonites, a fine grained, optically strain free, and equidimensional grain shape has been taken to indicate grain boundary sliding (GBS, White 1977, Vernon *et al.* 1983). However, similar microstructures could be obtained by static recrystallization which will generally destroy the crystallographic fabric unless the recovery is metadynamic (Jessel 1987), in which case it could result in oriented grain growth in high temperature tectonites (Culshaw & Fyson 1984). In the equigranular quartz ribbons, grain boundaries have not achieved an equilibrium configuration, which rules out a thermally activated static annealing after deformation. If accommodated by dislocation flow, as is often the case for relatively high temperature quartz rich tectonites (Etheridge & Wilkie 1979, Kronenburg & Tullis

1984) GBS could still preserve a crystallographic orientation. Equigranular quartz ribbons are therefore attributed to both GBM and GBS, the latter being aided by the presence of thin mica layers along some of the quartz boundaries which results in a decreasing intensity of a component of the c axis fabric (compare Figs. 8a & b). Similar conclusions were reached by Simpson (1983) for the ductile part of granitic shear zones of the Maggia Nappe in Switzerland.

DISCUSSION

c axis preferred orientations in quartzo feldspathic mylonites of the MLFZ point to the importance of dislocation creep as a deformation mechanism. Microstructures and related c axis fabrics are interpreted as the product of grain boundary migration and sliding. Experiments performed by Drury & Humphreys (1988) show that GBM and GBS develop together at relatively high temperatures, when deformation is still predominantly accommodated by intragranular dislocation creep. GBS is often equated with a superplastic regime of deformation (Boullier & Guegen 1975, Schmid *et al.* 1977, Behrmann 1985, Groshong 1988). Our interpretation of textures and crystallographic fabrics observed in mylonites from the MLFZ (Figs. 7 and 8) suggests that the strain conditions for superplasticity were present within the fault zone. c axis orientations in ribbons of equigranular quartz are viewed as altered imprints of those from ribbons of elongate quartz (Fig. 8). It has been argued that a fine grain size (< 10 μm) is necessary for GBS to occur (Behrmann 1985). However, Schmid *et al.* (1977) stressed that superplastic flow is mainly dependant on strain rate and temperature (i.e. superplasticity occurs at larger grain size (10 - 100 μm) at higher temperatures). Superplastic deformation in quartzite is normally associated with crystal-plastic strain at temperatures above 250°C (Behrmann 1985) or 400°C (Schmid *et al.* 1977, Etheridge & Wilkie 1979). Boullier & Guegen (1975) stated that a temperature limit of T/T_m > 0.5 (where T_m is the melting temperature of the mineral) is necessary. This is confirmed by Drury & Humphreys (1988) where $T' = 0.6$–$0.9\ T_m$.

The occurrence of plastic deformation overprinting the quartz veinlets in cataclastic rocks is worthy of discussion. In brittle-plastic shear zones, cataclastic deformation frequently succeeds a dominantly ductile deformation (Brock & Engelder 1977, Mitra 1984, O'Neill & Pavlis 1988), although the reverse is also described (Segall & Simpson 1986, Simpson 1986). Plastic over brittle deformation can be attributed to an increase in confining pressure and/or variations in fluid pressure (Simpson 1986), to a pre existing fracture dependency for the onset of ductile shear zones (Segall & Simpson 1986), or to fault reactivation.

In the Massawippi Lake area, tectonic movement along the La Guadeloupe fault could be interpreted as the cause for an increase in confining pressure. This is considered as an unlikely mechanism because the MLFZ and the La Guadeloupe fault are synchronous. Hence, tectonic overloading associated with the La Guadeloupe fault would be balanced by tectonic movement along the MLFZ. Fluids play an important role in the rheological response of rocks. A decrease in fluid pressure could result in an increase of both the strength and the ductility of the rock (Simpson 1986). In the MLFZ, most of the veinlet emplacement occurred after the formation of cataclastic breccia (Fig. 5a) which shows no evidence for the presence of extensive contemporary fluid circulation. In quartzo feldspathic rocks, cataclasite series rocks usually form at low temperature (< 250°C), below the greenschist grade of metamorphism (Watts & Williams 1979). The hypothesis retained here is that the plastic deformation is kinematically unrelated to the brittleness of the fault zone.

Significance of brittle and ductile fabrics of the MLFZ

Sibson (1977) attributes the coexistence of brittle and ductile fabrics within a single fault zone to textural variations related to the structural depth of formation. Ductile fault rocks could be uplifted to higher crustal levels, thus permitting mylonitic fault rocks to reach the brittle field of deformation. This implies that, in the footwall of a thrust such as the MLFZ, brittle deformation fabrics are superimposed on plastic fabrics. This is expressed by the presence of cataclastic rocks partly made up of fragments in which ductile fabrics are preserved (Sibson 1977, Gibson & Gray 1985, Stel 1986, O'Neill & Pavlis 1988). Obviously, this model cannot be applied to the MLFZ.

Another model attributes fabric heterogeneities to variable structural responses across the fault zone. The model predicts the existence of a dominantly ductile decollement zone bordered by less deformed country rocks (Eisbacher 1970, Mitra 1984, Simpson 1986). This model is incompatible with the observed structures and textures, and with the geometric relationships of regional deformation associated with the MLFZ.

Figure 9 shows the operative deformation mechanisms in a stress-temperature field diagram compiled from Groshong (1988). A deformation mechanism association is defined by deformation mechanisms active under the same environmental conditions in the same rock type. Various associations are here divided into low-temperature (fields I-III) and moderate temperature regimes (fields IV and V). The transition from low to moderate temperature regimes defines the tectonite front (TF, Groshong 1988). Groshong *et al.* (1984) found this transition to be at about 175°C. Transition between fields IV and V is here defined as the superplastic front (SF) which is also temperature dependant.

Fabrics found in granitic fault rocks of the MLFZ indicate that they occupy fields II, IV and V (Fig. 9). Cataclastic rocks of the MLFZ contain a well developed network of quartz filled fractures and intragranular and intergranular microfaults leading to cataclastic crushing (Figs. 5a and 6c), which places them within field II (Fig.

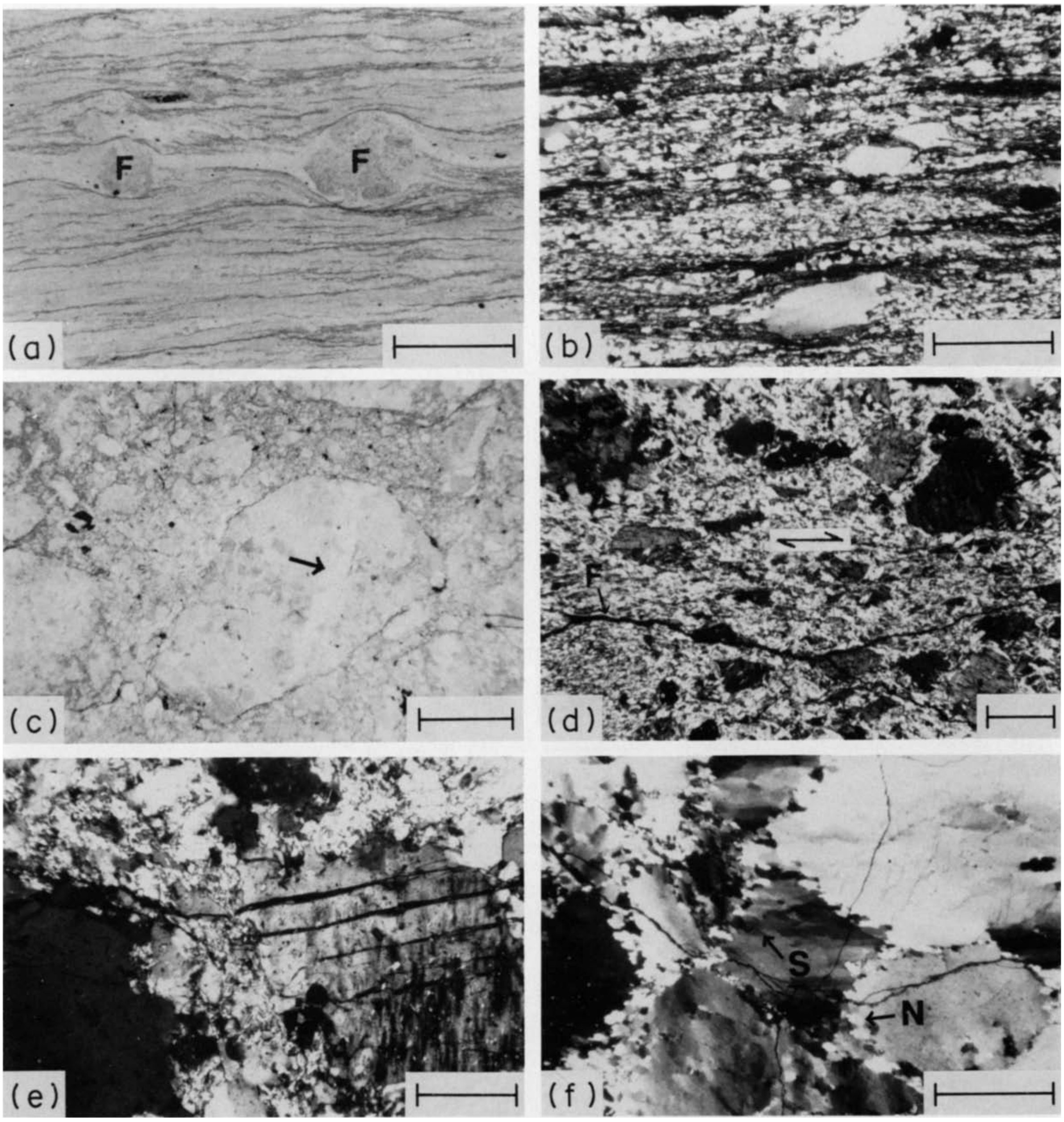

Fig. 6. Photomicrographs of fault rocks of the MLFZ. (a) & (b) are parallel to stretching lineation. (a) Felsic mylonitic schist showing σ-type asymmetrical feldspar porphyroclasts (F) suggesting a right-lateral shear sense. (b) Sheared sandstone of the Magog Group. Relict porphyroclastic quartz indicate a right-lateral shear sense. (c) Granitic cataclasite with angular fragments of microgranite in a crushed matrix. Note intragranular quartz veinlet (arrow). (d) Granitic cataclasite showing a crude foliation defined by fracture (F) and sericite flakes (double arrow). (e) Intergranular brittle fracture similar to that in (d) within a granitic cataclaste. (f) Core and mantle microstructure within a quartz veinlet cross-cutting the cataclasites as in Fig. 5(a). Quartz cores contain subgrains (S) and are mantled by new grains (N). (a) & (c) Plane light, (b), (d), (e) & (f) crossed polars. Scale bars for (a) & (b) = 1 mm, (c) = 2 mm, (d) = 260 μm, (e) = 150 μm, (f) = 500 μm.

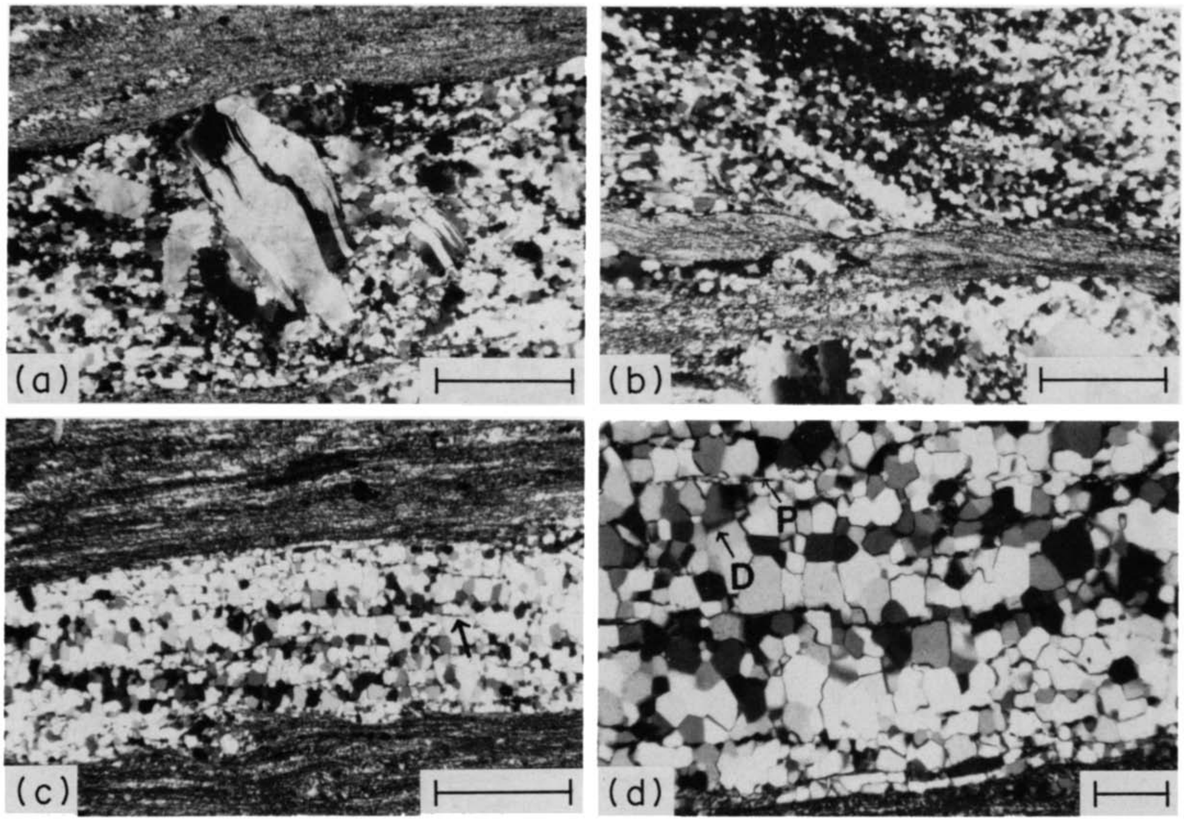

Fig. 7. Photomicrographs of granitic mylonites of the MLFZ, all with crossed polars and parallel to the stretching lineation. (a) Plastic deformation of a feldspar porphyroclast within an elongate quartz ribbon. Feldspar grain shows lattice bending and kinking. (b) Elongate quartz ribbon. Note well developed grain shape alignment indicating a left lateral sense of shear. (c) Equigranular quartz ribbon. Sub-rectangular quartz grains form composite foliation parallel ribbons separated by linear impurities (arrow). (d) Close view of the equigranular quartz ribbon shown in (c). Note that quartz crystals are strain free. 'Pinning' (P) and 'dragging' (D) microstructures point to a grain boundary migration mechanical process. Scale bars (a)-(c) = 500 μm, (d) = 100 μm.

9) Environmental conditions for cataclasis are those of the elastico frictional regime (Sibson 1977) which corre sponds to metamorphic temperatures and pressures lower than the greenschist grade (House & Gray 1982) Crystal plastic deformation seen in the quartz veinlets must originate from mechanisms characteristic of field IV (Fig 9) Intracrystalline plasticity in quartz crystals is not activated at temperatures below 250–300°C (e g Watts & Williams 1979) and there is no evidence that dislocation glide will permit large strain before and during cataclasis (White 1976). Textures found in mylo nites of the MLFZ belong to fields IV and V (Fig 9), as shown by well-developed crystallographic fabrics (Fig 8) and grain shape preferred geometry (Fig. 7b) as well as GBS and GBM microstructures. Moreover, textures found in feldspar grains within the mylonites are indica tive of a completely ductile bulk rock behavior which, according to Simpson (1985), is effective when feldspars begin to undergo plastic deformation at a middle to upper greenschist facies range of metamorphism.

If all these structures are related to a single faulting event, the dominant deformation-mechanism associ ation for quartzo feldspathic rocks evolved gradually from field II to field V, i.e. from temperatures < 175°C up to temperatures over 250–300°C. As the regional metamorphism is at greenschist grade (> 300°C, Winkler 1979), it is improbable that cataclastic deformation of field II (< 150°C) was developed at this metamorphic grade Cataclasis must have preceded the burial of these rocks Moreover, structural analysis of the area suggests that the MLFZ is a syn or a late metamorphic fault zone and there is no evidence for an extensive retrograde metamorphism along the fault zone Therefore, it is unsustainable that the youngest faulting event along the MLFZ began at temperatures below greenschist grade Thus, we suggest that cataclastic fabrics found within the MLFZ are relict textures preserved in tectonic slices of granitic rocks within a predominantly ductile fault zone

The existence of relict cataclastic fabrics within the MLFZ does not necessarily point to a fault reactivation history Cataclasites are not restricted to major fault zones (Groshong 1988). On the other hand, the MLFZ and other structural lineaments of the Quebec Appala chians have been suggested to be the locus of Taconian or Acadian strike slip faulting (Gauthier *et al* 1989). However, the kinematic analysis of the MLFZ indicates that it is an Acadian oblique-slip thrust fault Tremblay & St-Julien (1990) argued that Taconian deformation within Cambro Ordovician rocks of the Dunnage Zone is minimal and restricted to accretionary processes within the St Daniel Mélange (Fig 1a). West of the Baie Verte-Brompton Line, intense Taconian deformation is found and is attributed to the subduction of basement and cover rocks of a Cambro-Ordovician continental margin (Stanley & Ratcliffe 1985) Doolan *et al.* (1982) have proposed that an oblique collision along an irregu

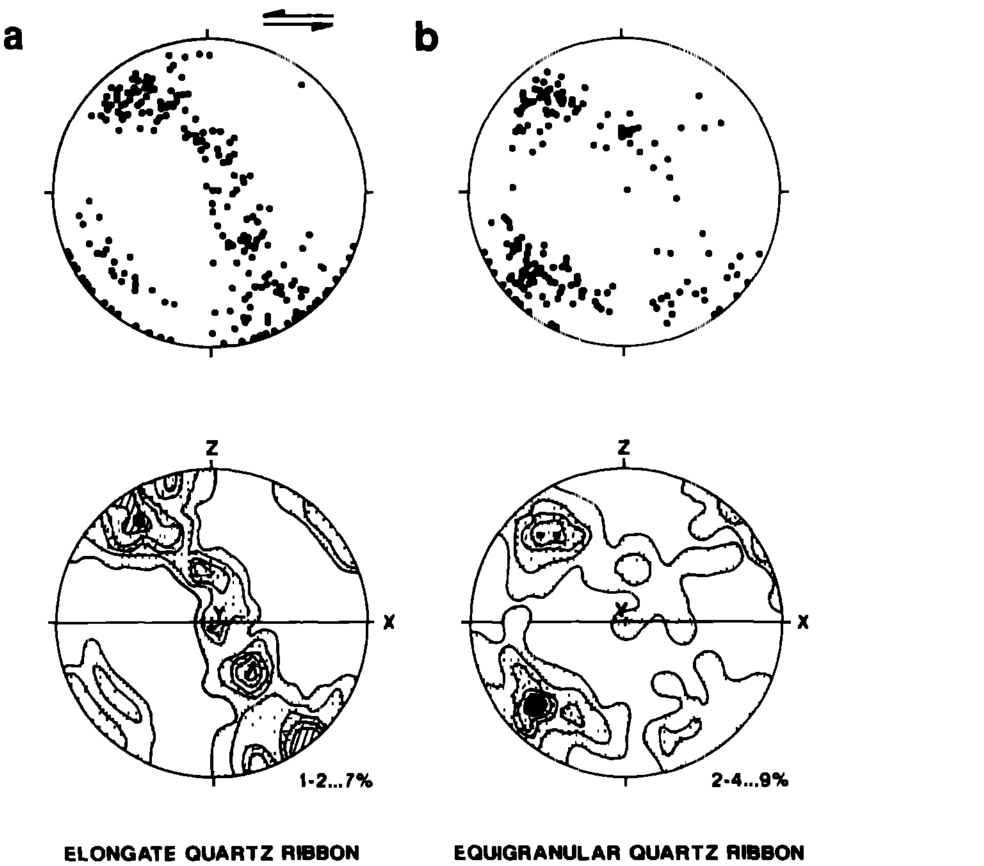

Fig 8 Lower hemisphere equal area stereographic plots of *c* axis measurements (above, $N = 250$) and contoured diagrams (below) from quartz ribbons of MLFZ granitic mylonites *X* axis (stretching lineation) horizontal and foliation plane vertical (a) Elongate quartz ribbon Note well defined type I girdle and the deduced shear sense (b) Equigranular quartz ribbon

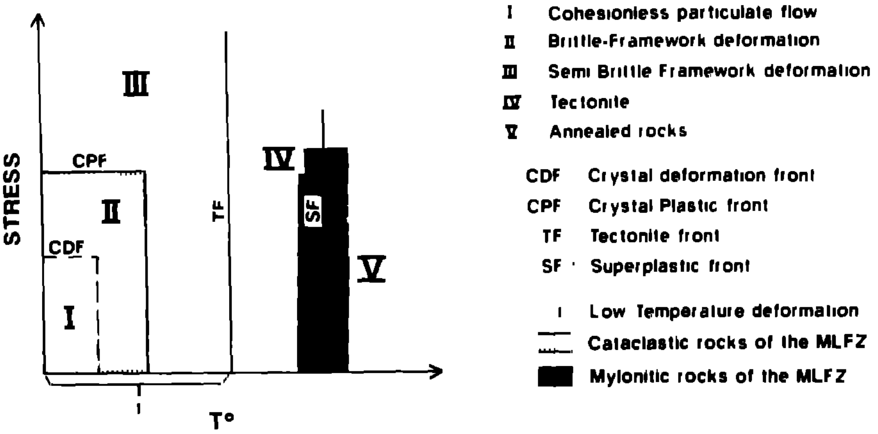

Fig. 9. Schematic deformation mechanisms association diagram for quartzo-feldspathic rocks compiled from Groshong (1988). Each association (I–V) is defined by the activity of specific deformation mechanisms (see text). Fault rocks of the MLFZ belong to fields II, IV and V.

lar continental margin could explain variations in folding and faulting styles of rocks of the Humber Zone. This irregular margin geometry could have been juxtaposed against tectonic fractures and discontinuities within the oceanic domain (Dunnage Zone). Multiple Taconian stages of fracturing and veining have been recognized within ultramafic and associated granitic rocks of the Asbestos and Thetford Mines ophiolitic complexes (O'Hanley 1987). This brittle behavior is probably associated with the Taconian tectonic accretion and obduction of oceanic crust–mantle. It is also significant that almost all the granitic bodies found with serpentinites along the Baie Verte–Brompton Line are brittley deformed (P. St Julien personal communication). Consequently, we think that if there was fault reactivation along the MLFZ, the earlier event could be interpreted as an Ordovician oceanic fracture activated prior and during the Taconic orogeny.

CONCLUSIONS

Fault rocks of the MLFZ show dynamic inconsistencies which are best resolved by interpreting them as the result of multiple stage faulting. Cataclastic rocks belong to a 'brittle-framework deformation' field characteristic of an elastico frictional regime below greenschist grade. Mylonitic rocks show microstructures and c axis orientations characteristic of the 'tectonite' and 'annealed tectonite' field of deformation. Mylonitic fabrics were developed under plastic and superplastic condition of strain in mid- to upper greenschist grade of metamorphism. The regional structural and metamorphic history is consistent with the deformational regime and deformation mechanisms found in mylonitic rocks, but is inconsistent with those found in cataclastic rocks.

We therefore conclude that quartzo-feldspathic fault rocks of the MLFZ contain coexisting, but unrelated, brittle and plastic fabrics. There is much more to learn about deformation mechanism paths and the reliability and stability of microstructures (Knipe 1989), however we believe that the interpretation presented here is a valuable way to analyse other so-called brittle–plastic transition shear zones.

Acknowledgements—We wish to acknowledge the Ministère de l'Energie et des Ressources du Québec which defrayed most of the field expenses and permitted the use of the data. We thank Donna Kirkwood, Pierre St Julien, Réal Daigneault and Marc Bardoux for their reviews of an earlier version. Formal reviews by J. Hadizadeh and H. Stel were useful and greatly improved the manuscript, we thank them both. Editorial review by C. Simpson has enhanced the quality of the manuscript. A. Tremblay acknowledges the Institut National de la Recherche Scientifique for a post doctoral scholarship during which this study has been conducted. NSERC of Canada is acknowledged for an operating grant (GP 1908) to M. Malo. Thanks are due to Luce Dubé and André Hébert for technical assistance.

REFERENCES

Behrmann, J. H. 1985. Crystal plasticity and superplasticity in quartzite, a natural example. *Tectonophysics* **115**, 101–129.

Bouchez, J. L. & Pêcher, A. 1981. The Himalayan main central thrust pile and its quartz rich tectonites in central Nepal. *Tectonophysics* **78**, 23–50.

Boullier, A. M. & Guegen, Y. 1975. SP mylonites: origin of some mylonites by superplastic flow. *Contr. Miner. Petrol.* **50**, 93–104.

Brock, W. G. & Engelder, T. 1977. Deformation associated with the movement of the Muddy Mountain overthrust in the Buffington window, southeastern Nevada. *Bull. geol. Soc. Am.* **88**, 1667–1677.

Cobbold, P. R. & Quinquis, H. 1980. Development of sheath folds in shear regime. *J. Struct. Geol.* **2**, 119–126.

Cousineau, P. A. 1988. Analyse tectonostratigraphique de la Zone de Dunnage, à l'est de la rivière Chaudière, Québec. Unpublished Ph.D. thesis, Laval university, Québec, Canada.

Culshaw, N. G. & Fyson, W. K. 1984. Quartz ribbons in high grade granite gneiss: modification of dynamically formed quartz c axis preferred orientation by oriented grain growth. *J. Struct. Geol.* **6**, 663–668.

Doolan, B. L., Gale, M. H., Gale, P. N. & Hoar, R. S. 1982. Geology of the Quebec Re-Entrant: possible constraints from early rifts and the Vermont–Quebec serpentine belt. In: *Major Structural Zones and Faults of the Northern Appalachians* (edited by St Julien, P. & Béland, J.). *Spec. Pap. Geol. Ass. Can.* **24**, 87–116.

Drury, M. R. & Humphreys, F. J. 1988. Microstructural shear criteria associated with grain boundary sliding during ductile deformation. *J. Struct. Geol.* **10**, 83–89.

Eisbacher, G. H. 1970. Deformation mechanics of mylonitic rocks and fractured granites on Cobequid Mountains, Nova Scotia, Canada. *Bull. geol. Soc. Am.* **81**, 2009–2020.

Etheridge, M. A. & Wilkie, J. C. 1979. Grainsize reduction, grain boundary sliding and the flow strength of mylonites. *Tectonophysics* **58**, 159–178.

Flinn, D 1977 Transcurrent faults and associated cataclasis in Shetland *J geol Soc Lond* **133**, 231-248

Gaudemer, Y & Tapponnier, P 1987 Ductile and brittle deformations in the northern Snake Range, Nevada *J Struct Geol* **9**, 159-180

Gauthier, M , Auclair, M , Bardoux, M , Blain, M , Boisvert, D , Brassard, B , Chartrand, F , Darimont, A , Dupuis, L , Durocher, M , Ganepy, C , Godue, R , Jebrack, M & Trottier, J 1989 Synthese géologique de l'Estrie et de la Beauce Ministere de l'Energie et des Ressources du Quebec, MB 89 20

Gibson, R G & Gray, D R 1985 Ductile to brittle transition in shear during thrust sheet emplacement, Southern Appalachian thrust belt *J Struct Geol* **7**, 513-525

Groshong, R H Jr 1988 Low temperature deformation mechanisms and their interpretation *Bull geol Soc Am* **100**, 1329-1360

Groshong, R H Jr, Pfiffner, O A & Pringle, L R 1984 Strain partitioning in the Helvetic thrust belt of eastern Switzerland from the leading edge to the internal zone *J Struct Geol* **6**, 5-18

House, W M & Gray, D R 1982 Cataclasites along the Saltville thrust, U S A and their implications for thrust sheet emplacement *J Struct Geol* **4**, 257-269

Jessel, M W 1987 Grain boundary migration microstructures in a naturally deformed quartzite *J Struct Geol* **9**, 1007-1014

Knipe, R J 1989 Deformation mechanisms—recognition from natural tectonites *J Struct Geol* **11**, 127-146

Kronenberg, A K & Tullis, J 1984 Flow strengths of quartz aggregates grain size and pressure effects due to hydrolytic weakening *J geophys Res* **89**, 4281-4297

Labbé, J Y & St Julien, P 1989 Failles de chevauchement acadiennes dans la region de Weedon, Estrie, Québec *Can J Earth Sci* **26**, 2268-2277

Laird, J , Lanphere, M A & Albee, A L 1984 Distribution of Ordovician and Devonian metamorphism in mafic and pelitic schists from northern Vermont *Am J Sci* **284**, 376-413

Lister, G S & Williams, P F 1979 Fabric development in shear zones theoretical controls and observed phenomena *J Struct Geol* **1**, 283-297

Mitra, G 1984 Ductile deformation zones and mylonites the mechanical processes involved in the deformation of crystalline basement rocks *Am J Sci* **278**, 1057-1084

Norrell, G T , Teixell, A & Harper, G D 1989 Microstructure of serpentinite mylonites from the Josephine ophiolite and serpentinization in retrogressive shear zones California *Bull geol Soc Am* **101**, 673-682

Obee, H K & White S H 1986 Microstructural and fabric heterogeneities in fault rocks associated with a fundamental fault *Phil Trans R Soc Lond* **A317**, 99-109

O'Hanley, D S 1987 The origin of chrysotile asbestos veins in southeastern Quebec *Can J Earth Sci* **24**, 1-9

Olsen, T S & Kohlstedt, D L 1985 Natural deformation and recrystallization of some intermediate plagioclase feldspars *Tectonophysics* **111**, 107-131

O'Neill, R L & Pavlis, T L 1988 Superposition of Cenozoic extension on Mesozoic compressional structures in the Pioneer Mountains metamorphic core complex, central Idaho *Bull geol Soc Am* **100**, 1833-1845.

Quinquis, H , Audren, Cl , Brun, J P & Cobbold, P R 1979 Intense progressive shear in Ile de Groix blueschist and compatibility with subduction or obduction *Nature* **273**, 43-45

Ramsay, J G 1980 Shear zone geometry a review *J Struct Geol* **2**, 83-99

Rutter, E H 1986 On the nomenclature of mode of failure transition in rocks *Tectonophysics* **122**, 381-387

Sanderson, D F J 1973 The development of fold axes oblique to the regional trend *Tectonophysics* **16**, 55-70

Schmid, S M 1982 Microfabric studies as indicators of deformation mechanisms and flow laws operative in mountain building In *Mountain Building Processes* (edited by Hsu, K J) Academic Press, London, 95-110

Schmid, S M , Boland, J N & Paterson, M S 1977 Superplastic flow in finegrained limestone *Tectonophysics* **43**, 245-280

Segall, P & Simpson, C 1986 Nucleation of ductile shear zones on dilatant fractures *Geology* **14**, 56-59

Shimamoto, T 1989 The origin of *S-C* mylonites and a new fault zone model *J Struct Geol* **11**, 51-64

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

Simpson, C 1983 Strain and shape fabric variations associated with ductile shear zones *J Struct Geol* **5**, 61-72

Simpson, C 1985 Deformation of granitic rocks across the brittle-ductile transition *J Struct Geol* **7**, 503-511

Simpson, C 1986 Fabric development in brittle to ductile shear zones *Pure & Appl Geophys* **124**, 269-288

St Julien, P & Hubert, C 1975 Evolution of the Taconian orogen in the Quebec Appalachians *Am J Sci* **275-A**, 337-362

St Julien, P & Slivitzky, A 1985 Compilation geologique de la region de l'Estrie Beauce Ministere de l'Energie et des Ressources du Quebec, Map 2030 from Report MM 85 04

St Julien, P , Slivitzky, A & Feininger, T 1983 A deep structural profile across the Appalachians of southern Quebec *Mem geol Soc Am* **158**, 103-111

Stanley, R S & Ratcliffe, N M 1985 Tectonic synthesis of the Taconian orogeny in western New England *Bull geol Soc Am* **96**, 1227-1250

Stel, H 1981 Crystal growth in cataclasites diagnostic microstructures and implications *Tectonophysics* **78**, 585-600

Stel, H 1986 The effect of cyclic operation of brittle and ductile deformation on the metamorphic assemblage in cataclasites and mylonites *Pure & Appl Geophys* **124**, 289-307

Sutter, J F , Ratcliffe, N M & Musaka, S B 1985 $^{40}Ar/^{39}Ar$ and K-Ar data bearing on the metamorphic and tectonic history of western New England *Bull geol Soc Am* **96**, 123-136

Tremblay, A 1990 Geologie de la region d'Ayers Cliff (moitie Est) Ministere de l'Energie et des Ressources du Quebec, MB 90 30

Tremblay, A In press Synthese geologique de la region de Sherbrooke Ministere de l'Energie et des Ressources du Québec

Tremblay, A & St Julien, P 1990 Structural style and evolution of a segment of the Dunnage Zone from the Quebec Appalachians and its tectonic implications *Bull geol Soc Am* **102**, 1218-1229

Tremblay, A , Hebert, R & Bergeron, M 1989a Le Complexe d'Ascot des Appalaches du sud du Québec petrologie et géochimie *Can J Earth Sci* **26**, 2407-2420

Tremblay, A , St Julien, P & Labbe, J Y 1989b Mise a l'évidence et cinematique de la faille de La Guadeloupe, Appalaches du sud du Quebec *Can J Earth Sci* **26**, 1932-1943

Tullis, J & Yund, R A 1987 Transition from cataclastic flow to dislocation creep of feldspar mechanisms and microstructures *Geology* **15**, 606-609

Vernon, R H , Williams, V A & D'Arcy, W F 1983 Grain size reduction and foliation development in a deformed granitoid batholith *Tectonophysics* **92**, 123-145

Watts, M J & Williams, G D 1979 Fault rocks as indicators of progressive shear deformation in the Guingamp region, Brittany *J Struct Geol* **1**, 323-332

Williams, G D 1978 Rotation of contemporary folds into the X direction during overthrust processes in the Lakesfjord, Finnmark *Tectonophysics* **48**, 29-40

Williams, H 1979 Appalachian orogen in Canada *Can J Earth Sci* **16**, 792-807

Williams, H & St Julien, P 1982 The Baie Verte Brompton Line early Paleozoic continent ocean interface in the Canadian Appalachians In *Major Structural Zones and Faults of the Northern Appalachians* (edited by St Julien, P & Beland, J) *Spec Pap Geol Ass Can* **24**, 177-208

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

White, S 1977 Geological significance of recovery and recrystallization processes in quartz *Tectonophysics* **39**, 143-170

Winkler, H G F 1979 *Petrogenesis of Metamorphic Rocks* Springer, New York