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Fault rocks as indicators of progressive shear deformation in the Guingamp region, Brittany

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Abstract—The North Armorican Shear Zone is a major structural feature running from the island of Molene in the west to Moncontour in the east of the Armorican Massif. In the region of Guingamp it cuts through a Precambrian migmatite complex and granitoid rocks of both Precambrian and Hercynian age. A variety of fault rocks are present in this part of the shear zone, and are thought to represent a time sequence in which deep level, ductile deformation gave way to higher level brittle displacements. Mylonite and cataclasite series rocks, and pseudotachylites are described and their conditions of formation considered. The Hercynian Quintin granite post-dates the main movement of the shear zone but is itself dextrally displaced during the late stages of shear movement.

INTRODUCTION

THE REGION of Guingamp lies in the north-western part of the Armorican Massif of north-west France (Fig. 1). Descriptions of the geological framework of the Armorican Massif are given by Cogné *et al.* (1974) and Roach (1972, 1977), who recognized two fundamental Precambrian units; the Pentevrian (Cogné 1959) which is a pre-900 Ma crystalline basement complex, unconformably overlain by the Brioverian (*ca.* 900 – 650 Ma), a greenschist facies supracrustal sequence. The oldest reliable dates for the massif (2018 ± 15 Ma) were produced by Calvez & Vidal (1978) from the basement Icart Gneisses of Guernsey. Both the Pentevrian and the Brioverian underwent metamorphism and deformation during the Cadomian Orogeny (700 – 500 Ma).

Numerous syn- and post-tectonic granites were intruded in northern Brittany during Hercynian times. Notably, in the Guingamp region the so called Quintin granite has been dated at 340 ± 5 Ma by Adams (1967) and the Guingamp granite dated at 315 ± 15 Ma by Leutwein *et al.* (1968).

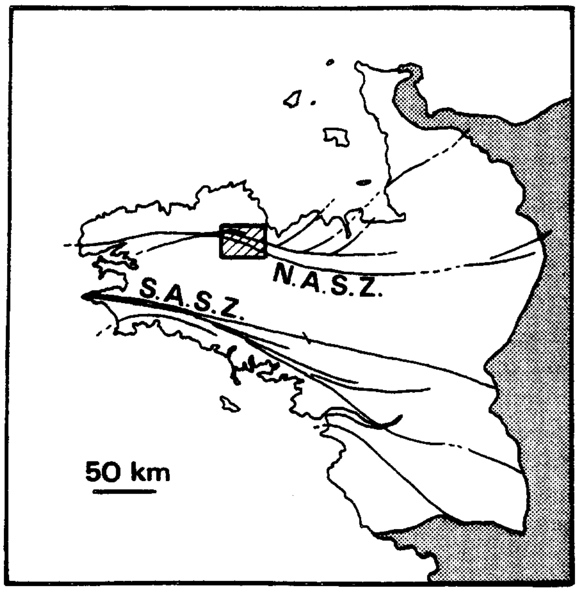


Fig. 1. Map of the North and South Armorican Shear Zone systems in the Massif, based on the maps of Cogné (1974); Guingamp area shaded.

Cogné *et al.* (1974) considered that the Armorican Massif was cut by two major systems of Hercynian shear zones, the North and South Armorican Shear Zones (Fig. 1). Chauris (1969) was the first person to map the extent of the North Armorican shear zone from the island of Molene, west of Brest, to the Hercynian Monocontour granite to the southeast of Guingamp. In this paper, rocks from a portion of the northern shear zone are described (Figs. 2 & 3), and their significance is discussed.

It is our aim to show how the fault rocks of Guingamp illustrate a history of deformation on all scales and at a variety of crustal levels. The fault rock classification proposed by Sibson (1977) is followed throughout; foliated and random-fabric fault rocks are taken to indicate generation in ductile (quasi-plastic) and brittle (elastico-frictional) regimes respectively.

PARENT ROCKS NOT DEFORMED IN THE SHEAR ZONE

The main shear zone comprises a belt 4 km wide within which many small tabular regions of high strain are found (Fig. 3). Fault rocks of the high strain regions are derived from a variety of parent rocks which are described below.

Migmatites of Guingamp

A variety of rock types are mapped as migmatites in the Guingamp area; true anatectic migmatites, injection magmatites and granitic gneisses are all described by this one term, and are considered to be Cadomian in age (Cogné 1976). Recently Brown *et al.* (1971) showed, on structural grounds that the St. Malo Migmatites, given a Cadomian age on survey maps, represent true Pentevrian basement. Brown (1978) has recognized two stages of Pentevrian migmatization of original metasediments

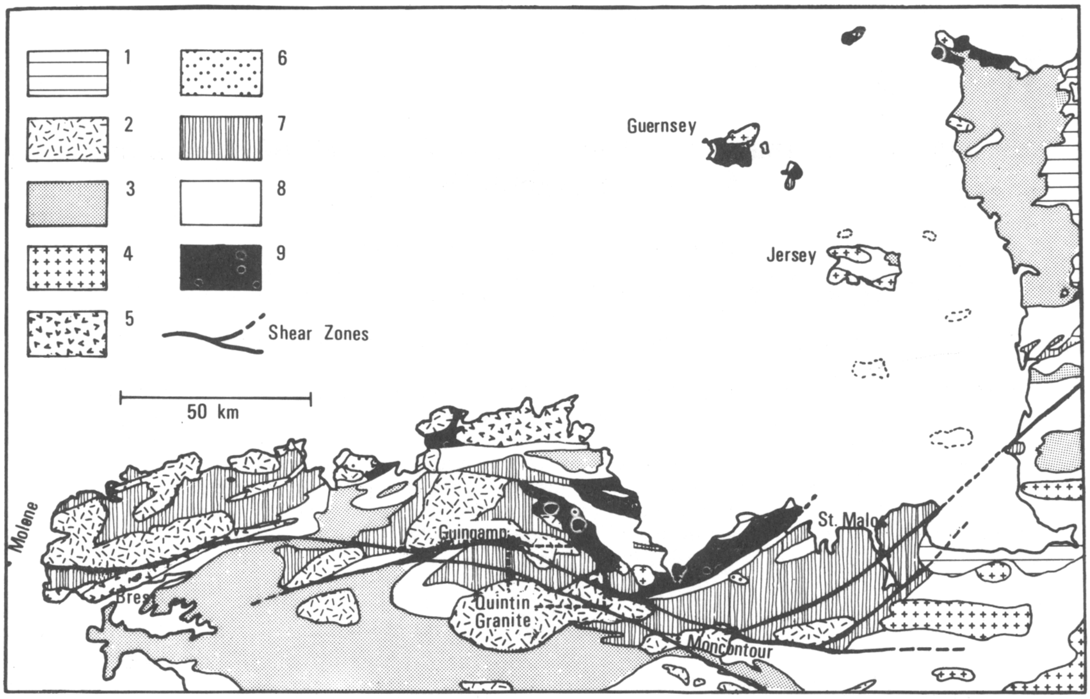


Fig. 2. Simplified geological map of the northern part of the Armorican Massif after Cogné (1974) and Roach (1977). Key: 1. Mesozoic sediments; 2. Hercynian granites; 3. Palaeozoic sediments; 4. Late and post-tectonic Cadomian igneous complexes; 5. Pre and syn-tectonic Cadomian intrusions; 6. Basic igneous masses; 7. Schists, amphibolites, gneisses and migmatites of uncertain age; 8. Brioverian; 9. Pentevrian basement.

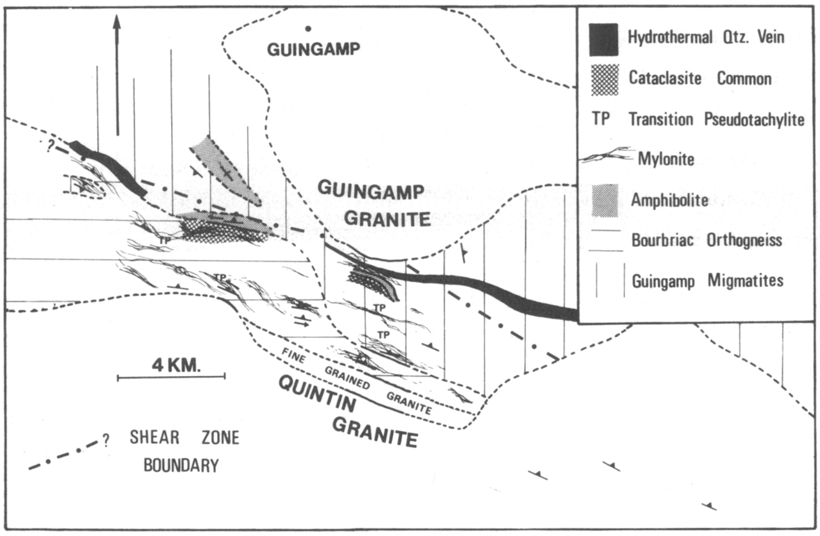


Fig. 3. Generalized geological map showing the distribution of the fault rocks throughout the Guingamp region.

before Brioverian sedimentation, although this conclusion is subject to debate as Brun & Martin (1979) interpreted the St. Malo Gneisses as a Cadomian dome.

Amphibolites

Schistose amphibolites derived from the metamorphism of basic intrusives are found within the migmatite complex. Generally the schistosity is steeply dipping and contains a sub-horizontal mineral lineation; the tectonic fabric was induced by a homogeneous deformation.

The amphiboles are magnesio-hornblendes (after Leake 1978), and their Al(vi)/Si ratios suggest that they were formed at a pressure of > 5 kbar, if the techniques of Raase (1974) are applied. It is thought that the Ti content of amphiboles is to some extent temperature dependent (Raase 1974), and on this basis, those of the Guingamp region appear to represent an upper greenschist/lower amphibolite facies metamorphism.

Granitoid rocks

Granitoid rocks included in the term 'Granite and Orthogneiss of Bourbriac' are assumed to be syntectonic Cadomian intrusives (Cogne 1976) and are usually porphyritic. They always contain a weak tectonic foliation. Hercynian two-mica granites are usually coarse grained and porphyritic, but finer grained varieties occur.

FAULT ROCKS OF THE GUINGAMP REGION

Fault rocks representing both ductile and brittle shear displacements are found in the Guingamp region. Mylonite and blastomylonite are indicative of ductile shear, whereas cataclasite and fault gouge result from brittle displacement. Pseudotachylite in this area possesses features common to both deformation regimes.

Mylonite series

Within the regional shear belt, protomylonites, mylonites and ultramylonites (Figs. 5a–c) derived from granitoid rocks lie within anastomosing shear zones which isolate pods of 'migmatite' and foliated Cadomian granite. Following Ramsay & Graham (1970), the mylonite foliation is thought to lie along the *XY* plane of the finite strain ellipsoid with a sub-horizontal stretching lineation tracing out the *X* direction (Fig. 4). This implies that the main ductile movement has been in a strike-slip sense.

Compositional banding of felsic and mafic layers often occurs in the mylonites and ultramylonites (Figs. 5b & c), and it is difficult to assess to what extent this is a relict gneissic banding or an induced banding formed during ductile deformation of a polymineralic rock with accompanying grain size reduction (Vernon 1974). The Guingamp mylonites, like the epidote amphibolite to greenschist facies mylonites of the Limpopo Belt (Wakefield 1977), reflect a ductility contrast between quartz and feldspar. Grain size reduction, characteristic of long-lived ductile shear zones (Watterson 1975) is accomplished by the dynamic recovery and recrystallization of strained quartz whose behaviour as a plastically flowing matrix leads to the brittle fragmentation and comminution of feldspar.

In contrast to the main mylonite development, deformation in the Quintin granite is restricted to a narrow belt and has given rise to protomylonite which has a similar structural trend to the mylonites of the main belt.

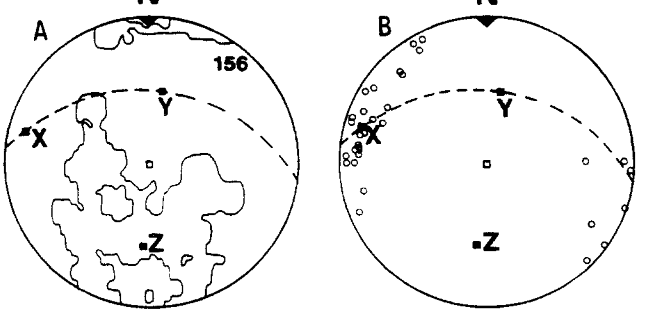


Fig. 4. (a) Contoured equal area plot of poles to the mylonitic foliation. (b) Equal area plot of stretching lineations. The mean orientations of *X*, *Y* and *Z* and the best fit great circle to the *XY* plane are shown.

In addition to the mylonitic foliation of the shear zone, other foliations recognized are:

S_1 Pre-mylonite tectonite foliations including gneissic banding, foliation in granitoids and schistosity in amphibolites, all being preserved in relict pods within mylonites (Fig. 5d). These may not be of equivalent age, but here are simply classed as pre-mylonite foliations.

S_{2A} Mylonite foliation.

S_{2B} A non-penetrative crenulation cleavage is sometimes developed at an angle of 30° to 45° to the mylonite foliation (Fig. 6). This feature has been described as shear band structure (White *et al.* 1978), and it appears to form after the development of the main mylonite foliation as a consequence of continued shear movement.

S_3 Commonly in ultramylonites, a penetrative crenulation cleavage is formed in micaceous layers at a high angle to the mylonitic foliation. This is sometimes seen to be axial planar to small scale tight folds which deform the mylonite foliation (Fig. 7).

Although quartz shows evidence of plastic flow in nature, and a mechanism for plastic deformation has been known since the thirties, it was not until the early sixties that such a mechanism was experimentally demonstrated (Carter *et al.* 1961, 1964). Specifically, the ductile deformation in shear zones was described by Ramsay & Graham (1970), and quartz and feldspar microstructures in mylonites were discussed in terms of plastic behaviour by Bell & Etheridge (1973) and White (1975). It is now fully appreciated that the property of quartz to deform plastically by intracrystalline slip and diffusion processes at higher temperatures plays a dominant role in the formation of quartzo-feldspathic mylonites (Wilson 1975, Nicolas & Poirier 1976, White 1976, 1977).

Throughout the Guingamp mylonite series rocks, quartz exhibits microstructural evidence of plastic behaviour; undulose extinction, deformation bands, sub-grains and elongate and ribbon quartz (Figs. 5e–g). The presence of new grains indicate that recrystallization has taken place, and using the criteria of Green *et al.* (1970), and White (1977), dynamically recrystallized grains can be distinguished from statically recrystallized grains.

In Fig. 5(e), a quartz grain shows inhomogeneous development of sub-grains formed during the recovery of a strained grain. The process of sub-grain rotation with increased lattice mismatch to produce discrete recrystallized grains (White 1977) has been operative as grains with high angle boundaries and of a similar size to the sub-grains are present. Core and mantle structure (White 1976) is seen (Fig. 5f) where undulose quartz of the core is replaced by a mantle of sub-grains and recrystallized grains. The progressive rotation of sub-grains within the highly strained outer edges of the quartz results in dynamic recrystallization. Other mylonites show less evidence of recovery dominated recrystallization, and have a higher proportion of ribbon

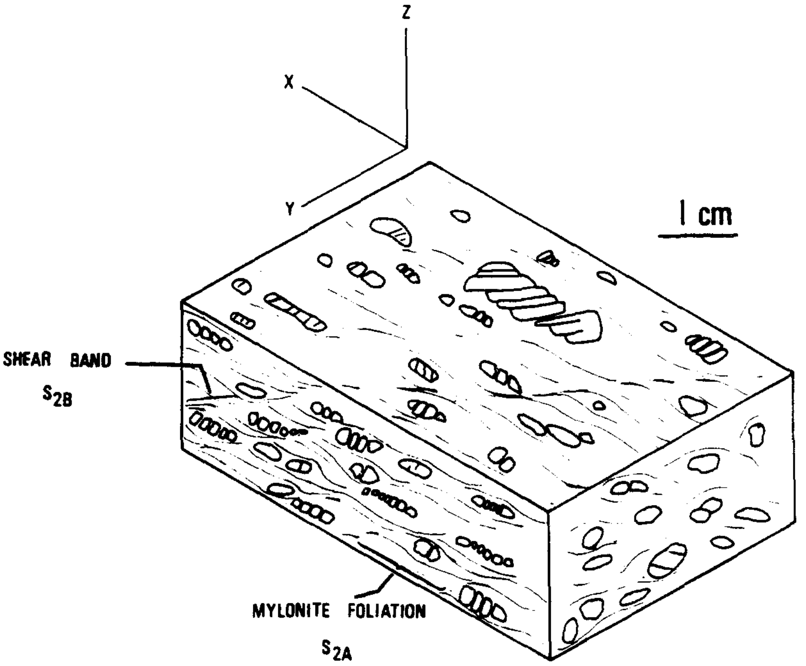


Fig. 6. Block diagram illustrating the mesoscopic fabric of a typical mylonite. The mylonite foliation S_{2a} contains a strong feldspar lineation and is cut by a weaker shear band foliation S_{2b}. The fractured felspar porphyroclasts show some post fracture rotation due to movement along S_{2b} planes.

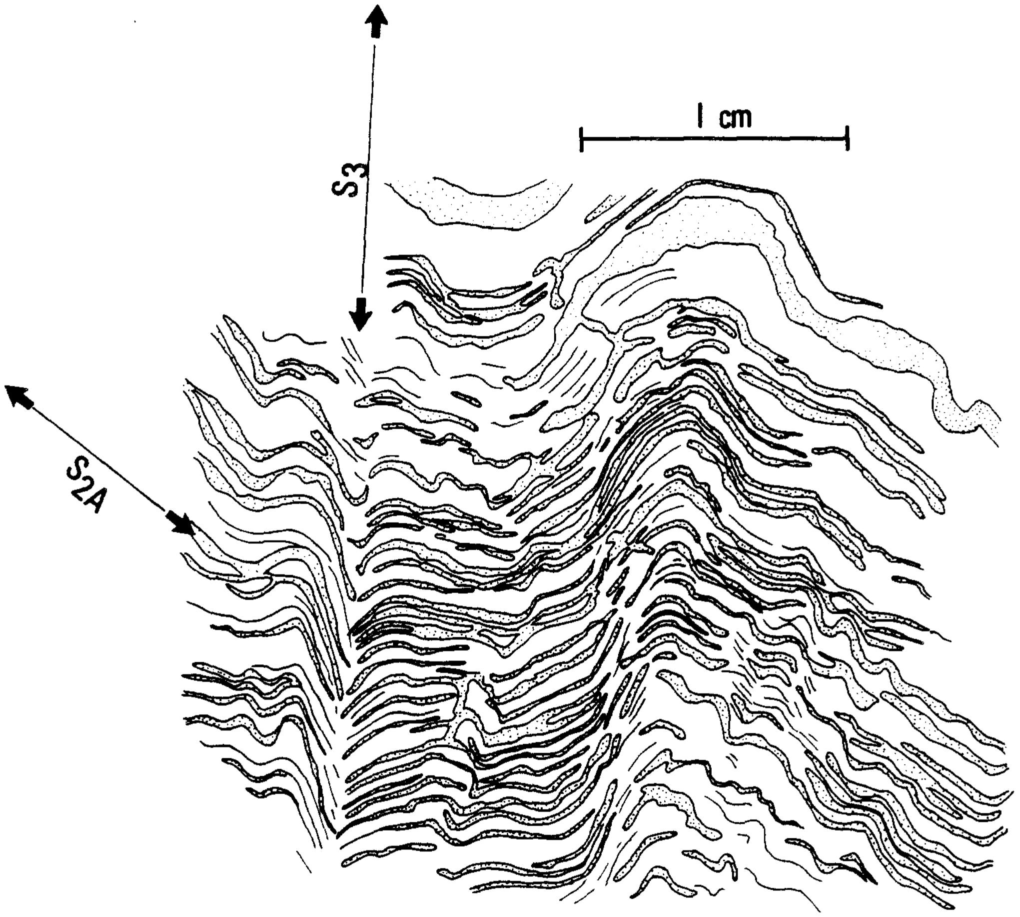


Fig. 7. Sketch of a folded ultramylonite, with thin mica rich and quartz rich (stippled) bands, showing the relationship between the mylonite foliation S_{2a} and the crenulation cleavage S_3.

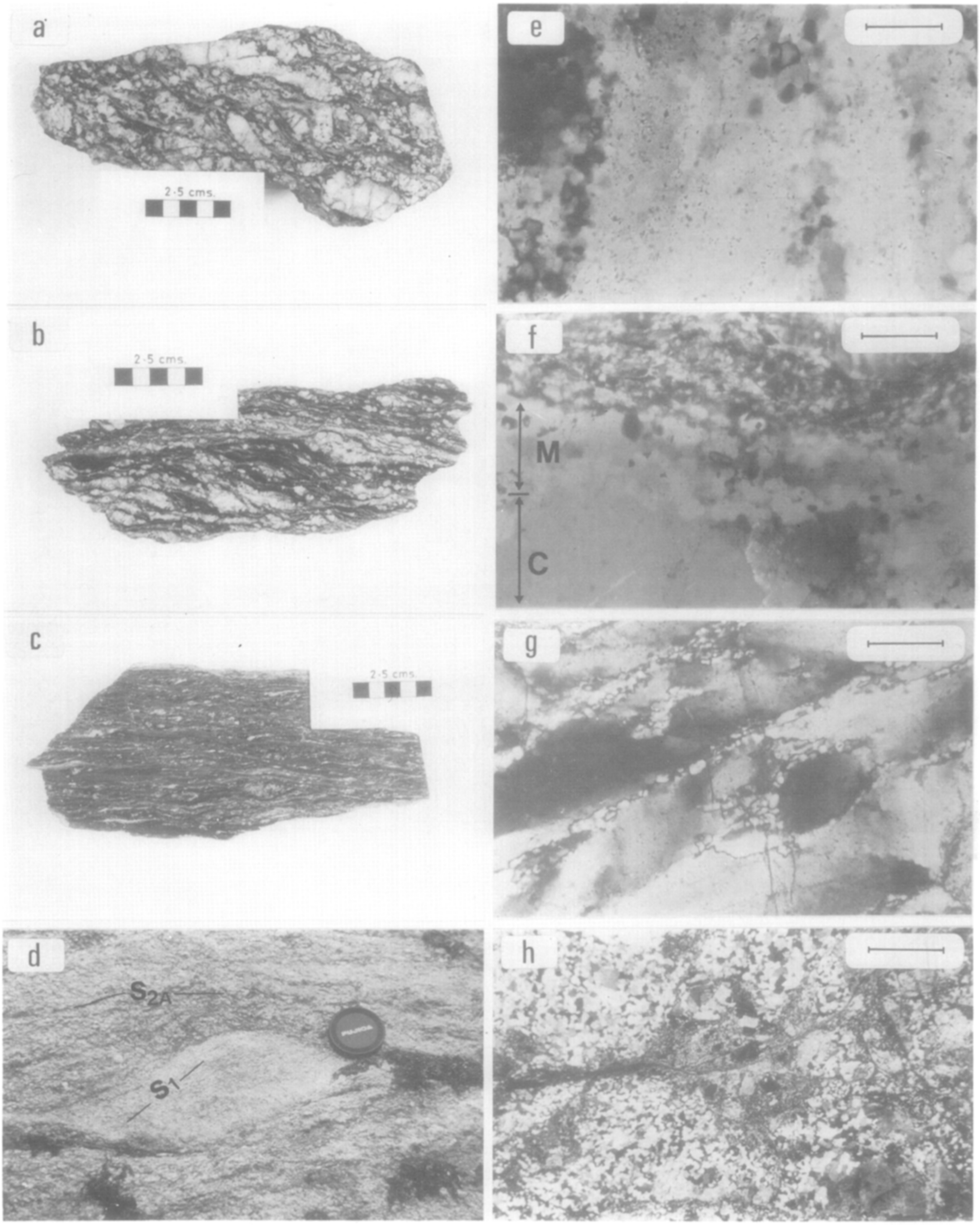


Fig. 5. (a) Protomylonite, (b) mylonite showing mafic and felsic bands, (c) ultramylonite, (d) pod of granitoid rock preserved in mylonite, (e) inhomogeneous subgrain development and dynamic recrystallization (sharp high angle boundaries) of quartz; note similar subgrain and new grain size, (f) core and mantle structure (C,M) in quartz, (g) nucleation recrystallization of quartz in high strain regions, (h) blastomylonite showing totally recrystallized quartz. Scale bars (e) and (f) 100μm; (g) 400μm; (h) 1mm.

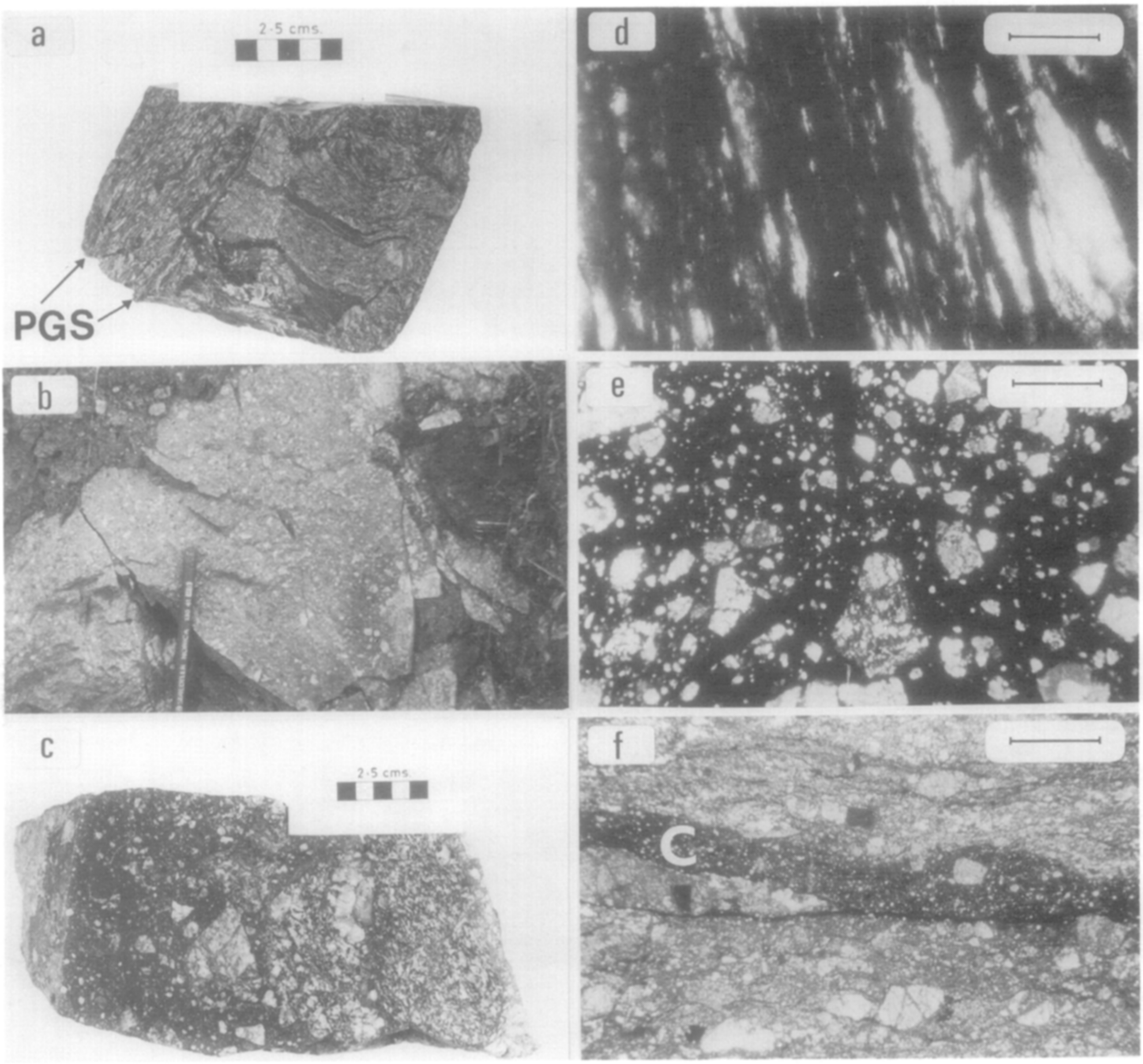


Fig. 10. (a) Pseudotachylite paired generation surfaces (PGS) with injection veins. (b) Cataclasite field occurrence. (c) Cataclasite enclosing a rounded fragment with a disrupted folitation. (d) Photomicrograph of an injection vein in (a). (e) Photomicrograph of cataclasite, with composite clasts. (f) Cataclasite band (C) in a mylonite. Scale bars (d) 100μm, (e), (f) 1mm.

quartz grains showing undulose extinction, which is characteristic when the rate of strain hardening dominates over the rate of strain softening (White 1977). Often quartz is seen to recrystallize in high strain regions such as deformation bands (Fig. 5g) by a process of nucleation recrystallization.

White (1973) has shown that quartz may be deformed and recrystallized in a cyclic manner. This is clearly seen in some Guingamp mylonites where different quartz grains can be seen at various stages in such a process; relict ribbon grains contain dynamically recrystallized grains, some showing grain growth. Representative quartz *c*-axis fabric diagrams (Fig. 8) show *c*-axis concentrations around *Z* (Fig. 8a) and both *Z* and *Y* (Figs. 8b & c). Such distributions are considered to be the result of the operation of intracrystalline slip on both the basal and prism planes (Wilson 1975, White 1976, Boullier & Bouchez 1978). The asymmetry in the diagrams produced by oblique fabric girdles is consistent with a dextral shear displacement about the *Y* axis (Burg & Laurent 1978).

Nicolas & Poirier (1976) and White (1976, 1977) have suggested that superplastic deformation can occur in rocks with sufficiently small grain size to allow diffu-

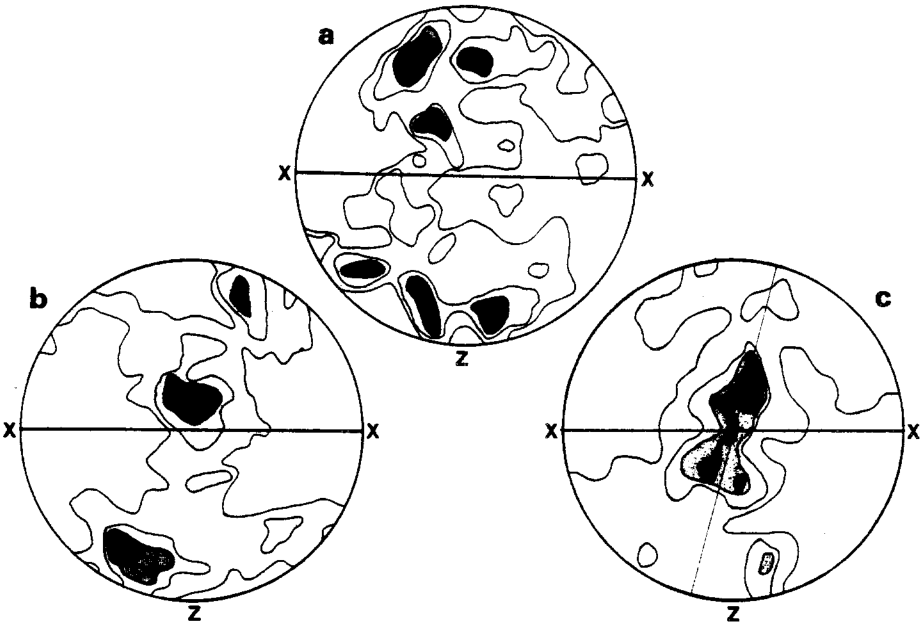


Fig. 8. Representative quartz *c*-axis fabric diagrams; measurements taken on strained grains with strong shape fabrics. For a discussion of the *c*-axis distributions see text. (a) Protomylonite/mylonite with ribbon quartz and limited dynamic recrystallization. 75 points, contours at 1.5, 3, 4.5, 6 and 7.5% levels. (b) Banded mylonite with strong quartz shape fabric development (aspect ratios up to 60:1), dynamic recrystallization by sub grain rotation common. 100 points, contours at 1, 3, 5, 7 and 9% levels. (c) Mylonite with quartz ribbons and dynamic recrystallization common. 100 points, contours at 1, 3, 5 and 7% levels.

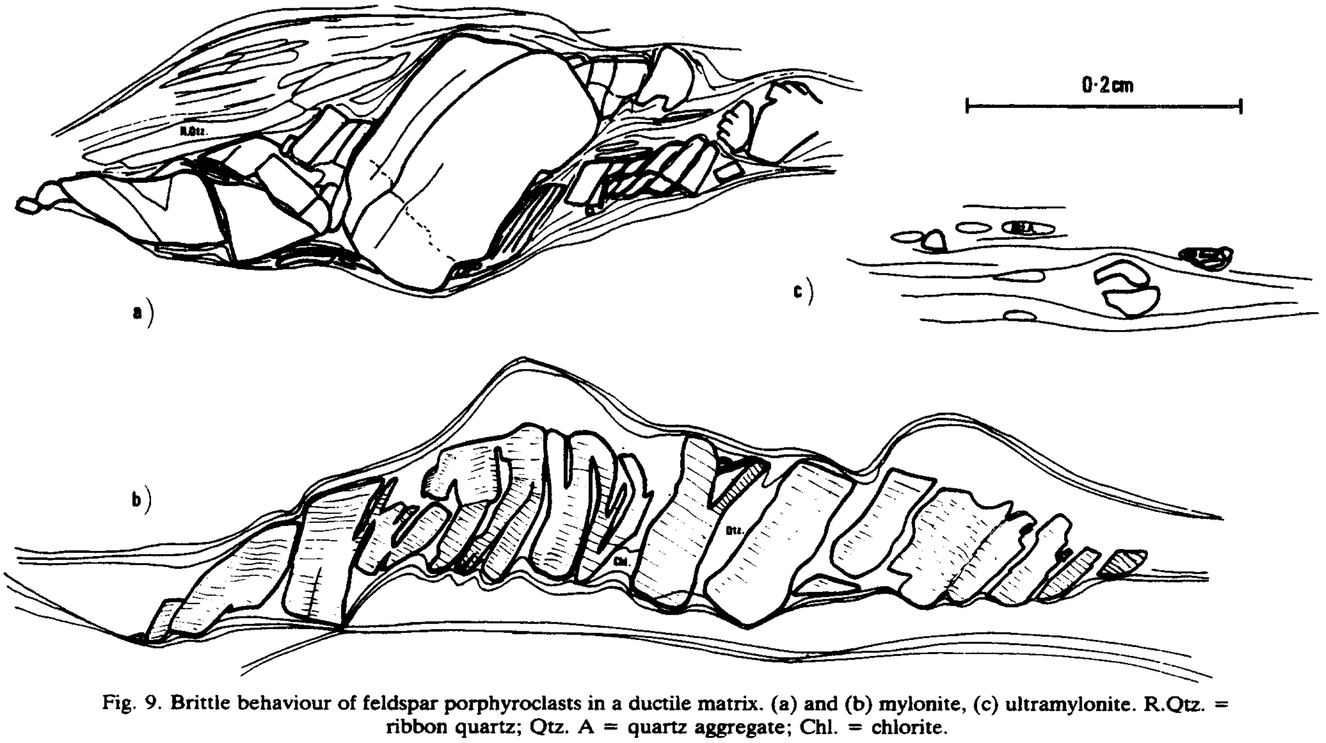


Fig. 9. Brittle behaviour of feldspar porphyroclasts in a ductile matrix. (a) and (b) mylonite, (c) ultramylonite. R.Qtz. = ribbon quartz; Qtz. A = quartz aggregate; Chl. = chlorite.

sion and grain boundary movement. In some mylonites, definite narrow bands with a grain size of < 100 μm are seen and this is thought to be the threshold requirement for superplasticity.

Although the dominant mode of deformation of feldspars is a brittle fracture, evidence of plastically deformed feldspar such as kinks and sub-grains (Bell & Etheridge 1973, Nicolas & Poirier 1976) may be found. The feldspar fracture is usually cleavage controlled (cf. Wakefield 1977), and antithetic internal movement of feldspars is shown in Fig. 9(a & b). Figure 9(a) is from a mylonite where strain hardening processes have been dominant but not exclusive; quartz ribbons abound with nucleation recrystallization in the high strain regions, with minor recrystallization by sub-grain rotation. In Fig. 9(b) dynamic recrystallization and dynamic recovery have more than balanced the strain hardening, and it appears that the quartz has been mobile as the tensional feldspar fractures are infilled with quartz showing plastic strain features. Figure 9(c) shows a relict feldspar porphyroclast broken and rounded in a foliated groundmass of mica, quartz and feldspar in an ultramylonite.

An order of deformation can be deduced from the minerals as they responded to changes in the deformational environment;

1. plastic deformation of quartz and feldspar;
2. cleavage controlled fracture of feldspar with plastic flow of quartz/mica into tensional feldspar fractures;
3. continued cyclic plastic deformation of quartz;
4. temperature reduction and 'freezing-in' of the microstructures.

Blastomylonites

Strictly, blastomylonites do not belong to the mylonite series defined by Sibson (1977) because they show pronounced grain growth, whereas mylonites show grain size reduction. They are, however, products of a ductile deformation mode. Figure 5(h) shows a blastomylonite in which the quartz is totally recrystallized. In several parts of the North Armorican Shear Zone blastomylonites are the typical fault rock products.

Pseudotachylite

Pseudotachylite found in fault zones is considered to be the product of frictional melting resulting from transient rapid sliding along a fault plane (Philpotts 1964, Sibson 1975). Recently, Wenk (1978) has shown that some pseudotachylites reveal textures which support an origin by extreme cataclasis as they represent a fine grained aggregate of strongly strained quartz and feldspar particles. Under the present nomenclature (Sibson 1975, 1977), we feel justified in using the term pseudotachylite where it applies to a fault generated vein which, under the optical microscope, possesses features resembling glass rather than an ultracataclasite.

Figure 10(a) shows a specimen of mylonitic rock containing pseudotachylite generated along two parallel surfaces some 30mm apart; a geometric characteristic that has been reported by Grocott (1977a) from W. Greenland. Pseudotachylite is injected into the mylonite at approximately 90° to the generating surfaces, a necessary requirement if a high frictional resistance is to be maintained on the fault (Sibson 1975).

In thin section, the pseudotachylite sometimes consists of aligned porphyroclasts of quartz and feldspar in an opaque matrix (Fig. 5d). The porphyroclasts are approximately ellipsoidal in shape giving an overall fabric to the veins. At first, this appears to be an anomalous feature in view of Sibson's (1975) statement that in the Outer Hebrides pseudotachylites "only on rare occasions are clasts oriented in other than a random fashion" which is to be expected by generation in the brittle deformation regime. The clasts are not flow aligned in the melt as they show evidence that they have undergone ductile deformation. This indicates that the pseudotachylite and included porphyroclasts have been deformed in a ductile manner after its brittle generation and injection.

Cataclasite series

By definition, cataclasites are random fabric fault rocks which are thought to be generated by brittle fragmentation in fault zones (Sibson 1977), (Figs. 10b & c). In Guingamp, they occur in zones up to 20 m wide but paucity of outcrop prevents the estimation of their lateral extent. Because of their lack of orientated clasts, it is impossible to tell the movement direction of the faults in which they are found.

In thin section, cataclasites contain clasts of feldspar (An_{10}) and quartz showing undulose extinction set in a fine groundmass containing laths of chlorite and titanomagnetite. All of the porphyroclasts are angular which suggests brittle comminution and fragmentation; in pseudotachylites rounding of porphyroclasts may result from processes such as decrepitation (Sibson 1975), thermal fragmentation (Sibson 1977) and thermal spalling (Grocott 1977a). Rounded porphyroclasts of disrupted mylonite and sub-angular clasts of pre-existing cataclasite are sometimes seen (Figs. 10c & e). Also thin cataclasite veins running parallel to the mylonite foliation are common (Fig. 10f), and these features testify to the polyphase nature of the shear movements.

Late fractures — incohesive fault gouge

Many late fractures showing only minor displacements cut the mylonites and cataclasites and represent the final movements of the zone. Fracture planes are usually iron stained, and the small fault zones sometimes contain blue–grey, incohesive fault gouge.

DISCUSSION

In the Guingamp region a number of different fault rocks are associated with the North Armorican Shear

Zone. These include rocks of the mylonite series, the cataclasite series, pseudotachylites and incohesive fault gouge, and it is thought that each is generated in a specific environment at a well-defined level (Sibson 1977, Grocott 1977b).

For the generation of mylonite, a dominant mineral constituent, e.g. quartz in a granitoid rock, must deform in a plastic manner. This is achieved when deformation takes place at > 250–350°C representing a depth in the crust of 10–15 km at normal geothermal gradients (Sibson 1977). The tendency for prismatic slip in quartz favours higher temperatures than this, or possibly low strain rate. Recrystallization of feldspars is not seen except in deformation bands/kink bands in crystals within blastomylonites. At temperatures of 250–350°C a relatively high strain rate *ca.* $10^{-12}s^{-1}$ (White 1975) is necessary for mylonite generation. Blastomylonite is likely to be formed under conditions of slightly lower strain rate or higher temperature.

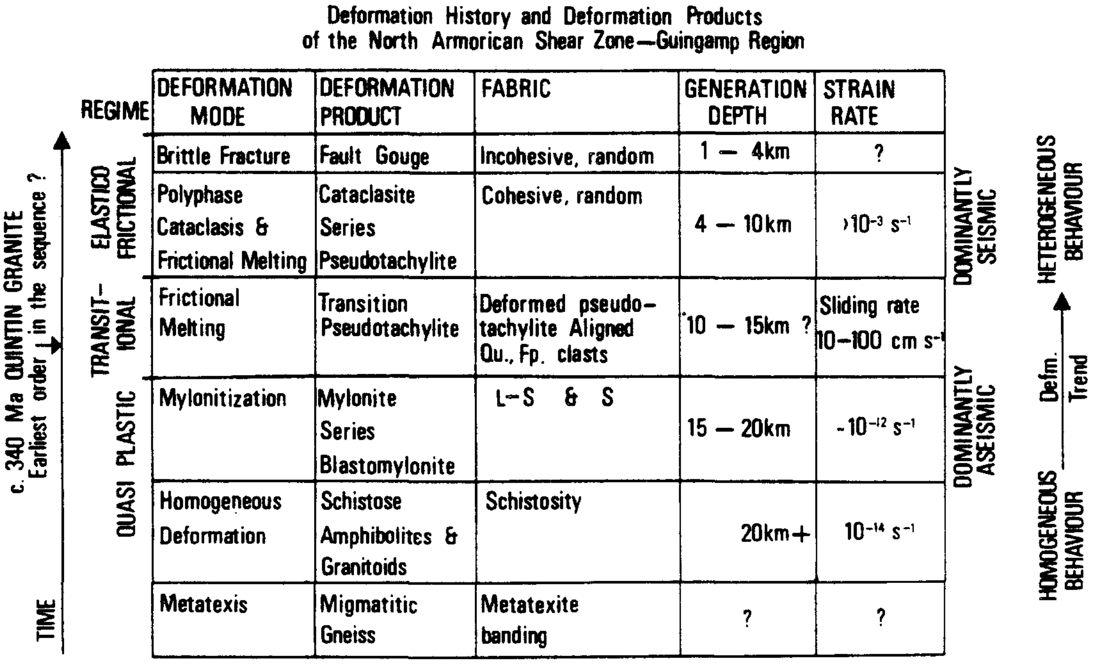
Pseudotachylite forms by the generation of molten material on a dry fault surface by frictional processes (Sibson 1975, 1977), and one may reasonably expect frictional melting to occur at typical seismic slip rates (10–100 cm s^{-1}). Pseudotachylites of the Guingamp region have been deformed in a ductile manner after their generation in a frictional regime giving rise to a rock fabric. This fabric could result from the generation of a pseudotachylite under brittle conditions within or near to the brittle/ductile transition zone in the crust, and the subsequent ductile deformation of the rock. This restriction to a specific zone for the generation of pseudotachylite may not be necessary as Sibson (in press) regards deformed pseudotachylite associated with the Outer Hebrides Thrust as a representative of transient discontinuities within ductile shear zones. It is believed that the pseudotachylite melt was generated by seismic slip at deep crustal level.

The Guingamp pseudotachylites no longer preserve their original textures, and therefore represent transitional fabrics with both ductile (ultramylonite) and brittle (true pseudotachylite) affinities. We have adopted the provisional term 'transition pseudotachylite' for its description.

Cataclasite series rocks are formed by cataclastic flow when temperatures are sufficiently low to prevent plastic deformation of minerals in the parent rocks, and deformation proceeds by brittle fragmentation and comminution of particles. In quartzo-feldspathic rocks this takes place at temperatures below 250°C. Incohesive fault gouge is thought to form at depths of ~1–4 km (Sibson 1977), although Wu *et al.* (1975) have predicted that clay gouge can exist down to depths of 10 km.

The fault rocks of the Guingamp region are thought to represent polyphase movements along a single strain softened zone. For all these products to be present at one level, fault movements must have taken place over a long period of time during which there was continued uplift and erosion of the area. Therefore, the fault rocks represent a sequence in time (from ductile to brittle rock products) which reflect a passage of the area from deep level through the brittle/ductile transition to the present day level. This model is analogous to that of Grocott (1977b) for the Ikertoq Shear Belt of W. Greenland which preserves both brittle and ductile rock products at present day level.

Although much of the early evidence for the dextral displacement of the shear zone came from the Quintin granite (Chauris 1969, Cogné 1976), the granite is generally undeformed. Only sporadic outcrops of foliated granite or protomylonite are present within the



Deformation History and Deformation Products
of the North Armorican Shear Zone—Guingamp Region

REGIME	DEFORMATION MODE	DEFORMATION PRODUCT	FABRIC	GENERATION DEPTH	STRAIN RATE
ELASTICO FRICTIONAL	Brittle Fracture	Fault Gouge	Incohesive, random	1 — 4km	?
	Polyphase Cataclasis & Frictional Melting	Cataclasite Series Pseudotachylite	Cohesive, random	4 — 10km	$>10^{-3}$ s^{-1}
TRANSITIONAL	Frictional Melting	Transition Pseudotachylite	Deformed pseudotachylite Aligned Qu., Fp. clasts	10 — 15km ?	Sliding rate 10–100 cm s^{-1}
QUASI PLASTIC	Mylonitization	Mylonite Series Blastomylonite	L–S & S	15 — 20km	~10^{-12} s^{-1}
	Homogeneous Deformation	Schistose Amphibolites & Granitoids	Schistosity	20km+	10^{-14} s^{-1}
	Metatexis	Migmatitic Gneiss	Metatexite banding	?	?

Fig. 11. Synoptic diagram of fault rock time sequence and conditions for their formation. The strain rates given are those typically used in discussions dealing with the formation of fault rocks (White 1975, 1976, Sibson 1975) and generation depths are taken from Sibson (1977).

granite along the trace of the shear zone, and no frictional products are found. Apparently, deformation of the granite represents only a minor part of the total shear displacement; a far greater part being recorded in the extensive development of mylonite series rocks outside the main granite outcrop. These mylonites predate the intrusion of the Quintin granite. Zones of foliated granite may be the result of ductile deformation of the hot granite body, and may be equivalent in age to the frictional rock products found elsewhere in this zone. The time sequence of fault rock development (Fig. 11) shows that the emplacement of the Quintin granite post-dates the main shear movement but pre-dates the final dextral displacement.

We conclude, therefore, that the main movement of the North Armorican Shear Zone represented by ductile rock products occurred between 600 Ma (Cadomian granites) and 340 Ma (Quintin granite), and only ~10 km of dextral displacement has occurred since this time. This conclusion supports the original theory of Chauris (1969) who suggested that the North Armorican Shear Zone was probably a Pre-cambrian rather than a Hercynian feature.

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REFERENCES

Adams, C. J. D. 1967. A geochronological and related isotopic study of rocks from NW France and the Channel Islands. Unpublished PhD. thesis, University of Oxford.

Bell, T. H. & Etheridge, M. A. 1973. Microstructure of mylonites and their descriptive terminology. *Lithos* **6**, 337–348.

Boullier, A. M. & Bouchez, J. L. 1978. Le quartz en rubans dans les mylonites. *Bull. Soc. géol. France.* **7**, 253–262.

Brown, M. 1978. The tectonic evolution of the Precambrian rocks of the St. Malo region, Armorican Massif, France. *Precambrian Res.* **6**, 1–21.

Brown, M., Barber, A. J. & Roach, R. A. 1971. The age of the Saint-Malo migmatite belt, Northern Brittany. *Nature Phys. Sci.* **234**, 77–79.

Brun, J. P. & Martin, H. 1979. The St. Malo migmatite belt: a late Precambrian gneiss dome — a comment. *Precambrian Res.* **8**, 137–143.

Burg, J. P. & Laurent, P. 1978. Strain analysis of a shear zone in a granodiorite. *Tectonophysics* **47**, 15–42.

Calvez, J. Y. & Vidal, P. 1978. Two billion years old relicts in the Hercynian Belt of Western Europe. *Contr. Mineral. Petrol.* **65**, 395–399.

Carter, N. L., Christie, J. M. & Griggs, D. T. 1961. Experimentally produced deformation lamellae and other structures in quartz sand (Abstract). *J. geophys. Res.* **66**, 2518–2519.

Carter, N. L., Christie, J. M. & Griggs, D. T. 1964. Experimental deformation and recrystallization of quartz. *J. Geol.* **72**, 687–733.

Chauris, L. 1969. Sur un important Accident Structural dans le Nord-Ouest de l'Armorique. *C. R. hebd. Seanc. Acad. Sci. Paris*, **268**, Ser. D, 2859–2861.

Cogné, J. 1959. Données nouvelles sur l'Antecambrien dans l'Ouest de la France: Pentévrien et Briovérien en baie de Saint-Brieuc (côtes-du-Nord). *Bull. Soc. géol. France.* **7**, 112–118.

Cogné, J. 1976. *Carte Géologique Détaillée de la France, St. Brieuc*, 3rd edn. 1 : 80,000.

Cogné, J., Durand, S. & Esteoule-Choux, J. 1974. Le Massif Armoricain. In: *Géologie de la France* (edited by Debelmas J.), Doin, Paris, Vol. 1, 105–161.

Green, H. W., Griggs, D. T. & Christie, J. M. 1970. Syntectonic and annealing recrystallization & fine-grained quartz aggregates. In: *Experimental and Natural Rock Deformation* (edited by P. Paulitsch) Springer, Berlin–Heidelberg.

Grocott, J. 1977a. The northern boundary of the Ikertoq shear belt, West Greenland. Unpublished Ph.D. thesis, University of Liverpool.

Grocott, J. 1977b. The relationship between Precambrian shear belts and modern fault systems. *J. geol. Soc. Lond.* **133**, 257–262.

Leake, B. E. 1978. Nomenclature of Amphiboles. *Can. Mineral.* **16**, 501–520.

Leutwein, F., Sonet, J. & Zimmermann, J. L. 1968. Géochronologie et évolution orogénique précambrienne et hercynienne de la partie nord-est du Massif Armoricain. *Sci. Terr. Mem.* **11**, 1–84.

Nicolas, A. & Poirer, J. P. 1976. *Crystalline Plasticity and Solid State Flow in Metamorphic Rocks.* Wiley, New York.

Philpotts, A. R. 1964. Origin of pseudotachylites. *Am. J. Sci.* **263**, 1008–1035.

Raase, P. 1974. Al and Ti Contents of Hornblende, Indicators of Pressure and Temperature of Regional Metamorphism. *Cont. Mineral. Petrol.* **45**, 231–236.

Ramsay, J. G. & Graham, R. H. 1970. Strain Variation in Shear Belts. *Can. J. Sci.* **7**, 786–813.

Roach, R. A. 1974. A Review of the Precambrian rocks of the British Variscides and their relationships with the Precambrian of N.W. France. La chaine varisque d'Europe moyenne et occidentale. *Coll. intern. CNRS, Rennes* **243**, 61–79.

Roach, R., Adams, C., Brown, M., Power, G. & Ryan, P. D. 1972. The Precambrian stratigraphy of the Armorican Massif, N.W. France. *24th Int. geol. Congr. Section 1*, 246–252.

Sibson, R. H. 1975. Generation of pseudotachylite by ancient seismic faulting. *Geophys. J. Roy. astr. Soc.* **43**, 775–794.

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

Sibson, R. H. in press. Transient Discontinuities in ductile shear zones. In: Int. Conf. on Shear Zones in Rocks, Barcelona, Spain, May, 1979. *J. Struct. Geol.*

Vernon, R. H. 1974. Controls of mylonitic compositional layering during non-cataclastic ductile deformation. *Geol. Mag.* **111**, 121–123.

Wakefield, J. 1977. Mylonitization in the Lethakane shear zone, eastern Botswana. *J. geol. Soc. Lond.* **133**, 263–275.

Watterson, J. 1975. Mechanism for the persistence of tectonic lineaments. *Nature Phys. Sci.* **253**, 520–521.

Wenk, H. R. 1978. Are pseudotachylites products of fracture or fusion? *Geology* **6**, 507–511.

White, S. 1973. Syntectonic recrystallization and texture development in quartz. *Nature Phys. Sci.* **244**, 276–278.

White, S. 1975. Estimation of strain rates from microstructures. *J. geol. Soc. Lond.* **131**, 577–583.

White, S. 1976. The effects of strain on the microstructures, fabrics and deformation mechanism in quartzites. *Phil. Trans. Roy Soc.* **A 283**, 69–86.

White, S. 1977. Geological significance of recovery and recrystallization processes in quartz. *Tectonophysics* **39**, 153–170.

White, S., Shaw, N. D., Burrows, S. & Norton, M. 1978. Shear bands in large strain deformation. Lecture given at T.S.G. meeting at Liverpool.

Wilson, C. J. L. 1975. Preferred orientation in quartz ribbon mylonites. *Bull. Geol. Soc. Am.* **86**, 968–974.

Wu, F. T., Blatter, L. & Robertson, H. 1975. Clay gouges in the San Andreas Fault System and their possible implication. *Pageophysics* **113**, 87–95.