Timing of deformation and the mechanism of cleavage development in a Newfoundland mélange

Paul F. Williams
Department of Geology, University of New Brunswick,
Fredericton, New Brunswick, Canada E3B 5A3

This paper is concerned with two aspects of folding and cleavage development in the "Carmanville Ophiolitic Mélange" of Newfoundland. First it is concerned with the problem of the timing of deformation relative to lithification and metamorphism, and second it is concerned with development of the cleavage.

According to previous interpretation the cleavage and associated folds are a product of soft-sediment deformation. The cleavage however, is a second generation axial-plane structure and microstructural observations, presented here, show it to be later than the peak of metamorphism. Further, F_2 folds occur in quartz veins where they are associated with a strong crystallographic fabric. The second generation structures are thus a product of hard-rock deformation. The origin of the F_1 folds however, remains equivocal.

Dyke-like structures previously attributed to migration of thixotropic mélange units are also post lithification and are a product of faulting and metamorphic differentiation.

The S_2 cleavage is largely a product of rotation. In the coarser sediments whole clasts are rotated into the cleavage plane. Within shale clasts and in shale beds the cleavage has developed principally as a result of transposition by microfolding. Folding fabrics have been modified by grain growth.

La présente étude traite de deux questions reliées au développement du plissement et du clivage dans le "mélange ophiolitique de Carmanville" (Terre-Neuve): 1) de la chronologie respective de la lithification et du métamorphisme par rapport à la déformation et 2) du développement du clivage.

Auparavant, on attribuait le clivage et les plis à la déformation de sédiments non-consolidés. Il appert cependant que le clivage correspond à un plan axial de deuxième génération et les éléments structuraux microscopiques indiquent que ce clivage est postérieur à la phase maximal du métamorphisme. De plus, les plis F_2 se retrouvent dans des veines de quartz où ils sont associés à une trame cristallographique très bien définie. Les structures de deuxième génération proviennent donc d'une déformation lithique. L'origine des plis F_1 demeure toutefois incertaine.

Des structures en forme de dykes que l'on attribuait jadis à des phénomènes de thixotropie au sein du mélange sont en fait le résultat de failles et de différentiation métamorphique survenues après la lithification des sédiments.

Le clivage S_2 est principalement dû à un mouvement de rotation et dans les sédiments grossiers, des fragments entiers ont été réorientés selon le plan de clivage. Le clivage qui affecte les lits et les fragments de schiste argileux est surtout le résultat d'une transposition par microplissement. Les trames de plissement ont été modifiées par la croissance des grains.

[Traduit par le journal]

INTRODUCTION

Minor folds and cleavage in mélange and associated sediments, in Notre Dame Bay and adjacent areas of Newfoundland (Fig. 1), have been attributed, in part, to soft-sediment slumping and deformation by a number of writers (eg. Helwig 1970; Horne 1970; Pajari *et al.* 1979; Arnott 1983). Others (eg. Nelson 1981; Karlstrom *et al.* 1982) have attributed the same structures to hard-rock deformation. The problem is a general one in areas of mélange (e.g. Hsü 1974; Naylor 1982); it is usually difficult to distinguish between "hard-rock" and "soft-sediment" structures in such areas. The problem is probably due to the presence, in such areas, of a complete spectrum of deformational conditions between those of soft-sediment, sub-aqueous slumping, through those of tectonic deformation of partially lithified sediments, to those of hard-rock deformation accompanied or preceded by metamorphism. However, if we are to understand the history and development of such areas it is important to determine the relative timing of lithification, metamorphism and deformation, and as part of the same problem, to be able to distinguish olistostromes and tectonic mélanges (see Hsü 1974).

Excellent examples of folds and cleavage that have been attributed to soft-sediment deformation (Pajari *et al.* 1979) occur on Green Island (locally known as Woody Island) in Hamilton Sound. A number of writers (for reviews see Williams

MARITIME SEDIMENTS AND ATLANTIC GEOLOGY
19, 31-48 (1983)

0711-1150/83/010031-18$3.70/0

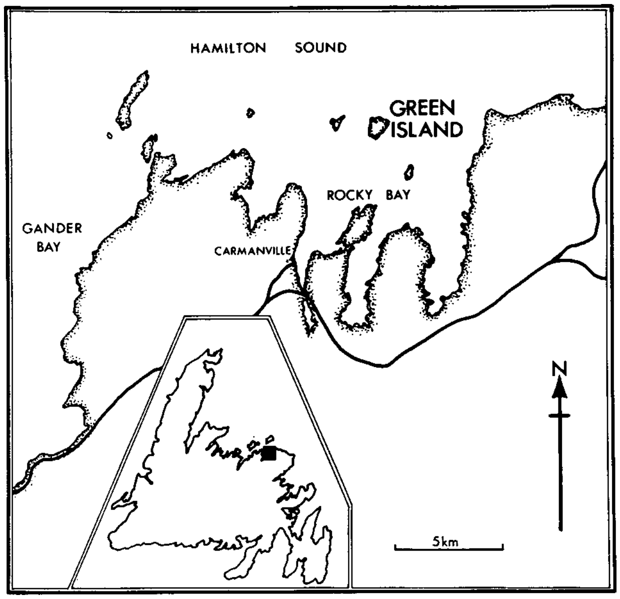


Fig. 1 - Locality map.

et al. 1969; Helqig 1970; Hobbs *et al.* 1976) have suggested criteria for separating soft-sediment and hard-rock structures and these criteria are applied here to the rocks of Green Island.

The purpose of this paper is twofold. One is to describe the structural history of the deformed mélange sediments of Green Island and to relate this history to that of lithification and metamorphism. The foliation and associated folds, described as soft-sediment structures by Pajari *et al.* (1979) are shown to be post-lithification and post-metamorphic but there remains a group of earlier structures of equivocal origin. The other is to discuss the processes involved in the development of the foliation. The principal mechanism is one of rotation on various scales. Locally the rotation involves individual clasts which behave independently. Elsewhere it is more structured and is associated with folding of an earlier foliation on one scale or another.

The rocks described lie in the Dunnage Zone (Williams 1979) and belong to the Davidsville Group. They are placed by Currie *et al.* (1980) in their units 7 and 9 which are of Caradocian age. They include bedded greywackes, sandstones, conglomerates, siltstones and shales and mélange. The coarser bedded sedimentary rocks are commonly graded and there are some complete Bouma sequences with micro-crossbedding and convolute lamination. The mélange comprises blocks of sedimentary rocks similar to the adjacent bedded sandstones, siltstones and shales in a dark matrix. In general the rocks do not look strongly metamorphosed, through locally, garnet and possible staurolite pseudomorphs are visible in hand specimen.

The various structures and their relationships are described first and then the nature of the rocks at the time of deformation and the development of the cleavage are discussed.

MESOSCOPIC STRUCTURE

Two generations of folds are recognized on the basis of overprinting relationships, but in general there is only one foliation and it is parallel to the axial-plane of the F_2 folds.

Fig. 2 - S_2 foliation defined by elongate sedimentary clasts. Knife is 10 cm long.

This is the foliation described by Pajari *et al.* (1979) and it is a particularly striking foliation in layers rich in shale fragments where the lenticular fragments are aligned parallel to the foliation (Fig. 2).

The area examined coincides with the southeast limb and fold closure of the large structure shown on the map of Currie *et al.* (1980), and it is convenient to describe the mesoscopic structures in the framework of this larger fold. In the limb of the large structure the mesoscopic folds are mostly open and asymmetrical with the prominent cleavage parallel to their axial-planes. The cross-cutting relationship of the cleavage and the large fold (Fig. 3) indicates that the latter is an F_1 structure and that the mesoscopic folds and related cleavage are F_2 structures. This is confirmed on the mesoscopic scale by the presence of a few tight to isoclinal folds that are overprinted by F_2 and the S_2 foliation (Fig. 4).

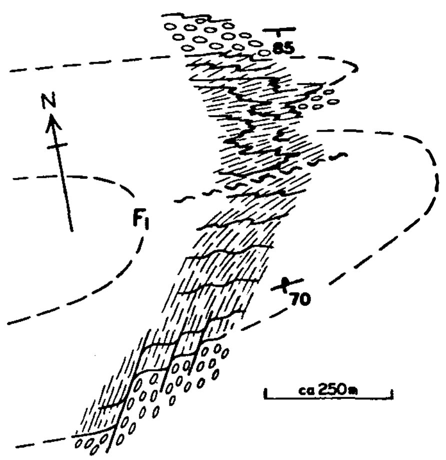


Fig. 3 - Diagrammatic sketch map of the east shore of the island showing the structural relationships. The elliptical symbols represent the mélange and the orientation of the ellipses indicates the local cleavage orientation as defined by elongate clasts. Heavy lines represent bedding and light lines the trace of cleavage in the bedded sediments.

Bedding is greatly disrupted and it has been suggested that this is due to sedimentary processes. While this seems a reasonable explanation for some of the disruption, there is evidence that at least part of the deformation is post lithification. Figure 5(a) shows one example. A competent sandstone layer is boudinaged and the neck between the boudins is occupied by vein-quartz indicating that the material was already lithified when the structure formed. The age of the boudinage is equivocal. However, the relationship between the boudins and the F_2 fold in the same layer indicates that boudinage in this example at least, is pre F_2.

In the rocks described as mélange (Pajari *et al.* 1979) many of the blocks are folded into dismembered isoclinal folds (Fig. 4). It is not possible to say whether these folds developed in isolated blocks or whether they developed in continuous or semi-continuous layers which later became disrupted by boudinage. Their shape suggests the latter and if boudinage occurred early it certainly continued during folding as can be seen from the development of post fold boudins such as illustrated in Figure 4(a). Again the presence of quartz veins indicates that at least some of the boudinage is post lithification so that even if the folds were "slump structures" they have been further modified after lithification. Some of these dismembered folds are oriented such that S_2 is parallel to their axial-surface (Fig. 4a) others are overprinted by S_2 (Fig. 4b). This could be interpreted as indicating two generations of isoclinal folds. However, both have the same style, which differs from the style of neighbouring F_2 folds (see Fig. 4e). Another possibility preferred by the writer is that all of the dismembered isoclines are F_1 structures but some of these, because they were "floating" in an incompetent matrix were able to rotate into the S_2 orientation during the development of F_2 structures.

One of the dismembered F_1 folds that is overprinted by S_2 (Fig. 4b) shows signs of an earlier foliation which may be an S_1. Unfortunately, the smooth flat nature of the outcrop made it impossible to sample this foliation.

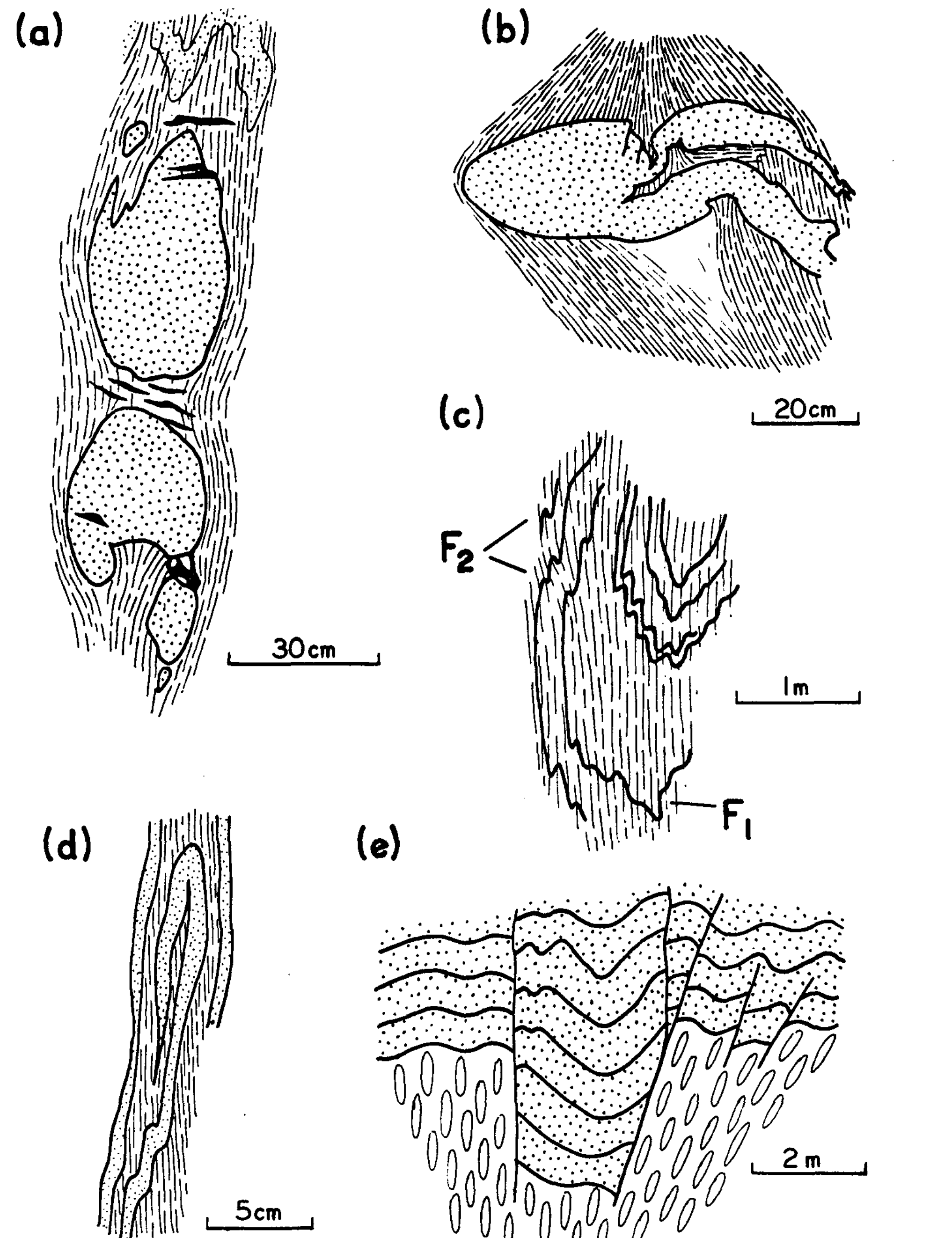


Fig. 4 - Folds and overprinting relationships. a) Dismembered fold in the mélange with S_2 parallel to its axial-plane. The flattened fold hinge is boudinaged and quartz veins (solid black areas) occupy the area between the boudins. This fold may be an F_2 structure or an F_1 fold rotated into the F_2 orientation. b) F_1 fold in the mélange overprinted by S_2 and gently folded by F_2. Note a locally developed foliation parallel to the F_1 axial-plane orientation. c) F_1 fold in shales overprinted by F_2 and S_2 which locally is almost parallel to the F_1 axial plane. d) F_1 fold in alternating sandstone and shale overprinted by F_2 and S_2. e) Fault related F_2 fold in sandstone and conglomerate with a well-developed foliation defined by preferred orientation of the clasts in the conglomerate.

The F_2 folds are fairly open (Fig. 2 and 4e) structures most of which are markedly asymmetrical. Locally they are parallel folds and there is evidence of flexural-slip having played a role in their development; quartz veins approximately perpendicular to bedding are displaced by bedding parallel faults (Fig. 6). The displacement is symmetrical on opposite limbs of symmetrical folds and the sense of displacement is consistent with a flexural-slip model. This again indicates a post-vein and therefore post-lithification origin for the F_2 folds.

The asymmetry of many F_2 structures reflects strong movement along zones parallel to alternate limbs of the folds. These zones were interpreted by Pajari *et al.* (1979) as liquefaction structures due to intrusion of liquified sediment into overlying beds. However, a hard-rock origin is indicated by the affect that the zones have on quartz veins, Veins entering the zones change direction abruptly in a manner consistent with the sense of movement across the zone. They become thinned and, in many places, boudinaged, indicating that they formed pre or syn-deformation. S_2 is particularly well developed within these zones parallel to the zone margins; the structure resembles a large scale crenulation cleavage with the movement zones representing the cleavage septae. The abundance of pelitic material noted by Pajari *et al.* (1979) in the movement zones can be explained by metamorphic differentiation, thus strengthening the analogy with crenulation cleavage which is very commonly differentiated (Hobbs *et al.* 1976, p. 218). Such zones, in which cleavage is more strongly developed than elsewhere, usually result in an increase in the layer-silicate content of a rock (e.g. Williams 1972 and 1979).

Veins inclined to S_2 at a large angle show evidence of being earlier than the foliation. For example the foliation bends around vein segments in a way that indicates that the veins were present as strong elements during development of the cleavage (Fig. 5b). Also where the veins are folded, S_2 is parallel to their axial-surface, suggesting that the folds are F_2 structures. Some of these folds are tight to isoclinal

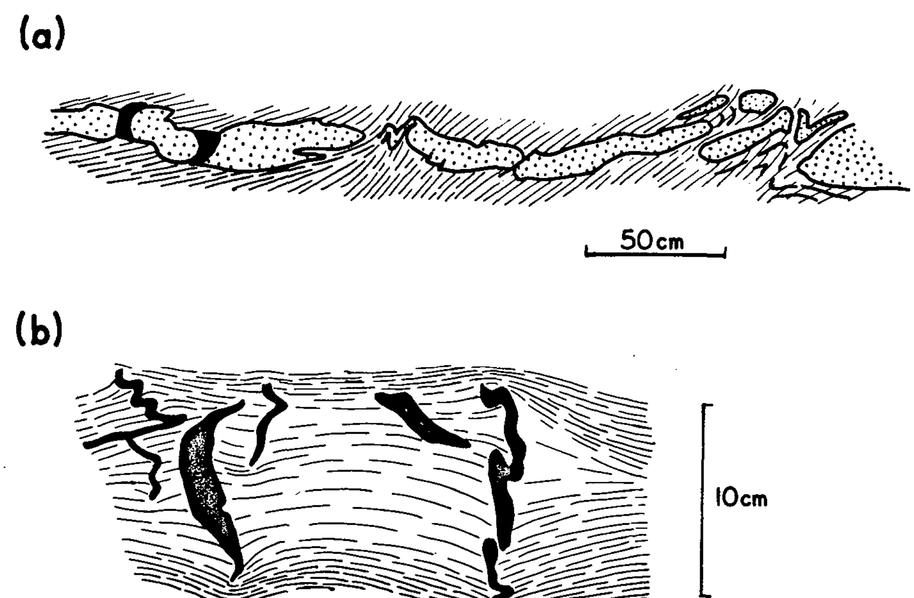


Fig. 5 - a) Disruption of a sandstone and a siltstone bed, in a thick shale horizon, by F_2 folding and boudinage. Vein-quartz (solid black) occupies the area between the boudins. S_2 is well-developed in the shale. b) F_2 folds in quartz veins (solid black). Note how the S_2 foliation is flattened around the quartz veins.

and therefore resemble F_1 more than F_2. However, in view of their consistent relationship to S_2 they are interpreted as F_2 folds.

In the hinge region of the large fold the exposure crosses into a different stratigraphic horizon, characterized by finer grained and more finely laminated rocks. There are abundant isoclinal folds in this area which can be shown by overprinting to be F_1 structures (Fig. 4c and d).

S_2, although approximately parallel to the axial-surfaces of F_1 folds, can be seen to transect the folds and locally small, more open F_2 folds, with S_2 parallel to their axial surface, overprint the F_1 structures (Fig. 4c and d).

MACROSCOPIC STRUCTURE

The large-scale structure of the island was previously recognized as a tight, steeply plunging fold that closes to the east northeast and faces, along its axial plane, to the west southwest (Pajari and Pickerill, pers. comm.). My observations were largely restricted to the hinge and southern limb of the fold but in that part of the structure I am in full agreement with the earlier interpretation.

The macroscopic relationships of the area studied in detail are shown in Figure 3. It can be seen from this diagrammatic sketch map that there are numerous symmetrical, mesoscopic, F_1 folds in the hinge of the large structure. These folds have axial-surfaces parallel to the axial-surface trace of the large fold and they have steep to vertical plunges. The large fold is therefore interpreted as an F_1 fold.

The penetrative S_2 foliation overprints the large F_1 fold. This relationship is best seen on the limb of the large structure where S_2 makes a large angle with the fold-limb (Fig. 3). In the hinge of the large structure S_2 is more nearly parallel to the F_1 axial surface orientation so that the overprinting is less obvious. Nevertheless, it can still be recognized (Fig. 4c and d).

Thus the macroscopic structure is a large, approximately vertical F_1 fold, overprinted by a younger, penetrative S_2 cleavage and related small folds.

MICROSTRUCTURE

Introduction

The rocks of Green Island are predominantly metasediments. They have been described as black slates, greywackes, siltstones and olistostromal units, the latter comprising shale, siltstone and greywacke olistoliths in a black pyritic, pelitic matrix (Pajari *et al.* 1979). Their sedimentary microstructure is still commonly preserved despite strong deformation, and metamorphism that is at least above the bio-

Fig. 6 - Segmentation and displacement of a quartz vein by bedding plane sliding during folding of sandstones and siltstones. The antiform is to the right of the photograph. Knife is 10 cm long.

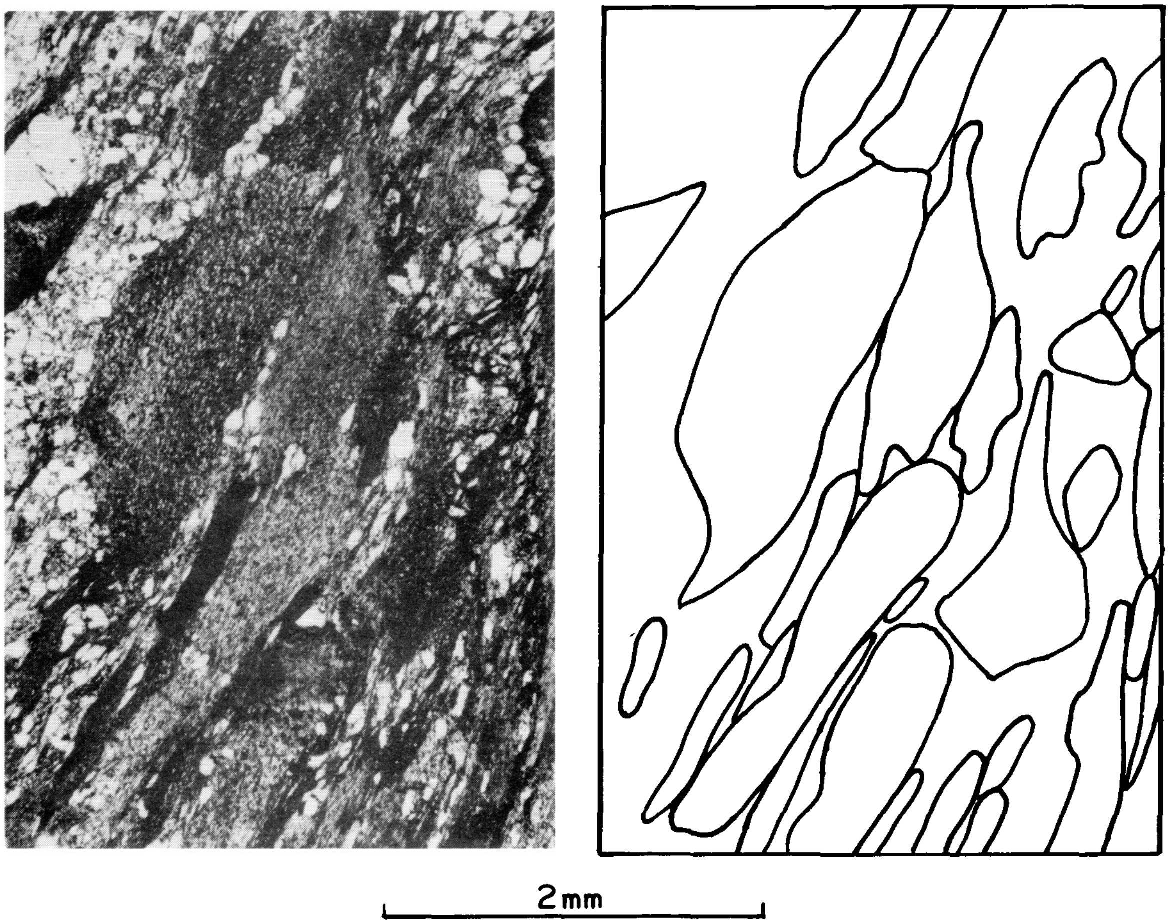


Fig. 7 - 'Muddy matrix' of mélange can be seen in this specimen to comprise shale clasts, coarse sand and grit. Individual shale clasts, where in contact, are difficult or impossible to distinguish in outcrop. Note that the matrix to the shale clasts is mainly fine sand or silt. The coarser portion of this rock comprises clasts of approximately 5 mm length.

tite isograd. In the shale beds the sedimentary microstructure is largely obliterated by the tectonic foliation. In the coarser beds however, cleavage development has not obliterated the earlier microstructure; original clasts, for example, can still be recognized (Fig. 7). Clasts composed of shaly material may be flattened and some quartz clasts are very long and thin and are aligned parallel to the cleavage (S_2) suggesting modification by solution processes. Other quartz clasts however are more equant and look sedimentary. It is significant that deformation of quartz clasts has, in general, only resulted in undulose extinction, indicating that internal strain of the grains is very low. In contrast, quartz in veins is strongly deformed as is quartz in some of the purer quartz siltstone and sandstone beds.

The metamorphic grade of the rocks is difficult to ascertain. The common mineral assemblage is quartz, biotite, white mica, garnet and porphyroblasts that are completely pseudomorphed by white mica and chlorite. The porphyroblasts are large (up to 0.75 mm) and are elongate hexagonal prisms. The shape and alteration suggest staurolite but no original material has been found to confirm this identification, they are therefore referred to in the following simply as the prophyroblasts; the smaller, garnet porphyroblasts are referred to by name.

Sedimentary microstructure of pebbly horizons

Some of the finer grained "olistostrome horizons" reported by Pajari *et al.* (1979) and depicted in their figures 13-15 have been examined in detail. No "pelitic" matrix in the sense of clay matrix was found. The finer grained examples (coarest clasts 5 mm) that I have examined comprise approximately equigranular clasts of quartz, feldspar, siltstone, sandstone and layer - silicate rich aggregates which may originally have been mudstones. One igneous clast was observed. Fine grained quartz and layer-silicates occur between the clasts but matrix is sparse and the

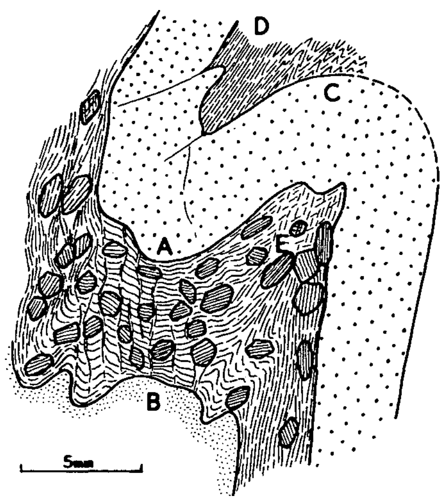


Fig. 8 - F_2 fold in garnet-rich (coarse stipple) and quartz-rich (fine stipple) layers. In the intercalated pelitic layers the hexagonal porphyroblasts are pseudomorphs after possible staurolites (see text) and the layering-parallel S_1 is seen in various stages of transposition into the S_2 orientation. Around A and B S_1 is crenulated but still essentially parallel to layering. Around C S_1 is tightly kinked giving a bimodal fabric symmetrical about the F_2 axial-plane. Around D there is a penetrative foliation with occasional kinks still recognizable but the fabric is no longer bimodal. Around E the mica fabric is again bimodal and some kinks can be recognized. Kinking however, is very irregular so that the impression is bimodal rather than kinked.

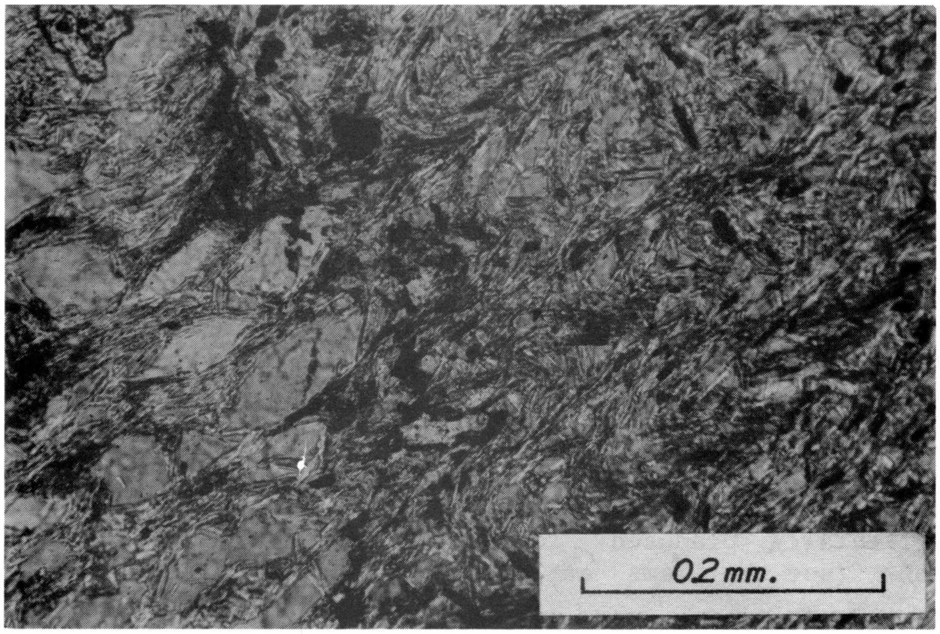
sediment is clast supported. In examples of slightly coarser rocks, in which layer-silicate rich clasts have long dimensions of up to 3 cms, grain size is strongly bimodal. The matrix is predominantly quartz sand and clasts are commonly in contact with one another even as seen on a planar surface (e.g. Fig. 7). There is no evidence of a clay matrix nor evidence that the larger clasts are matrix supported. Locally there are layer-silicate aggregates that could be interpreted as original clay matrix except that when they are examined in detail they are seen to be composed of many small clasts. These clasts can be distinguished by means of subtle variations in original composition, reflected mainly in their opaque mineral content. Thus these sediments are best described as clast supported grits and conglomerates in which shale fragments and quartz silt and sand predominate.

Foliation

The S_2 foliation varies considerably in microstructure in different lithologies. It is clearly seen to overprint an earlier foliation which may be of tectonic or sedimentary origin. The early foliation is defined now by metamorphic biotite as well as white mica, some of which could possibly be detrital. However, the biotite could be mimetic after a sedimentary foliation so that the origin of the foliation remains equivocal. The best evidence for an S_1 tectonic foliation is the mesoscopic evidence presented above (Fig. 4b).

In the pelitic layers the S_2 foliation is generally penetrative with most layer silicates oriented with (001) parallel to the cleavage, or it is domainal with mica films dominating the structure. In the domainal type, layer-silicates oriented with (001) parallel to the foliation are long and thin whereas those with (001) markedly inclined to the foliation are more equant (see Etheridge and Lee 1975) indicating growth during development of the foliation.

Figure 8 shows the S_2 foliation in the context of a small fold and it can be seen that the morphology varies considerably from place to place. On the concave side of a competent layer where the fold is tight the foliation is parallel to the axial plane and penetrative. In the corresponding position in another more open fold the layer-silicate microfabric is strongly bimodal with the maxima symmetrical about the axial-surface. On the convex side of competent layer fold-closures the layer-silicate fabric is either tightly kinked thus defining a bimodal fabric that is symmetrical about the axial-surface or crenulated with a preferred orienta-



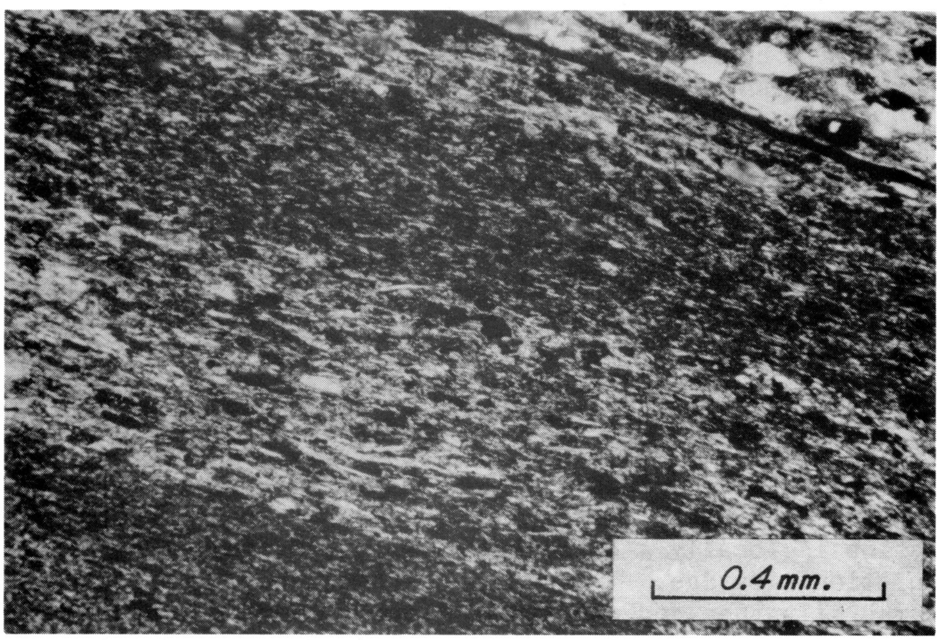


Fig. 9 - a) S_2 foliation cutting bedding. The foliation shows an anastomosing film type morphology in the coarse sediment and a kink or crenulation form in the finer sediment. b) Detail of a shale clast showing bedding and a bedding-parallel foliation defined by preferred orientation of layer-silicates.

tion parallel to bedding. On the limbs of the fold the foliation is penetrative and is parallel to bedding. In the latter situation the foliation is apparently inherited and is simply modified earlier bedding foliation, or bedding parallel tectonic foliation, that has been rotated into the S_2 orientation by the F_2 folding. The evidence that this is an earlier rotated foliation rather than a new one, apart from parallelism with bedding, is the fact that it can be traced around some F_2 folds. However, elsewhere this earlier foliation is transposed by kinks or crenulations into the S_2 foliation. The S_2 foliation is thus a composite surface in which transposition by folding is observed at different scales. Thus locally (principally in the hinges of small folds) there is transposition on a millimeter or finer scale, whereas elsewhere (principally in the limbs of small folds) the foliation is transposed on a larger scale. Similar morphologies are observed in experimentally produced folds and foliations (see Williams *et al.* 1977).

Kinks in the micas do not generally show strongly serrated boundaries suggesting that grain boundary migration has not been important (see Etheridge and Hobbs 1974). In general, however, the grains are not visibly deformed and it is fairly common for the kink-boundary to be occupied by a new grain oriented with (001) parallel to the kink-boundary. This also indicates that mica was still growing during development of S_2.

Where there are prophyroblasts in the pelites there are commonly crenulations associated with them and differentiation into quartz rich and layer-silicate rich domains. In that situation the S_2 foliation is locally a differentiated crenulation cleavage.

In the coarse siltstone and sandstone the S_2 foliation is defined principally by layer-silicate films which anastomose around clastic grains and quartz rich domains (Fig. 9a). The clastic grains generally show a preferred orientation which may be parallel to S_2 or parallel to bedding. Where the clasts, which are predominantly quartz, are parallel to bedding they invariably have a sedimentary appearance still with aspect ratios of the order of 2:1. Where the clasts are aligned parallel to S_2 they may have the same appearance or may be much more elongate, suggesting modification of shape by solution processes.

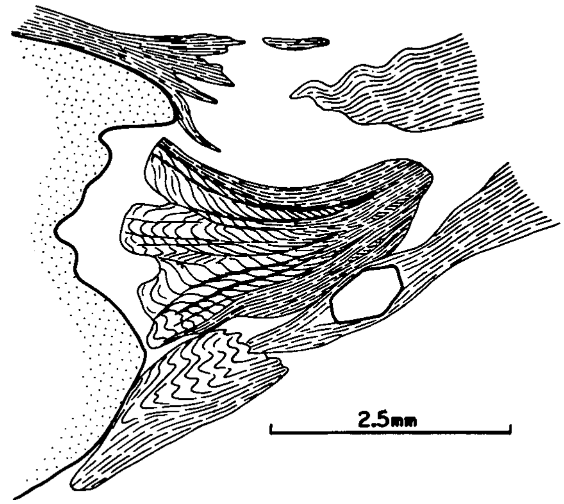


Fig. 10 - Heterogeneous foliation development in a shale clast adjacent to an igneous clast (stippled). Bending of the foliation and shale clasts around the igneous clast indicates that the latter is stronger than the material surrounding it. The shale clast in the centre of the sketch has apparently been partially protected from deformation by the igneous clast and thus has a broadly triangular form. In the weakly deformed portion there is a crenulation cleavage which can be traced into a slaty cleavage in the more deformed area (from left to right). A hexagonal porphyroblast of ?staurolite can be recognized in an adjacent clast.

In the conglomeratic sediments there is an S_2 foliation developed in the sediment as a whole and an S_2 foliation developed within some of the clasts. The former **is** defined principally by preferred dimensional orientation of the coarser clasts. There may or may not be preferred orientation of the finer matrix clasts, but, where a preferred orientation is present, the foliation is also defined by anastomosing layer-silicate films.

The foliation that is observed within clasts occurs in those clasts that are rich in layer-silicates and it varies considerably in morphology. There are elongate clasts in which the internal and external foliation and the length of the clast are all approximately parallel and the foliation is a penetrative cleavage defined by preferred orientation of layer-silicates. In some such clasts there is a layering defined

Fig. 11 - S_2 in adjacent shale clasts. The two clasts occupy the whole area of the photograph. In the left hand clast the foliation is a slaty cleavage and it is parallel to the long dimension of the clast although not parallel to the clast boundary as seen in this field of view. In the right hand clast the S_2 cleavage is a crenulation cleavage. This clast is elongate at a high angle to S_2.

by grain size and compositional variation. This layering is probably sedimentary and it is parallel to the cleavage suggesting that locally (i.e. within the clast), the "cleavage" is also of sedimentary origin (Fig. 9b). However there are other clasts where the foliation appears to be of tectonic origin. For example, the layer-silicate rich clast in Fig. 10 is adjacent to a stronger igneous clast and its funnel shape indicates that it has been protected locally from deformation by the stronger clast. In the broader, presumably less deformed portion of the funnel there is a differentiated crenulation cleavage that can be traced into the narrower part of the clast. As the narrower area is approached the crenulations straighten out and the microlithons narrow until there is no more evidence of crenulation and the mica preferred orientation is penetrative. A mechanism capable of explaining these observations has been proposed by Williams and Schoneveld (1981). It involves shear on the crenulated surface and unfolding of the crenulation at high strains.

Other elongate clasts with a penetrative foliation parallel to their length show relicts of kinked micas with the axial-surfaces of the kinks parallel to the foliation indicating that the foliation has developed by transposition of an earlier surface in the way described by Williams *et al.* (1977).

Another group of layer-silicate rich clasts is elongated in directions inclined to the external foliation. These clasts have two internal foliations. One is parallel to the length of the clast and is asymmetrically crenulated or tightly kinked such that the microfolds define a second foliation (Fig. 11). This S_2 is a crenulation cleavage, commonly differentiated, or a transposition foliation like the one described above, but in a less isoclinal state. There is a gradual transition from the latter foliation to the more intensely transposed foliation described above and the transition is accompanied by increasing parallelism of the clasts with the external cleavage. The S_2 foliation in these clasts is parallel to the external S_2 foliation.

Finally there is a group of more equant clasts that may have their long dimension parallel or perpendicular to S_2. These clasts are almost identical to the last group except that the S_2 kinks or crenulations are symmetrical and the enveloping surface to the internal S_1 is perpendicular to the internal and external S_2 (Fig. 12a). Both this group and the preceding oblique group tend to have a more ragged appearance than the slender smooth clasts of the first group. This ragged appearance is best developed in the last group and is due to the microfolding which involves the surface of the clast as well as the internal

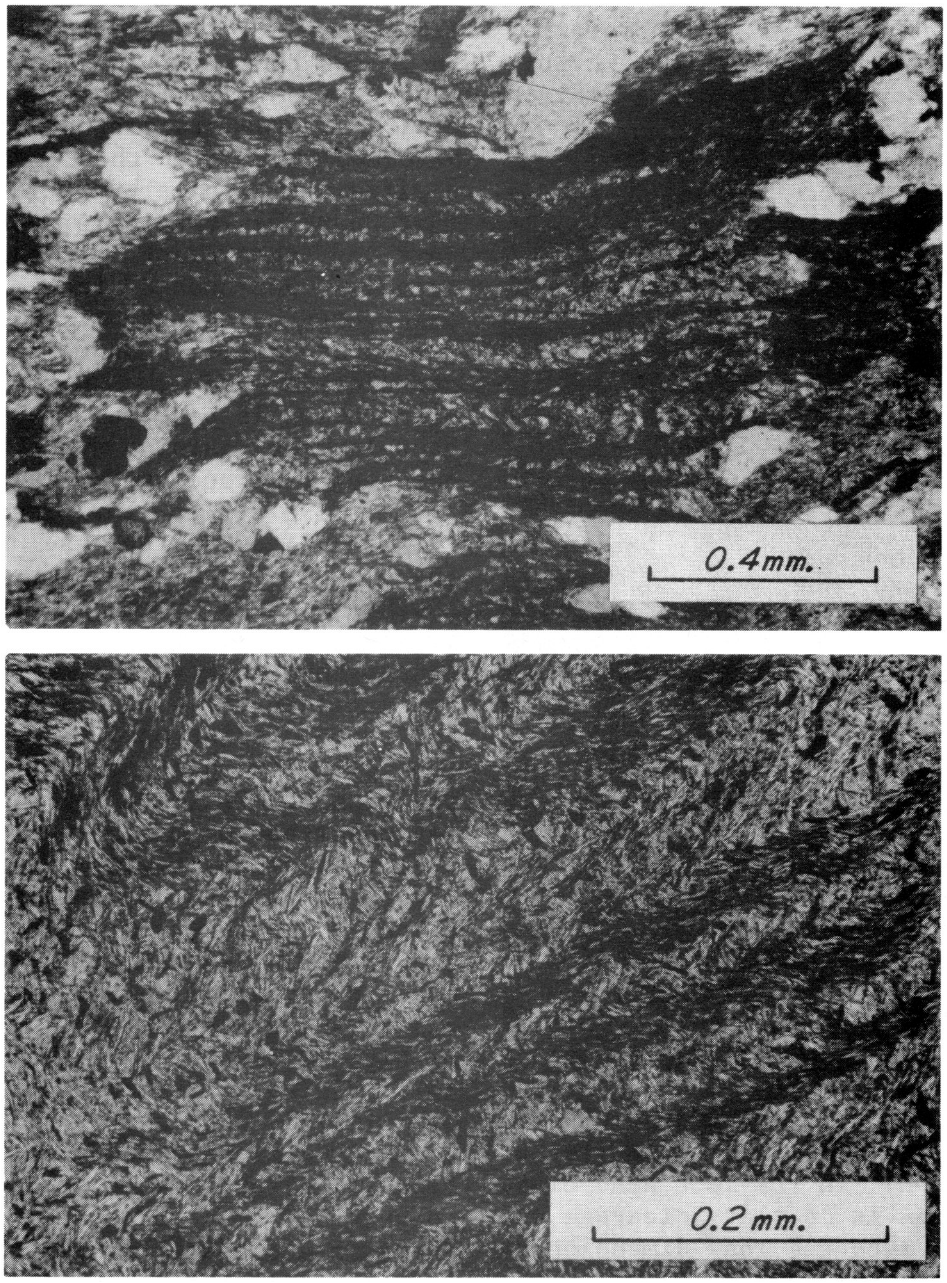


Fig. 12 - S_2 crenulation cleavage in a shale clast with the enveloping surface to the crenulations approximately perpendicular to S_2. a) general view of clast. b) detail of crenulations.

foliation. Again there is a complete transition between the more equant and the oblique group of clasts in shape and cleavage morphology.

Kinked micas in the various clasts appear identical to those already described from the pelite beds (compare Figs. 9a and 12b). Kink-boundaries are only weakly serrated but are commonly occupied by new grains oriented with (001) parallel to the axial-surface of the kink.

In addition to the S_1 and S_2 foliations described above there is a third weak crenulation cleavage that overprints S_2. Locally it is an asymmetric crenulation rather than a crenulation cleavage. It is developed in pelitic beds and in pelitic clasts and is inclined to S_2 at approximately 45°. It does not appear to be associated with any important folding in the area described.

Porphyroblasts

The porphyroblasts that are completely replaced by chlorite and white mica commonly have an internal foliation. They are found both in the layer-silicate rich clasts and in pelite beds and the same range of structures is found in both situations.

Where the S_1 foliation is preserved in the rock hosting the porphyroblasts, as for example in parts of Figure 8, the relationship between prophyroblast and matrix varies with the orientation of S_1. Where S_1 wraps around and F_2 fold closure, S_i (internal foliation) and S_e (external foliation) are continuous and where S_e is crenulated the porphyroblasts coincide exactly with one limb of the crenulations, which are parallel to S_2, in a way that suggests that they were present when the crenulation developed. They are never seen to overgrow a crenulation heliciticly. Where the prophyroblasts are numerous the crenulations are more variable in orientation than elsewhere suggesting that the presence of the porphyroblasts as abundant competent bodies increased the heterogeneity of strain associated with development of the crenulations. There are generally quartz rich pressure shadows associated with the porphyroblasts such that they are elongated in the plane of the crenulation cleavage, (i.e. the S_2 foliation). This also indicates that they were already present during the development of S_2.

In areas where S_1 is rotated parallel to S_2, as in the limbs of isoclinal F_2 folds, the S_1 foliation bends around the prophyroblast in a way that is commonly taken as an indication that the foliation is later than the porphyroblast (see Zwart 1960). In the rocks described here, S_i and S_e are generally continuous so that the structure does not indicate that S_1 is post porphyroblastesis but that development of the microstructure continued after growth of the porphyroblast. The most reasonable explanation is that the porphyroblasts overgrew S_1 which was later transposed by the F_2 folding into the S_2 orientation. At that time the transposed S_1 was flattened around the porphyroblasts (Fig. 13) and quartz pressure shadows grew in the plane of the transposed, old foliation.

In areas where there is a well developed S_2 transposition foliation S_i is generally inclined to $S_2(=S_e)$ by an angle approaching 90° and S_i and S_e are discontinuous. In such situations the S_2 foliation bends around the porphyroblasts and pressure shadows again extend the porphyroblast in the plane of S_2.

Lack of pressure shadows in S_1 and absence of bending of S_1 around porphyroblasts, except in areas where modificiation of S_1 by F_2 folding can be reasonably expected, suggests that the porphyroblasts overgrew S_1 either in the very late stages of S_1 development or after S_1 development. The various relationships between the porphyroblasts and S_2 crenulations, F_2 modified S_1 and S_2 transposition foliation all indicate consistently that S_2 developed later than the porphyroblasts although the possibility of overlap cannot be overruled. Thus, it is concluded that the porphyroblasts grew some time between

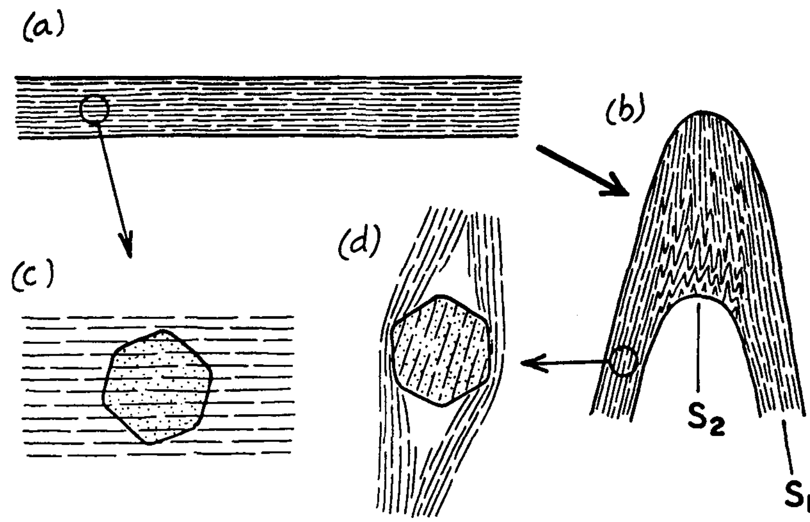


Fig. 13 - Model to explain timing of foliation and fold development and porphyroblastesis. The porphyroblast overgrows S_1 prior to F_2 folding (a & c). S_1 is then folded (b). In the F_2 hinge small-scale folding and transposition of S_1 produces a new axial plane foliation (S_2). In the F_2 limb S_1 undergoes bulk rotation and there is no new foliation developed. S_1 is however modified; it is flattened around the porphyroblast and a pressure shadow develops (d).

the late stages of the development of S_1 and the early stages of the development of S_2.

Relationships between the smaller garnet porphyroblasts and the foliations are more obscure. However, there are two consistent observations in areas where the surrounding foliation is an S_2 transposition surface. The foliation bends around the garnets and pressure shadows extend the garnets in the plane of S_2. Thus the garnets are pre or early syn S_2 and could be contemporaneous with the large porphyroblasts.

F_2 movement zones

The microstructure of the S_2 parallel movement zones comprises anastomosing mica-films and looks very much like a normal S_2 foliation. The difference is one of scale. The normal S_2 network has a film spacing measurable in tenths of a millimeter and film thickness measurable in hundredths. In the movement zones however, in addition to the normal S_2 network there are coarser films which are measurable in tenths of a millimeter and have spacings measurable in millimeters. There is no other difference, they are part of the anastomosing S_2 network. Some of the coarser films include large porphyroblasts. Layer-silicates in such films bend around the porphyroblasts and pressure shadows extend parallel to the film. Final deformation in the films is therefore post porphyroblast growth.

These are the zones interpreted by Pajari *et al.* (1979) as liquefaction structures (see above).

Veins

All veins examined by the writer are composed of quartz and have an equant polygonal grain shape. Grains show weak undulose extinction and large grains have well developed substructure. Grain boundaries are commonly pinned by fine grained micas but where the boundaries are clean good 120° triple point junctions are developed.

Inspection by means of a gypsum plate reveals a very strong crystallographic preferred orientation that results in the various colours sweeping around folds as the microscope stage is rotated.

This fabric and microstructure is typical of quartz that has recrystallized as a result of deformation and has undergone normal grain growth (see e.g. Hobbs *et al.* 1976, p. 110 and 111). It would not be expected of a weakly deformed quartz vein that was injected, along a folded surface in previously folded material.

DISCUSSION AND CONCLUSIONS

Timing of the deformation and the mechanism of mélange formation

There are structures, such as convolute bedding, that are clearly pre-lithification and others, such as dismembered folds, that may be pre-lithification. However, contrary to previous interpretation (Pajari *et al.* 1979) the S_2 foliation and associated folds are interpreted here as post-lithification structures. The evidence for this conclusion can be summarized as follows:

(i) S_2 is post-quartz veins and post-metamorphic porphyroblasts.

(ii) Quartz veins are folded by F_2 folds and the quartz fabric confirms that these folds really are deformational structures, as opposed to veins mimicking an earlier folded surface.

(iii) Development of pre F_2 boudins has been accompanied by formation of quartz veins. Thus demonstrating again that the F_2 deformation followed the formation of the quartz veins and was therefore presumably post- lithification Similarly, quartz veins are themselves boudinaged and faulted by movement associated with the development of S_2 and F_2 folds.

It is less clear as to whether F_1 folds are pre or post-lithification. They may be coeval with metamorphism and with formation of some quartz veins. However, the evidence is ambiguous and they may in fact pre-date both. Nevertheless all the structures observed on Green Island (excepting convolute bedding) can be explained in terms of hard-rock processes acting on a sequence of turbidites and associated conglomerates. The liquefaction structures of Pajari *et al.* (1979) have been shown to be hard-rock and all of the folding could be hard-rock; there is no contrary evidence. Further, the "olistostromal horizons" may also have been incorrectly identified.

The olistostromal horizons are not all simply matrix supported, coarse clastic sediments as they appear to be in the field and as they have been described by Pajari *et al.* (1979). The finer grained examples described here

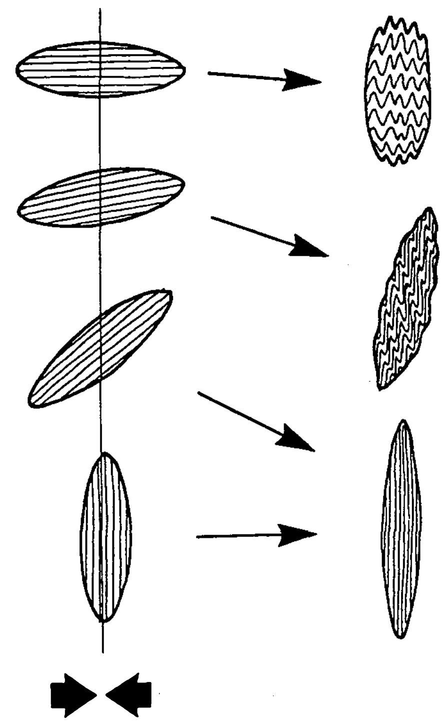


Fig. 14 - Summary of clast morphologies and model for their development. The morphology of the clasts and the nature of the foliation (represented by the right hand column) is a function of the initial orientation of the clast (left hand column). The foliation in clasts initially oriented parallel or almost parallel to the direction of shortening is folded to form a new crenulation cleavage. All other orientations are simply rotated and their initial foliation preserved. Note that this last group is likely to be numerically small since both the direction of shortening and the initial orientation of the long dimension of most of the clasts were approximately parallel to bedding. Thick arrows indicate directions of shortening.

are clast supported grits and conglomerates in which pelitic rock fragments are predominant. Pajari *et al.* (1979) believed that the pelitic fragments were essentially still unlithified at the time of deformation and, for this reason only, they believed that they were deposited as unlithified clasts. Since it has been shown here that they were lithified at the time of deformation (i.e. during the development of S_2) there is now no evidence that they were deposited as unlithified clasts. They may in fact have been deposited as shale or even as slate clasts. In view of the lack of pelitic matrix it is more reasonable to suggest that these sediments are simply clast supported conglomerates of the type commonly associated with turbidites (Walker 1979).

The coarser sediments, which contain blocks that may be measurable in tens of centimetres, are more enigmatic Some of these may also be clast supported with a preponderance of, hard to identify, pelitic fragments. Others however, certainly appear now to be matrix supported with sandstone blocks "floating" in silt. These sediments however, lack exotic blocks so that they could equally well be interpreted as boudinaged, thin sandstone beds, intercalated with thicker shale beds. The fact that many of the blocks are isoclinally folded and the folds themselves boudinaged (Fig. 4), with at least some of the boudinaging (e.g. Fig. 4a) post lithification, indicates that strain magnitudes are high and that the deformation environment is one that is compatible with the hard-rock, disruptive folding explanation. The writer favours this interpretation for these particular units in view of the general deformational picture but the evidence is so far inconclusive. This problem merits further work since it has important regional implications. As Hsü (1974) pointed out, if the mélange is of deformational origin, it implies that the sequence on one side of the mélange is allochthonous with respect to the sequence on the other side. That is not necessarily implied by an olistostrome.

Development of the S_2 cleavage

Several processes have operated in the development of the S_2 cleavage. They can be thought of as competing processes (cf. Williams 1977), the relative importance of which depend not only on metamorphic conditions but on such factors as initial fabric and microstructure and the kinematic history.

Rotation is perhaps the most important single factor in these rocks. It takes two forms. There is rigid body rotation of detrital grains and larger sedimentary clasts and there is the more ordered rotation of earlier foliations by microfolding. The rigid-body rotation is indicated by parallelism of bedding in pelitic clasts with S_2 (Fig. 9b) and by dimensional preferred orientation of only very weakly deformed quartz clasts. The large rotations observed indicate that grain and clast boundary sliding must have been important deformation mechanisms, probably facilitated by diffusion processes. This type of deformation mechanism has been named particulate flow by Borradaile (1981) following the terminology of soil scientists. Quartz veins, because of their aspect ratios, were unable to rotate without deformation and thus quartz within the veins is highly deformed and recrystallized whereas adjacent quartz clasts show only undulose extinction (cf. Williams 1972). This suggests that the fine grained matrix associated with the clasts has aided the rigid body rotation, as is to be expected if diffusion dependant deformation was dominant. Whether a pelitic clast underwent rigid body rotation or underwent deformation and internal rotation by microfolding, or a combination of both, presumably depended on initial orientation. A clast aligned parallel to the shortening direction would fold whereas one inclined to the shortening direction would rotate (Fig. 14).

Growth of new micas has been recorded and it played a definite role in the development of S_2. Its role however, was subordinate to rotation mechanisms. Differentiation has also featured in the cleavage development but well defined layering is not very common so that it does not appear to have been an important process. However, the absence of obvious differentiation does not mean that mass transfer processes have been unimportant; they are simply difficult to recognize.

Axial-plane cleavages have been made experimentally in salt/mica schist (Williams *et al.* 1977) and the deformation mechanism in that material is one of transposition of an earlier foliation by microfolding. It has been suggested (*ibid.*) that the same mechanism may be important in the development of natural foliations, especially in high-grade rocks. The foliations described here from the deformed shale clasts and from shale beds are very similar in appearance to those observed in the experimental materials. The main differences are that the natural micas look less deformed and they show a domain dependant variation in aspect ratios. Both differences can be explained by syn or post-deformational recovery and growth of the micas. In the advanced stages of transposition or after extensive growth (see Etheridge and Hobbs 1974) it is difficult to recognize evidence of this mechanism. However, it has obviously been important in the rocks described here and I would suggest that it is one of the more common mechanisms in mica rich rocks in general.

It has been stated that the morphology of the S_2 foliation in the coarser clastic rocks indicates a soft-sediment origin (Pajari *et al.* 1979). However, the evidence presented here shows that it developed in hard-rock and I am not aware of any morphological feature that is incompatible with such an origin or that cannot be found in rocks where foliation development is known to postdate metamorphism. The penetration of bedding by shale clasts was considered important evidence of soft-sediment deformation by Pajari *et al.* (1979). This microstructure however simply indicates that the clasts have deformed, it says nothing about the state of the clast at the time of deformation. This underlines the general point that morphology is rarely diagnostic of soft or hard-rock deformation.

ACKNOWLEDGEMENTS

I wish to thank George Pajari and Ron Pickerill for first introducing me to the rocks of Green Island. Karl Karlstrom, Ron Pickerill, Guy Plint, Ben van der Pluijm, Cees van Staal and Joe White are gratefully acknowledged for their criticism of various drafts of the manuscript. Sherri Townsend was responsible for word processing and Gary Landry and Bob McCulloch assisted with preparation of the figures. Fieldwork was supported by Memorial University and N.S.E.R.C. operating grant A7419.

ARNOTT, R.J. 1983. Sedimentology and stratigraphy of Upper Ordovician - Silurian sequences on New World Island, Newfoundland: Separate fault controlled basins? Canadian Journal of Earth Sciences, 20, pp. 345-354.

BORRADAILE, G.J. 1981. Particulate flow of rock and the formation of cleavage. Tectonophysics, 72, pp. 305-321.

CURRIE, K.L., PAJARI, G.E. and PICKERILL, R.K. 1980. Geological map of the Carmanville Map Area (2E/8), Newfoundland. Geological Survey of Canada Open-file map 721.

ETHERIDGE, M. A. and HOBBS, B.E. 1974. Chemical and deformational controls on recrystallization of mica. Contributions to Mineralogy and Petrology, 38, pp. 21-36.

ETHERIDGE, M.A. and LEE, M. F. 1975. Microstructure of slate from Lady Loretta, Queensland, Australia. Bulletin Geological Society of America, 86, pp. 13-22.

HELWIG, J. 1970. Slump folds and early structures, northeastern Newfoundland Appalachians. Journal of Geology, 78, pp. 172-187.

HOBBS, B.E., MEANS, W.D. and WILLIAMS, P.F. 1976. An outline of Structural

Geology, John Wiley & Sons, New York, 571p.

HORNE, G.S. 1970. Complex volcanic - sedimentary patterns in the Magog Belt of northeastern Newfoundland. Bulletin Geological Society of America, 81, pp. 1767-1788.

HSÜ, K. 1974. Mélanges and their distinction from olistostromes. *In* Modern and ancient geosynclinal sedimentation (*Edited by* Dott, R. & Shaver, R.). Society of Economic Palaeontologists and Mineralogists Special Publication No. 19, pp. 321-332.

KARLSTROM, K.E., van der PLUIJM, B.A. and WILLIAMS, P.F. 1982. Structural interpretation of the eastern Notre Dame Bay area, Newfoundland: regional post-Middle Silurian thrusting and asymmetrical folding. Canadian Journal of Earth Sciences, 19, pp. 2325-2341.

NAYLOR, M.A. 1982. The Casanova complex of the Northern Apennines: A mélange formed on a distal passive continental margin. Journal Structural Geology, 4, pp. 1-18.

NELSON, K.D. 1981. Mélange development in the Boones Point Complex, north-central Newfoundland. Canadian Journal of Earth Sciences, 18, pp. 433-442.

PAJARI, G.E. and CURRIE, K.L. 1978. The Gander Lake Davidsville Groups of northeastern Newfoundland: a re-examination. Canadian Journal of Earth Sciences 15, pp. 708-714.

PAJARI, G.E., PICKERILL, R.K. and CURRIE, K.L. 1979. The nature, origin and significance of the Carmanville ophiolitic mélange, northeastern Newfoundland. Canadian Journal of Earth Sciences, 16, pp. 1439-1451.

WALKER, R.G. 1979. Turbidites and associated coarse clastic deposits. *In* Facies Models (*Edited by* Walker, R.G.) Geoscience Canada, pp. 91-103.

WILLIAMS, H. 1979. Appalachian Orogen in Canada. Canadian Journal of Earth Sciences, 16, pp. 792-807.

WILLIAMS, P.F. 1972. Development of metamorphic layering and cleavage in low grade metamorphic rocks at Bermagui, Australia. American Journal of Science, 272, pp. 1-46.

——————— 1977. Foliation: a review and discussion. *In* Fabrics, Microtextures and Microtectonics (*Edited by* Lister, G.S., Williams, P.F., Zwart, H.J. and Lisle, R.J.). Tectonophysics 39, pp. 305-328.

——————— 1979. The development of asymmetrical folds in a cross-laminated siltstone. Journal of Structural Geology, 1, pp. 19-30.

WILLIAMS, P.F., COLLINS, A.R. and WILTSHIRE, R.G. 1969. Cleavage and penecontemporaneous deformation structures in sedimentary rocks. Journal of Geology, 77, pp. 415-425.

WILLIAMS, P.F., MEANS, W.D. and HOBBS, B.E. 1977. Development of axial plane slaty cleavage and schistosity in experimental and natural materials. Tectonophysics, 42, pp. 139-158.

WILLIAMS, P.F. and SCHONEVELD, C. 1981. Garnet rotation and the development of axial plane crenulation cleavage. Tectonophysics, 78, pp. 307-334.

ZWART, H.J. 1960. Relations between folding and metamorphism in the Central Pyrenees, and their chronological succession. Geologie Mijnbouw, 22, pp. 163-180.

Reviewers: W.D. Means
G. Borradaile