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Microstructural and fabric studies from the rocks of the Moine Nappe, Eriboll, NW Scotland

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Abstract—Microstructures and quartz c-axis fabric diagrams from mylonites and psammitic Moine schists, collected in traverses across the lower levels of the Moine Nappe in the Eriboll area, are presented. On approaching the Moine Thrust from the Kyle of Tongue, the following microstructural sequence is encountered: interlayered coarse grained biotite psammitic and schistose tectonites being in part mylonitic with two platy slide zones, one containing biotite and the other only muscovite and chlorite and both showing quartz microstructures indicative of post-tectonic relaxation; these pass into more mylonitic rocks nearer the thrust zone which in turn passes into the main chlorite-grade mylonite belt and finally, adjacent to the Moine Thrust, into reworked lower chlorite grade mylonites. Although there is some local variation, the overall quartz c-axis fabric is an incomplete asymmetric type I girdle. The main variation is the development of type II girdles in the reworked, ultrafine grained mylonites. The extent of the mylonitization is more extensive than previously reported. Studies of folds within the mylonite belt have revealed eye structures and small-scale folds; many are sheath folds. They cannot be unequivocally correlated with large-scale recumbent folds within the Moine Nappe. The results presented indicate that mylonitization is not limited to a single phase, and raises the possibility that there may be earlier Caledonian or possibly Precambrian structural elements present in the Eriboll region Moines prior to much of the mylonitization.

INTRODUCTION

THERE HAVE been several recent advances in our understanding of the processes of mylonitization and in the development of structural features within thrust and mylonite zones. These include:

(i) recrystallization producing quartz mylonites is usually syntectonic and not a post-tectonic grain growth of cataclastic fragments (White 1973a, Bell & Etheridge 1973);

(ii) mylonites are weaker than the country rock from which they are derived and consequently, deformation is preferentially concentrated in mylonite zones and may bear little relationship to the deformation within the country rocks (Watterson 1975, White 1976, 1977, White *et al.* 1980);

(iii) folds in mylonite zones develop and refold continuously during shearing (Carreras *et al.* 1977);

(iv) linear and planar fabrics rotate with increasing shear strain in shear regimes (Sanderson 1973, Escher & Watterson 1974, Carreras *et al.* 1977, Rhodes & Gayer 1977, Bell 1978, Quinquis *et al.* 1978, Williams 1978, Berthe & Brun 1980, Cobbold & Quinquis 1980) and

(v) the inference from (iii) and (iv) is that the orientation of structural elements within mylonite zones cannot be unequivocally correlated with similarly oriented structures outside mylonite zones.

The Moine Thrust Zone and its easternmost, highest level major thrust, the Moine Thrust (Fig. 1) extends for a distance of over 190 km and traditionally marks the northwest boundary of the Caledonian Orogen in Britain, although recent interpretations of the Outer Isles Thrust would question this (see e.g. Coward 1980). It is a belt of foreland thrusting and, in general, metasediments (Moines) of the Moine Nappe were thrust WNW over the foreland of early Proterozoic Lewisian gneiss, unmetamorphosed late Proterozoic Torridonian and Cambro-Ordovician sediments. The Moine Thrust Zone beneath the Moine Thrust varies in thickness and complexity, with several thrust-bound nappes containing imbricated sequences involving Lewisian and Cambro-Ordovician rocks.

Recent studies (Elliott & Johnson 1980, McClay & Coward 1981) indicate that in the Eriboll area, the lower thrusts, i.e. the Sole and Arnaboll, have generally 'piggy-backed' the higher Moine Nappe (and the Moine Thrust) into its present position. An implication is that the Moine mylonites within the Moine Nappe formed at the greatest depth. The sequential generation of, and movement along, thrusts within a thrust zone means that the mylonites directly associated with each thrust formed at different depths and times during the thrust zone history. It also means that mylonites may be reworked (remylonitized) at shallower levels.

In this contribution a reappraisal of the mylonitization in the Moine Nappe and its relationship to the Moine Thrust and Moine Thrust Zone, is made in the light of the outlined advances. A brief review of the Moine and Moine Thrust literature relevant to this study is provided first.

HISTORICAL REVIEW

The age relationships between mylonitization, thrusting and regional metamorphism and migmatization within the Moine schists have been a matter of debate for almost a century. From the earliest days, two views have

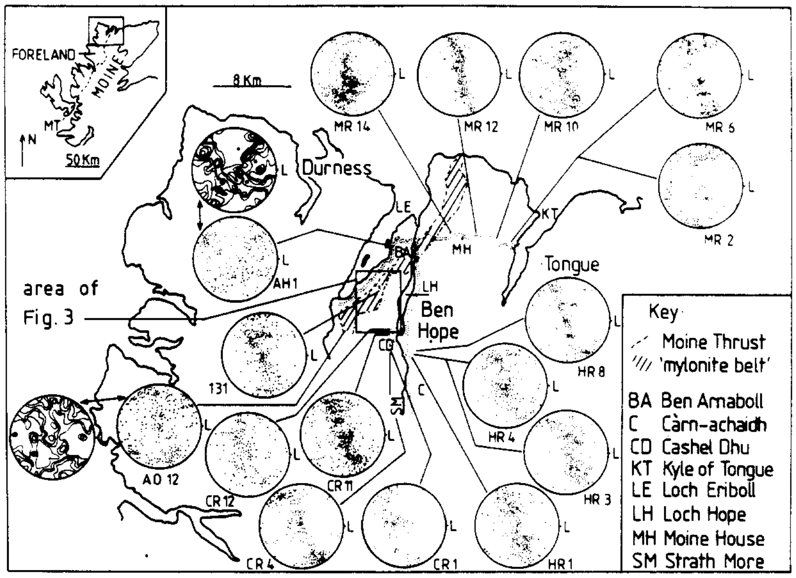

Fig. 1. Map of the area described in text showing the main traverses and quartz c-axis fabric diagrams across the lower levels of the Moine Nappe.

been held: (i) the mylonites formed in the thrust zone during Caledonian regional metamorphism of the Moine schists (Lapworth 1885, Peach in Peach & Horne 1930) and supported to varying degrees by Kennedy (1949, 1955), MacGregor (1952) and Christie (1960, 1963) and (ii) there was an earlier regional metamorphism with a later retrogressive metamorphism, associated with the thrusting, superimposed in the west, i.e. the mylonites postdate regional metamorphism (Horne in Peach & Horne 1930, Read 1931, 1934). The last view was supported by Phillips (1937, 1945, 1949) who concluded from quartz fabric studies that there was . . . 'a dislocation metamorphism undoubtedly superimposed upon the general Moine metamorphism'. . . and also by Wilson (1953) and McIntyre (1954).

The timing of events in (ii) have normally been regarded as Precambrian for regional metamorphism and Caledonian for thrusting and thermal metamorphism.

Kennedy (1949, 1955), whilst recognizing two metamorphisms, regarded them as separate early and late Caledonian events. Johnson (1956, 1957, 1960) and Barber (1965) suggested that the mylonites in the Loch Carron–Loch Alsh areas formed first as a result of cataclasis followed by a recrystallization and folding. Johnson (1956) also concluded that there was a western fabric associated with thrusting superimposed upon an eastern fabric and that both the regional metamorphism and large-scale folding of the Moine schists and subsequent Moine Thrust movements occurred during the Caledonian orogeny. Christie (1960, 1963) also proposed that there were two phases in the development of the mylonites which differed from the above. He postulated that primary mylonitization was contemporaneous with regional metamorphism of post-Cambrian age and was followed by local cataclasis of the primary mylonites to form secondary mylonites.

An additional controversy was whether there was a general decrease in metamorphic grade within the Moines from a zone of Caledonian migmatization towards the west, i.e. towards the Moine Thrust Zone, especially in the southwestern part of the Northern Highlands, and the relationship of the metamorphism to regional tectonics. The existence of a similar decrease in metamorphic grade in the northern (Sutherland) Moines was questioned (Watson 1963).

It was against this background of controversy that Soper and coworkers (Soper 1971, Soper & Brown 1971, Soper & Wilkinson 1975) set out to 'establish a regional metamorphic zonation' and to find the relationship in time between the thermal maximum in the interior of the orogenic belt and the westerly margin. They established the following four-phase history for the mylonites in the north Sutherland region: D1, the generation of the mylonitic foliation ($S1$) with intrafolial folds ($F1$), many of which were 'streaked out' (Christie 1963) and all of which predated a thermal maximum; D2, is coeval with the thermal peak which recrystallizes the $S1$ fabric (forming the blastomylonite fabric of the mylonites) and Moinian rocks above, and marked by a widely developed extension lineation ($L2$) superimposed upon the composite $S1$–$S2$ mylonite fabric and has associated with it folds ($F2$) about an ESE direction; D3, folding about N–S axes forming the belts of 'rucked up schists' with associated strong crenulations and D4, development of brittle conjugate structures, e.g. boxfolds and kink zones.

Soper & Wilkinson (1975) correlated similarly oriented structures within the Moine schists with those in the mylonites and suggested that both sets were produced during the same deformation phase. Thus arose a rigid framework for the deformation of the Moine schists in the north Sutherland area which precluded any Pre-Caledonian metamorphism or deformation. The key correlations upon which this was based were the extension lineation ($L2$) within the mylonite belt with the lineation in the Moine schists and the minor folds ($F2$) in the mylonites with the large folds present in the Moine schists (e.g. upon An Lean Charn). They also held the view that the mylonitization with its associated recrystallization and folding, because they affected Cambro-Ordovician sediments, must therefore, place a lower age limit of post Cambrian for the mylonitization, thrusting and regional metamorphism.

The exclusively Caledonian history of the west Sutherland Moines (represented as the Morar Division in Rathbone & Harris 1979), as far east as the Naver Slide and the migmatite complex, proposed by Soper and coworkers, did not agree with work in the southern and southwestern Moines (e.g. Giletti *et al.* 1961, Long & Lambert 1963, Pidgeon & Johnson 1973, van Breeman *et al.* 1974) which pointed to one or more Precambrian orogenic events. Long (1964) and Johnson & Shepherd (1970) believed that the main regional metamorphism of the Moinian in the Carn Chuinneag area was post-Cambrian and postdated an earlier low grade Moinian fabric of uncertain age prior to the intrusion of the Carn Chuinneag granite at 530 ± 10 Ma. It was suggested that Precambrian deformation could not be ruled out. Johnson *et al.* (1979), following a recalculation of the intrusion date which gave 560 ± 10 Ma, concluded that the pre-granite schistosity and folding in the surrounding Moines are unlikley to be Caledonian, thereby providing further evidence for Precambrian tectonism in the more northern Moines.

Barber & May (1976) in the Attadale to Arnisdale region recognised the effects of the Caledonian deformation upon earlier fabrics in the Moine and Lewisian. A late stage deformation (D4) within the Moine Nappe, correlated with the formation of the mylonites in the thrust belt, was shown to postdate an earlier phase of deformation (D3) and amphibolite facies metamorphism. As mylonitization in the thrust belt is the earliest phase of (post-Cambrian) Caledonian deformation, the earlier phases of deformation (D1–D3) are Precambrian, this interpretation is supported by the radiometric dating in the Moine.

More recent work (Powell 1974, Brook *et al.* 1976, 1977, Brewer *et al.* 1979) has supported the presence of Precambrian events in the southwestern Moines. It is becoming increasingly clear that, in this area, the Moine Thrust Zone is superimposed upon Moine rocks already deformed and metamorphosed to approximately garnet grade during the Precambrian.

In spite of the problems of correlation between the four-phase tectonic framework in the northern Moines with events in the southwestern Moines, the work of Soper & Brown (1971) has been the basis for later work on the Sutherland Moines (Soper & Wilkinson 1975, Soper & Barber 1979, Fettes 1979, Mendum 1979). The inability to correlate this sequence with that of the southwestern Moines has been explained by hypothesizing that a Precambrian metamorphic front lay between the two areas (Brook *et al.* 1977, Soper & Barber 1979) or that deposition of the west Sutherland Moines could postdate the Precambrian orogenic phase. On the other hand, some workers have inferred that Precambrian events are present in the northern Moines (Johnson 1975, Moorhouse & Moorhouse 1979a,b, Coward 1980, McClay & Coward 1981). This view has received support from Lintern *et al.* (1982) who described granitic masses in the Strath Halladale region of northeast Sutherland which cross-cut early structures possibly related to the regional D2 structures of Soper and coworkers to the west. These bodies have been dated at 649 ± 32 Ma. Difficulties also arise from the presence of slide zones within the northern Moines (Read 1931, Peacock 1975, Moorhouse & Harrison 1976, Mendum 1979, Soper & Barber 1982), and correlations of structures across these. It is a conclusion of this study that the tectonic framework of Soper and coworkers is in need of revision.

STUDY UNDERTAKEN

The detailed petrographical study was based upon field mapping and specimen collection in the lower levels of the Moine Nappe to southeast of Loch Eriboll, north Sutherland (Fig. 1).

The collection of oriented samples was made along four main traverses (see Fig. 1). The northernmost traverse sampled a series of freshly exposed Moinian rocks along the road over A'Mhoine. Another set of samples was obtained along the eastern shoreline of Loch Hope, in particular the lower cliff line and steep western escarpment of Ben Hope at the northern end of Strath More. A third traverse was made from the cliff lines just west of Cashel Dhu at the southern limit of the Creagan Road, northwards across the intervening ground into the mylonite belt outlined by Soper & Wilkinson (1975) at the base of the Moine Nappe, where sampling was more frequent. A parallel sequence of samples was taken from the summit of An Lean Charn down through the lower level Moines and weakly mylonitized Moines of Soper & Wilkinson (1975). Oriented thin sections were cut in the XZ plane (Fig. 2) i.e. parallel to the lineation, the morphology of which suggests it marks the maximim finite elongation (X), and perpendicular to the foliation which is assumed to represent the XY plane of the finite strain ellipsoid. These two fabric elements define a reference framework, to which the quartz fabrics were related.

In the following sections, the microstructures and fabrics will be described in two sections. They will be: (1) the mylonites and (2) the variably mylonitized and platy Moines. The division between the sections corresponds

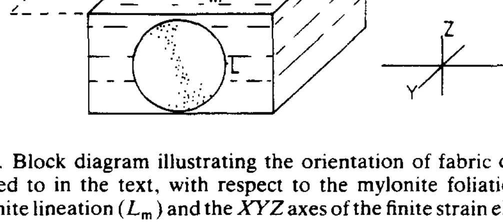

Fig. 2. Block diagram illustrating the orientation of fabric diagrams referred to in the text, with respect to the mylonite foliation (S_m), mylonite lineation (L_m) and the XYZ axes of the finite strain ellipsoid.

to the boundary between Soper & Wilkinson's (1975) distinction between mylonites and weakly mylonitized Moines. The descriptions will be followed by an outline of the fold structures in each area.

THE MYLONITES

Stratigraphy

The rocks forming the mylonite belt at the base of the Moine Nappe attain perhaps their best development in the ground immediately west of the Creagan Road (Fig. 3). The present study has been largely confined to the area lying southeast of the major chlorite phyllonite (Oystershell Rock) unit which occurs in the lower levels of the mylonite belt. The base of the Moine mylonites was taken as the top of the Oystershell Rock which along with the underlying quartzo-feldspathic mylonites are regarded as being of Lewisian origin (see also McClay & Coward 1981).

The sequence studied in this area (see Fig. 3) is described below in a structurally ascending order. The *chlorite phyllonite (Oystershell Rock)* is followed by *quartz mylonites* separated into two by a variably developed and laterally discontinuous green feldspathic crenulated schistose mylonite (phyllonite). The boundary with the chlorite phyllonite is complex, being interleaved, folded and associated with a large number of quartz veins. In addition to recrystallized quartz, the quartz mylonites contain small quantities of alkali and plagioclase feldspar and variable amounts of spindly white micas (±chlorite) aligned parallel to the mylonite foliation and which define a compositional layering. Granular epidote is present and appears as either strung-out aggregates or single grains. Little biotite and no garnets are found. A more extensive *green feldspathic schistose mylonite (phyllonite)* with a crenulated appearance is the next unit. It contains pods or areas of blocky gneissic material (e.g. NC 435535) and has within it a thin green chlorite phyllonite. The schistose mylonite has clasts of alkali feldspar (±plagioclase) set in an inosculating matrix of quartz, chlorite, muscovite and epidote. *Quartz mylonite,* banded in areas, reflecting changes in the phyllosilicate context (e.g. NC 423535) is the uppermost member of the main chlorite grade mylonite belt. It is dominantly recrystallized quartz and of the same mineralogy as the quartz mylonite units beneath, with areas rich in alkali feldspar (e.g. NC 433535). Rare pods of megascopically unmylonitized cross-bedded psammitic material are found in the uppermost quartz mylonites, e.g. NC 429538 and 418536. Locally (NC 422534), a discontinuous slice of schistose mylonite, similar to the above, occurs above the quartz mylonites.

All of the above mylonites exhibit a strong mineral lineation plunging to the ESE in the unfolded areas and their mineral assemblages are consistent with mid-greenschist facies metamorphism, as described by previous workers (MacGregor 1952, Johnson 1960, Barber 1965, Soper & Brown 1971, Soper & Wilkinson 1975).

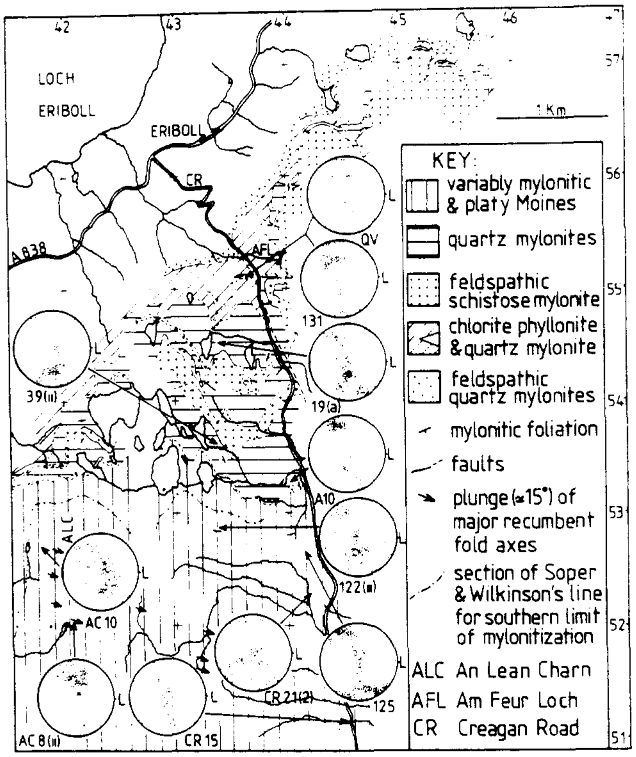

Fig. 3. Detailed map of the area marked on Fig. 1, illustrating the mylonite belt referred to in the text, with accompanying quartz c-axis fabric diagrams and the specimen locations. Abbreviation QV accompanying a diagram for specimen 131 denotes the fabric diagram for a mylonitized quartz vein.

It was found that the thickness of the individual units varied greatly when traced laterally from the Creagan Road.

Microstructures

The microstructures of the upper and lower *quartz mylonites* are shown in Fig. 4. In specimens showing little effect from late fold fabrics (see below and fold section), it is found that the quartz mylonites exhibit equidimensional (mean size 80–100 μm) and elongate asymmetrical grains, the sense of asymmetry being the same as the shear sense (Fig. 4a). Quartz grain boundaries are irregular and serrate with small recrystallized grains evident, many of which also exhibit internal strain features, namely undulose extinction, deformation bands and subgrains (see White 1973b). No remnant grains or quartz ribbons were identified, although there are signs of bands of grains with a similar orientation, suggesting ribbons were formerly present as observed by Soper & Wilkinson (1975). The microstructures and textures indicate that the quartz mylonites formed during dynamic recovery and recrystallization (White 1977).

Feldspar clasts in the quartz mylonites (perthitic alkali feldspar and albitic plagioclase) are generally small in size (up to 300 μm) and may exhibit shear and pull apart fractures leading to the separation of the fragments. In some phyllosilicate-rich layers, shear band structures (White *et al.* 1980, Platt & Vissers 1980) were noted.

All the quartz mylonites not exhibiting the later deformation fabrics (associated with the zones of 'rucking')

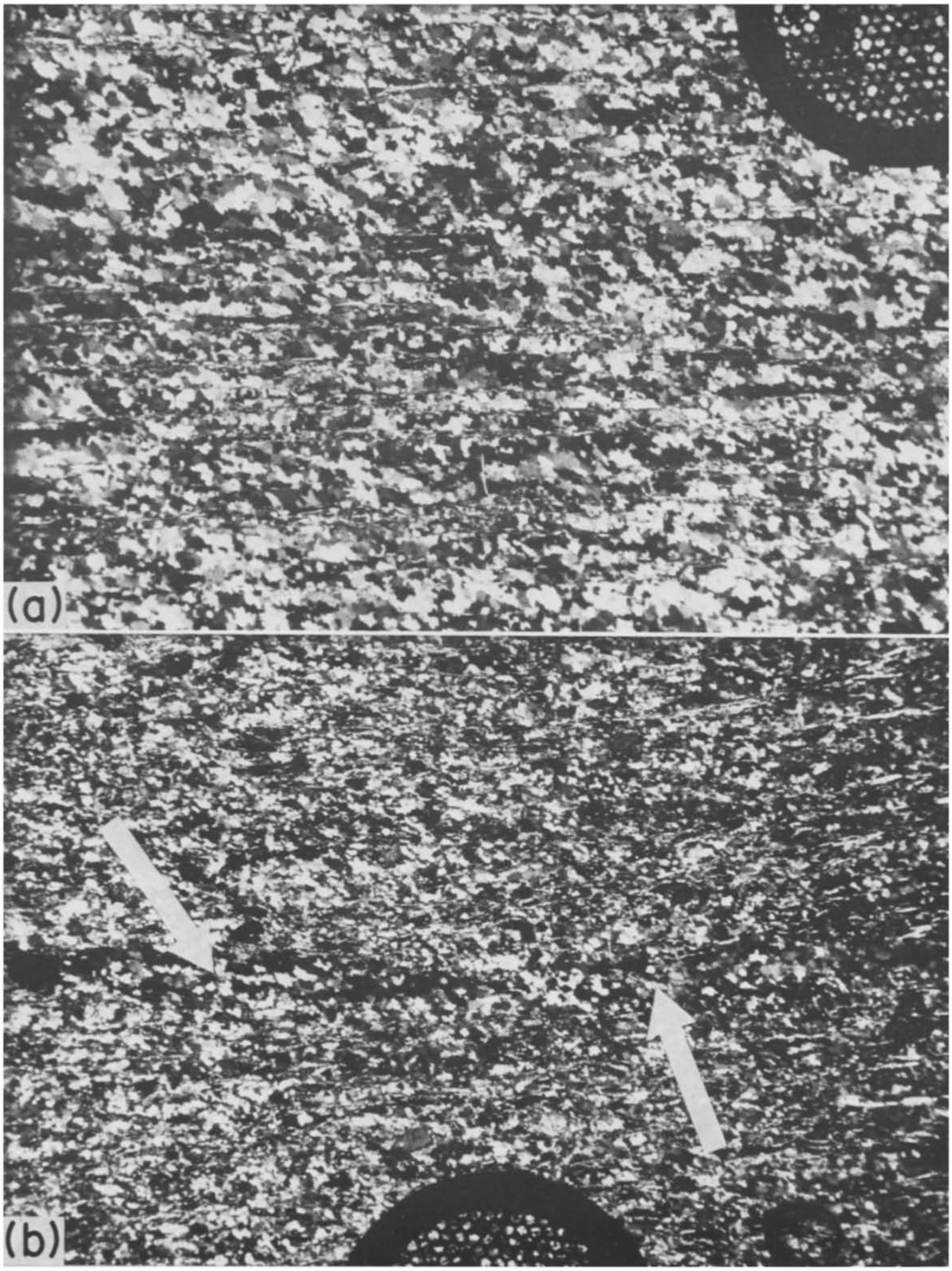

Fig. 4. Photomicrographs from quartz mylonites of the 'main mylonite belt' (a) General microstructures seen in the quartz mylonites, illustrating the completely recrystallized rock with irregular grain boundaries, strain shadows, inequidimensional asymmetric grains and areas of similar orientation, possibly representing former ribbons. (b) A mylonized quartz vein (arrowed) at a low angle to the mylonite foliation with identical microstructures to the host mylonite (see also Fig. 3, 131).

Scale note for micrographs in the figures

Unless otherwise stated, all photomicrographs show part of a copper grid 3mm in diameter for scale.

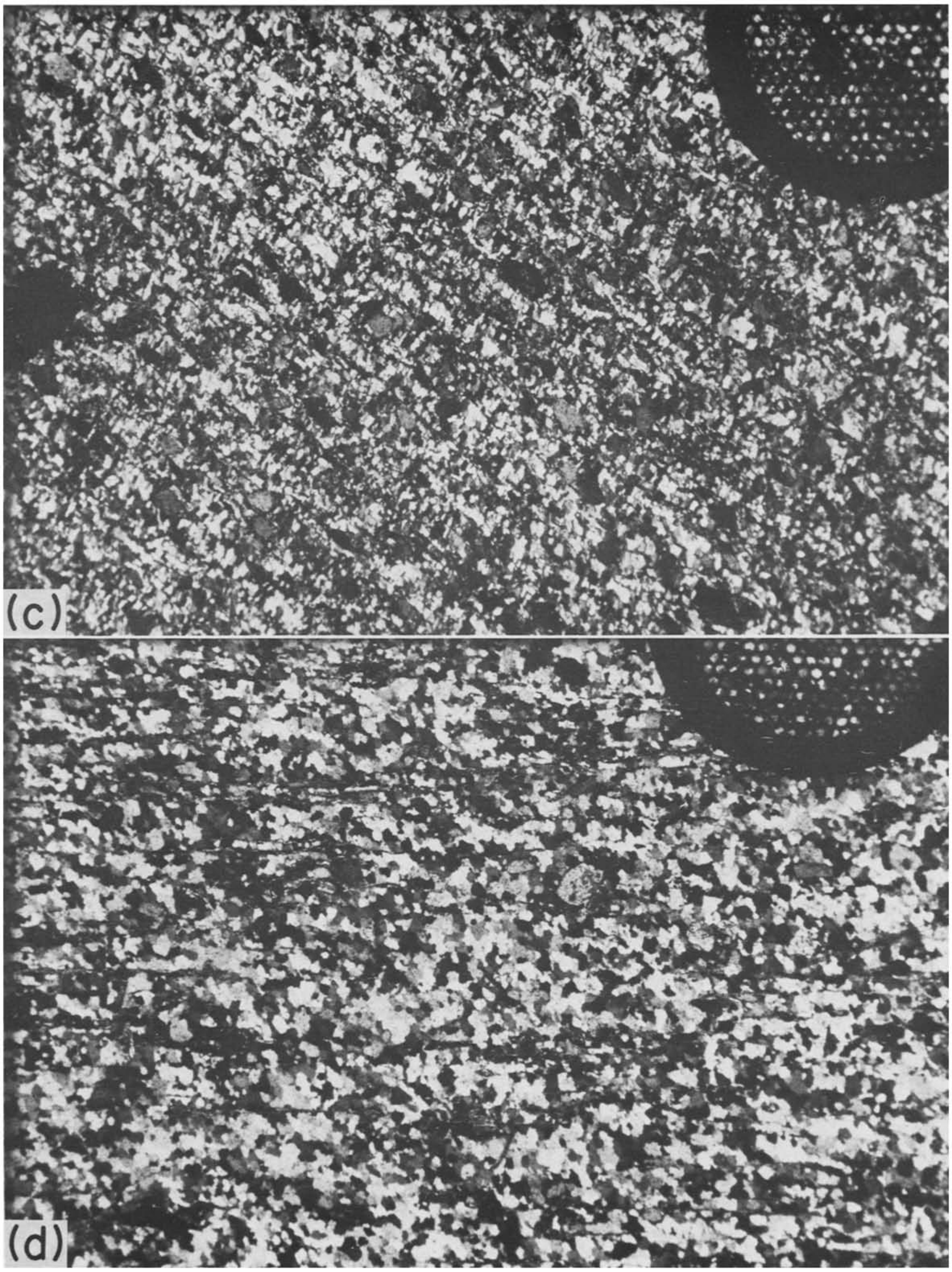

Fig. 4. (c) Strong axial planar grain preferred shape fabric (running from bottom left to top right) developed in areas of strong crenulations in the zones of rucked up mylonites (see Fig. 3, 39(ii)). The relict mylonite foliation runs from bottom right to top left. (d) Microstructures resembling those of the mylonites, from areas previously described as nonmylonitic Moine psammites (see Fig. 3, A10).

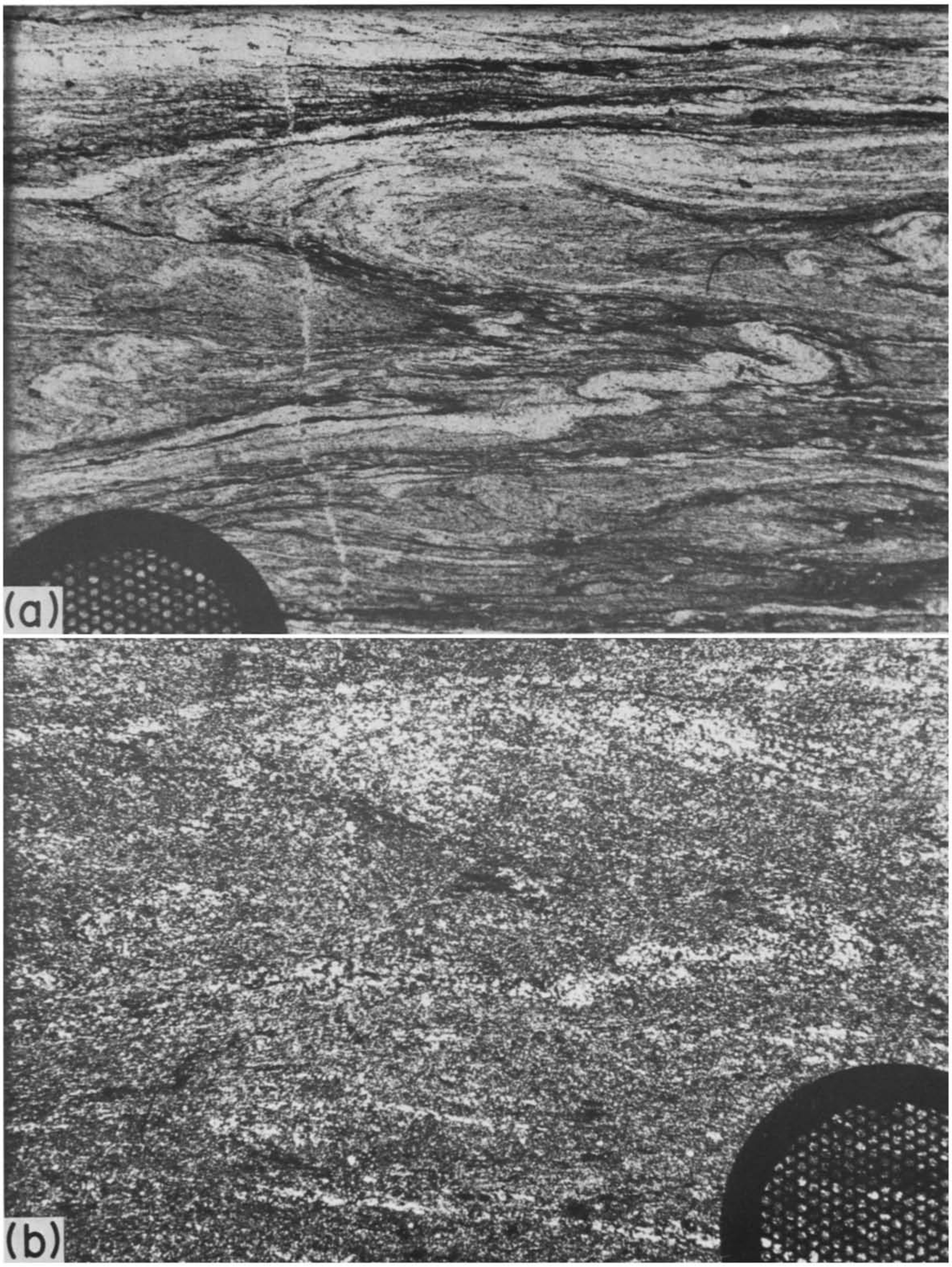

Fig. 5. Photomicrographs from the low-grade quartz mylonites at low structural levels in the Moine Nappe in Allt Odhsgaraidh (see Fig. 1, A012) showing: (a) & (b) the very fine grained nature, with intrafolial folds developed (PPL and XN, respectively).

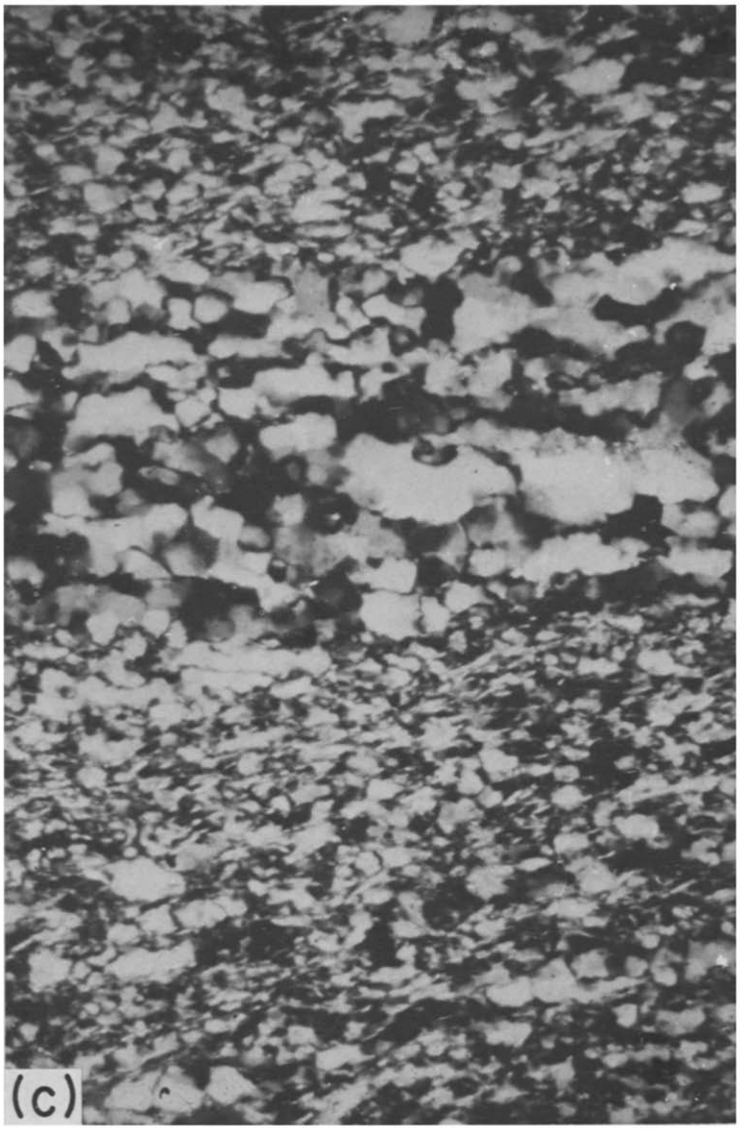

Fig. 5. (c) Higher magnification of the quartz bands illustrating the elongate quartz grain boundaries, parallel to the mylonite foliation.

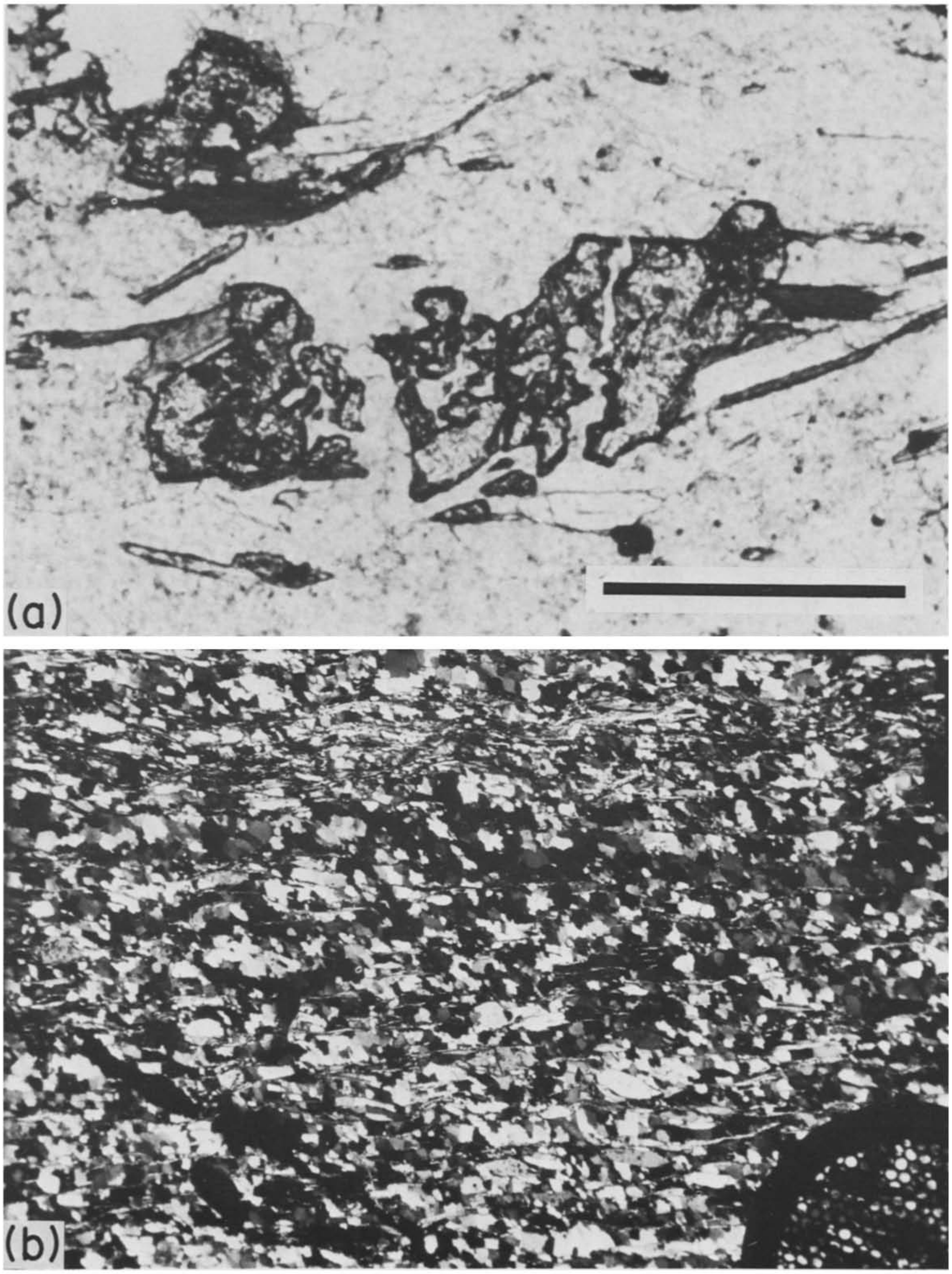

Fig. 6. Photomicrographs from the variably mylonitized, more obviously Moinian derived mylonites. (a) Fibre loaded garnet from within a quartz rich mylonitic rock (Fig. 3, 122(iii)). Scale bar 0.25 mm. (b) Recrystallized quartz layers exhibiting irregular grain boundaries and strain shadowing (see Fig. 3, 122(iii) & CR21(2) for fabric diagrams).

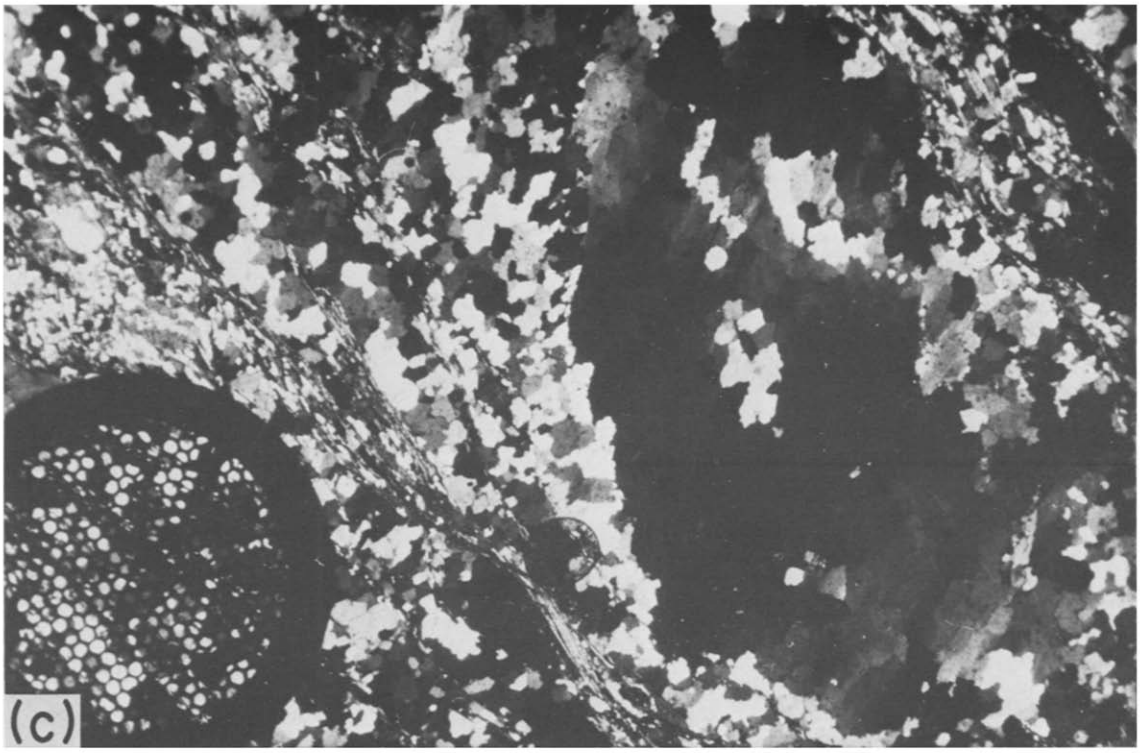

Fig. 6. (c) Core and mantle structure in a remnant quartz clast undergoing dynamic (syntectonic) recrystallization (Fig. 1. CR12(ii) region). Newly recrystallized grains sited along deformation band boundaries and areas of greater subgrain misorientations.

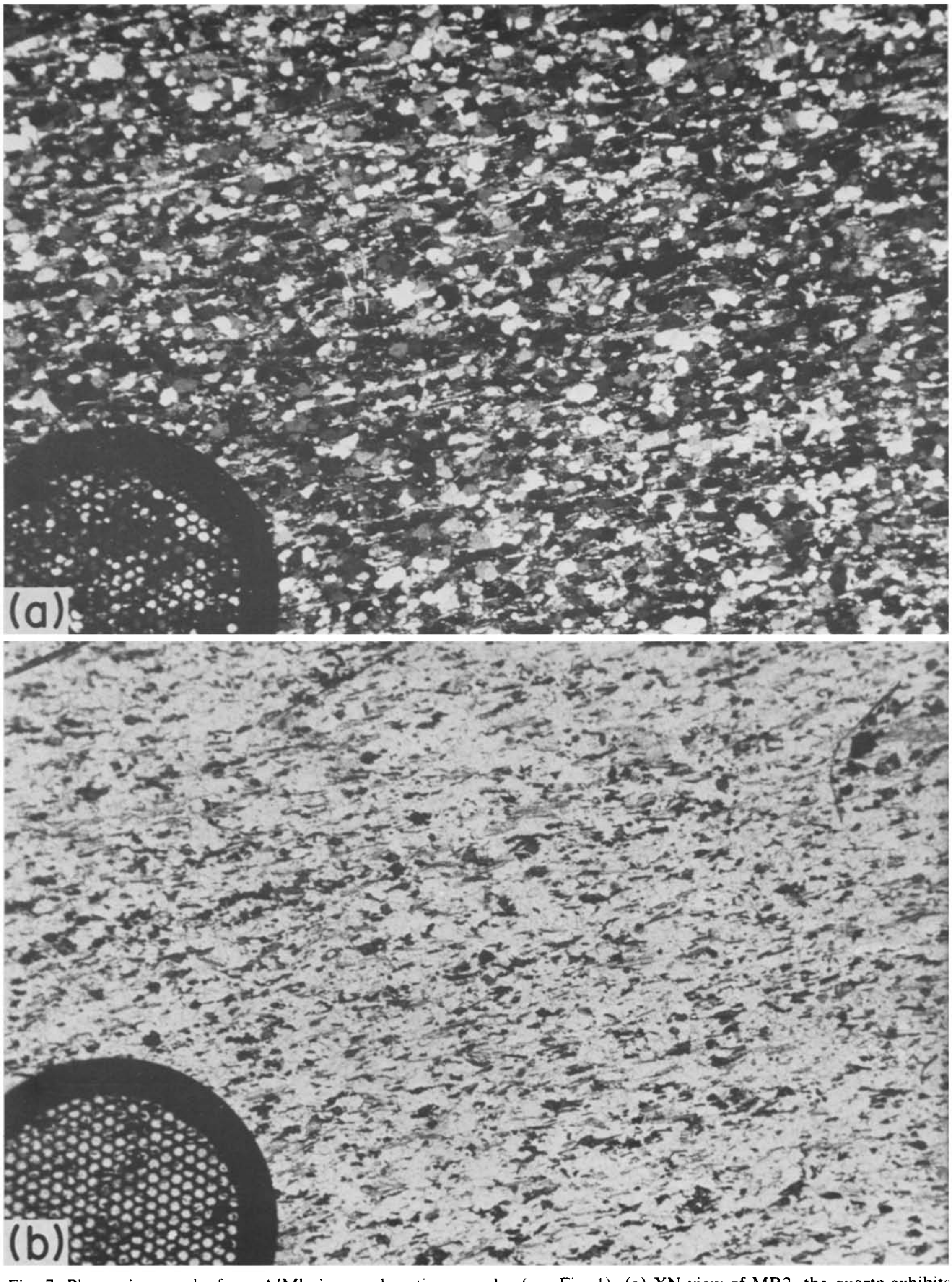

Fig. 7. Photomicrographs from A'Mhoine road section samples (see Fig. 1). (a) XN view of MR2, the quartz exhibits smooth grain boundaries, 120° triple junctions and very little strain shadowing. (b) PPL view of MR2, the most easterly micaceous psammite sample. The phyllosilicates are stubby in form and with the epidote and iron ore, show a uniform distribution throughout the rock.

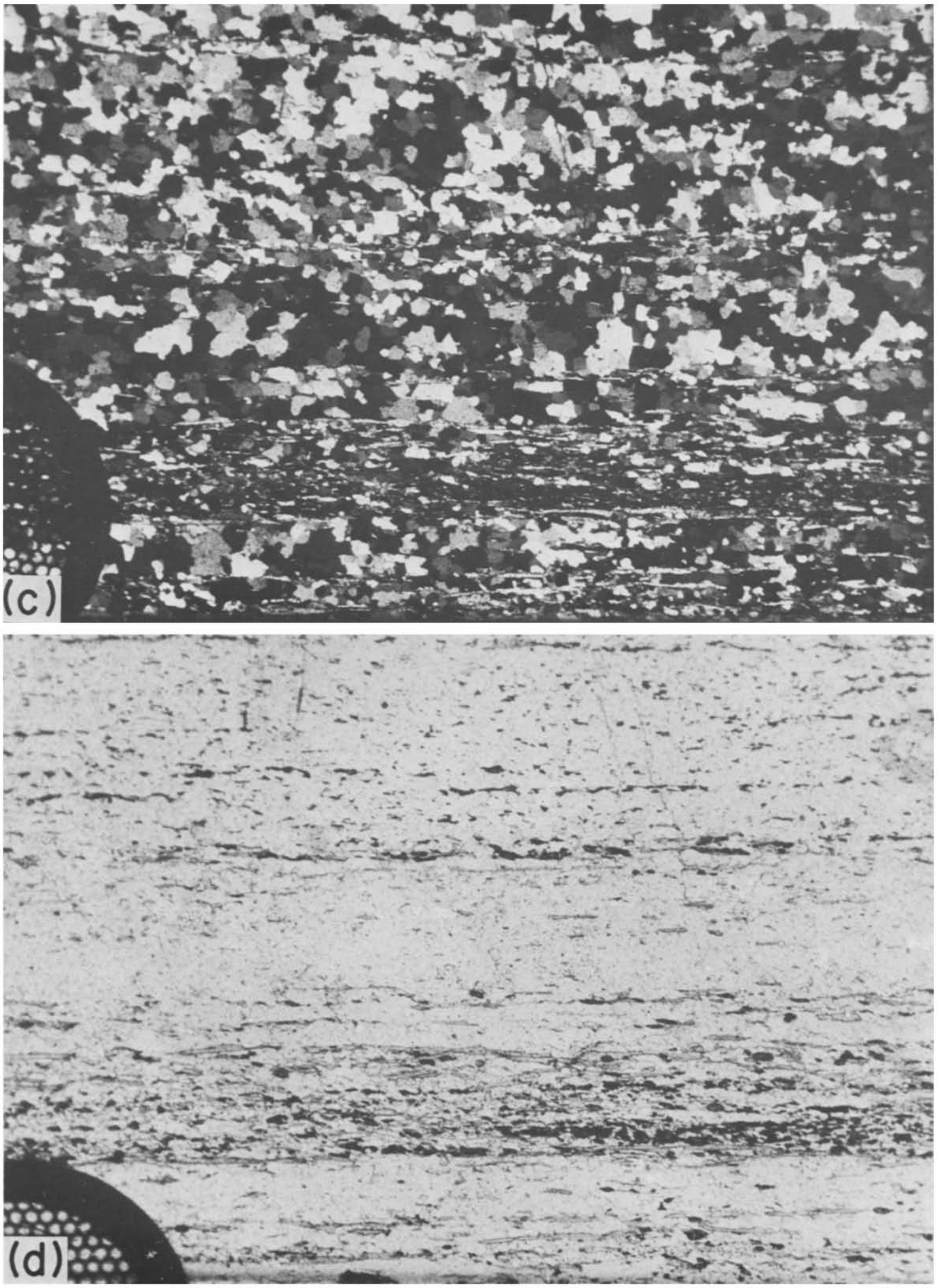

Fig. 7. (c) XN view of MR12(ii) illustrating the quartz microstructures including: complete recrystallization, smaller grain size in association with the micas and accessories, irregular grain boundaries, but with evidence of strain-driven grain boundary migration ('relaxation') as micas are enclosed by some grains. (d) PPL view of MR12(ii), illustrating the debris like trails of spindly white micas, epidotes and iron ore, within the strongly mylonitic flaggy rock.

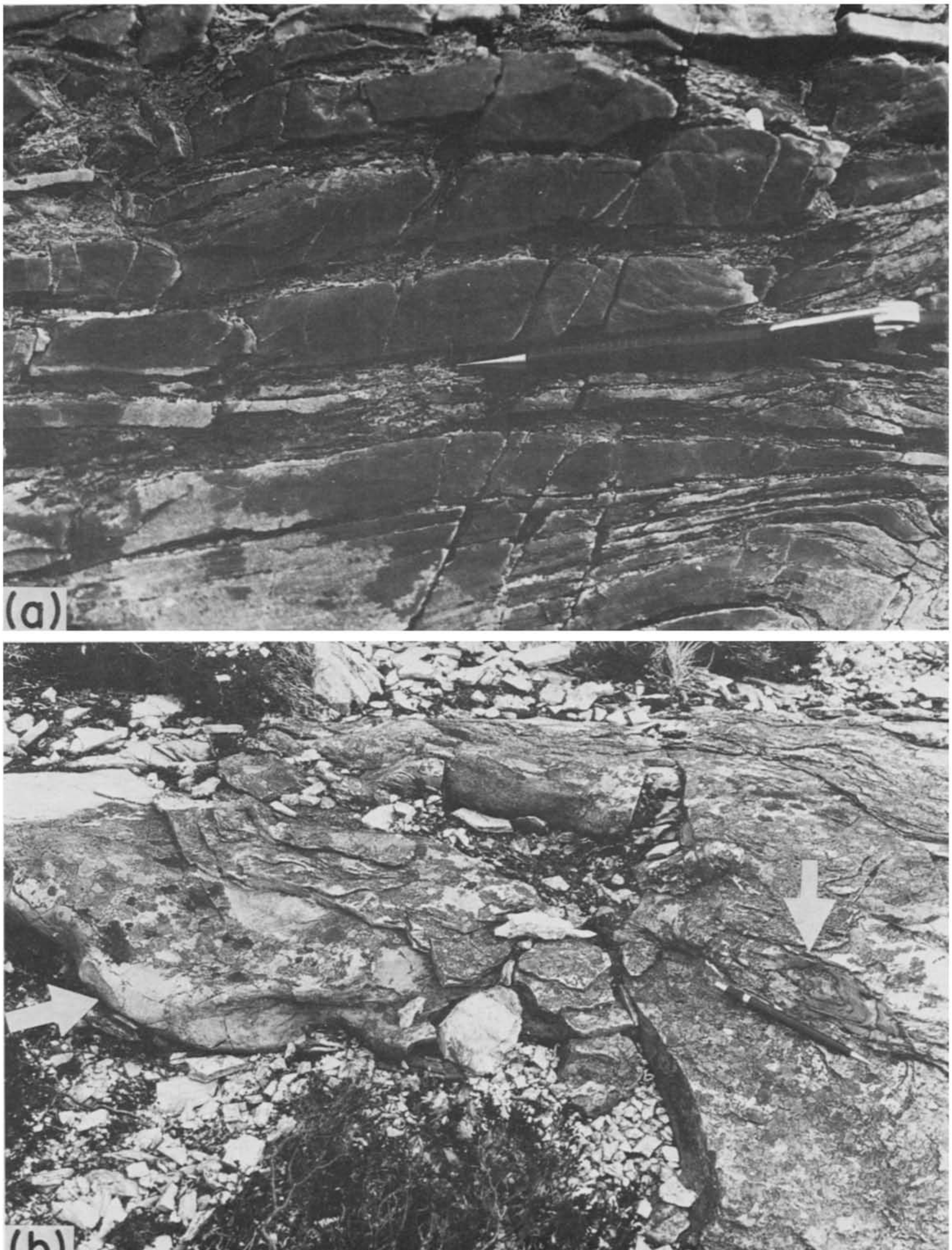

Fig. 9. Folds within the quartz mylonite above the Oystershell Rock. (a) A relatively early eye structure folding bands within the mylonite layering (above tip of pencil). (b) Non-cylindrical folding of the mylonite layers and a larger scale eyed structure close by (near the pencil). Pencil is at right angles to the mylonite extension lineation which has been folded. viewed ENE.

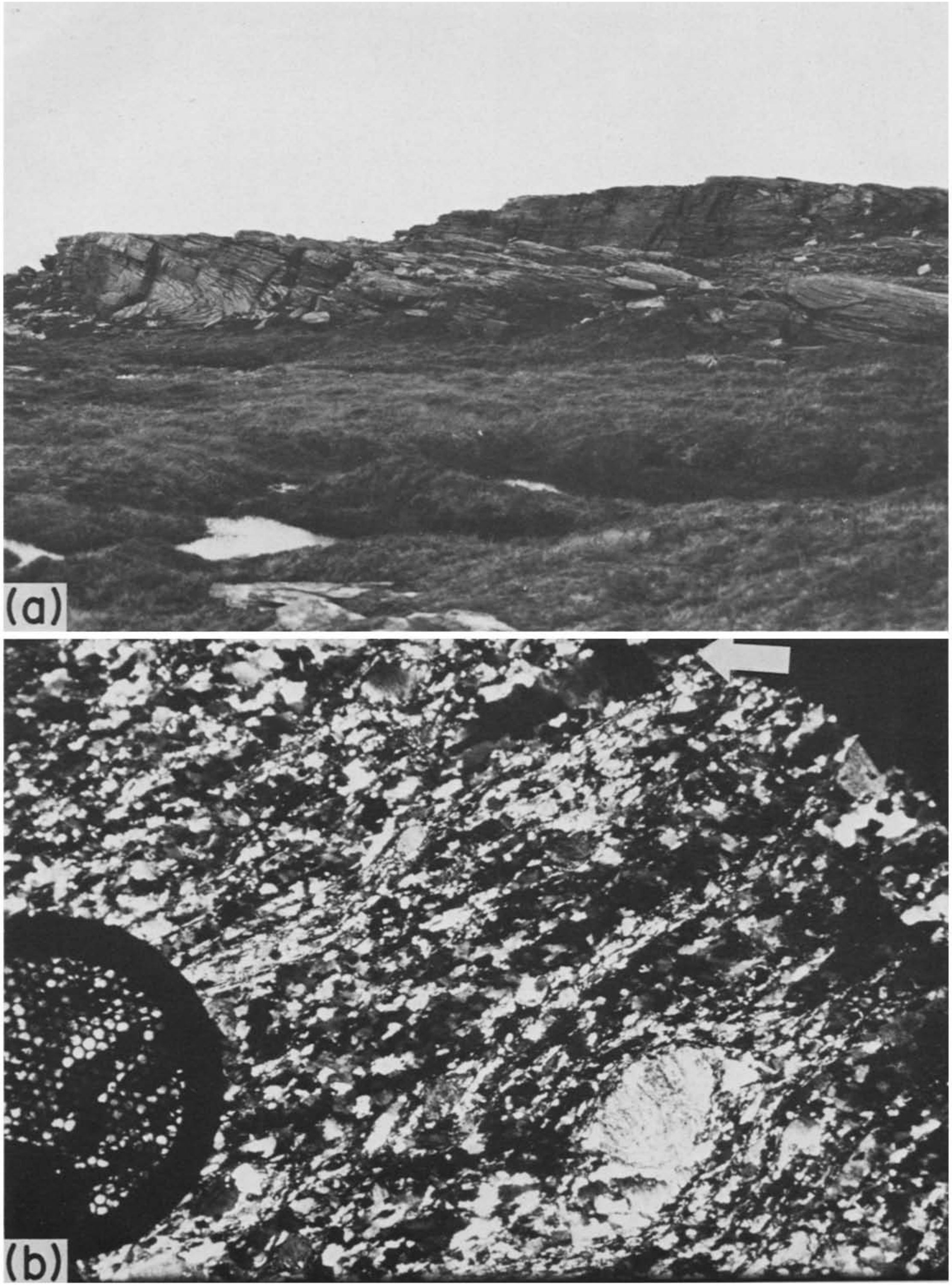

Fig. 12. (a) View of two large scale fold closures upon An-Lean-Charn (which plunge towards the ESE) separated by platy mylonitic rocks on the limbs, which in this case have obliterated the intervening closure. (looking NE, far cliff line is 4 m high). (b) Photomicrographs of the microstructures associated with the axial planar fabric and platy mylonitic fold limbs of the large-scale folds. Arrowed is a remnant quartz clast with deformation bands and undergoing syntectonic recrystallization.

sectioned to date, may be termed ultramylonites (Sibson 1977, White 1982). White (1982) stressed that the term should describe the amount of recrystallization and should not be a measure of, or dependent upon, the grain size. The quartz mylonites are in the main almost certainly a 'type r mylonite' (White 1982), i.e. the grain size reduction was by crystal plastic and neomineralization processes and not cataclastic processes accompanied by neomineralization.

Quartz veins are present in the mylonites and vary in appearance and form from early veins, commonly at low angles to the mylonite foliation and which are clearly mylonitized (Fig. 4b), to late white irregular veins often at high angles to the foliation and which may contain fragments of the host mylonite. These have suffered little, if any, deformation or mylonitization subsequent to their formation.

The only mylonite samples exhibiting different microstructures came from Allt Odhsgaraidh (see also White *et al.* 1982; NC 400515). These lower quartz mylonites have undergone a further grain size reduction and are juxtaposed upon cataclased and mylonitized Lewisian, and are thought to have been generated at higher crustal levels, possibly during imbrication and emplacement of the Moine Nappe as described by Butler (1982). These quartz mylonites are ultrafine grained, with grain sizes of 10–20 μm (Fig. 5) and they consist of alternating bands of white mica ± chlorite and quartz, with pressure solution like seams containing phyllosilicates and iron oxides. The finest grain sizes are found adjacent to the fault. Small-scale intrafolial folds (Figs. 5a & b) are common adjacent to the thrust contact, becoming less frequent away from the thrust. In quartz rich bands, the grain size is at its largest, with grain boundaries being elongate parallel to the mylonite foliation; the largest of these grains exhibit undulatory extinction and deformation bands (Fig. 5c). Feldspar clasts are small, fractured and pulled apart. The intensity of the extension lineation in these mylonites diminishes towards the thrust, i.e. where the grain size is smallest and where there has been maximum reworking. The microstructures and weak lineation of these very low grade greenschist facies mylonites suggest that a degree of superplasticity may have been attained.

Quartz c-axis fabrics

The quartz **c**-axis fabrics of the mylonite units are essentially similar and, as can be seen in Fig. 3, are an incomplete asymmetric type I crossed girdle (Lister 1977, Lister & Williams 1979, Lister & Hobbs 1980) with the sense of asymmetry being in the same sense as the mylonite grain preferred shapes and as the overall shear or thrust sense for the Moine Thrust Zone. However, there is often a slight displacement of the girdle from its position at the intersection of the Y-axis with the XZ plane. This displacement is apparently due primarily to the effects of late NNW–SSE zones of 'rucking' which locally exhibit strong crenulations and a variably developed grain preferred shape orientation associated with the development of axial planar fabrics (Fig. 4c). The **c**-axis fabric diagrams from such areas (Fig. 3, plots 19a & 39(ii)) show a weak new small circle maxima has been superimposed on the asymmetric type I girdle. The intensity of the small circle girdle increases with the strength of the axial planar foliation (Fig. 3, plots 39(ii) and Fig. 4c). Evidence for the continual recrystallization and reworking of the mylonites is found when the mylonitized quartz vein fabric is analysed. It is found (Fig. 3, plots QV and 131 and Fig. 4b) that they possess almost identical **c**-axis fabric patterns and textures to that of the host mylonite.

The ultrafine grained quartz mylonites exhibit a different fabric. They have a type II crossed girdle fabric (Fig. 1, plots A012 & AH1) which is more akin to the low temperature fabrics obtained from quartz bands in the Lewisian mylonites above the Arnaboll Thrust on Ben Arnaboll (see White *et al.* 1982).

VARIABLE MYLONITIC AND PLATY MOINES

As will be shown in the following section, the degree of mylonitization and its extent in rocks of more obvious Moinian parentage is far greater than reported by Soper & Wilkinson (1975) and includes areas mapped by them as unmylonitized and weakly mylonitized Moines.

The boundary between the mylonite belt and the mylonitized Moines is marked and transitional, i.e. a ductile structure, with no breccia or gouge noted. There is however, a marked change from grey and white banded and yellowish quartz mylonites to off-white, yellow and orange coloured, often feldspathic, mylonites. Thin discontinuous units, very similar to the green feldspathic phyllonites within the mylonite belt below, are closely associated with this boundary (e.g. NC 434532 and 423534).

Immediately south of the mylonite belt, the rocks show strong mylonitization with a well-developed quartz extension lineation parallel to that of the mylonites below. Mineralogically they contain the same assemblages and exhibit the same grain sizes as the quartz mylonites beneath.

Strong mylonitization continues to the south but there are now areas of lower strain in megascopically psammitic rock (e.g. NC 429527) which consist of recrystallized quartz, feldspar (both perthitic alkali and albitic plagioclase feldspars) spindly white micas ± chlorite, biotite, epidote, garnet and iron oxides. Feldspar clasts are often large (up to 2 mm) and there is generally a deflection of the fabric around them. Many of these, plus the garnet clasts show fractures (both shear and tensile fractures) and the separation of the fragments (Fig. 6a). These features were noted as far south as Cashel Dhu. The prevalence of psammitic schist horizons increases during the southwards traverse and these contain recognizable pelitic horizons with biotite and garnet. Some garnets reveal weak s-shaped inclusion trails and some a rather random inclusion fabric; both types exhibit chlorite beards.

Recrystallized quartz dominates the mylonitic Moines (Fig. 6b) and the more psammitic horizons. They have serrate boundaries, internal strain features and some inclusions of micas within larger grains. Relict quartz clasts are found exhibiting subgrains and a well-developed core and mantle structure (White 1976) (Fig. 6c). The internal strain features in the recrystallized quartz grains and the very irregular grain boundaries become less pronounced southwards. In general, the size of the recrystallized grains also tends to increase towards the south, though at any point, finer grained bands can occur. The above sequence reveals there has been active grain size reduction during dynamic (syntectonic) recrystallization and that the weaker internal strain features, less serrate grain boundaries and mica inclusions represent strain-driven grain boundary migration and recovery during the later, passive stages of an uplifted deep level shear zone while temperatures were still elevated.

The quartz c-axis fabrics (Figs. 1 & 3) show a remarkable and consistent, though more diffuse, similarity to those of the underlying quartz mylonites. Fabrics (Fig. 1) from the lower cliff line of Ben Hope are again similar to those just described and they represent levels approximately 1–1½ km above the thrust zone. The accessible areas of the precipitous 600 m western escarpment of Ben Hope have also been sampled and the area represents the highest structural levels within the Moine Nappe studied in this project. Siliceous psammites, often with a strong platy or flaggy appearance, form the greater thickness of the upper levels of the escarpment, within which thin flaggy highly deformed regions are common. Cross-bedding is preserved in the less platy rocks which exhibit a lineation plunging towards 105°. The grain sizes are 150–200 μm and the quartz grains exhibit internal strain features. Some c-axis fabrics are a little more symmetrical (Fig. 8, H1), but with a distinct asymmetric distribution, whilst others (Fig. 1, plots HR8), from beneath the level of slide a, but in a platy or flaggy rock, show a well developed asymmetric fabric.

The first indication of a different fabric and apparently lower strain microstructures are found just west of the Kyle of Tongue. Quartz bands are again recrystallized but now the grains have 120° triple junctions, very few internal strain features, smoother grain boundaries (Fig. 7a), and exhibit a more symmetrical diffuse c-axis girdle (Fig. 1, MR2 & 6). Biotite is common within these rocks but is partially altered to chlorite. The micas in general exhibit a rather stubby form and, along with epidote and iron oxides, are evenly distributed throughout the rock (Fig. 7b). The assemblages are indicative of upper greenschist facies conditions.

Two marked platy slide zones occur within the Moine outcrop. One is east of A'Mhoine house and the other some 450 m up Ben Hope (Fig. 1 & 8). The quartz microstructures in the first (Fig. 7c) are identical to those of previously described mylonitic flaggy/platy Moines, and the c-axis fabric is a strong, well-developed asymmetric type I girdle (Fig. 1, MR̃12). Very little biotite is found and any present is mainly altered to

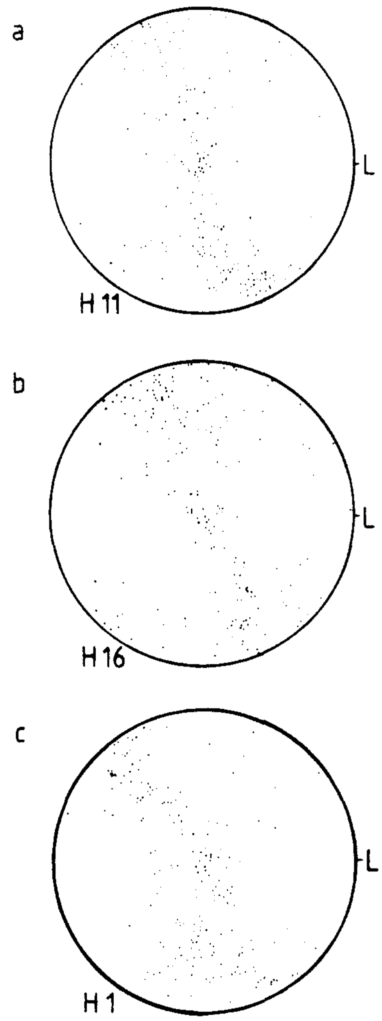

Fig. 8. c-axes fabrics for specimens collected on the western escarpment of Ben Hope. (a) Specimen H11 (NC 477511), thin flaggy fine grained siliceous Moines. (b) Specimen H16 (NC 474507), as above and in which skeletal and rotated garnets are present. (c) Specimen H1 (NC 476502), Moine psammite in an area exhibiting cross bedding, from just below the summit of Bĕn Hope.

chlorite. The epidotes and iron oxides form debris like trails (Fig. 7d). Grain sizes of the quartz associated with the phyllosilicates is smaller than in quartz rich layers. In the coarser grained less phyllosilicate rich layers, grains may enclose muscovites which may also pin quartz grain boundaries. This plus the weak optical strain features indicate that there has been some strain-driven boundary migration after the development of the slide zone. It represents a higher strain and retrogressive version of the previously described Moinian rock.

The second platy (slide) zone (the Ben Hope slide) occurs in the western escarpment of Ben Hope. There is a lineation plunging towards 105° and it differs from the first in that it contains a lens which is believed to be a Lewisian inlier (in addition to the Ben Hope Sill) associated with it (to be dealt with elsewhere). This inlier has, to the authors' knowledge, not been described before, the possibility only being referred to by Johnstone (pers. comm.) in Peacock (1975). Thin sections from the fine grained thin flaggy units (slide rocks) and the cross-bedded psammitic rocks from just below the summit, reveal totally recrystallized quartz bands. Within the slide rocks, muscovite-rich layers appear, with small amounts of biotite altering to chlorite. These micaceous layers exhibit weak shear band structures which indicate a northwestwards directed shear. Quartz microstructures are similar to the first slide zone, grain sizes being in the order 80–120 μm compared to 150–200 μm in the more massive psammitic material.

The quartz c-axis fabrics again exhibit a strong asymmetric type I girdle (Figs. 8a & b). The asymmetry and the shear band orientation indicate shear towards the northwest.

FOLD STRUCTURES

Two types of fold structures were distinguished, those outside the main mylonite belt, i.e. in the schists, but with which platy mylonitic rock generation is associated, and those within the mylonites. The latter will be described first.

Folds associated with flow within the mylonites

A series of folds are closely associated with mylonite generation and shearing and are best developed in the quartz mylonite above the chlorite phyllonite (Oystershell Rock) in the region north of Am Feur Loch (NC 438552). Evidence for the sequential development of these folds is shown by eye structures and tight, appressed folds folding the compositional banding within the mylonite layers (Figs. 9a and 10a) and which are themselves, in turn refolded (Fig. 10b). Many folds have the morphology of sheath folds (Carreras *et al.* 1977). Detailed measurements in the ground between NC 438552 and 441563 reveal a complete half girdle distribution of fold plunge directions about an axis containing the extension lineation (Fig. 11). The fold morphologies and orientations indicate a cyclic pattern of fold generation in which fold attitudes vary according to their stage of evolution (see Fig. 11). Folds at an early stage of development are transversely oriented, being close to or at right angles to the extension direction (lineation). They exhibit an overturning sense to the WNW, with gently dipping upper limbs, steep to overturned lower limbs and a variable state of noncylindricity, which is again related to their stage of evolution (Fig. 9b).

The folds plunging parallel or approximately parallel to the lineation vary from isoclinal intrafolial types, resembling Christie's (1963) 'streaked out folds' and Soper & Wilkinson's (1975) D1 type folds, through to tight, inclined to recumbent folds, in both 's' and 'z' configurations. They fold the mylonite layering which contains the earlier eye and intrafolial folds (see Figs. 10c & d). However, possibly not all the folds at an angle to the lineation were generated at a high angle. Some may have arisen at a low angle and were passively rotated around into the extension direction and did not develop 'noses'. In these cases, it will be virtually impos-

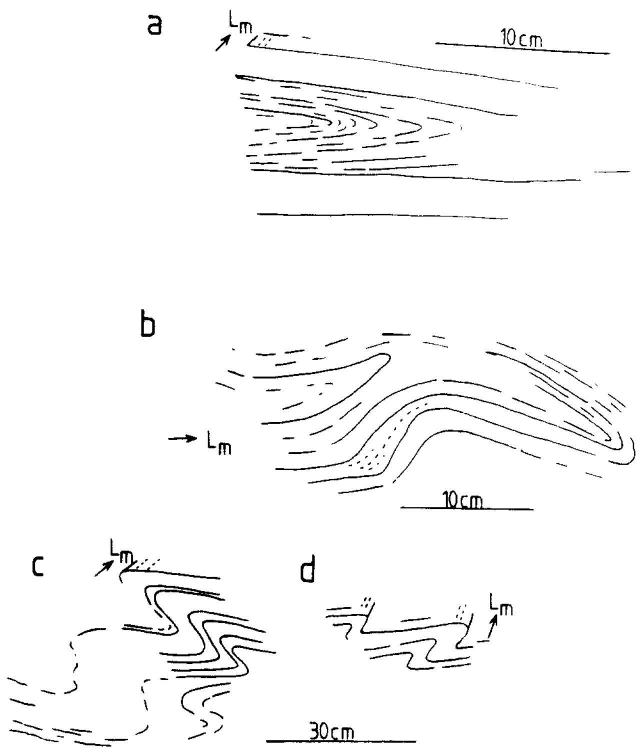

Fig. 10. Sketches of the minor fold structures observed in the quartz mylonites. (a) Isoclinal folding of the compositional banding within the mylonite layers. (b) Cyclicity of folds represented by the refolding of earlier isoclinal folds of the mylonite layering by more open, transverse fold. (c) & (d) 's' and 'z' fold configurations. Folds plunge ESE parallel to the extension lineation and are interpreted as fully rotated 'end member' types.

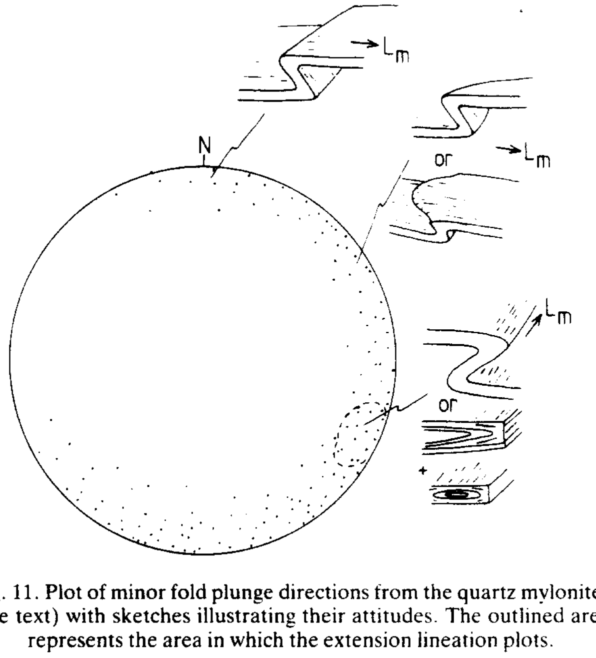

Fig. 11. Plot of minor fold plunge directions from the quartz mylonites (see text) with sketches illustrating their attitudes. The outlined area represents the area in which the extension lineation plots.

sible to differentiate them from sheath fold structures which have no visible 'noses'. This has been noted for similar fold sets elsewhere (e.g. La Tour 1981).

It can be seen that a continual sequential fold development is identifiable from intrafolial fold structures to folds at varying angles to the extension lineation through to later stage open, transverse folds, refolding earlier tight to isoclinal rotated folds representing higher shear strains. On this basis it was not possible to separate Soper & Wilkinson's (1975) D1 intrafolial folds from their D2 ESE plunging folds.

Late major, generally southward plunging monoclinal structures, which created the ridges of 'rucked up' mylonites refold the earlier sheath and rotated folds. They are associated with open flexures and cause a marked crenulation in phyllosilicate rich horizons.

The last fold structures in the mylonite belt are kinks and boxfolds (see also Soper & Wilkinson 1975) with brittle hinge lines (with quartz and haematite as a common crack infill) and a variable orientation and development, dependent upon the lithologies. They are concentrated in fissile, thinly layered rocks.

Folds in the schists

Large recumbent folds, plunging towards 115° occur on the summit of An Lean Charn and lower hillocks to the east. They have a very strong axial planar fabric, with flaggy/platy mylonitic rocks developed upon the fold limbs (Fig. 12a) and which when well developed, may obliterate some fold closures. Cross-bedding is preserved in some fold cores. The platy foliation exhibits an orientation close to that of the mylonite foliation lying structurally below and a lineation parallel to that in the mylonites. Microstructures from the core regions (Fig. 12b) show large deformed remnant quartz clasts (ribbons) which recrystallized to produce an initial quartz preferred grain shape related to the deformation bands within the host clasts. Smaller recrystallized grains are seen as grain size reduction continues, i.e. on the platy limbs and in higher strain regions. The platiness of the rock increases, with an accompanying decrease in mineral grain size, until flaggy mylonites, with no trace of the large folds, are formed.

Mendum (1976, 1979) has reported similar flaggy to platy mylonitic slide rocks in areas of the north Sutherland Moines where large-scale folds are developed. A rotation or reorientation of these folds into the 'X-direction' cannot be ruled out.

Fabric diagrams (Fig. 3, plots AC8(ii), AC10) for quartz from these flaggy mylonitic rocks reveal lopsided asymmetric girdles. The lopsided nature of the fabric is apparently related to the overprinting recrystallization not fully obliterating the previous fabric, i.e. there is an influence of a relict fabric upon the subsequent orientation of new, recrystallized grains as shown by Ransom (1971).

DISCUSSION

Extent of mylonites in the Moine Nappe

The field mapping and associated quartz microstructural and crystallographic fabric studies indicate that the mylonites in the Moine Nappe are much more extensive than previously suggested. The consistency of fabrics and quartz microstructures indicates that effects of mylonitization extend to at least 8 km from the thrust belt (see Figs. 1, 3 and 8). This seems to correlate with MacGregor's (1952) 'abnormal' Moines fabric which he described as being related to thrusting.

Mylonitization and deformation within the Moine Nappe at Eriboll

The mylonites studied can be divided, on the basis of mineralogy, microstructures, and quartz c-axis fabrics, into three types: two deeper-level mylonites having similar type I quartz c-axis girdles which are distinguished from each other by the presence of muscovite and chlorite ± biotite and quartz microstructures indicative of recovery; as against mylonites with muscovite ± chlorite and higher level low-grade mylonites adjacent to thrust structures at the base of the belt, and which display a type II girdle and ultrafine grained quartz. These arise from a later, higher level ductile reworking of earlier fabrics.

The origin of certain of the biotite mylonites can be seen upon An Lean Charn. A platy mylonitic foliation develops in the limb regions of folds and axial planar to large-scale recumbent isoclinal folds. Fold cores are preserved and contain cross-bedded psammitic material, but the folds plunge parallel to the lineation and may well have been rotated into their present position; they cannot confidently be used in establishing a deformation framework.

On Ben Hope, in addition to the Ben Hope Sill, highly deformed amphibolitic and gneissic material has been interpreted as a Lewisian inlier. It is bounded by thin flaggy or platy Moinian-derived mylonitic rock. The c-axis fabrics and shear band structures indicate similar deformation conditions were operative as for the mylonites to the west. Thus, it seems that a substantial component of shear was associated with the emplacement of the Lewisian inlier, suggesting it represents the allochthonous Lewisian inlier type of Moorhouse & Moorhouse (1977) and Rathbone & Harris (1979). At present, there is no basis on which to differentiate the fabrics from more westerly mylonites and the slide rocks and it seems reasonable to associate the emplacement of the Lewisian along a slide with the very earliest (Caledonian) mylonitization events at deep levels in a slide zone. During the evolution of the nappe the deeper-level (i.e. the more easterly) slide zones became inactive, and in mylonites associated with these zones, quartz grains underwent strain induced grain boundary relaxation.

All the mylonitic rocks developed by ductile deformation mechanisms, there is no evidence of extensive cataclasis followed, by static thermal annealing. However, the presence of quartz veins in the higher level main mylonite belt and in the later, low-grade reworked mylonites attest to periods of high stress, brittle deformation. The varying states of mylonitization of these veins suggests that episodic brittle events occurred during the ductile deformation and add weight to the argument that mylonitization was spread over a longer period than has previously been suggested.

Quartz c-axis fabrics

An asymmetric type I fabric about the lineation and foliation was found in the specimens studied. The sense of asymmetry of the fabrics is consistent with the sense of inferred thrusting (see also White *et al.* 1982). A similar relationship has been recorded in several mylonite belts (Bouchez 1977, Burg & Laurent 1978, Bouchez & Pecher 1981, Behrman & Platt 1982). Although several workers (Carreras *et al.* 1977, Lister & Price 1978, Simpson 1980, Schmid *et al.* 1981, Passchier 1983) have shown that the sense of asymmetry may not always be consistent with the sense of shearing, no inconsistencies were found in the Moine Nappe.

Some sections show a displacement of the girdle in a direction, normal to the lineation within the foliation plane. It was found that the displacement increased and a new small circle girdle appeared nearer to the zones of 'rucking' and it is concluded that the late rucking has led to the fabric modification, but not to the obliteration of the existing fabric skeleton. Modelling experiments (Lister & Williams 1979) show that asymmetric fabrics are stable and will remain as a skeletal form to late stages of subsequent deformation, whilst Phillips (1945) and Simpson (1980) also report that an existing fabric has lasting effects upon subsequent fabrics in naturally deformed samples. This will be especially so when the deformation conditions do not induce major changes in the operative quartz slip systems. The effect of such changes is seen in the reworked mylonites adjacent to the thrust structures at the base of the nappe. These samples exhibit a type II girdle and most likely reflect a change from dominant prism **a** and basal **a** slip with subsidiary rhomb slip in the muscovite ± chlorite and biotite and muscovite ± chlorite mylonites to dominant basal **a** slip with subsidiary rhomb and prism slip. This change is consistent with a change in the temperature during deformation.

Folds within the mylonites

Folds are common within the main belt of greenschist facies mylonites and most post-date the onset of mylonitization and mark the time when the mylonites were the active deforming medium. Their occurrence in the lowest grade greenschist facies mylonites is thought to be further evidence for a longer deformation history involving the reworking of the upper level mylonites. Sheath folds are present. As in other descriptions of shear zones (Sanderson 1973, Escher & Watterson 1974, Rhodes & Gayer 1976, Bell 1978, Williams 1978), fold axes are not found exclusively normal or parallel to the lineation. It is suggested that the minor folds seen in the mylonites were generated and rotated during and by continued dominantly simple shear deformation and evolved in a cyclic fashion.

Regional implications

Within the zone studied there is evidence of platy mylonitic rocks generated axial planar to, and on the limbs of, isoclinal folds within the Moine schists. It would appear therefore, that there was substantial deformation prior to mylonitization. The highest grade mylonites encountered during the study were those of biotite grade. The preservation of megascopically psammitic rocks in the biotite grade regions suggests that, in the area studied, the mylonites first developed at this grade and were reworked at chlorite–muscovite grade and then parts of these were in turn reworked at lower temperatures adjacent to the thrust zone. The main mylonite belt greenschist facies (chlorite grade) mylonites underwent a phase of intrafolial fold development followed by rucking. The lowest grade mylonites underwent a further phase of intrafolial folding, at which time the chlorite grade mylonites had probably become passive. Late brittle folds affected all chlorite and lower grade mylonites.

The only mylonites in the Moine Nappe that can be unequivocally associated with post-Cambrian movements are the low temperature mylonites adjacent to the thrust zone. The chlorite and biotite mylonites with associated minor folds reflect earlier deformation phases. Soper and coworkers postulated a purely Caledonian (post Cambrian) deformation for the Moine Nappe in this region eastwards, to at least the Naver Slide. This was based upon the correlation of small-scale folds within the mylonites with large-scale folds, from which mylonites were being generated, in the western part of the Moine Nappe. These large-scale folds may have suffered rotation into their present position. The deformation framework was then tied into the regional metamorphism whereby textures resulted from a recrystallization during an MS2–MP2 metamorphism. Hence, the whole of the nappe across to the Naver Slide and the migmatite complex was believed to have had a purely post-Cambrian deformation and metamorphism because the structures used in the correlation affected Cambrian rocks in the thrust zone. Such structures cannot be unequivocally used to establish a deformation framework, and although an early Caledonian age for the large-scale folds in the Moine Nappe is favoured, it is possible that some are Precambrian structures which have suffered deformation and reorientation during the Caledonian.

To summarize, the deformation sequence in the Moine Nappe may be: folding of the Moine schists and psammites with platy biotite grade mylonite zones (slide zones) developing upon the fold limbs and incorporating slices of Lewisian in the nappe along early slide zones in the east; reworking of the biotite grade mylonites under mid-upper chlorite grade conditions and the subsequent development of folds within these; reworking of the western mid-upper chlorite grade mylonites under low grade chlorite conditions adjacent to the thrust zone. Only the latter mylonites which form along the Moine Thrust can be safely equated with a Caledonian deformation. It is concluded that the Moine Thrust (s.s.) is located in an older slide zone.

The conclusion is that although the Soper & Brown (1971) and Soper & Wilkinson (1975) deformation sequences have been useful in the Sutherland region, they are too restrictive and based upon tenuous structural correlations. We suggest that the structures within the mylonite and thrust zones should not be correlated with similarly oriented structures outside the zone, as deformation concentrates within mylonite zones often to the exclusion of outside areas (Watterson 1975, White *et al.* 1980). It is also concluded that microstructural, fabric and field evidence neither unequivocally supports nor disproves an exclusive post-Cambrian, Caledonian history for deformation and metamorphism in the Moine Nappe in the Eriboll region. Precambrian deformation proposed by Moorhouse & Moorhouse (1979a, b) to the east cannot be ruled out and, therefore, there may be no differences in the tectonic history of the northern and southern Moines.

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