" A structural and metamorphic study of Moine rocks between Loch Eil and Loch Eilt, Inverness-shire."

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ABSTRACT OF THESIS.

" A structural and metamorphic study of Moine rocks between Loch Eil and Loch Eilt, Inverness-shire."

Research between Loch Eil and Loch Eilt, Inverness-shire on rocks from all three Divisions of the Moine Succession within the Caledonian Orogen, above and to the east of the Moine thrust zone, has revealed that, within the area, all three Divisions share a common sequence of polyphase deformation and metamorphism. All three Divisions have been metamorphosed during Precambrian and Caledonian tectono-thermal events.

The Ardgour granitic gneiss, a ≈ 1000 Ma. pre-D_1 intrusion, sub-parallel but transgressing the junction between the central Glenfinnan and eastern Loch Eil Divisions, has undergone partial anatexis during D_1 , unlike the adjacent migmatitic metasediments. Major F_1 folds have not been recognised.

Grenvillian (≈ 1000 Ma.) D_2 deformation and high grade metamorphism has almost completely transposed the D_1 granitic gneiss foliation but there is no unequivocal evidence for D_2 partial anatexis. F_2 major folds occur in the vicinity of the granitic gneiss but not elsewhere.

Caledonian D_3 deformation has produced major F_3 sub-recumbent folds with curvilinear hinge lines, within the Glenfinnan and Loch Eil Divisions, above a flat lying major D_3 simple shear zone, the Sgurr Beag slide. Large scale WNW transport has carried these rocks over the underlying Morar Division rocks. Only at the highest levels of strain within the shear zone has the Grenville metamorphic assemblage re-equilibrated. Elsewhere the early isograds are folded during D_3.

Post-D_3 microdiorite sheet intrusions have suffered medium grade (amphibolite facies) metamorphism during D_4 deformation. D_4 heterogeneous, horizontally directed pure shear deformation, possibly above a flat lying decollement plane such as the Moine Thrust, has produced major upright folds of the D_3 Sgurr Beag slide and rotated Glenfinnan Division F_3 sub-recumbent isoclines towards vertical, to form the D_4 "steep belt".

D_5 deformation has produced sporadically distributed minor folds possibly related to regional warps of the earlier major structures.

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CHAPTER 1.

Introduction.

1.1) Geographical setting.

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1.1) Geographical setting.

The area under investigation is centred on Glenfinnan village and covers an area of approximately 65 sq. km. (25 sq. miles). It stretches for 20 km. in an E-W direction along the A 830 road from Fort William to Mallaig and extends northwards 3 to 4 km. from the road (Fig.1).

The area rises from sea level at Loch Eilt to over 800 metres and is physically rugged and mountainous. It is dissected by a number of N-S trending glens which divide the area into a series of hills all over 600 metres high and bounded by deep glens or lochs on three sides (Map 8). The area is all within the old county of Inverness-shire.

1.2) Regional geological setting.

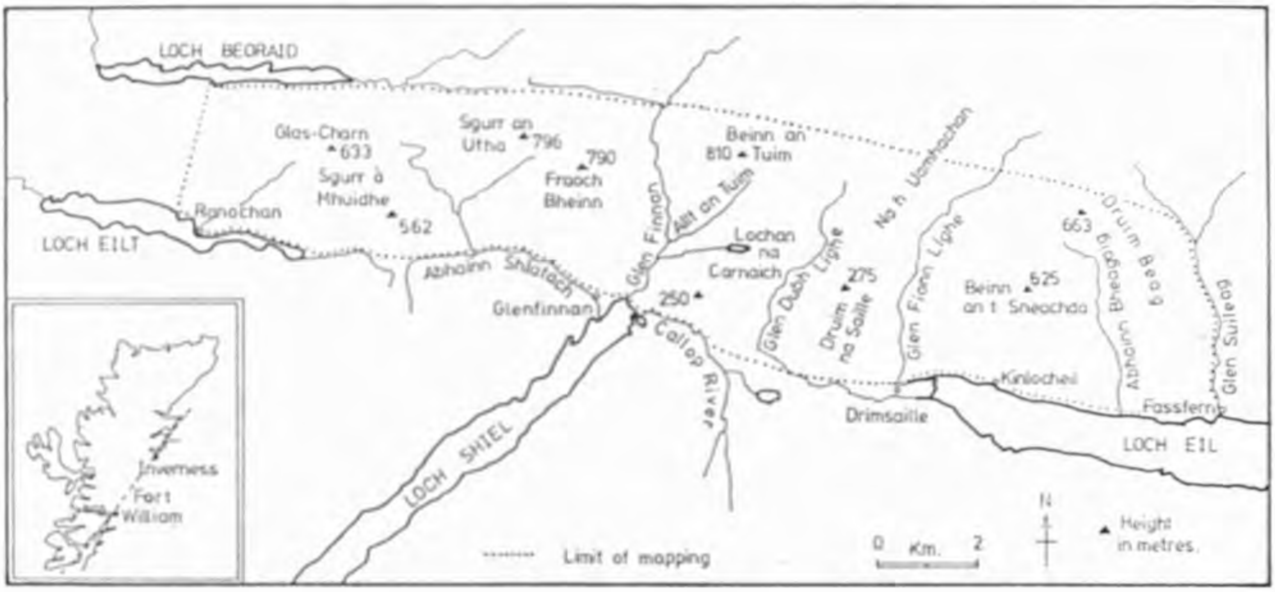
The area lies within the Northern Highlands of Scotland, that is to the NW of the Great Glen fault (Fig.2). The eastern end of the area is 9 km. W of the Great Glen fault zone. The Moine Thrust zone is 30 km. NW of the western end of the area and probably extends eastwards to underlie the area at depth (see Chapter 9).

The Moine rocks of SW Inverness-shire occur in three N-S trending structural - stratigraphical units termed, from W to E, the Morar, Glenfinnan and Loch Eil Divisions (Johnstone et al. 1969. see Fig 2). East of the Great Glen fault lithologically similar rocks, generally termed the Central Highland Granulites, have been divided into the Central Highland Division and the Grampian Group (Piasecki, 1980). No direct structural correlations have been made across the Great Glen fault. A variety of sinistral and dextral displacements on the fault have been proposed (eg. Kennedy, 1946; Winchester, 1974; Phillips et al.,1976; Van der Voo & Scotese, 1981; Estang & Piper, 1984).

The Moine rocks or " Moine nappe " has been carried westwards over the stable foreland of Lewisian Archaean-Proterozoic basement gneisses, unconformably overlain by Torridonian Proterozoic sediments which are in turn unconformably overlain by Cambro-Ordovician shelf sediments. The junction between the stable cratonic foreland and the Moine rocks of the Caledonian orogen is marked by the complex imbricate Moine Thrust zone.

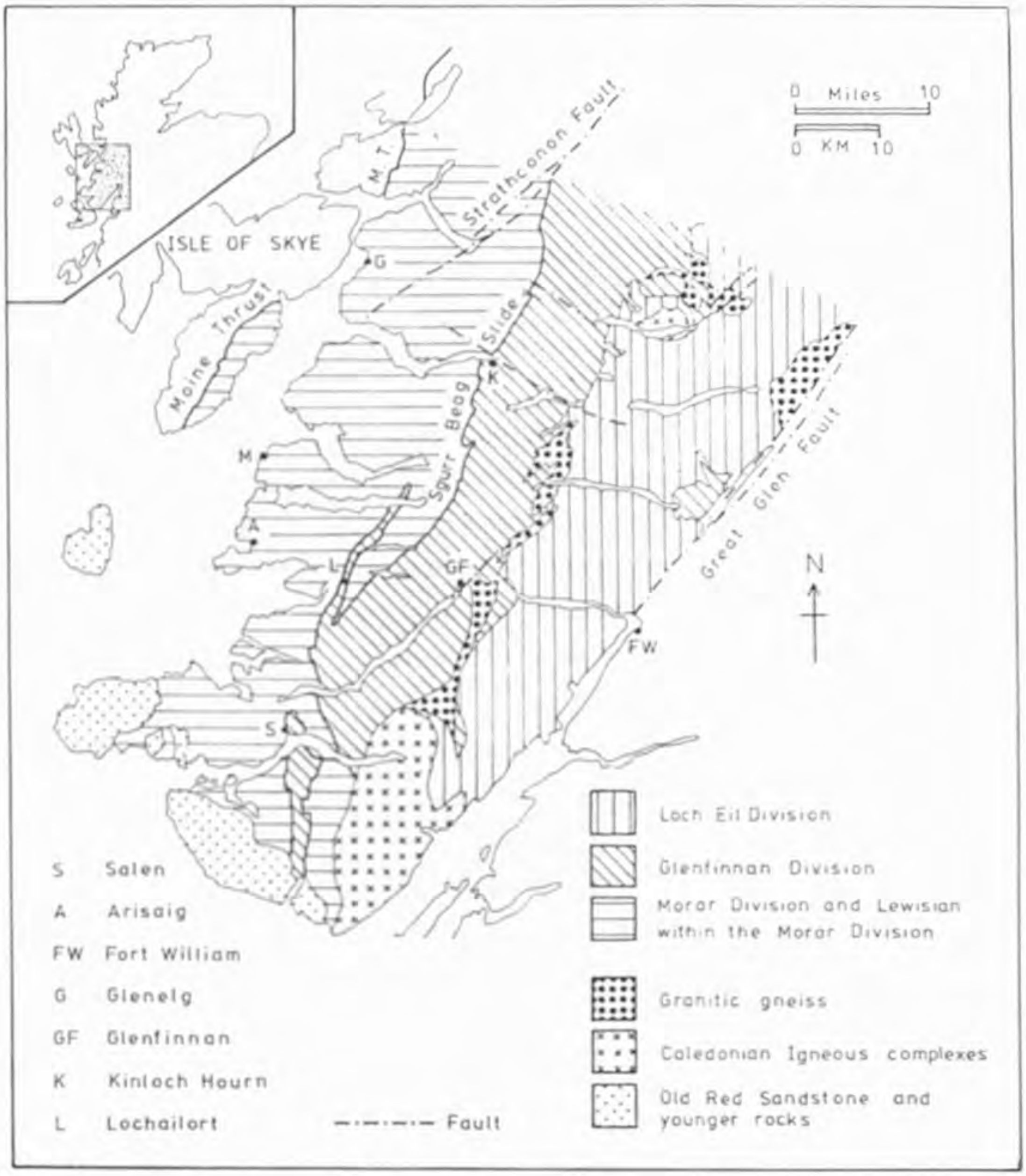
A stratigraphy has been established for the Morar Division (Powell,1974) which places the Lochailort pelite at the top of the Division, stratigraphically in contact with, and above the Upper Psammite of the Morar Division.

Figure 1.



Map to show the extent of the area.

Figure 2.



Map to show the major geological divisions of the southwestern Moine rocks.

The Lochailort Pelite is very similar to rocks of the Glenfinnan Division (Johnstone et al.,1969,p.163). However in the Kinloch Hourn area the junction between the Upper Psammite and the Lochailort Pelite is held to be tectonic, termed the Sgurr Beag slide (Tanner,1971).

East of this junction the Glenfinnan Division comprises a series of striped and banded psammitic and pelitic rocks which are characterised by their general steep disposition and the migmatitic segregations often seen in the pelitic and banded lithologies.

A series of migmatitic granitic gneisses outcrop in the E of the Glenfinnan Division near the contact with the Loch Eil Division. The migmatitic granitic gneiss in the area under investigation is called the Ardgour granitic gneiss (Harry,1953).

The junction between the Glenfinnan and Loch Eil Divisions has been called the "Loch Quoich line" (Clifford,1957). Dalziel (1966) interpreted it as a stratigraphical junction, with the Loch Eil Division psammites being younger than the Glenfinnan Division, whereas Piasecki & van Breemen (1979) imply that the junction is a tectonised unconformity between the older highly deformed basement, the Glenfinnan Division, and a cover sequence, the Loch Eil Division.

Previous researchers have erected structural sequences for various parts of the Western Highlands. In general these tend to be remarkably similar, commencing with an early phase of small scale isoclines (F_1), followed by a phase of tight to isoclinal small and large scale folds (F_2) associated with a very strong penetrative axial planar fabric (S_2) in pelitic and semi-pelitic rocks. The third phase of deformation is a strong crenulation (S_3) and folding (F_3) of the pre-existing fabrics.This is followed by various later open crenulations and warps.

A suite of late syn- to post-metamorphic intrusions, microdiorites, cut all three Divisions in the area under study.

The metamorphic grade rises from the W coast where it is lower greenschist facies (Kennedy,1949) to upper amphibolite facies in the Glenfinnan area (Johnstone et al.,1969). The increase in metamorphic grade is reflected in the rocks by a coarsening of grain size and the development of migmatitic segregations. Metamorphic grade is difficult to establish in the Loch Eil Division, a group of dominantly psammitic rocks.

It is generally believed that the main metamorphism is coeval with the second phase of deformation (D_2)(Powell,1974). The metamorphic pattern is complicated by retrogression and folding of the isograd surfaces by later deformation events.

Correlations between various areas and tectono-metamorphic interpretations associated with geochronological evidence have often proved equivocal.

Former correlations of the top of the Moine succession with the base of the Dalradian succession have led to the suggestion that all the deformation is Caledonian in age (Kennedy,1955). More recent geochronological work (Giletti et al.,1961; Long & Lambert,1963; Brook et al.,1976,1977) has produced a number of isochrons and mineral ages of Precambrian age indicating that at least part of the deformation and metamorphism occured in the late Precambrian.

1.3) Objects of the investigation.

The investigation was initiated in this area for a number of reasons. The area offers a short section across all three Divisions of the Moine rocks of the NW Highlands. Much of the neighbouring district had already been studied in detail by Howkins (1961), Clark (1961), Dalziel (1963a), Powell (1963) and Brown (1964) (see Fig.3). All parts of the area were reasonably accessible.

Attempts at structural correlations (Brown et al.1970; Powell,1974) between neighbouring regions had highlighted several problems and geochronological work by Long & Lambert (1963), van Breemen et al.(1974) and Brook et al.(1976) only seemed to complicate these.

It was hoped that mapping the area from the head of Loch Eil westwards to Loch Eilt could shed light on the nature of the junctions between the Morar, Glenfinnan and Loch Eil Divisions. More specifically,if the junctions turned out to be tectonic rather than stratigraphical it was hoped that it would be possible to establish the structural age of the junctions with respect to the Divisions on either side of each junction. Thus, armed with a knowledge of the polyphase deformation sequences within each Division and their inter-relationship, in conjunction with geochronological evidence it was hoped that it would become possible to differentiate between Precambrian and Caledonian structural elements.

The study therefore presents an initial survey of some of the area, togather with a re-examination and re-interpretation of the work done in other parts of the area.

Figure 3.

Map to show the areas covered by
previous pieces of research.

1.4) Methods of Investigation.

Basic geological mapping was carried out on a scale of 1:10,000 on parts of the published Ordnance Survey sheets numbers NM88SW : NM88SE : NM98SE NN08SW : NM97NW : NM97NE : NN07NW.

Aerial photographs were also used occasionally but these were not found to be particularly useful.

Measurements of junctions, bedding, schistosities, foliations, minor fold axial planes, minor fold hinge lines, minor fold vergence and fault planes were plotted directly onto the field slips in the field. Other observations of fold style, metamorphic condition, sample localities etc. were logged in field notebooks.

All measurements of the orientation of planar and linear elements were made using a Silva 15 TD CL compass. Difficulties were encountered in the measurement of the plunge direction of linear elements on steeply dipping planes. These were overcome using the techniques of stereographic projection.

1.5) Degree of exposure.

With the exception of the floor of Glen Finnan and Drimsallie forest the area is very well exposed. Exposures are rarely more than 50 m. apart and in the west there are areas where the exposure is almost total.

Unfortunately Drimsallie forest obscures some of the critical junctions of the Ardgour granitic gneiss,and a newly planted strip of forest west of Glenfinnan village will make access onto the hills of Sgurr an Utha and Fraoch Bheinn more difficult in the future.

C H A P T E R 2.

Previous stratigraphical and structural research.

2.1) The Morar Division.

This section records the development of ideas concerning the stratigraphy of the Morar Division and the relationship of the Morar Division to the Glenfinnan Division further east. Some detail of the structures involved is included

In the early 1930's officers of the Geological survey were engaged in mapping parts of Sheet 52 (Sunart) and Sheet 61 (Mallaig) (Geological Survey of Great Britain [Scotland] 1" to 1 mile). Reports of their findings (Richey et al.1930's) can be found in the Geological Survey of Great Britain 1931 - 1939. Summary of Progress, (West Highland Division).

The first record of a recognisable stratigraphical succession within the psammitic and pelitic rocks in the Morar area is reported in the summary of progress 1935, p.78. In 1936 (p.74) this sequence was termed the Morar Succession and consisted of:

(Youngest) 3) Upper Psammitic Group.

2) Striped and Pelitic Group.

(Oldest) 3) Lower Psammitic Group.

This succession was reported as forming the outer envelope af a N-S major anticline, the core of which is occupied by "much crumpled and folded sedimentary and igneous gneisses, distinct from the Morar Succession". In 1938 (p.71) Dr. Kennedy reported that the rocks of the lower (core) series contained overfolding not present in the upper (envelope) series and he proposed that the lower series be called "the Sub-Moine Series of Morar". The upper stratigraphical series was termed "the Moine Series of Morar" and both series were included in the term "The Morar Succession".

In 1938 (p.71 & 72) Dr. Simpson extended the mapping eastwards, recording a psammitic unit (the Ardnish Psammite) which he correlated with the Upper Psammitic Group of Morar. He noted that it passed eastwards into a belt of striped and pelitic schists (the Lochailort Pelite).

The first geological sketch map of the Morar area was published by Richey & Kennedy (1939). It showed the regional extent of the Moine Series and the outcrop of the Sub-Moine Series in the core of the major Morar anticline (see Fig.4). The Moine Series was considered as having suffered little or no destructive deformation and containing abundant sedimentary structures whereas the Sub-Moine Series was recognised as a highly folded complex of paragneisses and orthogneisses which retain very few of their original sedimentary structures.

These two series "can usually be separated without difficulty both in the field and under the microscope"(p.27). Richey & Kennedy concluded that Sub-Moine rocks belong to a series older than the Moine schists for the following reasons.

1) There are recumbent folds in the Sub-Moine Series of the core which are not found in the Moine Series of the envelope. It is argued that these folds are recumbent anticlines which close to the east (cf. Kennedy,1955).

2) Intense crumpling is characteristic of the Sub-Moine Series but not of the adjacent Moine Series.

3) The Moine and Sub-Moine rocks are in different metamorphic states.

4) At a number of places along the western limb of the Morar anticline the Lower Psammitic Group at the base of the Moine Series transgresses various members of the Sub-Moine Series. It is stated (p.43) that there is no evidence for tectonic sliding along this junction.

Whilst trying to prepare the maps and memoirs on the One Inch Sheets 61 & 52 MacGregor (1948) expressed doubts about some of the conclusions reached by Richey & Kennedy (1939). Both he and Mr V.A.Eyles, mapping in the Loch nan Uamh area could find no mappable line to define the base of the Moine Series. He examined the criteria used by Richey & Kennedy to distinguish between the Sub-Moine and Moine Series' and found that none of them was diagnostic of either of the series. His observations are summarised below:

1) False-bedding can be found within units of the Sub-Moine Series and

Figure 4.

Map to show the Moine-Sub Moine Series.
(after Richey & Kennedy,1939).

the Moine Series.

2) Although calc-silicate ribs are not found in the Lower Psammitic Group and the lower part of the Striped and Pelitic Group, "calc-silicates of Arnipol type" are found in both the Sub-Moine Series and the Lower Psammitic Group and the Lower Striped schists of the Moine Series.

3) The proportion of mica increases gradually outwards from the base of the Sub-Moine Series.

4) Minor folds of the same style are found in both the Moine and Sub-Moine Series.

5) Oblique foliations can be found in both the Sub-Moine and the Moine Series.

6) The sutured quartz-mosaic textures of the rocks in the Sub-Moine Series decrease gradually on passing southeastwards out of the Sub-Moine Series.

He concluded that there is a possibility that all the metamorphic " Sub-Moine sediments of Morar constitute a lower part of the Moine Series associated with hornblende orthogneisses".

The description by MacGregor of a stratigraphical passage between the Sub-Moine Series and the Moine Series caused Kennedy to re-examine the area. The results of his examination were published in 1955. Kennedy concluded that the lower part of the Moine Succession had been re-duplicated within the core of the Morar anticline. The envelope of the Moine Series rocks (the Morar nappe) had been driven westwards over a broken autochthonous basement. The stratigraphy of the Sub-Moine Series was given as:

(Youngest) 3) Pelitic Group.

2) Psammitic Group.

(Oldest) 1) Hornblende Group (Lewisian).

The Psammitic and Pelitic Groups of the core were correlated with the Lower Psammitic Group and the lower part of the Striped and Pelitic Group of the Moine Succession respectively.

The structure of the core was described as a broken recumbent anticline overturned and closing to the west (cf. Richey & Kennedy,1939). This was ascribed to westward thrusting of the envelope relative to the core.

In a comparison of the structures in Morar with those further west in the Caledonian foreland area of Skye he correlated the recumbent folding of the crystalline basement together with the overlying Torridonian and Cambro-Ordovician sediments of Skye with the recumbent folding and nappe formation in Morar. This phase of deformation had, therefore, to be post-Arenig in age. Kennedy correlated this phase of deformation with the mid-Ordovician phase of deformation seen elsewhere in the SE Highlands of Scotland. He noted that it was pre-metamorphic whereas the Moine thrust is post-metamorphic and thus subsequently assumed to be a late Silurian structure.

Thus it is seen that the nature of the Moine - Sub-Moine boundary was highly equivocal, Richey & Kennedy (1939) had concluded that it was an unconformable junction, MacGregor (1948) stated that it could be stratigraphical and Kennedy (1955) concluded that it was a tectonic junction.

Lambert (1958) postulated that the Moine schists of Morar had been affected by progressive and subsequent retrogressive metamorphisms. However, only the core schists were considered to be affected, on a major scale, by this retrogression and the outer limit of re-working was regarded as the core - envelope boundary which was described as transgressive to lithological boundaries.

In 1964 Powell, considering the area at the SE end of the Morar anticline eastwards through the Ardnish, Lochailort and Loch Eilt environs found that the outermost unit of the Sub-Moine Series, which he termed the Beasdale Pelitic Group, passed upwards stratigraphically through the Lower Psammitic Group of the Moine Series. He gave these Groups local names, the Loch nan Uamh Pelitic Group and the Ardnish Psammitic Group respectively (see Fig.5 taken from Powell,1964)

To the east of the Ardnish Psammitic Group he identified a unit, the Lochailort Pelitic Group, which he argued was younger than the Ardnish Psammitic Group and therefore formed a fifth group in the stratigraphical succession.

He identified a fold, the Glenshian synform, within the Lochailort Pelitic Group which he believed repeated the western succession to the east of its core. Passing eastwards through the Lochailort Pelitic Group, the Arieniskill Psammitic Group was correlated with the Ardnish Psammitic Group. Further east the Arieniskill Pelitic Group was correlated with the Loch Mama Pelitic Group and in the extreme of his area he correlated the Loch Eilt Psammitic Group with the Loch nan Uamh Psammitic Group (see Fig.5).

From the Lochailort Pelitic Group eastwards the rocks are in the "Moine Injection Belt".

In 1966 Powell published the first detailed evidence of the polyphase nature of the deformation in the SE Morar - Lochailort area. He stated that the Moine schists had suffered at least four phases of folding; folds of the first phase being extremely difficult to recognise. He reported major second phase and fourth phase folds with NE-SW trending axial planar traces. Major folds of a third phase were found to be restricted to the west of the area and represented by open warps with NW-SE trending axial

Figure 5.

WESTERN ASSEMBLAGE (stratigraphical succession)	EASTERN ASSEMBLAGE
LOCHAILORT PELITIC GROUP	LOCHAILORT PELITIC GROUP
ARDNISH PSAMMITIC GROUP	ARIENISKILL PSAMMITIC GROUP
LOCH MAMA PELITIC GROUP	ARIENISKILL PELITIC GROUP
LOCH NAN UAMH PSAMMITIC GROUP	LOCH EILT PSAMMITIC GROUP
BEASDALE PELITIC GROUP	

Table to show the stratigraphy of the Lochailort area.
(taken from Powell,1964. table 2).

planar traces. The temporal relationship between the recorded third phase and fourth phase is difficult to establish and Powell (pers. comm.) would now interpret the third phase as the fourth phase and vice versa in this F_1 to F_4 sequence.

Johnstone et al.(1969) divided the Moinian assemblage north of the Great Glen fault into three major divisions. The oldest rocks in the west were now termed the Morar Division, succeeded eastwards by the Glenfinnan Division, in turn succeeded eastwards by the Loch Eil Division.

Whilst accepting that the Lochailort Pelitic Group is at a higher stratigraphical level than the Ardnish Psammite (Upper Morar Psammitic Group), Johnstone et at. stated that "the junction is marked at least locally, by a slide (p.161)". The Lochailort Pelitic Group was grouped within the Glenfinnan Division with which it shares certain lithological characteristics.

Working in the Kinloch Hourn area of NE Knoydart, Tanner (1797) established two successions, a Runival succession which he correlated with part of the Morar Division, and a Kinloch Hourn succession which he correlated with the Glenfinnan Division. He noted that the junction between these two successions was in places occupied by small pips of Lewisianoid rocks and that the Runival succession "younged" upwards into this junction. Consequently he argued the junction must be tectonic in origin and he termed it the Sgurr Beag slide. Tanner noted that this junction can be traced 16 km. southwestwards towards Loch Eilt.

As cataclastic textures were not found anywhere along the slide zone he concluded that movement occurred when the ductility of both basement and cover had been considerably enhanced. Structural considerations led him to conclude that the slide movement occurred during the second folding episode although he noted the presence of two separate pre-F_2 folding episodes.(The structural age of the Sgurr Beag slide is considered in more detail in Chapter 7).

Working on allochthonous inliers of Lewisian within the Moine succession Rathbone & Harris (1979), used strain indicators to establish qualitative strain states adjacent to the Sgurr Beag slide in several areas. They reported that high strain is indicated by loss of sedimentary structures in psammites, an increase in the proportion of platy psammites, the tightening of minor folds, a re-orientation of quartz veins and a reduction of grain size.

In Glen Shiel and Garve in northern Inverness-shire the presence of Lewisian inliers along the slide permit the inference that the high strain zones are also tectonic slides. In the Lochailort area they demonstrated high strain along the junction of the Ardnish Psammitic Group and the Lochailort Pelitic Group but as Lewisian pods are not present along this junction they made the assumption that the junction is a tectonic break (or slide) as well as a zone of high strain.

2.2) The Ardgour granitic gneiss.

In this section the development of ideas concerning the nature, origin, stratigraphical position, extent, structure and metamorphism of the Ardgour granitic gneiss will be examined. The absolute age of the Ardgour granitic gneiss will be considered in Chapter 3.

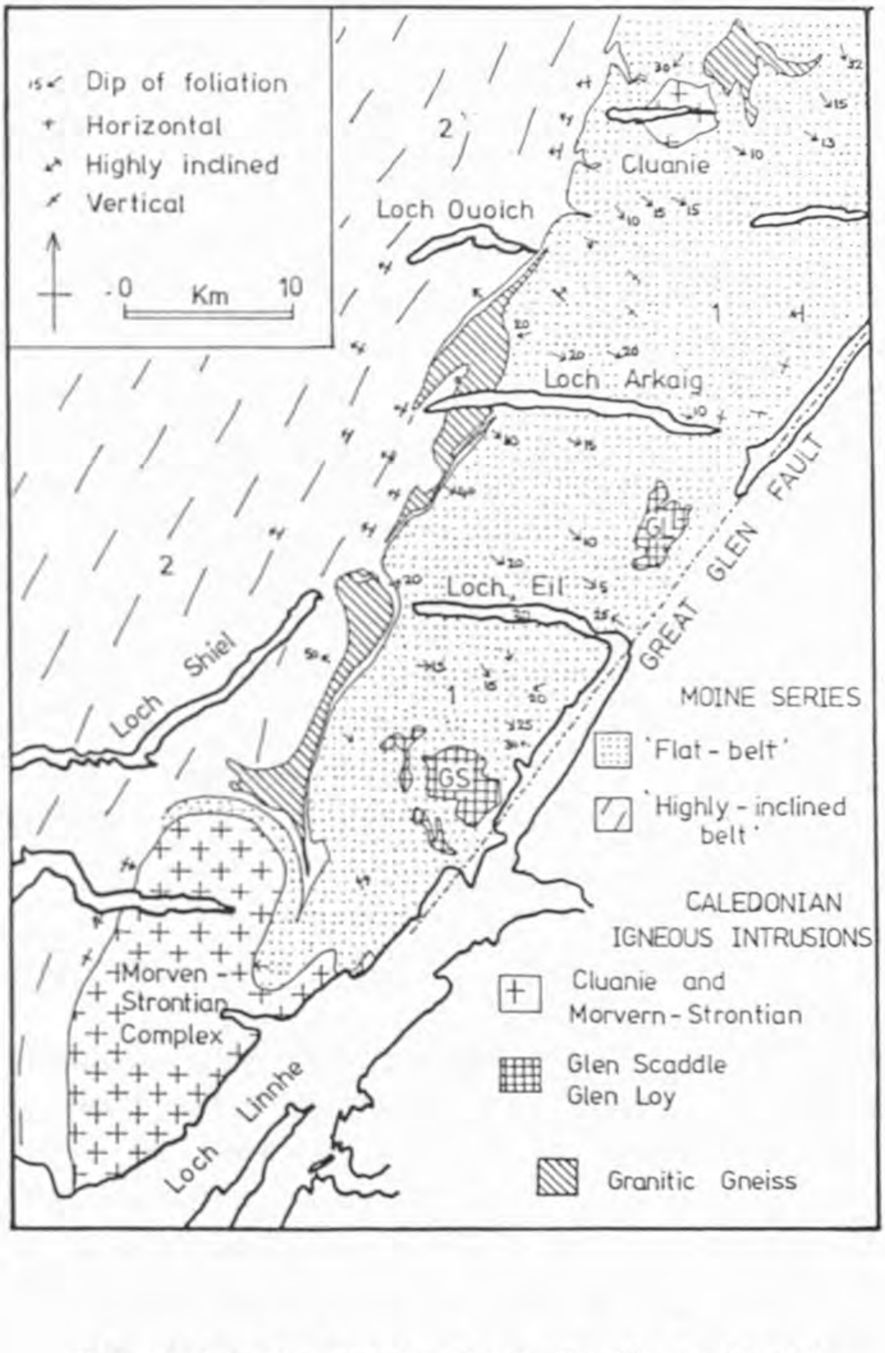
The gneiss was first recognised in the Strontian area by Peach & Wilson (1904), who described it as a foliated granite or augen gneiss comparable to the pre-orogenic Inchbae augen gneiss of Ross-shire. The gneiss was shown as "augen gneiss, Sgurr Dhomhnuill" on Sheet 53 (1" to 1 mile) published by the Geological Survey in 1948.

Leedal (1952) traced the outcrop of the "Acid orthogneiss and related injection gneiss of Sgurr Dhomhnuill" from the northern end of the Morvern-Strontian complex northeastwards through western Ardgour and further northeastwards to north of the Cluanie Complex (see Fig.6, taken from Leedal,1952 p.38).He noted that this line of discontinuous outcrops divided a "highly inclined" belt of striped and pelitic rocks in the west from a "flat belt" composed of psammites in the east. He emphasised "that this outcrop line does not mark a continuous tectonic structure"(p.37). He correlated the psammites east of this discontinuous line with the Upper Psammitic Group of Morar.

Harry (1953) published the first detailed petrographical and petrogenetic study of the gneiss and noting that the " augen gneiss, Sgurr Dhomhnuill" does not outcrop on the summit of Sgurr Dhomhnuill, he renamed it " the composite granitic gneiss of western Ardgour. He noted that the composite granitic gneiss outcroped within the oligoclase - biotite - quartz gneisses (otherwise referred to as the regional injection gneisses) and to the west of a wide belt of psammites which he correlated with the Upper Psammitic Group of the Morar succession. The gneisses, which he correlated with the Striped and Pelitic Group of Morar, thus have to occupy the core of a major anticline..

According to Harry, the composite granitic gneiss contains quartz, micro-

Figure 6.



Map to show the extent of the "highly inclined" and "flat belt" in the SW Moine rocks.

(after Leedal,1952).

cline, oligoclase and biotite and it develops sub-ordinate amounts of augen-gneiss due to growth of microcline porphyroblasts. The composite granitic gneiss and its augen facies pass insensibly and irregularly into adjacent regionally developed oligoclase - biotite - quartz gneisses.

He argued that the oligoclase - biotite - quartz gneisses formed from pelitic schists by feldspathisation(the metasomatic introduction of Na, Ca, Mg). This front of feldspathisation was followed by a front of silica-potash metasomatism which converted some of the oligoclase - biotite - quartz gneiss into composite granitic gneiss. He states that "there is good reason to believe that in Ardgour the granitic gneiss is a replacement that never became separated from its roots to form an independent, far travelled intrusive body"(Harry,1953.p.303).

Harry's evidence that soda metasomatism produced the oligoclase - biotite quartz gneiss from pelitic Moine schists is based on a comparison of two specimens of normal pelitic Moine schists against his analyses of the gneisses. However Butler (1965) published analyses of 22 pelitic rocks from Ardnamurchan and states that these pelites are essentially isochemical with the regionally injected (pelitic) gneisses.

While mapping the area east of Loch Duich on the west coast, Clifford (1957) postulated the existence of a nappe structure which he termed the Kintail Klippe. In searching for a root zone for this nappe he noted certain petrological similarities between some of the rocks in the nappe and the orthogneisses and injection gneisses found in the Loch Arkaig district which are contiguous with the composite granitic gneiss of western Ardgour, further south. He noted that the junction between the "highly inclined" belt and the "flat" belt was also the eastern limit of intense regional injection and permeation in the Moine. He termed this junction the "Loch Quoich line".

Dalziel & Johnson (1963) examined the structural setting of the granitic gneiss and tried to establish its age in relation to the structures within the gneiss and the surrounding metasediments (they emphasise that their use of the terms granitic gneiss and metasediment was merely a convenience; they did not wish to imply necessarily that the granitic gneiss precursor was a granite. Using such criteria as rodding of quartzo - feldspathic material, veining along fold axial planes, folding of migmatitic veins and crystallographic orientation of minerals: these authors attempted to analyse the chronology of the events relating to the generation of the granitic gneiss and its subsequent deformational history.

On a larger scale they concluded that the foliation of the gneiss had been folded by major F_3 folds. They noted that the gneiss contains numerous folds which they correlated with F_2 folds in the metasediments, wherein they fold microcline bearing pegmatites. The main foliation of the gneiss was noted to be axial planar to these folds and in zones of strong (D_2) deformation the pegmatites are very nearly co-planar to the axial planes of the F_2 folds. They concluded that the pegmatites are pre- or early F_2 in age. In some places a foliation in the gneiss is seen to be folded round the cores of F_2 folds and thus they argued that the gneiss contains at least two foliations; pre-F_2, and syn-F_2.

In conclusion Dalziel & Johnson favoured an origin for the granitic gneiss involving migmatisation of the limbs of a primary F_1 nappe which was subsequently folded into major F_2 folds.

Dalziel (1963b) examined zircons from the granitic gneiss and compared them to zircons from the surrounding metasediments, the Strontian Granite and the orthogneisses of Carn Chuinneag and Inchbae.

In comparison to those in the surrounding metasediments the granitic gneiss zircons are slightly less rounded, have more outgrowths and overgrowths and different elongation- and length-frequencies. The Carn Chuinneag and Inchbae mass, which Dalziel accepted as a pre-metamorphic magmatic intrusion, contains zircons which are mainly euhedral and have not been rounded during the subsequent deformation and metamorphism. Dalziel concluded that the Ardgour gneiss is not a pre-metamorphic magmatic granite, rather it formed by migmatitic processes from Moine metasediments. The slight differences in the morphology of zircons in the granitic gneiss and metasediments he attributed to either initial source differences or to slight outgrowths and overgrowths onto original metasedimentary zircon grains.

Dalziel (1966) reviewed the stratigraphical position of the Ardgour gneiss, noting that the gneiss occupies the same horizon as the Beinn an Tuim Striped group in the area immediately northeast of Glenfinnan village. He established a local structural succession,given below;

(Highest) 3) Glen Garvan Psammitic Group.

2) Druim na Saille Pelitic Group.

(Lowest) 1) Beinn an Tuim Striped Group.

Previous workers (Leedal,1952;Harry,1953) had correlated the Glen Garvan Psammitic Group with the Upper Psammitic Group of Morar and consequently the Druim na Saille Pelitic Group and the Beinn an Tuim Striped Group were correlated with the Striped and Pelitic Group of Morar. Dalziel compared stratigraphical thicknesses and noted that the Druim na Saille Pelitic

Group and the Beinn an Tuim Striped Group frequently contain amphibolite bands and lenses, as does the Lochailort Pelitic Group but not the rocks of the Morar succession. he therefore proposed a correlation of the Lochailort Pelitic Group with the Beinn an Tuim Striped Group and perhaps also the Druim na Saille Pelitic Group. An outcome of this correlation is to place the Glen Garvan Psammitic Group stratigraphically higher than the entire Morar and Lochailort Succession (see Fig.7, after Dalziel,1966).

Dalziel also summarised the structural geology of the granitic gneiss and its surrounding metasediments. He stated that the granitic gneiss and the associated metasediments had been affected by four principal fold episodes termed F_1 to F_4 , with the development of major folds of F_2 and F_3 age. Dalziel noted that the central outcrop of granitic gneiss, Beinn an Tuim Striped Group and Druim na Saille Pelitic Group is flanked on both sides by the Glen Garvan Psammitic Group indicating that the central belt lies in the core of a major anticline. On both sides of the central belt major F_2 synforms (and synclines) were mapped. In the central belt the F_2 folds are much tighter and more variable in orientation. The plunge of their axes frequently swings through vertical, with antiforms turning into synforms along the axial plane trace. Dalziel attributed the fold variability to lithological control - a mobile pelitic and migmatitic core being surrounded by massive rigid psammites.

Within the granitic gneiss the strong S_2 foliation has largely obliterated any pre-existing S_0-S_1 foliation, so that the geometry of the major folds in the central belt is, of neccessity, based on inferences and extrapolations. Figure 8 (taken from Dalziel,1966.p.142) shows the distribution of lithologies and the position of major F_2 folds in the Glenfinnan area

In an unpublished geochemical and mineralogical study of the granitic gneiss of western Ardgour, Gould (1966) attempted to discover the origin of the granitic gneiss.

He stated that the composition of the granitic gneiss is very similar to the eutectic point in a $NaAlSi_3O_8$ - $KAlSi_3O_8$ - SiO_2 - H_2O system at pH_2O 5Kb. This eutectic would have a melting temperature of 660°c - 670°c. The similarity in composition of the granitic gneiss and the eutectic he took as evidence that the source material did not get very much hotter than the minimum melt composition and that only small percentages of partial melt were produced.

Gould examined Dalziel's conclusions concerning zircon populations (Dalziel,1963b,see above),that the granitic gneiss had never been through a magmatic stage and pointed out that zircons inherited from a metasedimentary host, may have remained almost insoluble in a low temperature anatectic

Figure 7.

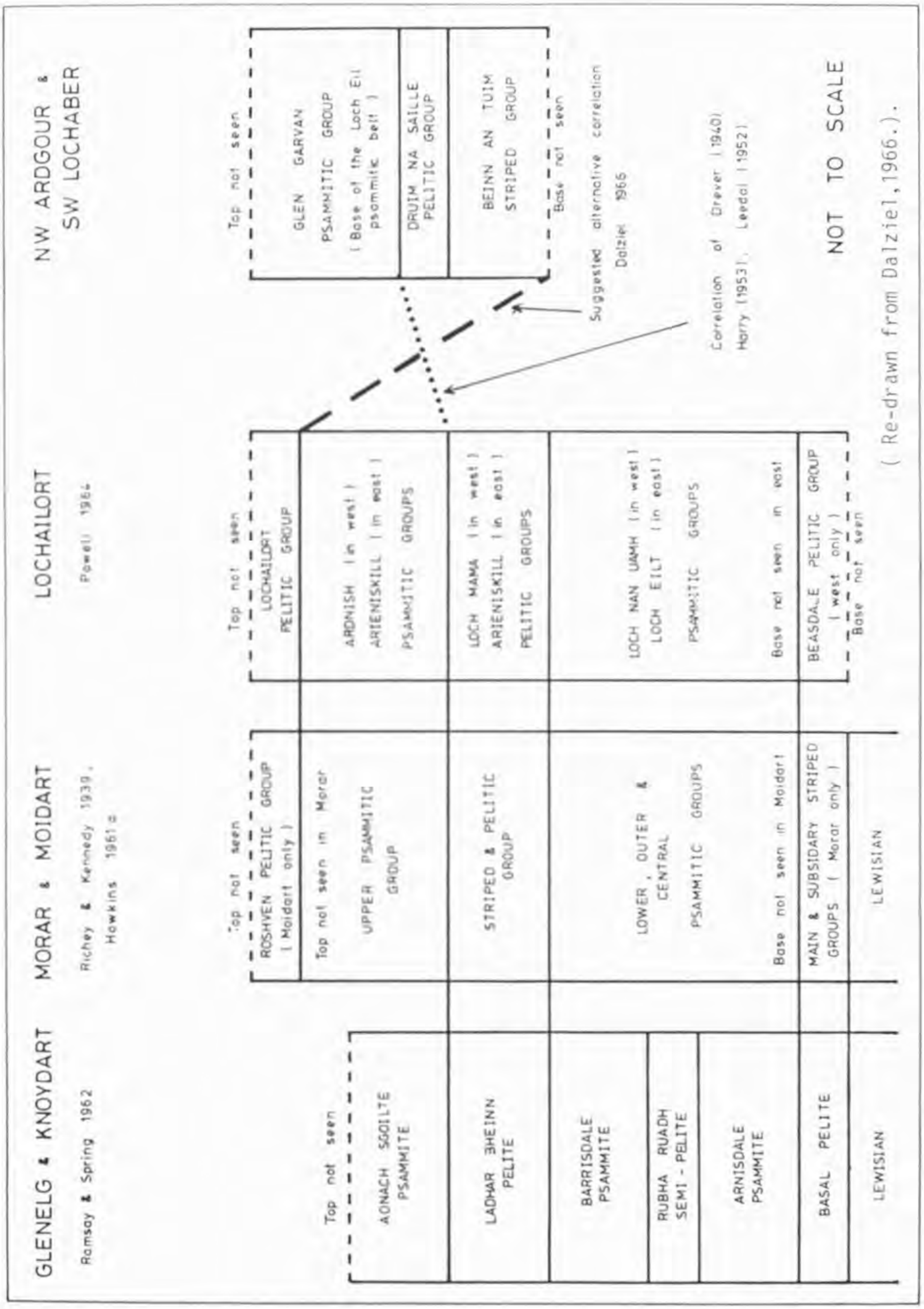
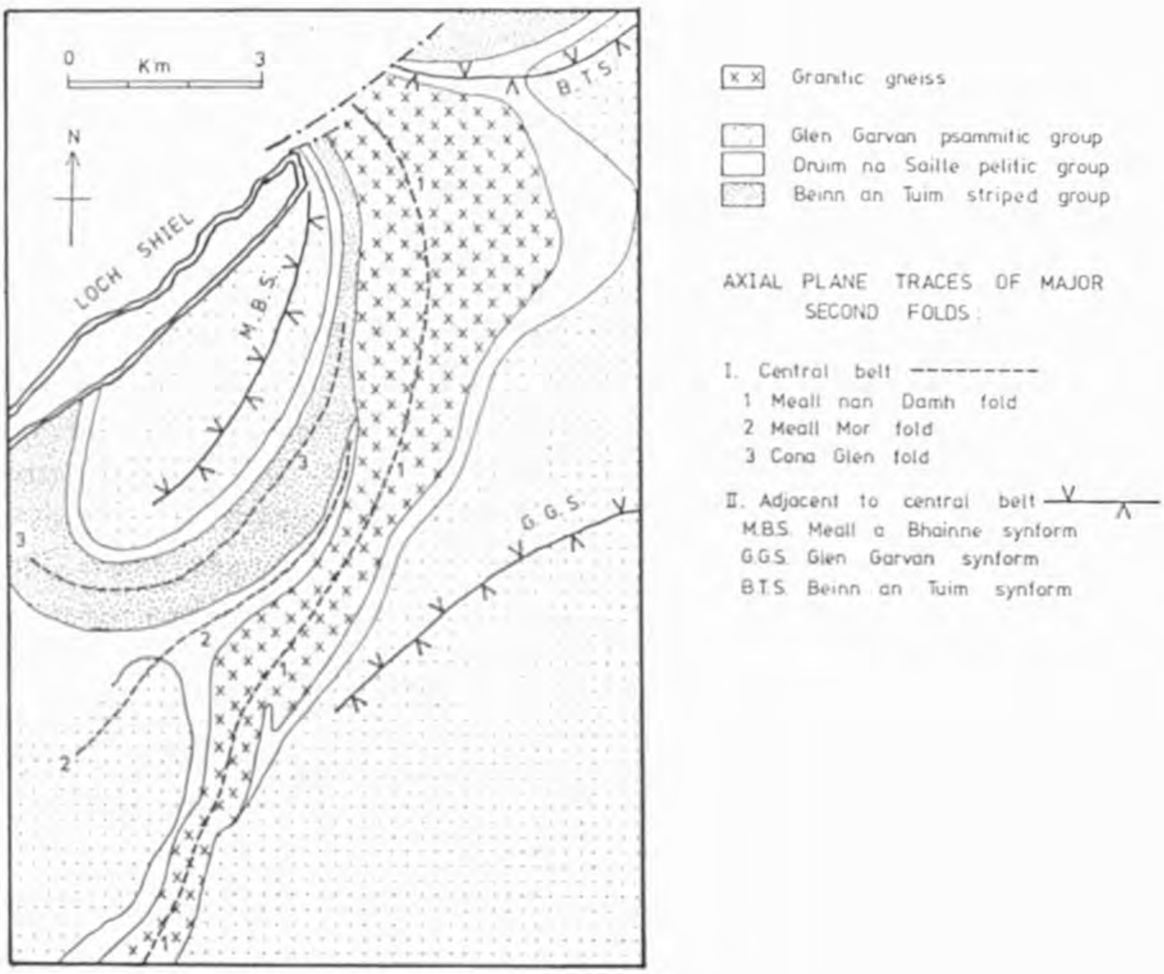


Table to show possible correlations of lithological units in NW Ardgour.and SW Lochaber with those in Morar,Lochailort, Moidart, Glenelg and Knoydart.

Figure 8.



Map to show the distribution of lithologies and major fold axial plane traces southeast of Glenfinnan-Loch Shiel.

(atfer Dalziel,1966).

melt.

Gould's final hypothesis involved the intrusion of thin veins or sills of granitic of granitic liquid into the top of the Beinn an Tuim Striped Group followed by partial homogenisation during subsequent episodes of folding and metamorphism. To fit the structural evidence presented by Dalziel (1963a, 1966) the intrusions have to be pre-F_2 in age.

2.3) The Loch Eil Division.

This section reviews the relatively small amount of work which has been done on the Loch Eil Division rocks in the Loch Eil-Glenfinnan area and also work from other areas which indirectly has a bearing on the geological history of the Loch Eilt-Loch Eil area.

Drever (1940) mapped the Glen Scaddle complex of Ardgour, near Loch Linnhe (see Fig.6) and concluded that the Glen Scaddle complex intruded a sequence of limestones and pelites which rest discordantly on normal Moine granulites (= Loch Eil Division). He noted that the granulites, away from the aureole effects of the intrusive complex, contain rare pelitic gneisses with fibrolitic sillimanite. Calc-silicate bands associated with the "granulites" are composed of basic plagioclase, quartz, pyroxene or hornblende and some garnet. Both of these observations indicate that the rocks have undergone high grade metamorphism.

It has already been noted that Leedal (1952) drew a map showing the extent of the "flat" belt east of the "highly inclined" belt, noting that the "flat" belt was composed of psammites which he correlated with the Upper Psammitic Group of Morar, a correlation reiterated by Harry (1953) and Clifford (1957). Clifford first used the term "Loch Quoich line" to describe the junction between these "flat" and "highly inclined" belts.

H.M Geological Survey reports pertaining to Sheet 62 (Loch Eil) were published in Summ.Proc.Geol.Survey (Lawrie et al,1954-1964) and are here summarised: It was noted that the broad belt of flaggy psammitic granulites with moderate easterly dips structurally overlie the Glenfinnan Injection zone on its eastern side (Sum.Proc.Geol.Surv. for 1954 p.48). The psammitic granulite belt is in general separated from the gneisses to the west by a zone of injection pelite with psammitic intercalations. Locally the gneiss is in direct contact with the granulite belt;"there is no evidence to indicate that the gneiss cuts across the marginal schist/granulite zone like an intrusive igneous rock" (for 1955,p.48). In the psammitic granulites an early phase of recumbent folds was followed by near northerly or southerly gently plunging hinge lines and a set of open folds with east-west axes.

It has already been noted that Dalziel (1966) revised the correlations of the Loch Eil Division with the Upper Psammitic Group of Morar, placing the Glen Garvan Psammite (the base of the Loch Eil Division) above the Lochailort Pelitic Group and the Glenfinnan Division. He regarded the succession as stratigraphical and, importantly in the present context, recognised the same deformation sequences in the Glen Garvan Psammitic Group and in the granitic gneisses etc.

In contrast Lambert et al.(1979) claimed that the structural history of the Loch Eil Division was simpler than that of the Glenfinnan and Morar Divisions which have been affected by a "Morarian (1000 to 750 Ma.) event". They further claimed that calc-silicate lenses in the Loch Eil Division show negligible retrogressive effects. Thus they concluded that the Loch Eil Division is post-Morarian (1000-750 Ma.) and pre-Grampian. In addition they correlated the Loch Eil Division with the Grampian Group (Harris et al.1975) east of the Great Glen fault.

Without detailed evidence, Lambert et al.(1979) claimed that the simplest explanation of these "facts" is to consider the Loch Eil Division as unconformable on earlier metamorphosed Glenfinnan Division.

Piasecki & van Breemen (1979) divided the Central Highland Granulites lying east of the Great Glen fault into two major units; a lower Central Highland Division composed of gneissose rocks and an upper Grampian Group of younger metasediments. The two Divisions are described as separated by a tectonic break (the Grampian Slide) and as having different metamorphic and structural histories. The divisions correspond respectively to a basement complex and a cover sequence.

Piasecki & van Breemen attempted regional correlations of these units based on lithological similarities. The basement complex (Central Highland Division), they correlated with the Glenfinnan Division and the cover sequence (the Grampian Group) with both the Morar and Loch Eil Divisions. Such correlations require, in the Glenfinnan-Loch Eil area, the "Loch Quoich line" to be a tectonic break separating divisions with different structural and metamorphic histories.

Strachan (1982) in mapping the area around the head of Loch Eil divided the rocks near the base of the Loch Eil Division into a number of formations which vary in thickness across strike; a variation which he attributed to complex tectonic sliding during an early D_1 deformation event. Rocks on both sides of this slide zone have subsequently undergone the same tectonic history. Subsequently Strachan (1985) presented an outline of the lithology and structure of the Loch Eil Division in the Ardgour-Loch Eil-Loch Arkaig area, in which he considered the effects of sliding to be negligible. He mapped sub-recumbent minor F_1 and F_2 folds and believed them to be of Precambrian (Grenville) age. Caledonian re-working produced upright D_3 to D_5 structures in the area, with upright N-S trending major F_4 folds being the dominant structures. He believed that the more intense D_4 deformation within the Glenfinnan Division further west produced the "steep" (ie. highly inclined) belt.

Concluding remarks.

Previous stratigraphical and structural research has established an E-W tripartite division of the Moine Succession. A local stratigraphical succession has been established within the western Morar Division. but not within the central Glenfinnan or eastern Loch Eil Divisions. Direct stratigraphical correlations between the three divisions cannot be made.

The junction between the Morar and Glenfinnan Divisions has been established as tectonic and termed the Sgurr Beag slide. Its age relative to the structural evolution of the adjacent divisions has not been established. The Glenfinnan-Loch Eil Division boundary is most probably a stratigraphical junction although indirect correlations have been used to suggest that the junction should have a basement/cover relationship.

The origin of the Ardgour granitic gneiss and its precise structural setting has not yet been established.

C H A P T E R 3.

Previous Geochronological Research.

3.1) Rubidium-Strontium half-life values.

3.2) Previous research.

3.1) Rubidium-Strontium half-life values.

In the last twenty years the accepted value for the decay constant of Rb has changed. The value used for the constant in the papers quoted in this chapter are listed below.

Giletti et al.(1961) quoted a half-life of ^{87}Rb of 4.7 x 10^{10} yr. (or λ^{87}Rb=1.47 x 10^{-11} yr^{-1}). Long & Lambert (1963) used λ^{87}Rb=1.47 x 10^{-11} yr^{-1} but they revised a Sr spike calibration and changed 22 of the 77 Rb/Sr ages quoted by Giletti et al.(1961) downwards by 2.5%. The only value relevant to this summary which was changed was 740±25 Ma. to 720±25 Ma. A full list of the changes is given in Long & Lambert (1963,p.245.Appendix c). Van Breemen et al.(1974), Brook et al.(1976) and Brook et al.(1977) used λ^{87}Rb=1.39 x 10^{-11} yr^{-1}. Van Breemen et al.(1978) calculated or re-calculated all the ages given in this paper using the value for the decay constant recommended by Steiger & Jager (1977) of λ^{87}Rb=1.42 x 10^{-11} yr^{-1}.

Subsequently Brewer et al.(1979), Aftalion & van Breemen (1980) and Powell et al.(1983) have used λ^{87}Rb=1.42 x 10^{-11} yr^{-1}.

3.2 Previous research.

Since the earliest geological work in the NW Highlands, the age equivalence and metamorphic history of the Moine schists has always been a source of debate.

It has been suggested that the Moine schists may be part of the Lower Palaeozoic succession, metamorphosed during the Caledonian orogeny (Frodin, 1922). Alternatively the Moine series may comprise Precambrian sediments metamorphosed either during a Precambrian orogeny (J.Horne in Peach & Horne, 1930.p,200-1) or during the Caledonian orogeny (Bailey,1950) or during both (Read,1934.p308). They have been considered as equivalents of the Torridonian sediments of the foreland region (Peach,1892,1913. Bailey,1950. p.235-6) and indeed sedimentary equivalents of the Lewisian gneisses(Barrow, 1925 in discussion of Tilley,1925).

Virtually all permutations of these possibilities have been suggested at some time (for a short summary see Phillips,1951).

Towards the end of the period before the application of techniques of isotopic dating the majority of workers favoured a Caledonian age for the metamorphism of the Moine sediments (eg. Bailey(1950),Kennedy(1955),Ramsay (1963), Sutton & Watson(1953), Read(1934))

The first published isotopic dates from the Moine schists were presented by Giletti et al.(1961). Ages from biotites in Moine metasediments (using both Rb/Sr and K/Ar techniques) lay in the range from 435±10 Ma. to 405±10 Ma., indicating that metamorphism of the Moine rocks occurred during the Silurian or early Devonian (according to the time scales of Kulp(1959) and Holmes(1960)).

Rb/Sr ages from muscovites in pegmatites from Knoydart and North Morar gave ages in the range 740±25 Ma. to 665±15 Ma. and these analyses were interpreted as relict ages of the pegmatites. As the pegmatites apparently are not related to any other major igneous intrusions it was assumed that they were produced by a Precambrian regional metamorphism of an age greater than 740 Ma.

Long & Lambert(1963) generally confirmed and elaborated on the results of Giletti et at.(1961). They reduced slightly the age of the late-Caledonian recrystallisation from 415±15 Ma. to 390 Ma.,an age which they noted is in agreement with the ages of 390±5 Ma. found in the Caledonian granites of the Southern Uplands and the Lake District.

Pegmatite mineral ages (Long & Lambert,p.288) ranging from 740±15 Ma. to 365±10 Ma. were recorded but the "agreement" of ages at 730 Ma. appeared to the authors to be grounds upon which to infer Moinian sedimentation took place not long before.

A muscovite age of 550±15 Ma from a specimen of Moine schist from SW Morar was interpreted as indicating a pre- late-Caledonian metamorphism, the isotopic signature of which was retained because the area at the western border of an area affected by the Caledonian metamorphism.

In a study of pegmatites from eight localities along the Lochailort-Glenfinnan road section. van Breemen et al.(1974) identified five types of pegmatite related temporally to the structural history of the area, these are:

1) Early pegmatites folded by F_2 folds.
2) Pegmatites axial planar to F_2 folds.
3) Pegmatites post-F_2 , folded by F_3 folds.
4) Pegmatites post-F_3 but pre-dating a suite of lamprophyres.
5) Pegmatites which post-date the lamprophyres.

Using Rb/Sr analyses of muscovites from early (pre-F_2) pegmatites they obtained ages of 727±19 Ma.,730±17 Ma. and 735±45 Ma. This last age was from muscovites taken from the centre of a pegmatite whilst finer and orientated muscovites from the margins of the same pegmatite yielded an age of 546±105 Ma. Another possible pre-F_2 pegmatite yielded an age of 445±30 Ma. All the remaining pegmatites were structurally post-F_2 and yielded ages of 450±10 Ma., an age which was interpreted as relating to the D_3 deformation. In correlating F_2 folding in the area with the deformation sequence seen in the Moine thrust belt, the D_2 and later deformation was regarded as post-Arenig in age.

Importantly, these authors noted that Morarian pegmatites are found in the Morar and Glenfinnan Divisions, an observation which leads to the conclusion that the Sgurr Beag slide which separates these divisions cannot be a "front" dividing belts of Morarian and Caledonian tectono-metamorphic activity.

It will be shown later that the fold chronology adopted by van Breemen et al.(1974) has to be revised (Chapter 5) with important consequences to the regional correlations and thus the "structural" significance of these isotopic ages.

In their investigation of the Rb/Sr systematics of the Ardgour granitic gneiss, Brook et al.(1976) using large whole rock samples obtained an isochron age of 1050±46 Ma. with an initial $^{87}Sr/^{86}Sr$ ratio of 0.708±0.002, thus improving on the "preliminary isochron" of 810 Ma. reported by Lambert (1969). Brook et al. accepted Dalziel's (1966) interpretation of the origin and structural context of the gneiss and thus interpreted the age as that of the last Sr isotope homogenisation during the migmatisation. Thus the age of 1050±46 Ma was assumed to represent the age of the early main metamorphism in the area.

Brook et al. further argued that the parental material of the Ardgour granitic gneiss would have had a mantle $^{87}Sr/^{86}Sr$ ratio of 0.700 at ca. 1250 Ma. and therefore concluded that sedimentation of the Moine Series occured between 1250 Ma. and 1050 Ma. Thus direct correlations between the Moine Series and the Torridonian sediments of the foreland are untenable because of the 995±24 Ma. and 810±17 Ma. ages of the Stoer and Torridon Groups respectively (Moorbath, 1969).

In 1977 Brook et al. published the first Rb/Sr whole rock isochron obtained from a metasediment within the Moine Schists. An age of 1024±96 Ma. with an initial $^{87}Sr/^{86}Sr$ ratio of 0.709±0.001 was obtained for the Morar Pelite (= Striped and Pelitic Group of previous workers) sampled near Druimindarroch on the extreme west coast of Inverness-shire. Noting that the rocks of the area underwent their highest grade of metamorphism (upper greenschist to lower amphibolite facies) during an event which partly preceded and overlapped the local D_2 deformation, the authors interpreted the age of 1024±96 Ma. as the age of the main (M_2) regional metamorphism. The tectono-metamorphic history of the Morar area thus appears to involve;

a) Grenville events: isoclinal folding (F_1), low grade metamorphism (M_1) and intense penetrative deformation (D_2) accompanied by medium grade metamorphism (M_2)

and

b) Caledonian deformation and mild retrogression.

The authors noted that the acceptance of the isotopic age for the Morarian pegmatites suggests an even more complex history on a regional scale.

Van Breemen et al.(1978) attempted a regional synthesis of Precambrian and Caledonian events in the NW Highlands of Scotland and Western Ireland. They accepted that ages of 1070±30 Ma. and 1000±30 Ma. from the Annagh gneiss complex of western Ireland and the 1020±50 Ma. age of the Ardgour granitic gneiss indicate the effects of a Grenville orogenic episode. A 555±5 Ma. U/Pb zircon age from a segregation pegmatite in the Ardgour granitic gneiss is considered to date a period of migmatisation later than the formation of the granitic gneiss and further U/Pb monazite analyses for early pegmatites in the Moine Series yielding ages of 815±30 Ma. and 730±20 Ma. were reported.

Whole rock Rb/Sr data for two semi-pelites from the Morar succession are presented which do not define isochrons, rather, van Breemen et al. argue that "best fit" lines corresponding to ages of 700±100 Ma. and 720±130

Ma. are, in view of the similar ages from the Morarian pegmatites, geologically meaningful. On the basis of this data these authors claim, firstly that the Morar Division is older than the 815±30 - 730±20 Ma. early pegmatites and, secondly that the Morar Division was not involved in the Grenville orogeny (arguing that the 1024±96 Ma. whole rock Rb/Sr isochron for the Morar Pelite (Brook et al.1977) is possibly less convincing than their whole rock regression ages of 700±100 Ma. and 720±130 Ma.).

Thus they conclude that there is evidence for a Grenville Orogenic cycle ca. 1070-1000 Ma. a Morarian "episode" ca. 800-700 Ma. and a Caledonian Orogenic cycle.

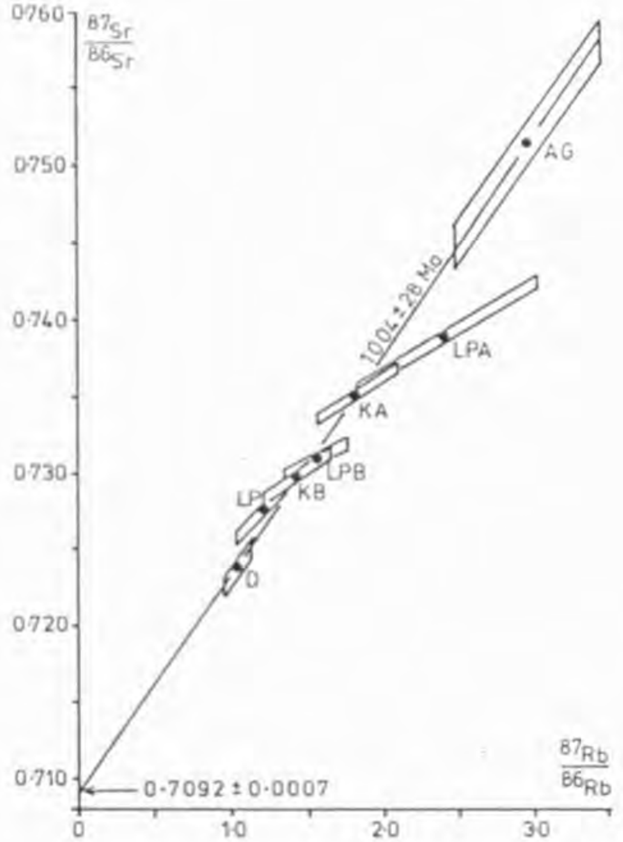
Using new Rb/Sr whole rock data together with previous analyses Brewer et al.(1979) reviewed the tectono-metamorphic history of the Morar - Glenfinnan area. Samples from small areas or "domains" produced three new isochrons and two regression line ages for pelites in the Morar-Knoydart area. The isochron ages are 459±32 Ma.,416±20 Ma. and 389±58 Ma. The regression line ages are 479±24 Ma. and 769±50 Ma. The high $^{87}Sr/^{86}Sr$ initial ratios of these isochrons and regression lines indicate that the source material for these pelites had a considerable pre-Caledonian crustal history. Using these results togather with the data determined for the Morar Pelite (Brook, et al.1976) these authors showed that a graph of $^{87}Sr/^{86}Sr$ ratio at present and at times during the past, plotted against time (see Fig.9) shows that all the samples had a common initial $^{87}Sr/^{86}Sr$ ratio of 0.7092 at 1004±28 Ma. They conclude that this common $^{87}Sr/^{86}Sr$ ratio was produced by initial isotopic homogenisation during sedimentation and/or diagenesis,and that subsequent Precambrian and Caledonian activity has reworked isotopically similar material. Since the Ardgour gneiss falls on the array of Moine pelite domains, the authors argued that the precursors of the gneiss was an integral part of the Moine succession.

On a regional scale Brewer et al. concluded that the Moine rocks of the Morar Glenfinnan area have undergone Grenville and Caledonian orogenic events, the latter causing, in places, resetting of the earlier dates. The mineral isotopic data, representative of cooling ages, indicate that the Moine Series cooled after the Caledonian orogeny "by a process of sequential uplift along discrete zones (Slides) commencing in the west.

Aftalion & van Breemen (1980) examined Rb/Sr data from the Ardgour granitic gneiss and surrounding metasediments and compared these with the apparently conflicting U/Pb data from zircons and monazites in the granitic gneiss.

They obtained a Rb/Sr regression line age of 455±60 Ma. from adjacent 5 - 8 cm. slabs of semi-pelite in the Glenfinnan Division near the Ardgour granitic gneiss. The "errorchron" obtained implying that Sr migration has

Figure 9.



Domain isochron diagram for the southwestern Moine, AG-Ardgour granitic gneiss, D-Morar Pelite at Druim-indarroch, KA & KB-Morar Pelite at Knoydart, LP, LPA & LPB-Lochailort Pelite.

(taken from Brewer et al.1979).

occured on a scale of 10 cm. or greater during the Caledonian orogeny.

The 1030±50 Ma. isochron from the granitic gneiss (Brook et al,1976) was obtained from large (20 cm.+) whole rock samples and this age is taken to indicate that Sr migration did not occur over distances of 20 cm.+ during the Caledonian thermal event, thus a Grenville age for this rock has been preserved.

U/Pb zircon data obtained by Aftalion & van Breemen from a segregation pegmatite in the granitic gneiss yielded discordant points which appear to be alligned along a chord on a choncordia diagram with upper and lower intersection ages of 1517±30 Ma. and 556±8 Ma. respectively. Neither of these ages corresponds to a geologically recognisable event. The U/Pb results were modelled mathematically by the authors using a theory of multi-episode lead loss during periods of thermal activity occuring after the initial or primary age of zircon formation. Using these models the best solution involves a 1700-1800 Ma. age for the primary zircons, Grenville metamorphic events at ca. 1100-980 Ma. and prograde Caledonian events at ca. 465-490 Ma.

A model involving Grenville, Morarian and Caledonian metamorphic events could not be made to fit the data, indicating that no or only a very weak interference by the Morarian "event" has occured.

Thus these conclusions, based solely on isotopic data, are generally in agreement with those of Brook et al.(1976 and 1977) and Brewer et al. (1979). Powell et al.(1983) obtained Rb/Sr ages of 776±15 Ma. to 746±15 Ma. for muscovites from a "Morarian" pegmatite at Ardnish in the Morar Division and showed that the host metasediments had been deformed and metamorphosed prior to the injection of the pegmatite.

They suggested that the Rb/Sr muscovite ages for "Morarian" pegmatites should not be taken as proof of the existence of an orogenic episode at this time.

C H A P T E R 4

Lithology, Mineralogy and Petrology.

4.1) Introduction.

In this section the lithology and petrography of the various metasedimentary rock types are discussed. Similar lithologies in each of the major divisions of the Moine rocks (Morar, Glenfinnan and Loch Eil) are compared and contrasted wherever possible.

Metamorphic textures and mineral reactions are discussed in Chapter 6. Discussion of the petrology and mineralogy of basic and metabasic rocks forms part of Chapters 6 & 8.

Detailed work on the metamorphism and geochemistry of the calc-silicates around the Ranochan-Loch Eilt area is discussed in Chapter 7.

4.2) Psammitic rocks.

4.2a) Introduction.

The Loch Eil Division is composed almost entirely of psammitic rocks. Within the Glenfinnan Division,psammites are interbanded with large porportions of pelites and semi-pelites. The whole assemblage was often mapped as Striped Group. The psammites of the Morar Division outcrop only in the extreme west of the area mapped.

4.2b) Loch Eil Division psammitic rocks.

(1) Lithology.

The psammites of the Loch Eil Division are exposed in a large number of small exposures generally less than 20 metres long. In the field, lithological and stratigraphical correlations between exposures proved impossible, consequently each exposure was described and placed into one of four lithological sub-groups;

A) Thick psammites 50-100 cm. thick with fewer thin pelitic units 15-20 cm. thick.

B) Very thick pure quartzites with subtle colour banding, micaceous psammites and few, if any, pelites.

C) Dominantly semi-pelites with occasional thin ribs of quartzite (a few cms. thick).

D) Flaggy psammites/ semi-pelites, 10-15 cm. thick, plus thin quartzites.

Figure 10 shows lithologies typical of the Loch Eil Division.

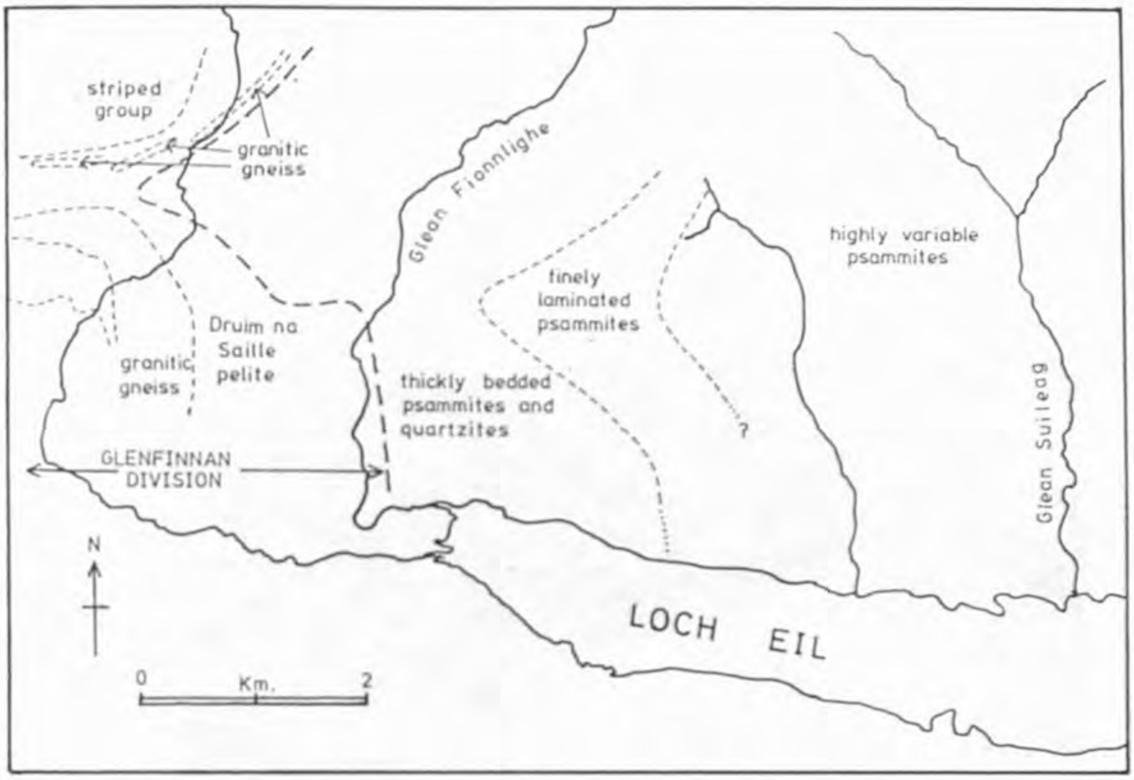
Initially individual exposures were placed fairly easily into one of the sub-groups. Later exposures were often gradational between the subdivisions described above. However, all exposures were placed, as best as possible, into one of the sub-groups. A map of these Loch Eil psammite lithological sub-groups shows little stratigraphical sense but when sub-groups A and B, and sub-groups C and D were combined then some suggestion of litho-stratigraphy was obtained (fig.10). However little significance is placed on this map.

No mappable pelitic horizons were found in the area. The junction of the Loch Eil Division psammites with the rocks of the Glenfinnan Division to the west is very poorly exposed. Its nature and significance is described in Chapter 5.

In the field the psammites are generally well bedded with thin layers of micas along the bedding planes. Bedding is nearly always planar although cross bedding was noted in a few places (Fig.11). Abundant cross bedding has been reported from the Loch Eil psammites a few Kms. to the south, along strike from the western end of the Loch Eil psammites in the area mapped (Strachan,1982).

In hand specimen the Loch Eil psammites are typically cream or light grey coloured, with an average grain size of about 1 mm. They vary from homogeneous massive quartzitic psammites (eg. Fig.12,section 191/1033) to more heterogeneous banded psammites (Fig.12, section 97/376) where the banding is less well defined by mica layers. Where the banding is less well defined and the micas more evenly distributed, the rocks grade into semi-pelites.

Figure 10.



Map of possible lithological sub-divisions at the northwestern end of Loch Eil.

Figure 11.

Photograph to show cross bedding in the Loch Eil Division psammites.
(Exp.252 Grid Ref. NM 96218092).

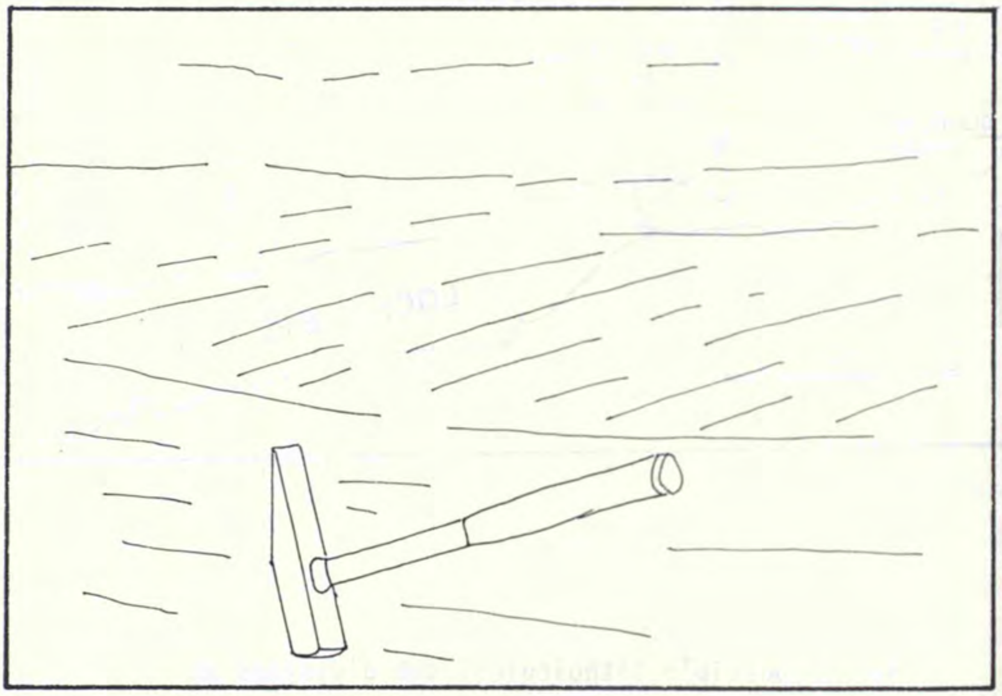


Figure 12.

Modal analyses of Loch Eil Division Psammites.

	18/109	29/180	35/229	38/229	39/229	86/276	87/279	90/301	95/335
QUARTZ	45.3	61.8	79.5	56.8	80.2	64.0	77.1	87.2	79.2
FELDSPAR	45.4	30.2	9.2	10.5	10.8	3.1	4.4	6.4	8.8
MUSCOVITE	0.4	0.0	3.3	17.4	5.9	25.7	18.4	2.1	0.1
BIOTITE	8.7	3.6	7.5	14.9	2.5	3.5	0.1	0.3	1.6
GARNET	0.0	0.0	0.0	0.0	0.0	2.0	0.0	0.0	0.0
EPIDOTE	0.1	3.3	0.0	0.0	0.1	0.0	0.0	3.1	9.2
CHLORITE	0.0	0.6	0.6	0.0	0.2	1.2	0.0	0.4	0.0
CALCITE	0.0	0.0	0.1	0.0	0.3	0.0	0.0	0.0	0.0
ACCESSORIES	0.1	0.5	0.0	0.4	0.0	0.5	0.0	0.5	1.1

	97/376	102/436	111/482	114/503	120/520	123/539	130/572	190/1032	191/1032
QUARTZ	37.1	25.2	59.2	50.4	41.4	71.5	53.1	65.8	81.7
FELDSPAR	34.5	30.9	32.1	31.0	46.1	19.0	20.3	25.1	2.3
MUSCOVITE	4.7	21.0	0.7	0.3	1.0	3.5	7.4	5.2	14.6
BIOTITE	23.5	19.0	5.9	2.6	9.4	5.8	13.1	3.2	1.2
GARNET	0.0	0.0	0.0	0.8	0.0	0.0	0.0	0.0	0.0
EPIDOTE	0.0	0.2	0.8	12.2	0.7	0.0	0.2	0.1	0.0
CHLORITE	0.0	0.0	1.2	0.8	0.2	0.0	5.5	0.1	0.0
CALCITE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
ACCESSORIES	0.2	3.7	0.1	1.9	1.2	0.2	0.4	0.5	0.2

	192/1032	193/1035	194/1036	195/1037	196/1038	197/1039		Average
QUARTZ	50.0	72.7	20.6	65.8	66.6	79.0		61.30
FELDSPAR	18.4	5.6	77.2	14.1	18.8	16.2		22.01
MUSCOVITE	22.8	14.6	0.8	19.2	3.4	4.1		8.20
BIOTITE	7.2	6.6	0.3	0.5	9.2	0.3		6.27
GARNET	0.0	0.0	0.0	0.0	0.0	0.0		0.17
EPIDOTE	0.0	0.0	0.0	0.0	0.0	0.0		1.25
CHLORITE	0.3	0.2	0.7	0.3	0.3	0.0		0.53
CALCITE	0.0	0.0	0.0	0.0	0.0	0.0		0.02
ACCESSORIES	1.3	0.3	0.4	0.1	1.7	0.4		0.61

500 Points per section

(2) Mineralogy and petrography.

Modal analysis of some Loch Eil psammite samples are given in Figure 11. An average of the analyses is also given. It is clear from the analyses that the rocks are slightly micaceous, feldspathic quartzites and psammites.

Quartz.

Quartz is usually the most abundant mineral in thin section. Grain size varies from 0.2-2.0 mm. with most sections having quartz crystal sizes of 0.5-1.0 mm. Grain shape is variable and usually anhedral, except in some of the more semi-pelitic sections where there is often a slight elongation of crystals within the mica foliation. Individual grains can show internal deformation lamellae and the development of sub-grains, whereas other crystals in the same section may be unstrained.

Feldspar.

Both plagioclase and K-feldspar have been found in many thin sections. Grain size varies from 0.2-3.0 mm. Crystal shapes are always anhedral. Partial alteration to saussurite and sericite is common. Optical estimates of the anorthite content of plagioclase in different sections range from An.19 to An.47. The presence of K-feldspar is indicated by occasional microcline twinning (Fig.13a), perthitic textures (Fig.13b) and myrmekitic textures (Fig.13c). Many of the feldspars are slightly altered, untwinned and without myrmekitic or perthitic textures. These could be either plagioclase or K-feldspar. It is estimated that there is more plagioclase than K-feldspar in the sections.

Biotite.

Biotite occurs as laths varying in length from 0.3 to 1.5 mm. long. It shows variable pleochroism.

α = Straw yellow, pale green.

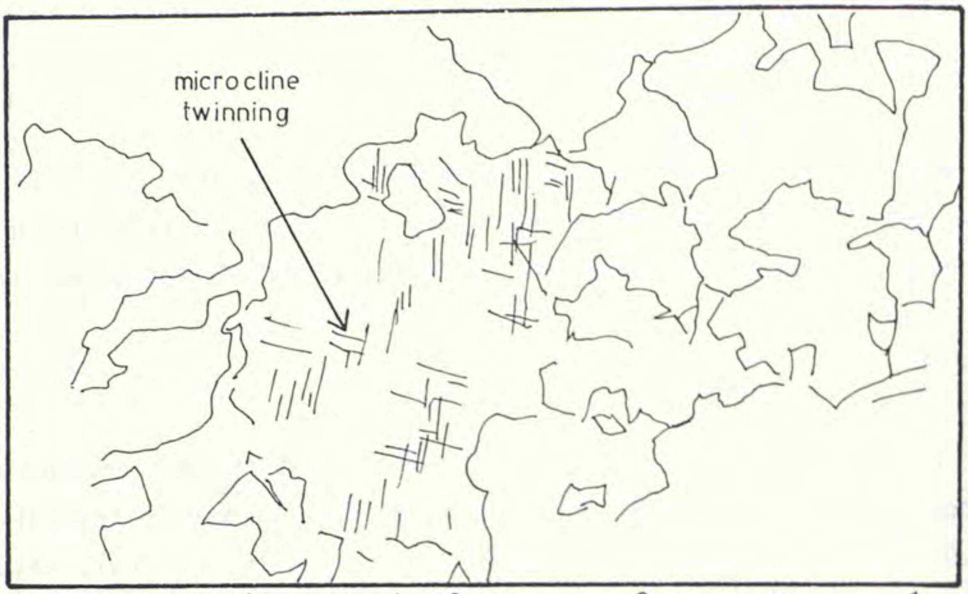
$\beta = \gamma$ = Dark brown,blood red,dark green.

These variations are presumed to reflect compositional differences, possibly in the relative amounts of Ti, Mg and Fe ions.

Biotite laths are either evenly disseminated throughout the rock or concentrated into micaceous foliae. In both cases the laths define a strong foliation.

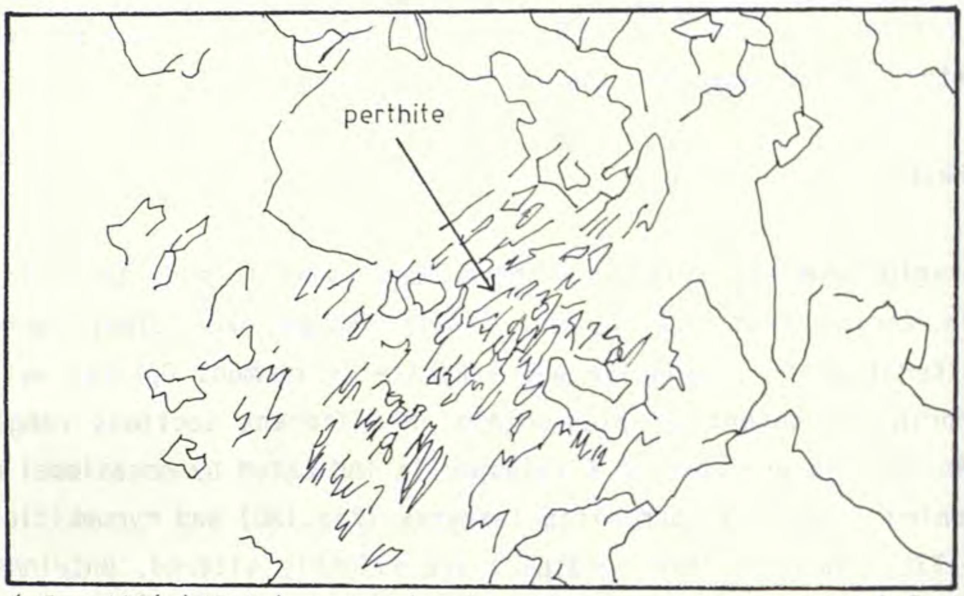
Figure 13. Photomicrographs of feldspar textures in the Loch Eil Division psammites (see text for discussion).

(a)



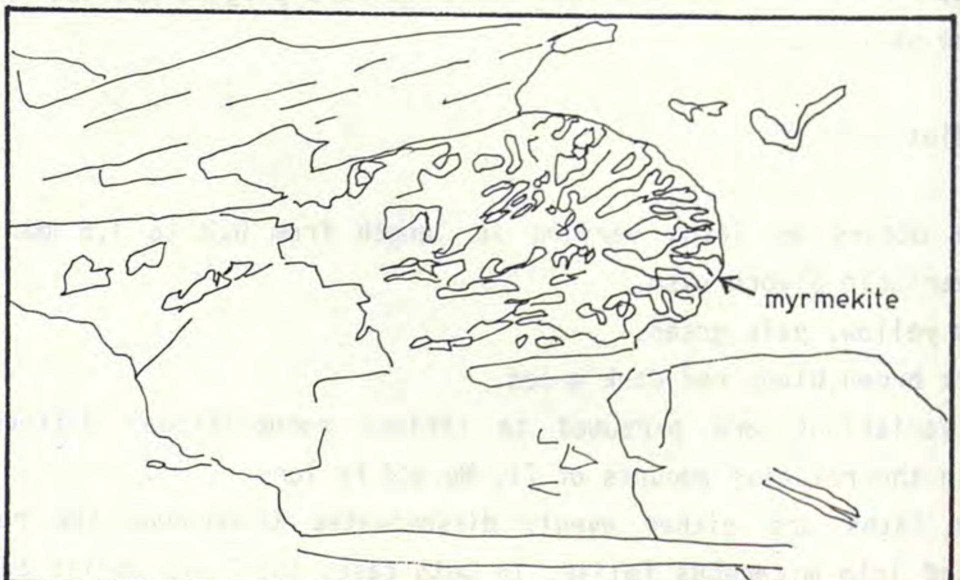
(Exp.194/1036). crossed polars. 0 mm. 1

(b)

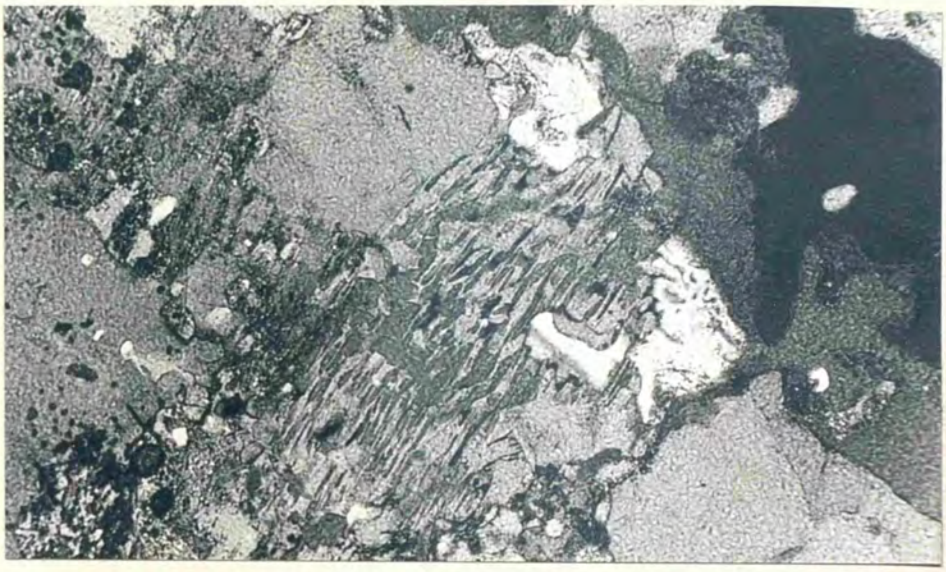


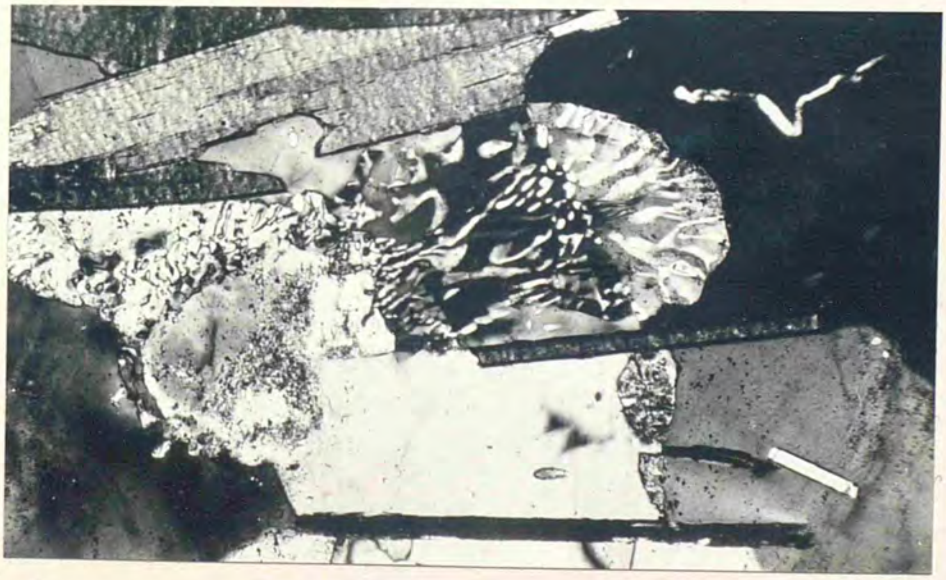
(Exp.123(c)/539).crossed polars.0 mm. 0,5

(c)



(Exp.123(b)/539).crossed polars0 mm. 0,5





Muscovite.

Muscovite occurs in two main forms. There is a suite of small elongate muscovite laths in which the grain size varies from 0.3-2.0 mm. The laths enhance the biotite foliation described above. Larger muscovite laths which do not show any preferred orientation occur as porphyroblasts which are not elongate along the cleavage and are up to 4-5 mm. long.

Other minerals.

Other minerals found in the Loch Eil psammites include garnet, epidote, chlorite, calcite iron ore, sphene, zircon and apatite.

Garnet was found in two sections (86/276 and 114/503), it occurs as small anhedral skeletal crystals up to 0.5 mm. diameter containing some small quartz inclusions.

Epidote is quite common in some of the thin sections (eg.114/503, 95/355, 90/301, 29/180). It occurs as anhedral crystals 0.2-0.7 mm. long. Cleavage is well developed, pleochroism is faint, birefringence is in bright second order colours (δ =0.035) and the optic axial angle is negative. It is concluded that the epidote is very iron rich (Deer et al.1966).

Chlorite occurs as an alteration product of biotite. Calcite is found in some of the more altered sections. Iron ore, sphene, zircon and apatite occur as minor accessory minerals.

4.2c) Glenfinnan Division psammitic rocks.

(1) Introduction.

The rocks of the Glenfinnan Division in the area outcrop over an E-W distance of 13 Km. between a junction in the west termed the Sgurr Beag slide and an eastern junction termed the "Loch Quoich line" (see Chapters 5 and 7).

Exposure within the Glenfinnan Division is generally excellent. Outcrops are often 50-100 metres long separated from adjacent exposures by a few metres or tens of metres. There is poor exposure in the floor of Glen Finnan, in the floor of Allt Feith a Chatha and on the steep slopes of the summit of Sgurr a Mhuidhe which are tree clad. Recent re-afforestation N of the road, W of Glenfinnan village, may make access onto Fraoch Bheinn and Sgurr an Utha difficult in the future.

(2) Lithology.

Most of the rock types within the Glenfinnan Division are banded psammites and pelites often called the Striped Group, together with more homogeneous semi-pelites. In the east of the Glenfinnan Division a strongly gneissose rock, the Ardgour granitic gneiss, is found (see Chapter 4.6).

For the most part the lithologies in the Glenfinnan Division were mapped as Striped Group, Psammites and Pelites, although in the east around Beinn an Tuim more detailed lithological sub-divisions were made (Fig.14). The local lithological sub-divisions used during mapping were:

A) Typical "striped group". Psammitic quartzites up to 50 cm. thick with thinner pelites.

B) Uniform, fine grained micaceous psammite or semi-pelite, gradational into lithology (D).

C) Thinly banded psammitic and pelitic material, 1-3 cm. thick, developing augen-like textures in the pelites.

D) Thinly banded micaceous psammites, 2-3 cm. thick, with micaceous partings. Very well banded.

E) Fairly coarse grained psammites/quartzites togather with biotite rich pelites, non-augen textured. (Could be typical striped group(A)).

General sub-divisions of lithology within the Glenfinnan Division are given on Map 1.

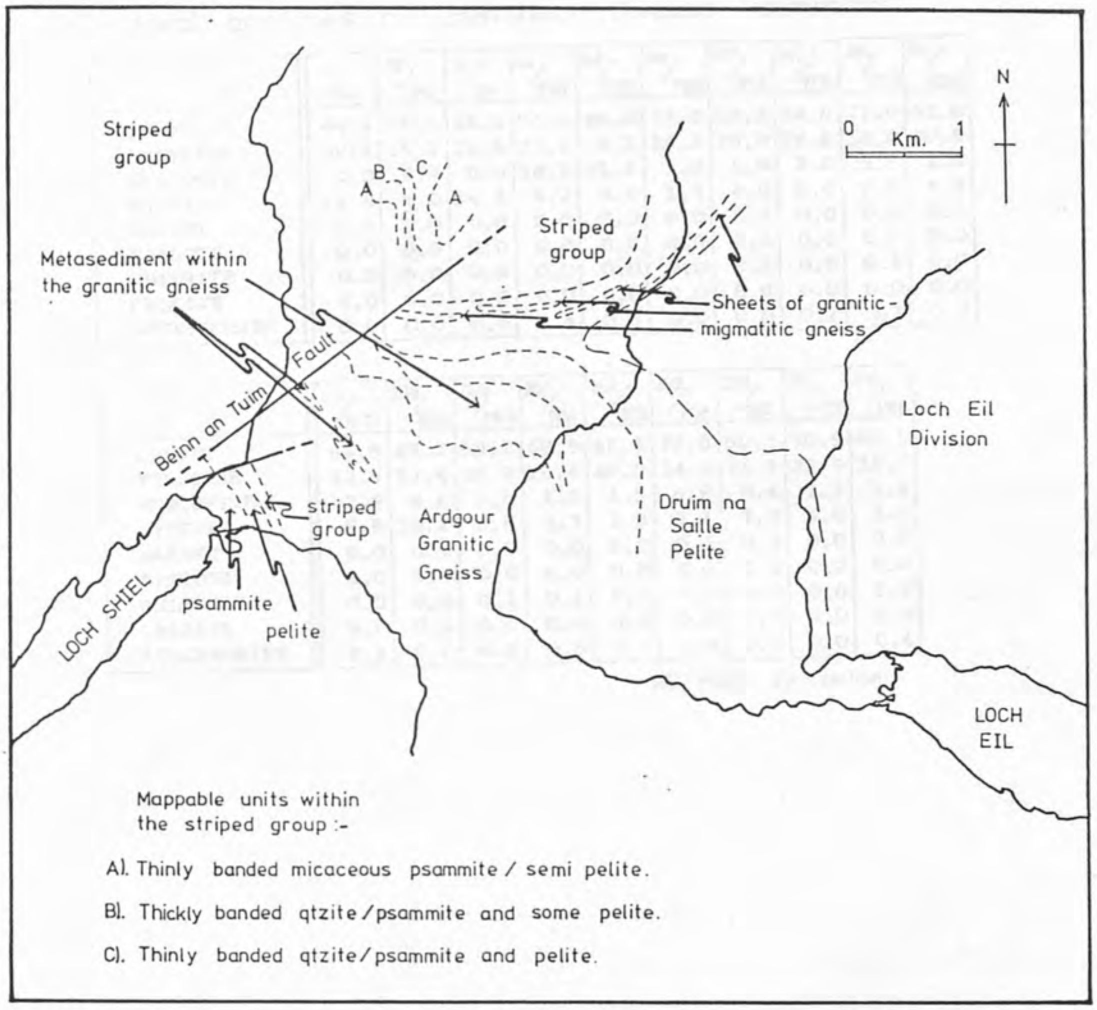
In the field the strongly banded nature of the Glenfinnan Division is characteristic and whilst this certainly reflects original bedding, no unequivocal examples of cross-bedding, graded bedding or other sedimentary structures were observed.

The Glenfinnan Division psammites have an average grain size of 1-2mm. They are light-medium grey to cream coloured. Biotite is common, usually finely disseminated. Occasionally a striped appearance is developed with psammitic layers of 2-3 cm. and migmatitic pelitic layers of 1-2 cm. The rocks in the east tend to be more evenly semi-pelitic whereas the psammites in the west appear to be more quartzitic.

(3) Mineralogy and petrography.

Modal analyses of some of the Glenfinnan Division psammites are given in Figure 15. The rocks are composed of four main minerals; quartz, feldspar, muscovite and biotite, togather with a number of accessory minerals.

Figure 14.



Map to show the main lithological sub-divisions in the Glenfinnan Division rocks east of the Beinn an Tuim fault.

Figure 15.

Modal analyses of Glenfinnan Division Psammites.

	9/60	26/164	31/214	44/230	148/672	161/783	177/963	188/1031	205/1065	214/1087
QUARTZ	44.6	76.8	48.6	54.0	40.0	73.0	65.8	54.6	71.0	63.8
FELDSPAR	38.6	19.2	36.6	23.2	28.2	24.2	29.0	29.8	18.0	30.8
MUSCOVITE	0.0	0.0	0.0	18.2	21.6	1.0	1.0	9.0	5.8	1.0
BIOTITE	16.8	4.0	14.4	4.2	9.6	1.8	4.0	6.4	2.6	4.4
GARNET	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
EPIDOTE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
CHLORITE	0.0	0.0	0.0	0.0	0.0	0.0	0.2	0.0	0.4	0.0
CALCITE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
ACCESSORIES	0.0	0.0	0.4	0.4	0.6	0.0	0.0	0.2	0.2	0.0

	236/1093	238/1093	260/1163	261/1164	262/1165	263/1166	264/1167	265/1168	266/1169
QUARTZ	64.8	22.2	58.1	68.5	68.5	75.6	50.7	70.9	67.5
FELDSPAR	23.4	57.4	32.7	26.4	26.8	14.5	41.5	22.9	24.7
MUSCOVITE	2.8	0.0	0.0	1.2	1.5	6.6	0.4	2.2	2.4
BIOTITE	8.8	20.2	8.4	3.3	2.8	3.1	7.0	4.0	3.0
GARNET	0.0	0.0	0.0	0.0	0.0	0.0	0.3	0.0	0.8
EPIDOTE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
CHLORITE	0.0	0.0	0.1	0.1	0.1	0.2	0.0	0.0	1.2
CALCITE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
ACCESSORIES	0.2	0.2	0.8	0.5	0.3	0.0	0.1	0.0	0.4

500 Points per section

Quartz.

Quartz is usually the most abundant mineral seen in thin section. Grain size varies from 0.2-3.0 mm. with most being from 0.5-1.0 mm. Grain shape is highly variable, grain boundaries are usually curved and sutured. In some sections the quartz crystals are slightly elongate within the mica foliation. Within the grains deformation bands and sub-grains are common but not ubiquitous.

Feldspar.

Feldspar is found in all thin sections, comprising 14.5% to 57.4% of the modal analyses in different sections (Fig.15).

Both plagioclase and K-feldspar are found, although K-feldspar is not observed in some of the thin sections. Grain size varies from 0.2-3.0 mm. Plagioclase shows multiple albite twinning. Optical determinations of the % anorthite content vary from 22% An. to 43% An. in different sections. K-feldspar often shows weak microcline twinning and can show myrmekitic and perthitic textures. Untwinned feldspar could be either K-feldspar or plagioclase. Where both feldspars are present in a thin section plagioclase always seems to be the more common.

Biotite.

Biotite occurs as laths from 0.2-4.0 mm. long exhibiting a variety of pleochroic schemes:

α = Straw yellow.

β = γ = Dark brown, blood red, dark green.

The smaller laths occur as finely disseminated crystals in the more quartzitic sections. Larger biotite laths tend to occur in the more pelitic or semi-pelitic material, often associated with muscovite laths. The biotite laths are always orientated into a well developed foliation which in some sections is highly folded and crenulated.

Muscovite.

Muscovite occurs as laths from 0.3-5.0 mm. long. There seem to be two modes of occurence, commonly muscovites are strongly alligned and associated with similarly sized biotites in the micaceous foliae within the psammites. A less common mode of occurence within the micaceous foliae, is as large unorientated and non-elongate muscovite crystals.

Other minerals.

Garnet is found occasionally, usually in the micaceous foliae. It occurs as small round crystals sometimes with small inclusions, but without good inclusion trails or zoning. Chlorite is found occasionally growing at the expense of biotite. Calcite is very occasionally found where plagioclase crystals are very highly saussuritised. Other accessory minerals include iron ore, apatite, sphene and zircon.

4.2d) Morar Division psammitic rocks.

(1) Introduction.

The term Morar Division psammite is used here to describe the psammitic and striped lithologies occurring in an area of 2-3 sq. Km. west of the Sgurr Beag slide around Ranochan (see Chapter 7). The term is not intended to cover all the psammitic rocks which outcrop in the type area around Morar-Arisaig. Figure 16 shows a lithological map of the Ranochan area

The work of Powell (1964) indicates that the lithology to the west of the mapped area is a thick psammitic unit termed the Loch Eilt Psammitic Group which he interprets as younging eastwards. Thus the Ranochan Pelite should be the eastern equivalent of the Arieniskill Pelitic Group and the Ranochan Psammitic and Striped Group the equivalent of the Arieniskill Psammitic Group (see Fig.5).

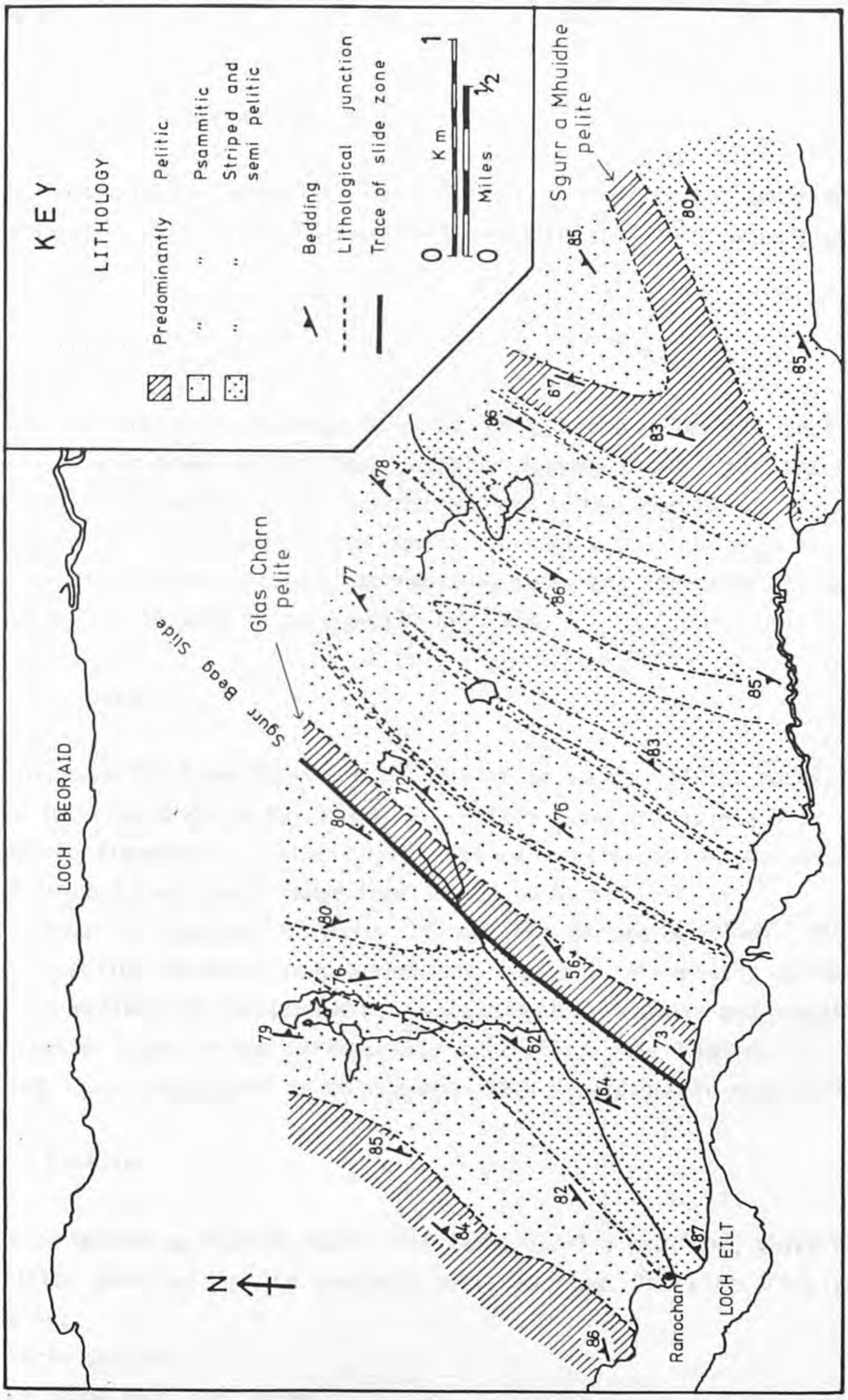
Until the area between Ranochan and Arieniskill has been mapped in detail the attempts at correlations cited above can be little more than speculation.

(2) Lithology.

Most of the rocks mapped are psammites, banded psammites and striped lithologies. Only in the extreme west did this pass gradationally into a pelitic unit (Ranochan Pelite) which served to delimit the area. Exposure is good, excellent on the higher ground where nearly continuous exposure is found. The rocks mapped as psammites are fairly uniform, banded quartzo-feldspathic rocks with very few if any micaceous or pelitic laminae. The striped group is a banded psammite and semi-pelite/pelite lithology. All samples which were collected are psammitic rather than quartzitic.

In hand specimen the rocks are well banded, cream or light grey coloured. Grain size is about 1 mm. Fine biotite laminae occur at 1-2 cm. intervals.

Figure 16.



Map to show the distribution of lithologies at the eastern end of Loch Eilt.

Some of the samples are fairly pure quartz-feldspar rocks. The specimens seem less pelitic and micaceous than the adjacent Glenfinnan Division psammites.

(3) Mineralogy and petrography.

Modal analyses are given in Figure 17, which indicate that the psammites are composed of quartz and feldspar with smaller amounts of mica plus accessory minerals.

Quartz.

Quartz is the most abundant mineral in all the psammites sectioned. Grain size varies from 0.2-3.0 mm., usually between 0.5-1.5 mm. Grain size and shape is variable, grain boundaries are often curved or irregular. There is no obvious elongation of quartz crystals within the mica foliation of some of the rocks. Internal deformation, sub-grain formation and deformation bands can be seen in many quartz crystals.

Feldspar.

Plagioclase and K-feldspar can be identified in many of the rocks. Grain size varies from 0.2-4.0 mm. Plagioclase often shows multiple albite twinning and is frequently saussuritised. Optical estimates of the anorthite content (Michel Levy test) range from An 22% to An 52%.

K-feldspar is present in most, if not all of the sections. It often shows microcline twinning and myrmekitic textures. Myrmekitic texture is not always present in sections which contain both K-feldspar and plagioclase; its presence seems to be structurally controlled (see Chapter 7). It is estimated that plagioclase is more common than K-feldspar in most sections.

Biotite.

Biotite occurs as fairly small thin laths 0.2-1.5 mm. long, which define a foliation parallel to the compositional layering. The pleochroic scheme is usually;

α = Straw yellow.

$\beta = \gamma$ = Blood red, dark green.

but α = Straw yellow.

$\beta = \gamma$ = dark brown, is also found.

Figure 17.

Modal analyses of Morar Division Psammites.

	240/1094	244/1097	267/1170	268/1171	269/1172	270/1173	271/1174	272/1175	273/1176
QUARTZ	54.2	61.8	76.8	61.2	61.9	68.3	60.3	58.4	68.2
FELDSPAR	32.8	29.4	20.8	27.6	33.8	27.7	35.6	34.9	22.3
MUSCOVITE	2.4	1.6	0.9	0.4	2.4	0.1	1.0	0.0	3.5
BIOTITE	10.6	6.6	0.7	10.1	1.9	3.6	3.1	5.2	5.5
GARNET	0.0	0.0	0.1	0.3	0.0	0.0	0.0	0.0	0.1
EPIDOTE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
CHLORITE	0.0	0.2	0.4	0.1	0.0	0.3	0.0	1.5	0.3
CALCITE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
ACCESSORIES	0.0	0.4	0.3	0.3	0.0	0.0	0.0	0.0	0.1

500 Points per section.

Biotite is more common than muscovite in most of the sections and is usually fairly evenly disseminated throughout the specimens.

Muscovite.

Muscovite crystals tend to be slightly larger than biotite crystals in the same rock, with grain sizes from 0.2-2.0 mm. Muscovite usually enhances the biotite foliation. Some muscovites, although elongated parallel to the mica foliation have cleavage strongly oblique to the foliation.

Other minerals.

Chlorite is sometimes seen growing at the expense of biotite. Accessory minerals include iron ore, sphene, garnet, epidote and clinozoisite. Not all of these minerals are found in any one section.

4.2e) Comparison of the psammitic rocks.

In this comparison it must be stressed that the term Morar Division psammite refers only to those psammites in the mapping area, west of the Sgurr Beag slide.

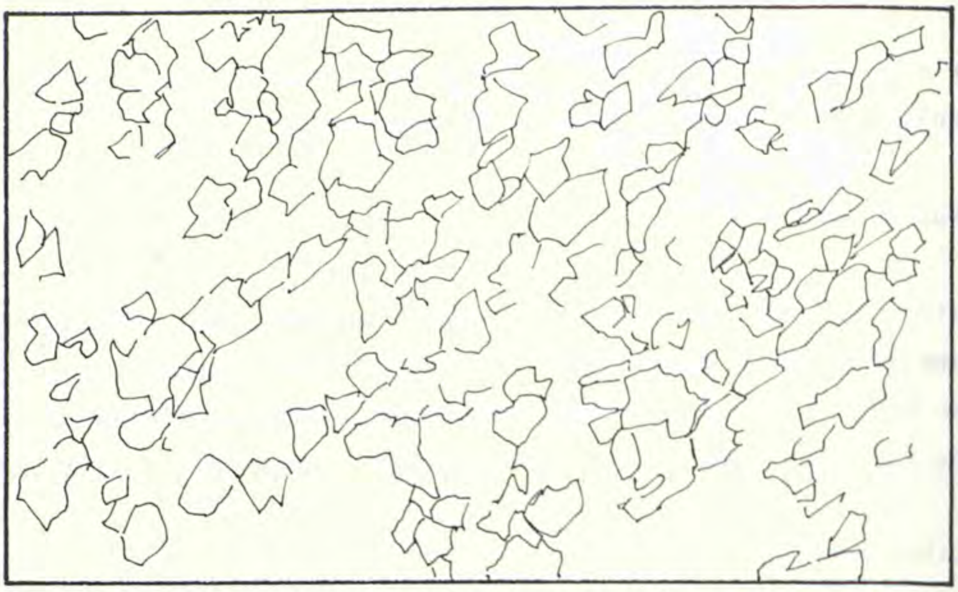
The lithological descriptions above indicate that, on a large scale, the Loch Eil Division psammites can be distinguished from the psammites of the Glenfinnan and Morar Divisions. The Loch Eil Division contains fairly uniform psammites with very few pelitic or semi-pelitic horizons. An examination of the mineralogical and petrographic descriptions and the modal analyses (Figs.12,15,17,) of the psammites indicate a number of points. Firstly, no mineral species is specific to any of the groups of psammites. Secondly, grain size, textures etc. do not serve as a means of distinguishing between the groups. Thirdly, the modal analysis of any individual section of psammite will not serve as a means of placing it in one of the psammitic groups.

Figure 18 shows photomicrographs of "typical" Loch Eil, Glenfinnan and Morar Division psammites. The diagrams in Figure 19 have been drawn to show any possible differences in modal analysis between the Loch Eil, Glenfinnan and Morar Division psammites. While no individual modal analysis would categorically place a psammite into one of the groups, some general trends can be seen.

Given the evident limitations of sampling density it seems that both the Morar and Glenfinnan Division psammites tend to contain a lower percent-

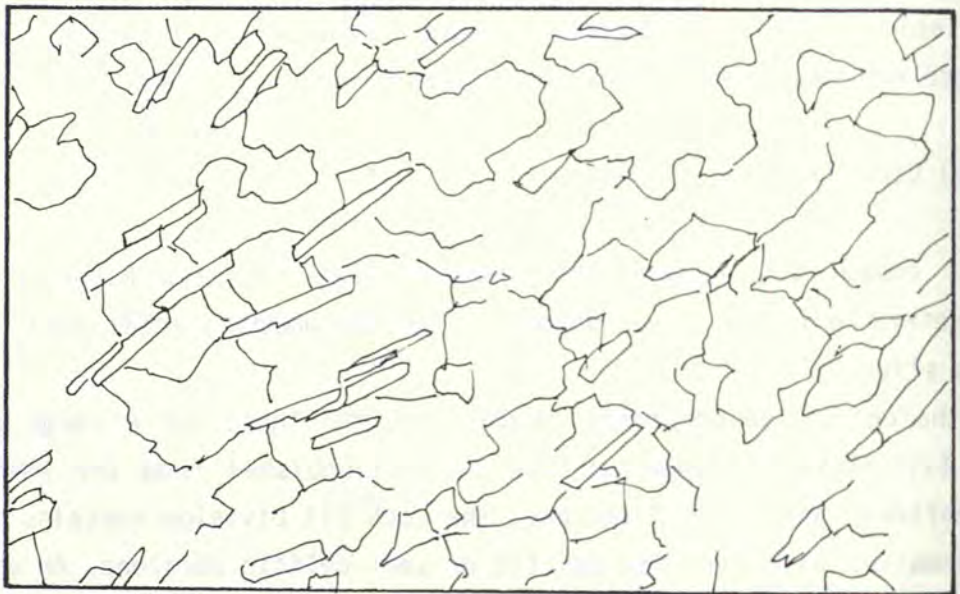
Figure 18. Photographs of "typical" Loch Eil, Glenfinnan and Morar Division psammites.

(a)



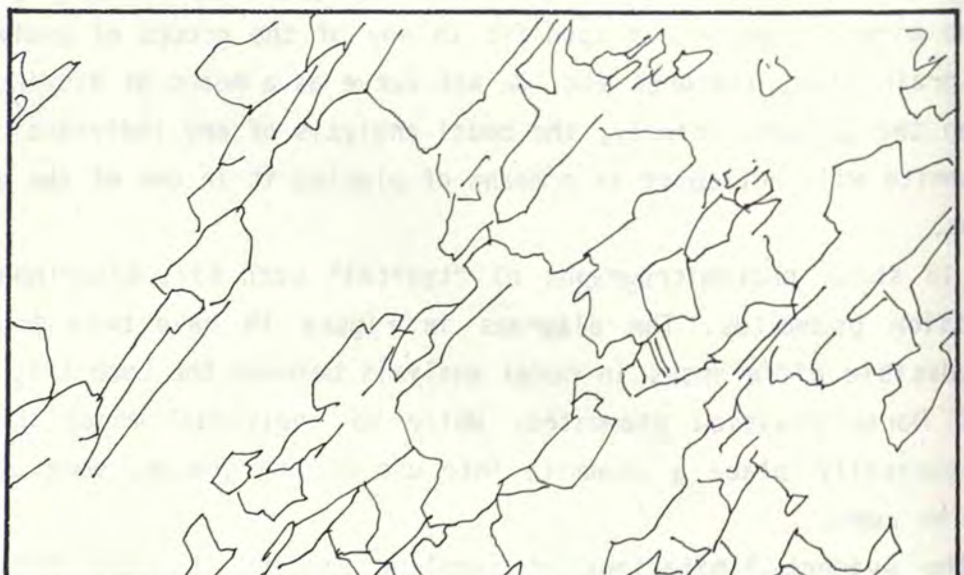
(Exp.190/1036). crossed polars. 0 mm. 1

(b)

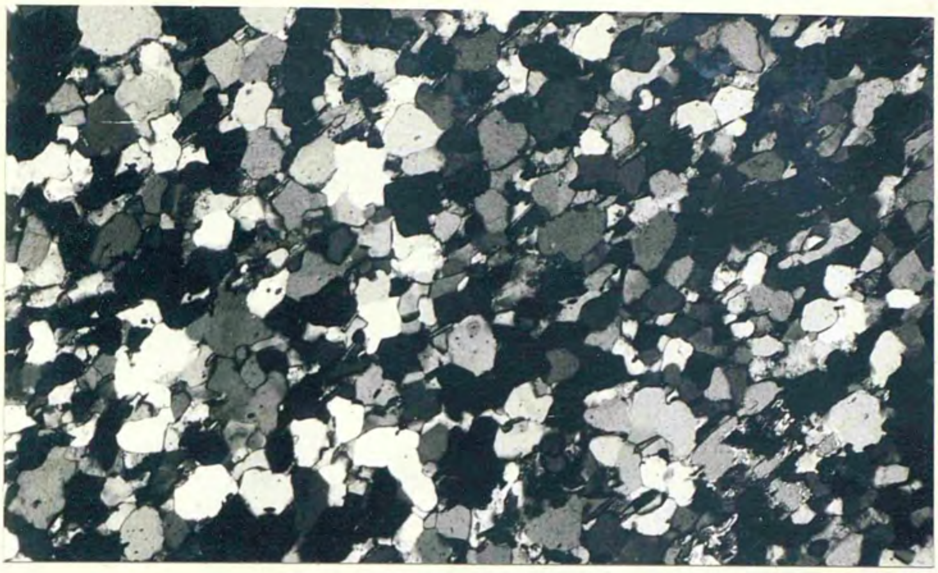


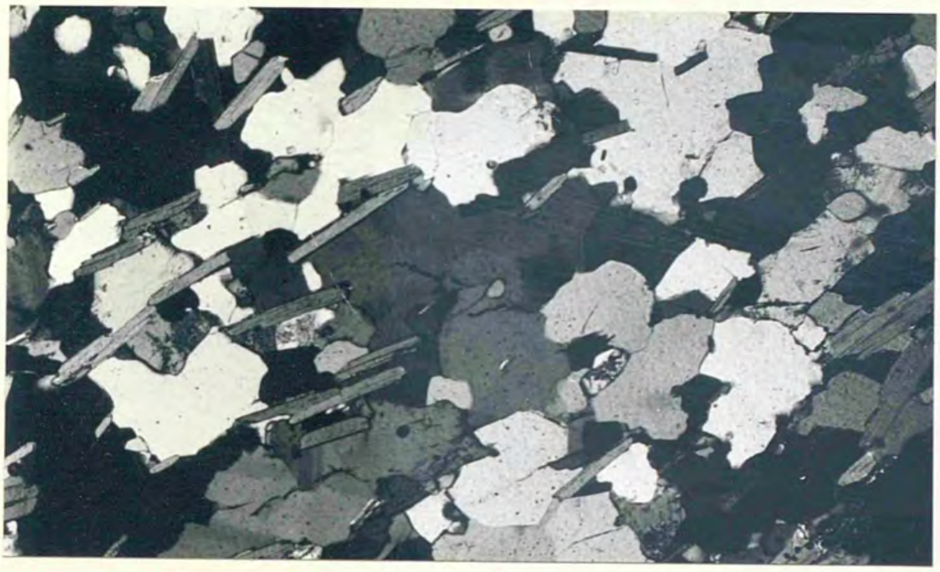
(Exp.45(b)/230). crossed polars. 0 mm. 1

(c)



(Exp.269(b)/1172). crossed polars. 0 mm. 1





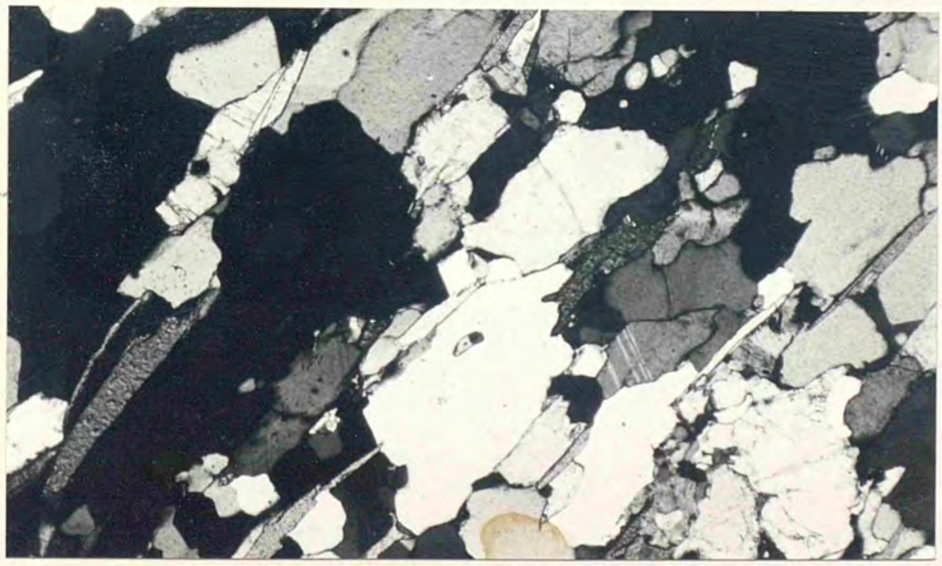
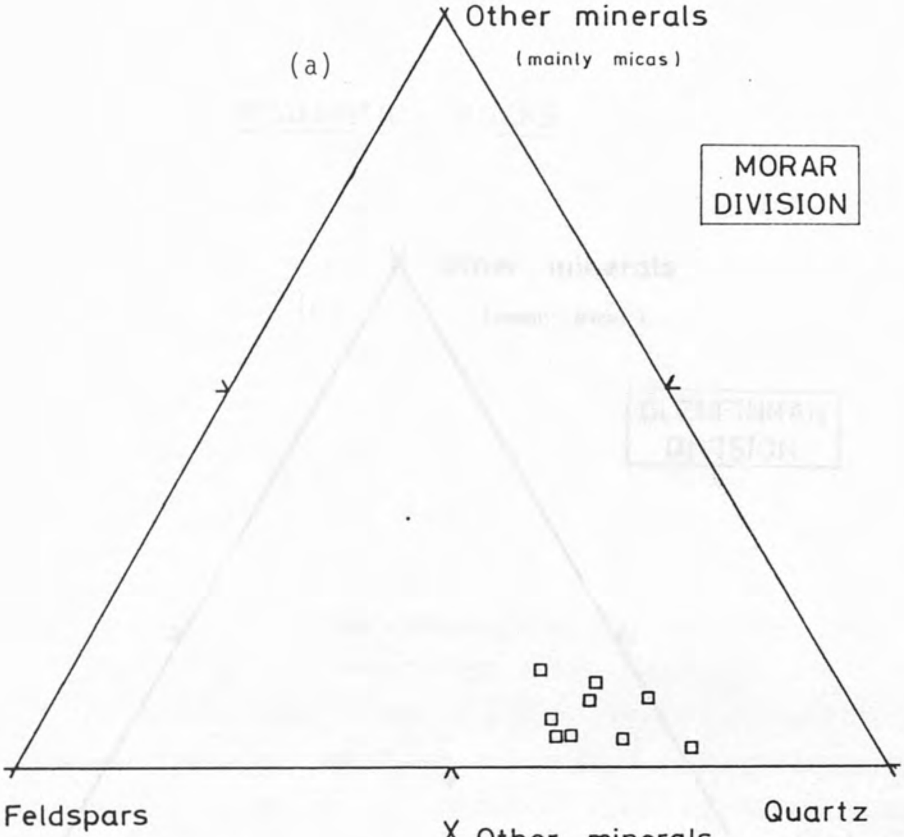


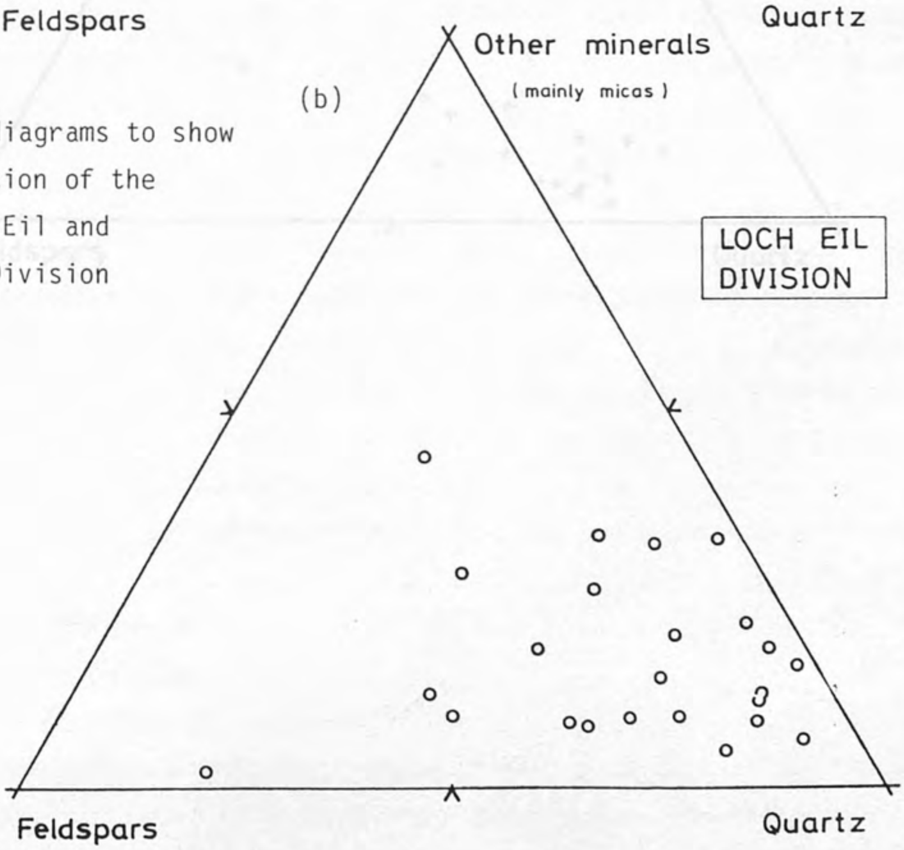
Figure 19.

PSAMMITIC ROCKS

(a)



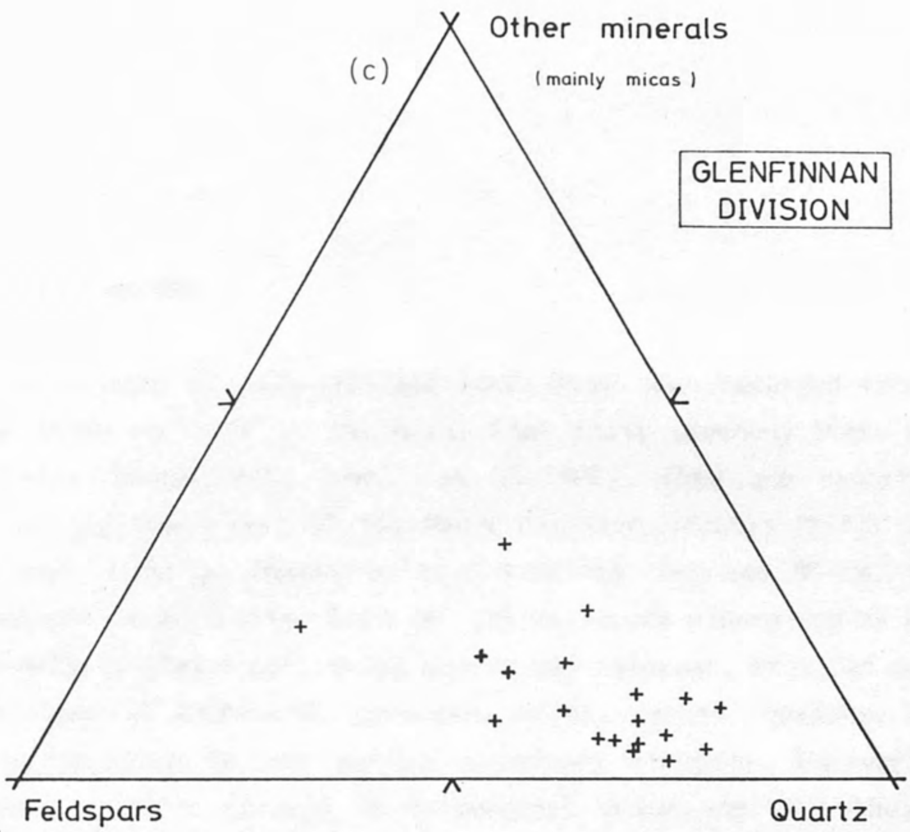
(b)



Triangular diagrams to show the composition of the Morar, Loch Eil and Glenfinnan Division psammites.

Figure 19.
(continued).

PSAMMITIC ROCKS



age of minerals other than quartz and feldspar. Within the Loch Eil Division psammites fairly pure quartzites can occur. Wide variations of the ratio of feldspars to other minerals occur within the Loch Eil Division whereas the Glenfinnan and Morar Division psammites tend towards a constant ratio of feldspar to other minerals.

Using the data from psammites collected from all three divisions of the Moine rocks it has not been possible to draw major conclusions relating to the sedimentology, sediment source, sediment maturity etc.

4.3) Calc-silicate rocks.

4.3a) Introduction.

Sporadic occurences of calc-silicate rocks have been recorded from many areas of the Moine rocks NW of the Great Glen fault (Kennedy,1949; Winchester,1972,1974a; Tanner,1976; Powell et al.1981). They are reported to be absent from the lower part of the Morar Division (Powell,1974). In the area mapped they occur as lenses up to 3-4 metres long and 30 cm. thick, although thickness is more often 5-10 cm. Calc-silicate mineralogy is highly variable; as well as always containing quartz and feldspar, they can contain smaller quantities of amphibole, pyroxene, micas, garnet, epidote, clinozoisite, chlorite, calcite and various accessory minerals. The variation in mineralogy is partly related to metamorphic grade and thus they have been used in attempts to define metamorphic grade in areas which otherwise are lacking in lithologies which develop minerals usable as metamorphic grade indicators. Kennedy (1949) divided the Morvern-Knoydart area (see Chapter 6,Fig.59) into four zones identified by the index minerals listed as follows in a prograde metamorphic sequence:

1) Zoisite-(calcite)-biotite zone.
2) Zoisite zone.
3) Anorthite-hornblende zone.
4) Anorthite-pyroxene zone.

Winchester (1972) used the geochemical ratio of CaO/Al_2O_3 at which biotite is replaced by hornblende to produce an "isograd" map of the Fannich Forest area. Powell et al.(1981) have used plagioclase anorthite content in calc-silicates as an indicator of metamorphic grade.

In the area mapped a total of 32 calc-silicates were found and collected for petrological and geochemical study. Calc-silicates from all three divisions were collected and no striking differences were noted that might relate to stratigraphical or spatial controls.

In the Ranochan area of eastern Loch Eilt (Fig.1) the metamorphic reactions and textures observed in the calc-silicates togather with structural mapping have been used to elucidate the structural and metamorphic history of the Sgurr Beag slide (Chapter 7).

4.3b) Lithology.

All the calc-silicates were collected from psammitic lithologies, mostly from fairly thick homogeneous psammites or occasionally from the psammitic portions of the striped lithologies. This contrasts with the observations of Winchester (1972.p,405) who reported that the calc-silicates from Fannich Forest are normally associated with dark biotite schists.

The calc-silicate layers are always parallel to bedding planes in the surrounding psammites. Laterally calc-silicates grade into psammites. The change from calc-silicate to psammite across the bedding plane is abrupt.

From these observations it appears that the calc-silicates in the mapping area are more likely to have been slightly calcareous sedimentary lenses rather than post- or late-diagenetic concretions (in concretions one might expect to find abrupt lateral boundaries). However Tanner (1976.p,100) suggests that calc-silicates probably originated as late diagenetic calcareous concretions and Dalziel (1963a) has described rare calc-silicates which have grown across the foreset laminations of current bedded units.

In hand specimen the calc-silicates are distinctive and quite variable. They are usually white, light grey or cream coloured, mottled with rusty brown garnets up to 5 mm. in diameter and thin green amphibole laths up to 10 mm. long. The amphibole laths define an obvious foliation which is often slightly oblique to the bedding planes. Most of the samples are fairly homogeneous across the bedding planes,but some samples eg, 37/229 and 207/1068 are banded, containing a central band and equally thick margins. Sample 37/229 has a feldspar-amphibole-quartz core mantled by quartz-garnet-clinozoisite margins. Sample 207/1068 has a quartz-garnet-clinozoisite core mantled by feldspar-amphibole-quartz margins.

Modal analyses of the calc-silicates are given in Figure 20. These analyses were compiled by point counting across areas which were thought to be representative of the complete thickness and average composition of the calc-silicate lenses.

Figure 20.

Modal analyses of Calc-silicate rocks.

	34/229	36/229	37/229	40/229	103/440	104/451	105/451	119/519
QUARTZ	79.1	75.3	55.6	77.5	79.0	37.3	42.7	44.3
PLAGIOCLASE	10.4	18.4	22.8	5.5	15.2	18.7	18.8	32.3
AMPHIBOLE	0.0	4.0	8.4	1.2	0.0	10.9	13.1	5.0
BIOTITE	0.0	0.0	0.0	0.0	0.4	0.0	0.0	0.0
MUSCOVITE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
GARNET	0.1	0.0	7.2	0.3	0.0	4.1	1.9	3.4
PYROXENE	0.0	0.0	1.9	1.3	0.0	0.0	0.0	0.0
CALCITE	0.0	0.0	0.4	0.7	0.0	0.0	0.0	0.0
EPIDOTE	0.0	0.0	0.0	0.0	5.2	0.0	0.0	0.0
CLINOZOISITE	9.9	1.5	2.5	13.2	0.0	26.8	21.6	11.7
CHLORITE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	1.7
ACCESSORIES	0.5	0.0	1.2	0.3	0.0	2.2	1.9	1.6

	131/573	134/582	48/231	81/250	167/883	170a/948	170b/948	174/956
QUARTZ	51.8	56.4	83.8	68.2	48.9	52.8	47.8	70.9
PLAGIOCLASE	36.9	12.2	8.3	24.2	26.5	33.0	29.7	15.7
AMPHIBOLE	2.5	0.0	1.7	0.4	10.2	7.3	5.6	5.1
BIOTITE	0.0	0.0	0.1	0.9	0.0	0.2	0.1	0.0
MUSCOVITE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
GARNET	3.8	0.0	0.3	0.0	3.7	1.8	12.5	6.5
PYROXENE	0.0	0.0	0.1	0.0	0.0	0.0	0.0	0.0
CALCITE	0.3	0.2	0.1	0.0	0.0	0.0	0.0	0.0
EPIDOTE	0.0	30.6	0.0	0.0	0.0	0.0	0.0	1.2
CLINOZOISITE	3.4	0.0	5.4	5.4	9.3	1.9	3.4	0.0
CHLORITE	0.0	0.2	0.0	0.0	0.4	0.0	0.0	0.0
ACCESSORIES	0.6	0.6	0.2	0.9	1.0	3.0	0.9	0.6

	175/961	187/1031	200/1050	204/1065	207/1068	208/1071	212/1087	213/1087
QUARTZ	43.5	66.0	63.3	60.5	50.4	53.4	58.2	59.3
PLAGIOCLASE	37.9	23.1	26.5	28.0	36.0	20.0	32.9	29.1
AMPHIBOLE	9.5	3.6	2.7	5.6	1.9	3.7	0.1	0.1
BIOTITE	1.1	0.1	0.0	2.5	0.0	0.0	0.0	1.7
MUSCOVITE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.7
GARNET	4.5	3.7	4.7	1.2	5.7	16.4	2.1	3.1
PYROXENE	0.0	0.0	0.4	0.0	0.0	1.6	0.0	0.0
CALCITE	0.1	0.8	0.0	0.0	2.0	0.0	0.0	0.4
EPIDOTE	0.5	0.0	0.0	0.0	0.0	0.0	0.0	0.9
CLINOZOISITE	0.0	1.7	0.5	0.4	1.3	0.5	1.3	0.0
CHLORITE	1.3	0.3	0.3	0.5	1.5	1.7	3.5	3.1
ACCESSORIES	1.6	0.7	1.6	1.3	1.2	2.7	1.9	1.6

	215/1088	225/1091	252/1110	253/1115	254/1119	255/1123	256/1126	259/1145
QUARTZ	54.3	52.3	51.3	70.2	47.9	74.2	69.5	56.2
PLAGIOCLASE	36.7	37.9	41.1	24.8	36.7	20.3	21.2	33.1
AMPHIBOLE	0.7	0.1	2.8	0.3	6.7	0.3	0.4	3.5
BIOTITE	1.3	4.4	0.7	0.3	0.0	0.0	0.0	0.0
MUSCOVITE	1.4	1.7	0.5	0.3	0.3	0.3	0.2	0.6
GARNET	2.8	0.8	0.8	0.7	6.6	2.0	4.0	2.6
PYROXENE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
CALCITE	0.0	0.4	0.0	0.1	0.0	0.0	0.0	0.0
EPIDOTE	0.0	0.0	0.0	2.5	0.0	0.0	2.1	0.0
CLINOZOISITE	0.1	0.0	0.5	0.0	0.4	1.3	0.0	2.0
CHLORITE	1.9	1.5	0.7	0.3	0.9	1.1	2.0	1.1
ACCESSORIES	0.8	0.9	1.6	0.5	0.5	0.5	0.6	0.6

1000 Points per section.

4.3c) Mineralogy and petrography.

Quartz.

Quartz is the most abundant mineral in all the calc-silicates (Fig.20), comprising 37.3% to 83.8% of each of the samples. Grain size varies from 0.1 to 3.0 mm. with most sections having grain size variations from about 0.5 to 1.5 mm. Some of the sections contain equidimensional quartz grains whereas others often show a well developed quartz shape fabric (see Fig.21a) with aspect ratios of up to 3:1. In some sections quartz crystals contain internal deformation bands whereas other sections contain quartz crystals with no apparent internal deformation. In sections with deformation bands it tends to be the larger crystals which show the internal deformation. Occasionally sections show the development of equidimensional sub-grains rather than deformation bands within the quartz crystals. Some sections show quartz crystals with a weak crystallographic "c" axis orientation (as seen using a sensitive tint plate, eg.Fig.21b). Sections showing prefered crystallographic orientations are most commonly, but not exclusively, found in the Ranochan area (see Chapter 7: fabric development around the Sgurr Beag slide).

There is no correlation between the calc-silicates showing; internal deformation, quartz shape fabrics, quartz crystallographic fabrics, or calc-silicates collected from the psammites of the different divisions of the Moine schists. Figure 22 is a table showing subjective estimates of quartz textures.

Feldspar.

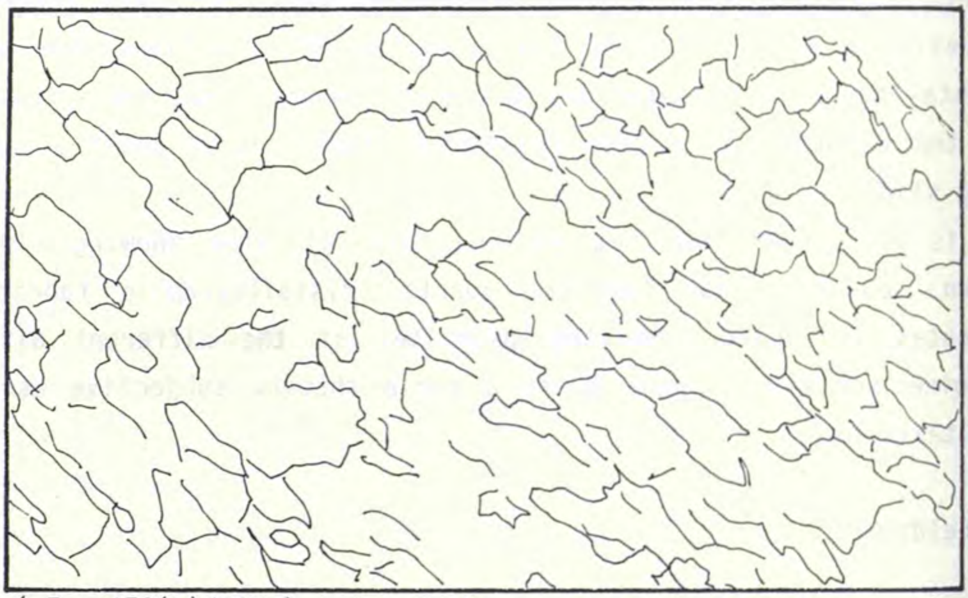
Feldspar and quartz are the only two minerals which were found in all of the calc-silicates. Feldspar is usually much less abundant than quartz. Relatively small percentages of feldspar seem to coincide with relatively large amounts of clinozoisite (see Fig.20).

Grain size varies from 0.1 to 1.5 mm. with average sizes from 0.5 to 1.5 mm. Feldspar crystals often occur as aggregates in which, when highly altered, it is difficult to see grain boundaries. In most sections the feldspar crystals do not show any preferred orientation. however in the Ranochan area, in some of the sections the elongate shapes of feldspar crystals are alligned parallel to the quartz shape fabric.

All the feldspar is plagioclase. In some heavily altered sections it is difficult to see good albite twins. Plagioclase composition ranges from

Figure 21a.

Photomicrograph to show the quartz crystal shape fabric in a calc-silicate.



(Exp.170(b)/948). plane polars. 0 mm. 1

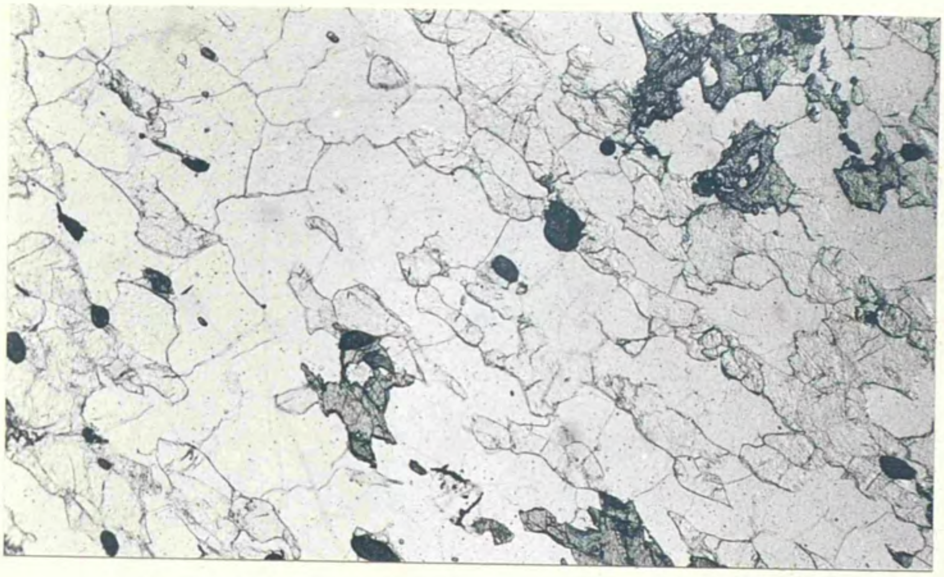
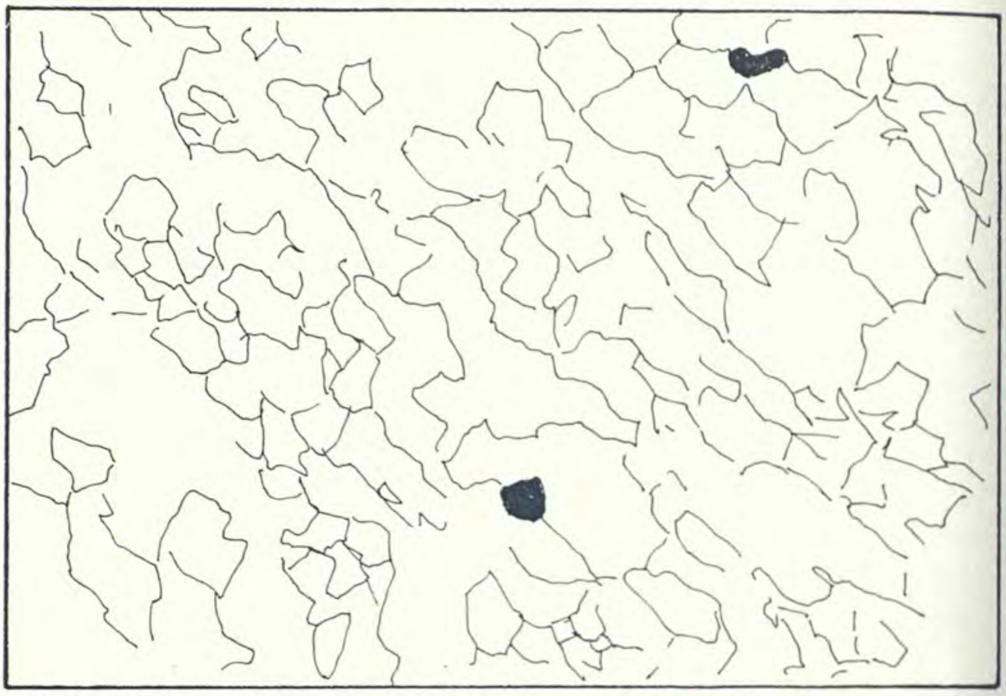


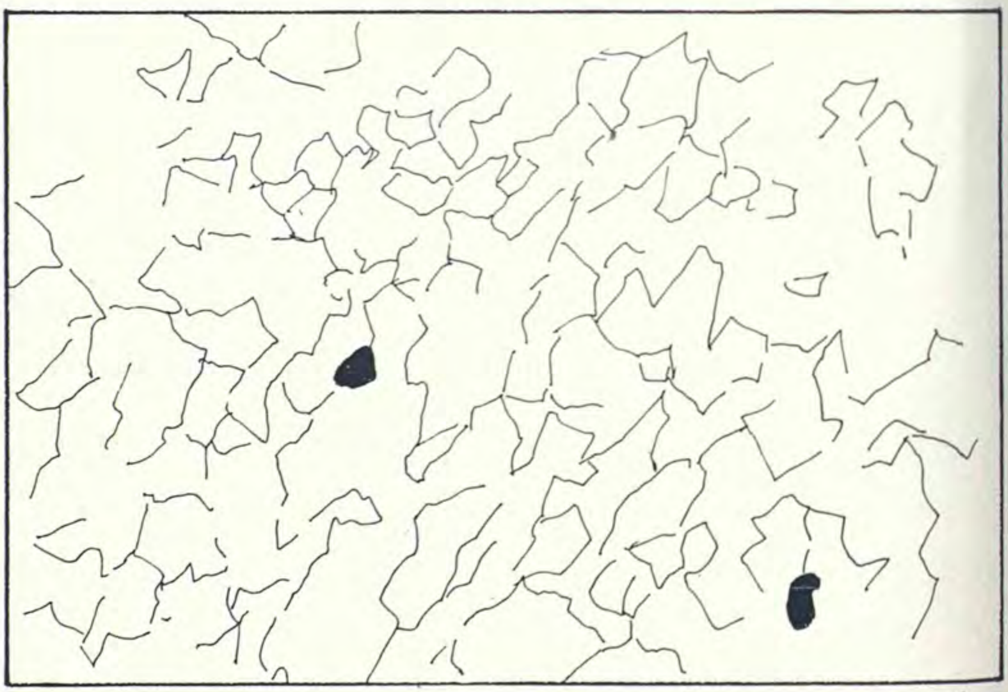
Figure 21b. Photomicrographs of the quartz crystallographic C-axis fabric in a calc-silicate (Exp.204(b)/1065).



Cross polars plus gypsum plate

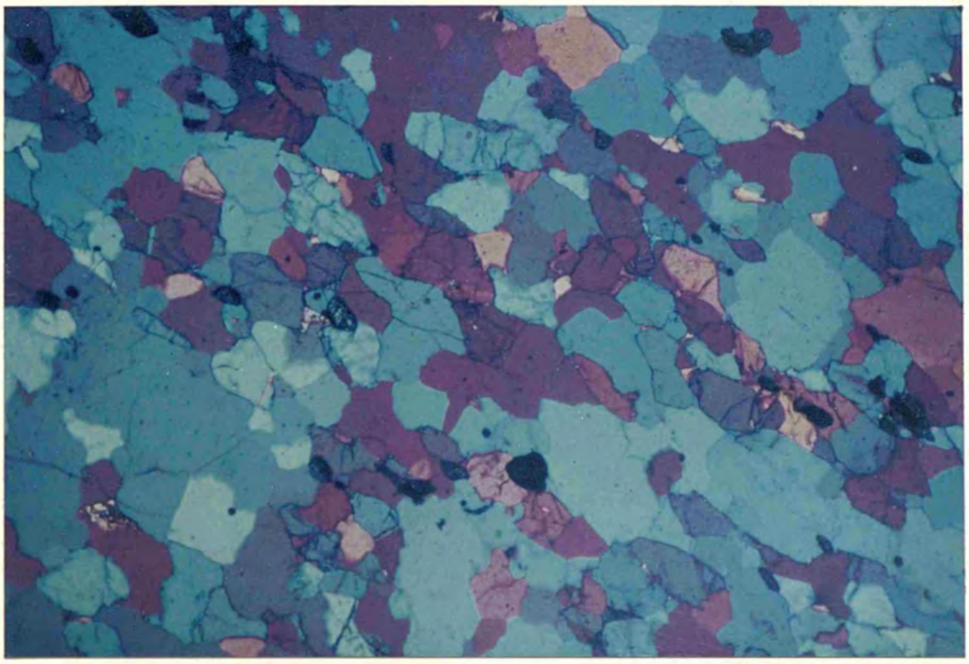
0 mm. 1

Same view, with rotation of 90° on the microscope stage.



Crossed polars plus gypsum plate.

0 mm. 1



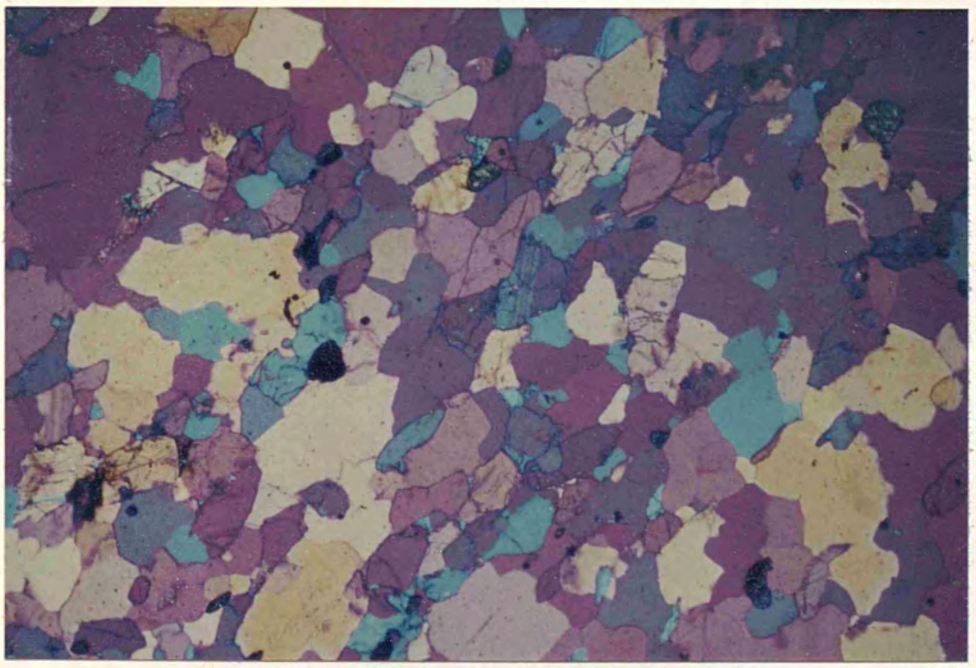


Figure 22.

Table to show the qualitative development of quartz deformation textures in calc-silicate rocks.

Calc-Silicates: Quartz textures.

Div.	Sample	Grain size		Shape fabric			Apparent cryst. fabric			Internal deformation		
		min. (mm)	max. (mm)	NONE	WEAK	STRONG	NONE	WEAK	STRONG	little or none	weak def. bands	occas. sub grains
? LE	34/229	0.5	3.0	—			—				—	
? LE	36/229	0.5	3.0	—				—			—	
? LE	37/229	0.1	3.0	—			—				—	
? LE	40/229	1.0	2.0	—				—			—	
? LE	48/231	1.0	2.0			—			—	—	—	
Gf	81/250	0.5	4.0			—		—		—	—	
LE	103/440	0.3	2.0	—			—			—		
LE	104/451	0.1	1.0	—			—			—		
LE	105/451	0.1	1.0	—			—			—		
LE	119/517	0.3	1.0	—			—			—		
LE	131/573	0.3	1.0		—			—			—	
LE	134/582	0.5	2.0		—		—			—		
Gf	167/883	0.3	1.5		—				—	—		
Gf	170/948	0.5	1.5			—			—	—	—	
Gf	174/956	0.3	1.5	—				—		—	—	
Gf	175/961	0.2	1.0	—			—			—	—	
Gf	187/1031	1.0	2.0	—				—			—	
Gf	200/1050	0.3	2.0		—				—		—	
Gf	204/1065	0.3	1.5		—				—		—	
Gf	207/1068	0.3	1.5	—					—			—
Gf	208/1071	0.2	1.5	—				—			—	—
Gf	212/1087	0.5	1.5		—			—			—	
Gf	213/1087	0.4	1.0	—				—			—	
Gf	215/1088	0.5	2.0		—			—			—	—
Gf	225/1091	0.5	2.0	—				—			—	
Morar	252/1110	0.2	1.0		—				—		—	
Gf	253/1115	0.3	2.0		—				—		—	
Morar	254/1119	0.2	1.0		—			—			—	—
Morar	255/1123	0.2	2.0			—			—		—	
Morar	256/1123	0.3	2.5		—				—		—	—
Morar	259/1145	0.3	3.0	—				—			—	—

an anorthite content of 42% to 97% (see Appendix 2). An. content was determined optically using the Michel-Levy technique and care was taken to measure the angle X^C at high An. values, since these angles can exceed 45°. The lower values of anorthite content were recorded from sections containing very highly altered or poorly twinned plagioclase. The An. content of the calc-silicates in the Ranochan area is considered more fully in Chapter7.

The plagioclase crystals show highly variable amounts of alteration. Some sections are only very slightly saussuritised whilst, more commonly, in others the feldspars are all or nearly all totally altered to fine grained light brown saussurite (eg.section 134/582). Occasionally relatively large sericitic mica and calcite crystals can be seen. Saussuritisation of plagioclase is found in calc-silicates from all three litho-stratigraphical divisions in the area.

Occasionally the plagioclase has a texture resembling an exsolution or symplectic intergrowth texture; both portions of the symplectite being plagioclase (see Fig.23). Where this texture, which is developed in calc-silicates from all three divisions, is found it is common to find vermicular intergrowths of clinozoisite associated with the feldspars. The significance of this texture is discussed in Chapter 6.

There is no spatial relationship between feldspars which are saussuritised and those which show symplectic textures.

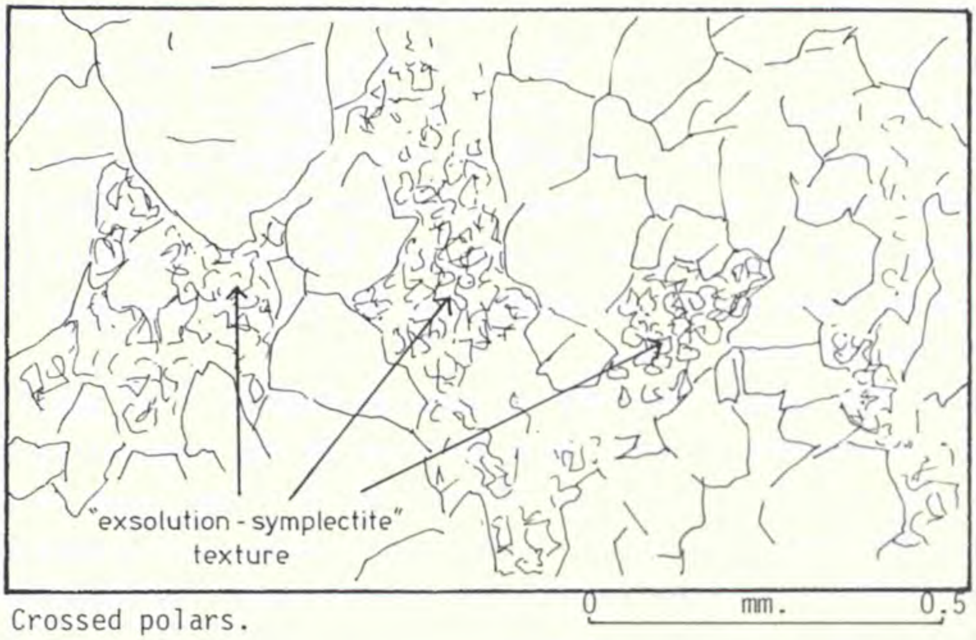
Epidote-Clinozoisite.

Relatively small quantities of clinozoisite or epidote (for a summary of optical properties see Fig.24) were found in all of the sections with the exception of section 225/1091. Clinozoisite is much more common than epidote. No zoisite was found in any of the sections (see Fig.20).

Clinozoisite is most commonly found associated with plagioclase, in which it is usually embedded. It always occurs as anhedral crystals or vermicular intergrowths, never as euhedral crystals. Grain size varies from small worms 0.1 mm. long to crystals 3 mm. It has already been noted that clinozoisite is often found associated with two-plagioclase symplectites (Fig.25b Across some sections there is a gradation from plagioclase through plagioclase/clinozoisite into clinozoisite (eg.Fig.25a, section 34/229). The large clinozoisite crystals contain bands of crystallographically differently orientated material. The texture can be interpreted as reflecting growth at the expense of plagioclase containing albite twins (see Chapter6). Clinozoisite is also found intergrown with poikiloblastic garnets. Epidote was recorded from sections 212/1087 and 213/1087 where it is found assoc-

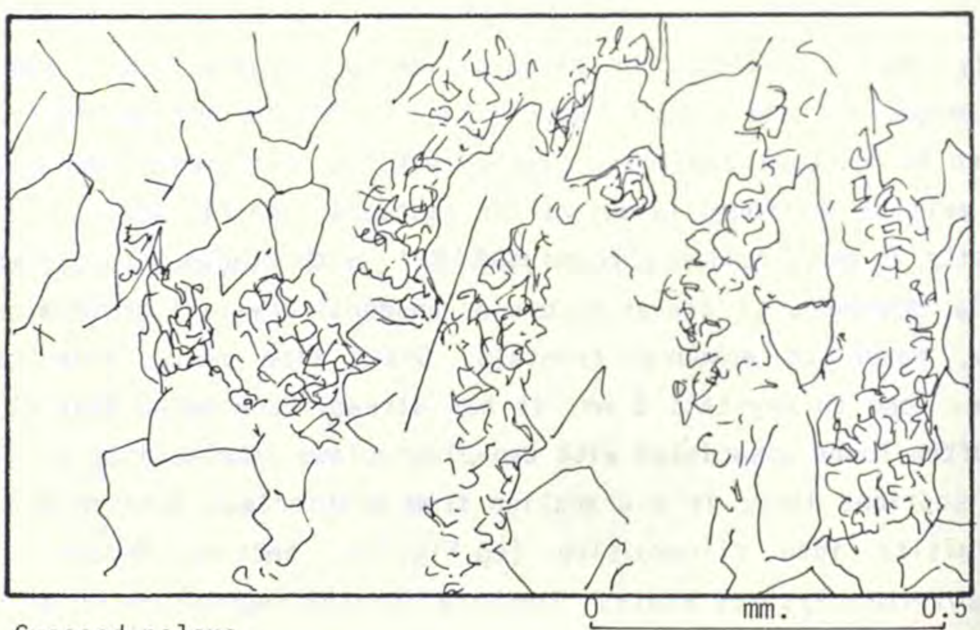
Figure 23. Photomicrographs to show plagioclase "exsolution-symplectite" texture in a calc-silicate. (Exp.167(b)/883).

(a)



Crossed polars.

(b)



Crossed polars.

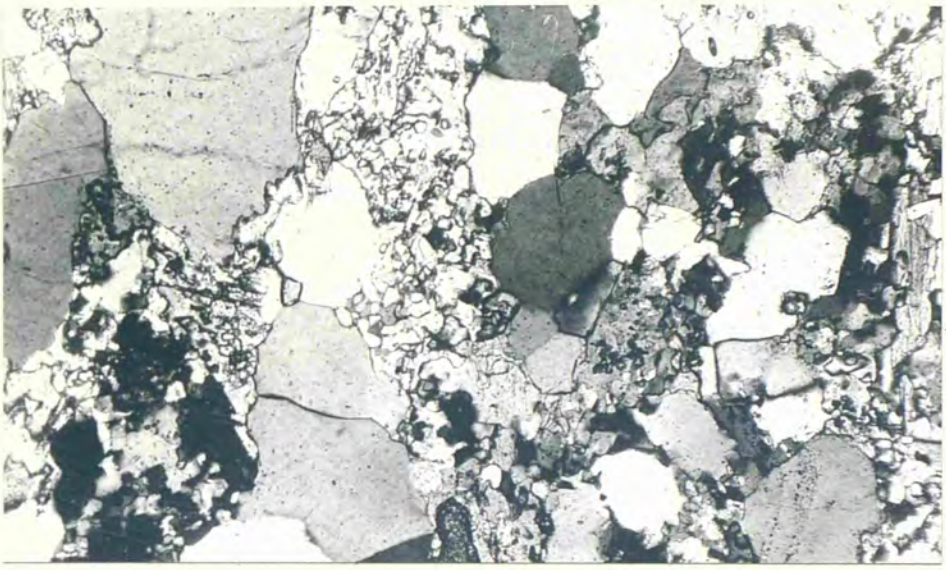


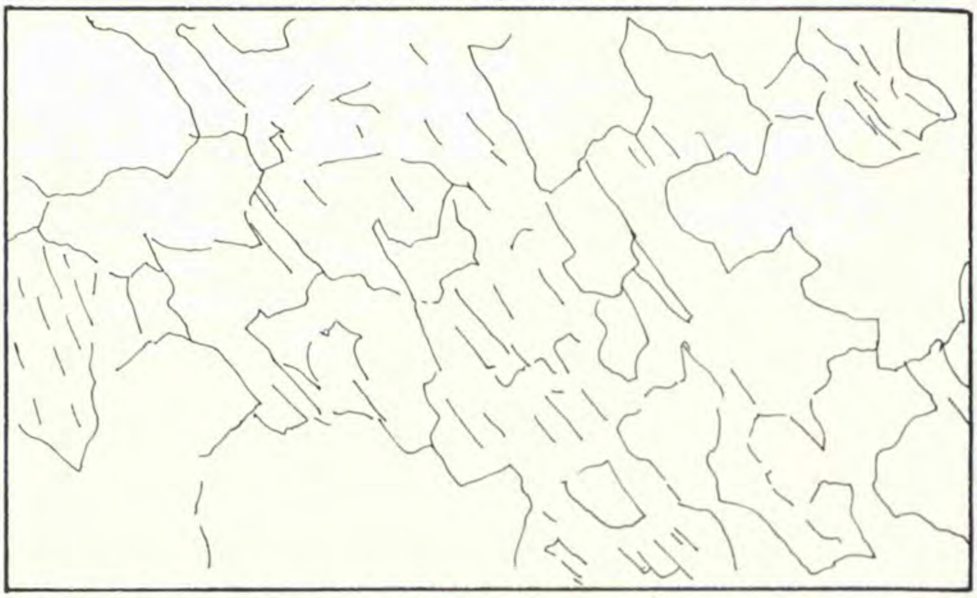
Figure 24.

	CLINOZOISITE		EPIDOTE
Composition	$Ca_2 Al Al_2O OH (Si_2O_7)(SiO_4)$	$\leftrightarrow$	$Ca_2Fe^{3+}Al_2O OH(Si_2O_7)(SiO_4)$
Colour in thin section	Colourless	$\longleftrightarrow$	pale green / yellow
Birefringence	0.005	$\longleftrightarrow$	0.049
$2V_\gamma$	14°	$\longleftrightarrow$ 90° $\longleftrightarrow$	116°
Cleavage	very weak	$\longleftrightarrow$	often well developed

Some optical properties of the solid solution series
Clinozoisite —— Epidote .

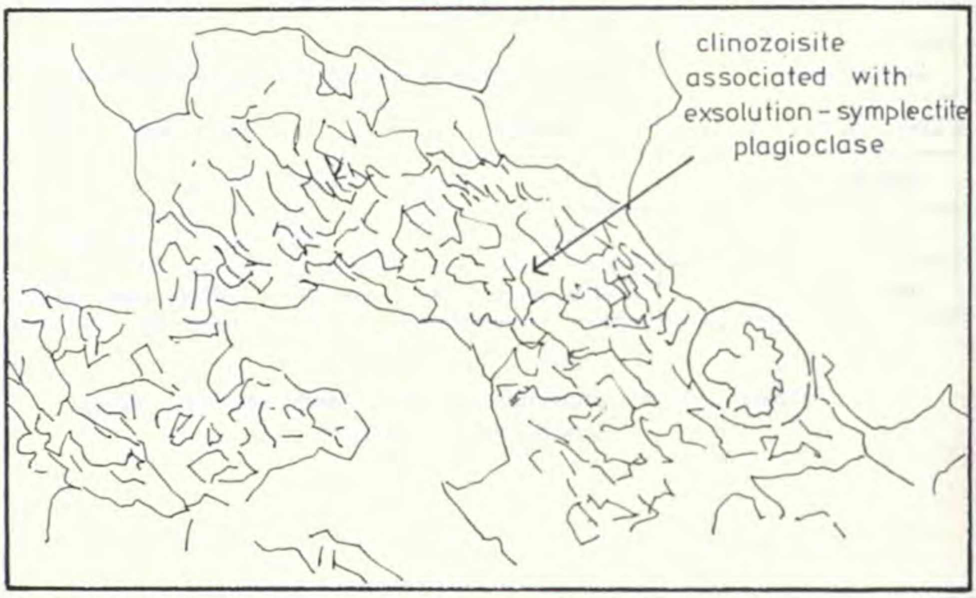
Figure 25. Photomicrographs of Epidote-Clinozoisite textures in calc-silicates (see text for discussion).

(a)



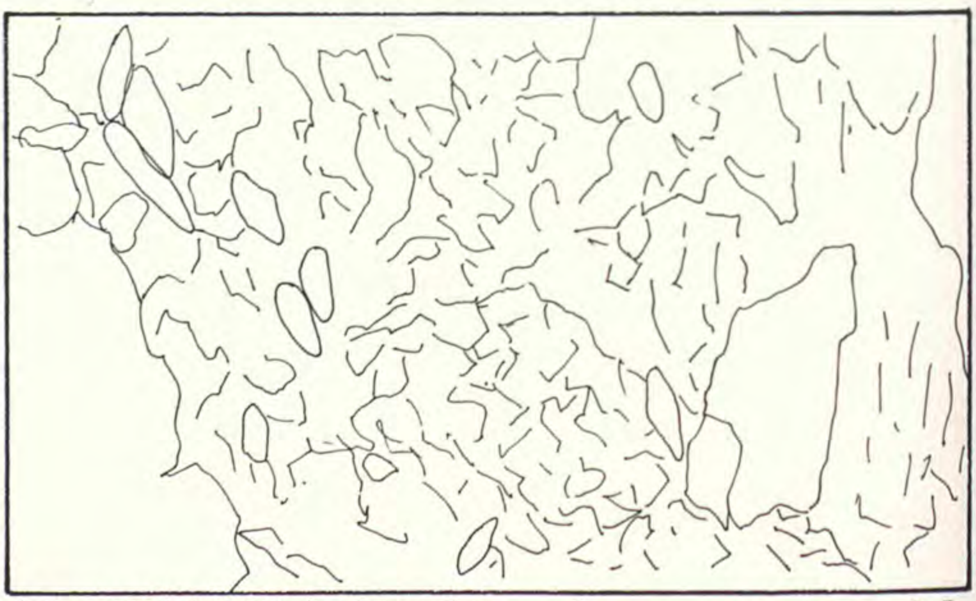
(Exp.34/229).crossed polars. 0 mm. 1

(b)



(Exp.167(b)/883). plane polars. 0 mm. 0.5

(c)



(Exp.259(c)/1145). plane polars. 0 mm. 0.5

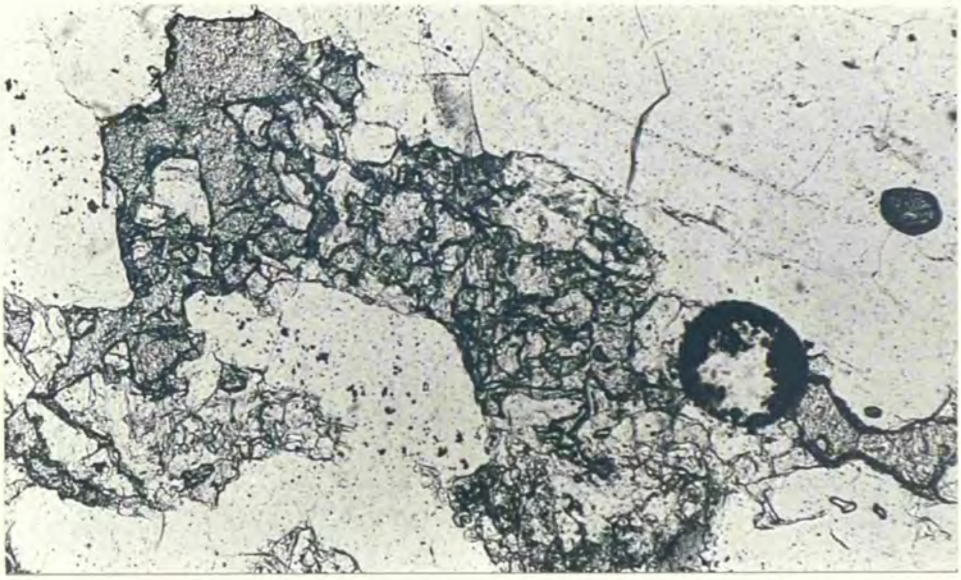
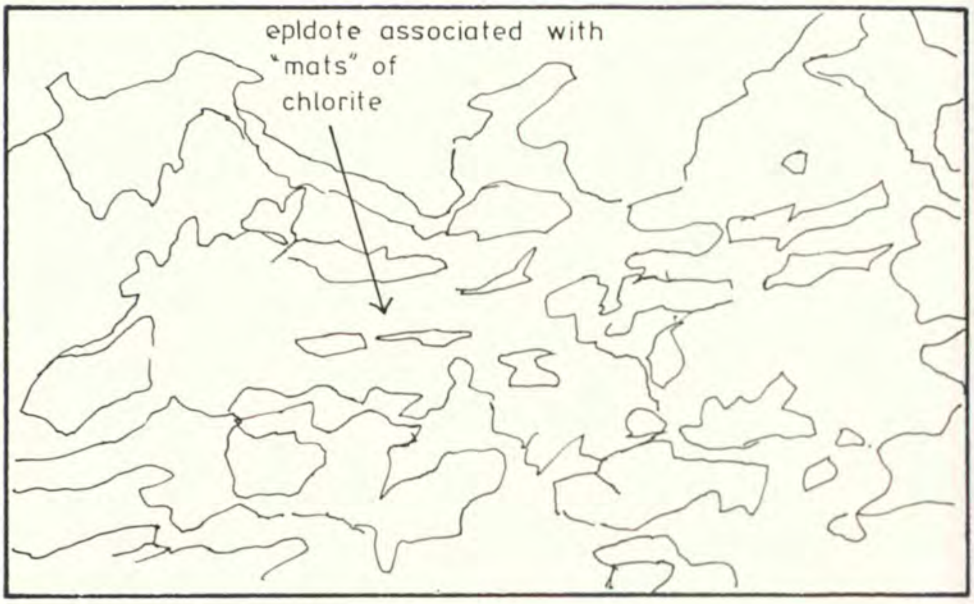


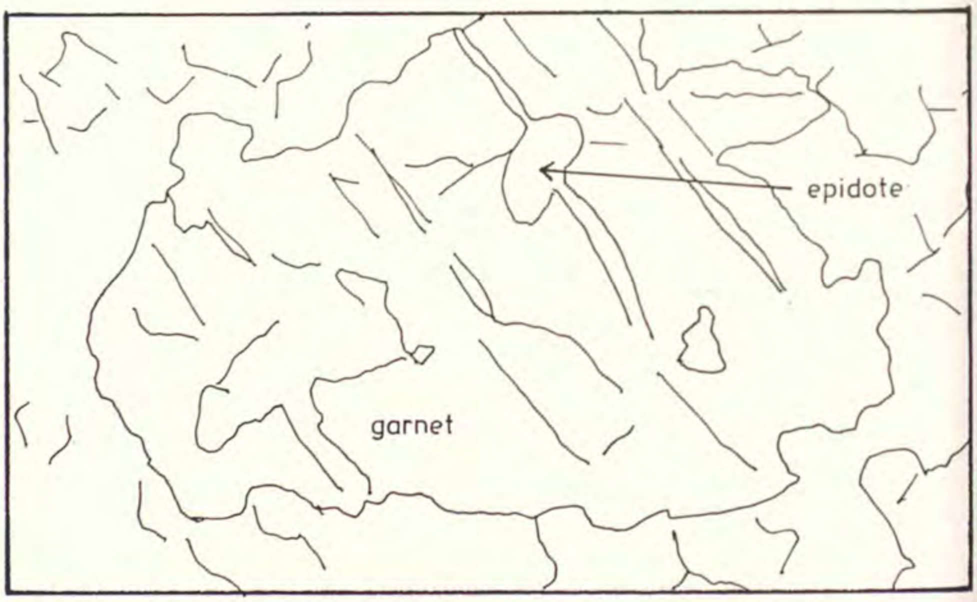
Figure 25 (cont). Photomicrographs of Epidote-Clinozoisite textures in calc-silicates (see text for discussion).

(d)

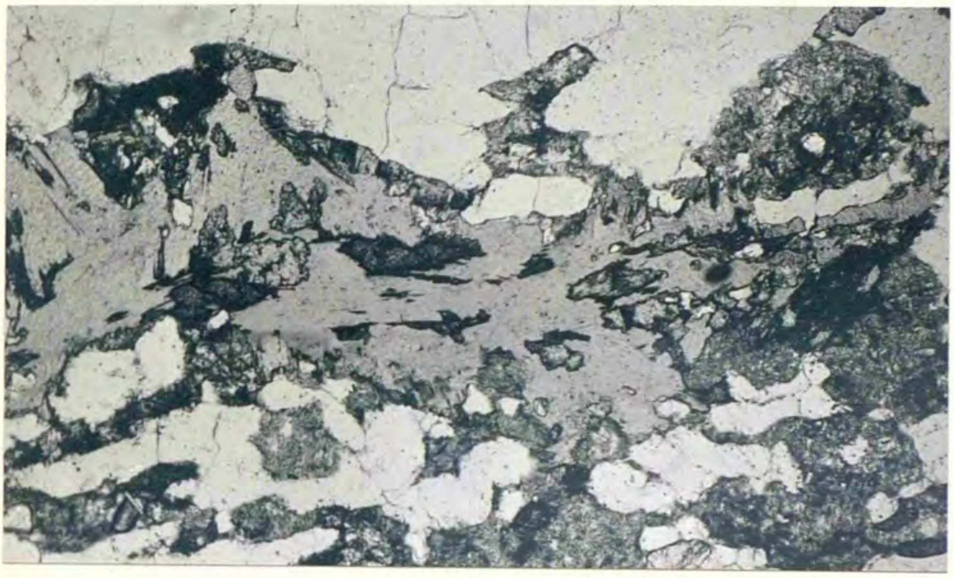


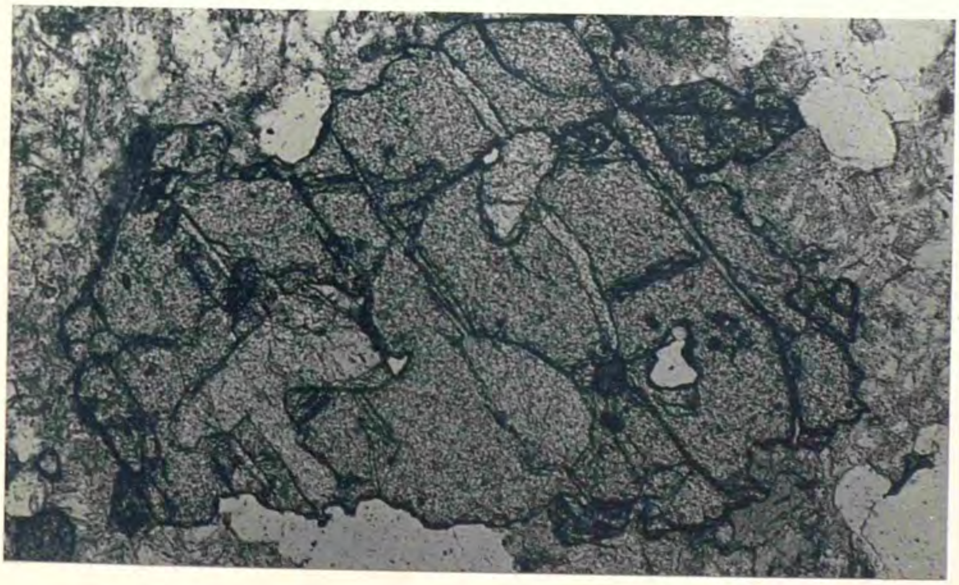
(Exp.212/1087). plane polars. 0 mm. 1

(e)



(Exp.213/1087). plane polars. 0 mm. 0.5





iated with mats of chlorite (Fig,25d) and also intergrown with garnet (Fig.25e).

Optical identification of Epidote Group Minerals.

Zoisite, clinozoisite and epidote have all been recorded from Moinian calc-silicates. On textural evidence some of these have been regarded as minerals from low grade progressive assemblages whilst others have been regarded as products of retrogressive mineral reactions (see Powell et al.1981). If such textural and mineralogical differences exist then it becomes important to be able to identify the minerals with some certainty.

In all the sections examined zoisite was not found. Clinozoisite-epidote is a nearly complete solid solution series (Raith,1976) and some of the properties of this series are listed below (Fig.24).

An increase of $2V_\gamma$ in clinozoisite corresponds to a change in highest observed birefringence colours. These change from; anomalous blue/grey-blue/yellow- lemon yellow- yellow/pink- pink/purple. This change of birefringence and $2V_\gamma$ is correlated with an increase of Fe^{3+} in the lattice at the expense of Al^{3+} . The change from clinozoisite to epidote has been set arbitrarily at $2V_\gamma$ 90°. In the sections examined there seems to be a range of $2V_\gamma$ from about 40°-50° to 100°-110°.

Clinozoisite shows strong dispersion, particularly noticeable near the extinction position which is consequently diffuse. In sections of calc-silicates from the Arisaig area, zoisite crystals show similar birefringence to co-existing clinozoisite crystals but they always show sharp straight extinction and never show dispersion (monoclinic crystals show a variety of dispersive features whereas orthorhombic crystals do not).

Amphibole.

Amphibole is a common component of most of the calc-silicates (see Fig.20). Grain size varies from 0.2 to 3 mm.,most sections have grain sizes ranging from 0.5 to 1 mm. Crystal shapes are anhedral, occasionally elongate, defining a foliation which is often slightly oblique to the bedding in the sample, but more often the grain shape is fairly equidimensional. Some crystals contain many inclusions of small crystals which are usually slightly smaller than the quartz crystals in the groundmass. A few sections contain some amphibole crystals which are more acicular in shape and it is possible that these are a different generation of amphibole (see Charnley, 1976).

The amphiboles have a characteristic pleochroic scheme;

α= pale straw.

β= pale/mid green.

γ= pale mid green/blue.

The extinction angle Z^C ranges from 20°-33°. $2V_{\alpha}$ 50°-70°. Birefringence colours are generally upper 1st order. Therefore it is concluded that the amphibole is Hornblende, possibly one containing relatively little iron.

Amphibole is often seen partly replaced by chlorite, and less frequently by both chlorite and calcite. Replacement is often along the amphibole cleavage. Globular clinozoisite crystals often seem to grow into amphibole, very occasionally replacing it along its cleavage. The relationship of amphibole to other mineral phases is more difficult to ascertain. Garnet and amphibole are often closely associated but textural evidence for their relative age is equivocal. Amphibole is often found within saussuritised plagioclase and again their relative age is equivocal.

Pyroxene.

Pyroxene is found in very small quantities in a few calc-silicates (see Fig.20). Its significance lies in its possible use as a metamorphic grade indicator and as such it is discussed in Chapter 6.

Generally it is found associated with amphibole, both the massive form and the smaller acicular crystals. Pyroxene grain size varies from 0.1 to 1 mm. It is colourless in thin section and has moderate relief. Crystals show good cleavage, basal sections with two cleavages are relatively common. Birefringence colours rise to mid 2nd order. Maximum extinction angle Z^C is 44°. Extinction is always sharp, there is no dispersion. $2V_{\gamma}$ 50°-60°. The pyroxene is presumed to be either diopside or augite (Winchester & Whittles,1979 call it salite). It is most easily distinguished from clinozoisite-epidote which has similar relief and can have similar birefringence, by its lack of dispersion.

Garnet.

Garnet is found in most of the calc-silicates (see Fig.20). it occurs in two main forms; equidimensional anhedral crystals 0.5-2.0 mm, containing quartz inclusions often smaller than the quartz crystals in the matrix and much more commonly as larger sponge textured masses of garnet up to 10-15 mm. in diameter. Individual isolated fragments of garnet in these "sponges" are generally 0.2 to 0.5 mm. long and they frequently contain inclusions of quartz.

Garnet is frequently altered to, or replaced by, a variety of minerals, eg. chlorite, clinozoisite-epidote, amphibole, plagioclase, iron ore and biotite. Of these minerals chlorite and clinozoisite are the most common and they are also the minerals showing fairly unequivocal replacement textures of garnet. Neither zoning nor inclusion trails was seen in any garnet in the calc-silicates.

Biotite.

Biotite occurs in small quantities in some of the specimens (see Fig.20). When considered along with whole rock major element geochemistry it has been used as an indicator of metamorphic grade (see Chapter 6).

Biotite crystals range from 0.1 to 1.0 mm. long; some are thin laths, others more equidimensional. Pleochroic schemes vary in different sections, the variations are listed below;

α = Pale straw.

$\beta = \gamma$ = Dark brown, blood red, dark green.

Biotite is occasionally associated with garnet and also amphibole and in such cases reaction relationships are not apparent. Biotite is frequently altered along its cleavage to chlorite.

Muscovite.

Very small quantities of muscovite are found in the calc-silicates collected from the western side of the area (see Fig.20 and Map 6).

It occurs in two modes; growing as a product of the saussuritisation of plagioclase where the small unorientated crystals range in size up to 0.5 mm.; and intergrown with chlorite, where the muscovite laths range in length from 0.1 to 0.5 mm. In these intergrowths the chlorite and muscovite cleavages are parallel and age relationships are impossible to ascertain. As these intergrowths are nearly always associated with the breakdown of either biotite or garnet it is possible that chlorite and muscovite have formed at the same time.

Other minerals.

Calcite occurs in small quantities in nearly every sample where it is a product of the saussuritisation of plagioclase.

The minerals termed accessory in Figure 20 are iron ore, sphene, apatite and zircon. All of these occur in most of the sections, iron ore being the most abundant.

4.4) Semi-pelitic and Striped Group Lithologies.

4.4a) Introduction.

The term semi-pelite is used to describe homogeneous quartz-feldspar-biotite-muscovite rocks intermediate between pelites and psammites, whereas the term striped group lithology is used for hand specimens which are banded with alternating thin psammitic and pelitic bands.

4.4b) Lithology.

In the field semi-pelites are mid grey/brown coloured rocks. Quartz and feldspar grain sizes vary from 0.1 to 3.0 mm. Mica crystals 0.2 to 3.0 mm. long define a strong foliation which is usually folded into small crenulations, occasionally tight but usually open to close (Fleuty,1964). Within homogeneous semi-pelites bedding is usually impossible to detect, but frequently it can be seen in the adjacent psammitic beds where bedding is nearly always close in orientation to the mica foliation.

The striped group lithologies are composed of thin cream coloured psammites 3-20 mm. thick containing quartz and feldspar crystals with grain sizes from 0.2 to 2.0 mm,interbedded with dark brown/grey pelitic layers 2-20 mm. thick, predominantly composed of micas, with grain sizes from 0.2 to 3.0 mm. long.

4.4c) Mineralogy and petrography.

Figure 26 shows the modal analyses of all the samples collected as semi-pelites and striped group lithologies. In general they are seen to be quartz-feldspar-biotite-muscovite rocks with subordinate amounts of other minerals.

Quartz.

Quartz is found in all sections, Figure 26 gives an indication of its

Figure 26.

Modal analyses of striped and semi-pelitic rocks.

← Glenfinnan Division →

	8/60 ●	13/90 ●	19/112 ●	21/138 ●	32/217 ●	136/588 ●	138/643 X
QUARTZ	31.2	22.2	29.8	32.6	24.5	16.2	49.2
FELDSPAR	28.4	45.8	33.2	37.2	43.2	47.0	14.8
BIOTITE	20.8	23.4	25.0	10.2	1.4	25.6	9.6
MUSCOVITE	14.0	8.2	11.2	10.6	9.4	10.2	26.0
GARNET	3.8	0.0	0.0	0.0	0.0	0.0	0.0
IRON ORES	1.2	0.4	0.4	0.4	0.2	0.2	0.4
CHLORITE	0.0	0.0	0.2	8.8	20.8	0.0	0.0
ACCESSORIES	0.6	0.0	0.2	0.2	0.0	0.8	0.0

← Glenfinnan → ← Loch Eil →

	146/664 X	163/799 X	201/1086 ●	110/481 ●	121/525 ●	129/550 ●
QUARTZ	55.8	41.8	29.6	29.4	23.0	43.6
FELDSPAR	27.8	18.8	46.0	36.2	12.4	14.2
BIOTITE	6.6	18.4	17.8	32.0	35.4	21.2
MUSCOVITE	9.6	17.6	5.6	1.2	28.4	13.8
GARNET	0.2	0.2	0.0	0.0	0.2	2.4
IRON ORES	0.0	0.0	0.0	0.4	0.4	4.8
CHLORITE	0.0	0.0	0.8	0.0	0.0	0.0
ACCESSORIES	0.2	0.6	0.2	0.8	0.2	0.0
SILLIMANITE		2.6				

500 Points per section

● Semi-pelitic rock.

X Striped lithology.

abundance, ranging from 16.2% to 55.8% in different sections. Grain size varies from 0.1 to 2.0 mm. with average grain size in most sections ranging from 0.5 to 1.0 mm. Crystal shapes are anhedral, mostly equidimensional, but in some sections quartz crystals are elongated within the mica foliation. A weak C-axis fabric, detected using a gypsum plate, is present in a few sections. Deformation bands and sub-grains can be seen in some crystals. Crystal boundaries vary from polygonal mosaics to bulging, crenulated intergrowths.

Feldspar.

Feldspar is found in all sections, it is more abundant than quartz in some of the sections. Plagioclase can be seen in all the sections. K-feldspar showing perthitic and myrmekitic textures can be identified in a few sections though it may be present, untwinned, in more sections. Feldspar grain size varies from 0.2 to 3.0 mm. with most crystals in the range 0.5 to 1.5 mm. Crystal shape is usually anhedral, but with slight elongation within the mica foliation in some sections. Estimates of the plagioclase An. content in different sections vary from An.25 to An.51. Most of the feldspar is at least slightly saussuritised.

Biotite.

Biotite occurs in all sections, although in section 32/217 it is almost totally replaced by chlorite. The laths are from 0.1 to 3.0 mm. long and are always very thin. Biotite in different specimens shows three pleochroic schemes:

α = Straw yellow.

$\beta = \gamma$ = Dark brown, blood red. dark green.

Biotite laths are always aligned into a strong planar foliation which is usually crenulated by micro-folds. The laths never show undulose extinction around the crenulations. Occasionally the foliation undulates around small quartz-feldspar augen and again the laths do not show undulose extinction.

Muscovite.

Muscovite occurs in all specimens. All sections contain thin muscovite laths ranging from 0.1 to 3.0 mm. long which enhance the biotite foliation described above. They never show undulose extinction around the crenulations. The laths, although co-planar with biotite, often seem to overgrow it. In a few sections there are relatively large (1-3 mm.) non-elongate muscovites which show random orientation. These crystals grow over the composite biotite-muscovite foliation.

Garnet.

Garnet is recorded in only four of the samples collected (see Fig.26), although it seemed to be more common in the field. It occurs in minor quantities and its importance lies in its use in metamorphic and structural considerations. Grain size ranges from 0.5 to 2.0 mm. The mica foliation wraps around garnet crystals. Textures within the garnets vary considerably, zoning can be seen in section 8/60 in which one garnet has a central portion containing small globular quartz inclusions surrounded by garnet without inclusions, this in turn is rimmed with garnet containing many inclusions (Fig.27a). Section 129/500 contains zoned garnets with "sponge textured" rims (Fig.27b) but there are also similar sized garnets in this section which are completely "sponge textured" (Fig.27c). Figure 27d shows a possible curved inclusion trail in a garnet in section 8/60.

Other minerals.

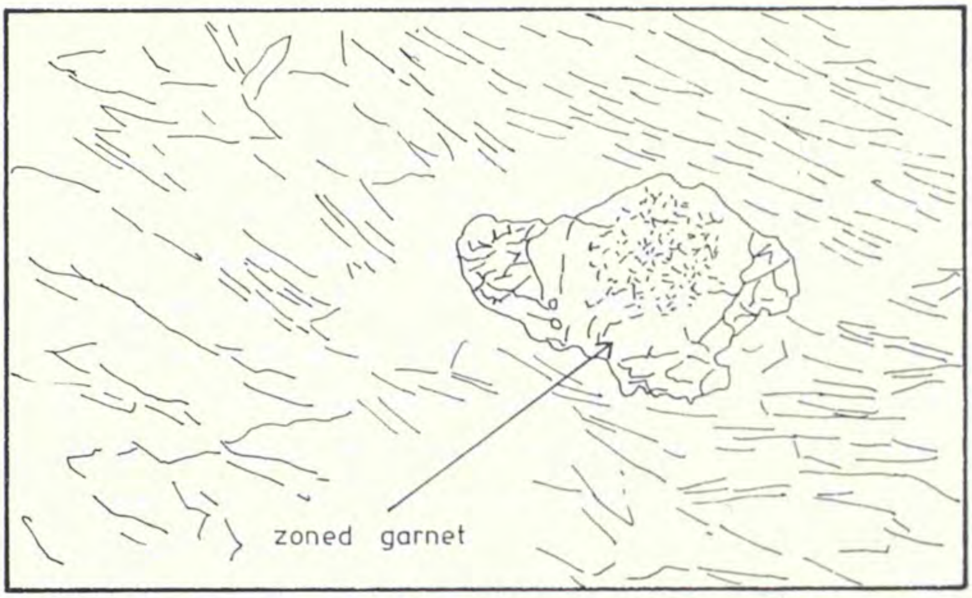
Pale green pleochroic chlorite is found in a few sections where it is seen to replace biotite. Small quantities of iron ore, apatite and zircon are found in most sections. Sillimanite was found in relatively small quantities in section 163/799 where it overgrows large poikiloblastic muscovite crystals.

4.4d) Comparison of semi-pelites and striped group lithologies.

The mineralogy and mineral distribution within semi-pelites and striped group lithologies show some subtle differences which are outlined below. In the semi-pelites the modal analyses are of relatively homogeneous rocks, so that quartz, feldspar, biotite and muscovite are found throughout the whole rock. In the striped group lithologies the modal analyses are calculated for traverses across bands of psammite and micaceous pelite. The psammites are quartz rich with small quantities of plagioclase and biotite. Within the pelitic bands muscovite is more common than biotite

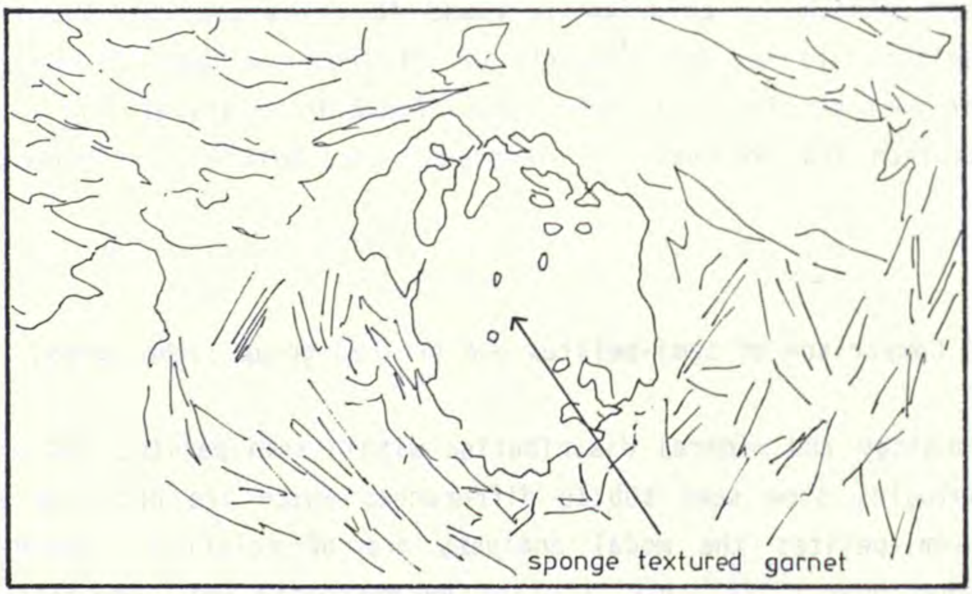
Figure 27. Photomicrographs of garnet textures in semi-pelitic and striped group lithologies.

(a)

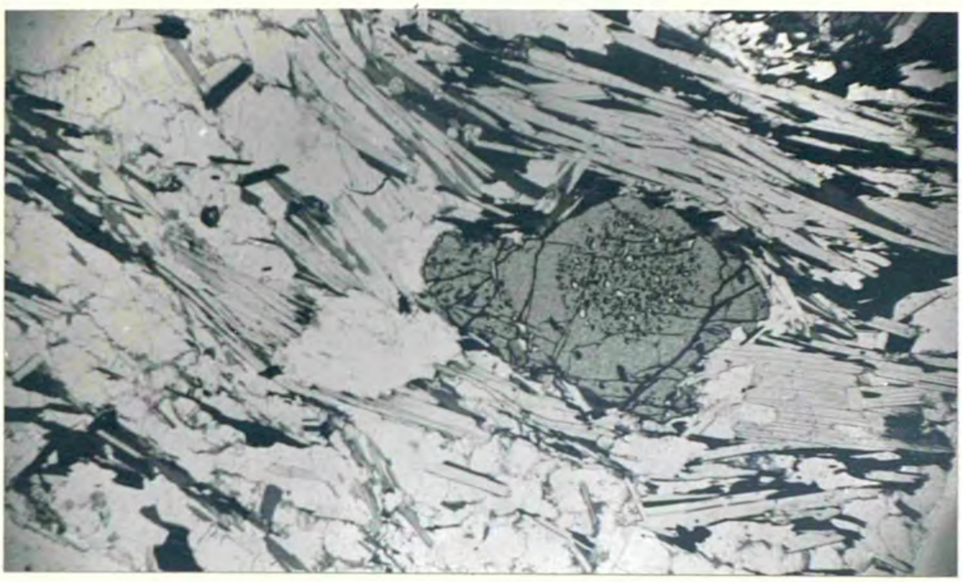


(Exp.8/60). plane polars. 0 mm. 1

(b)



(Exp.129(c)/550). plane polars. 0 mm. 1



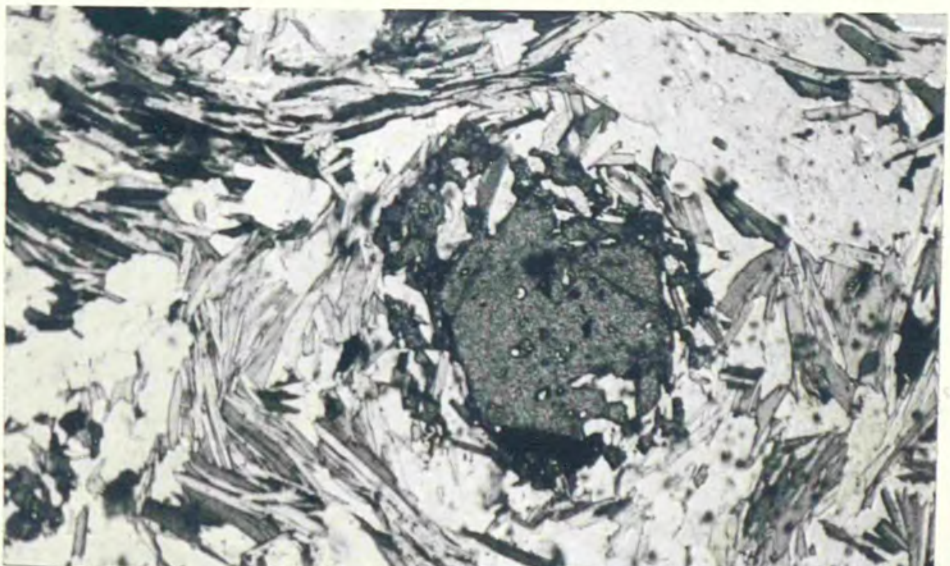
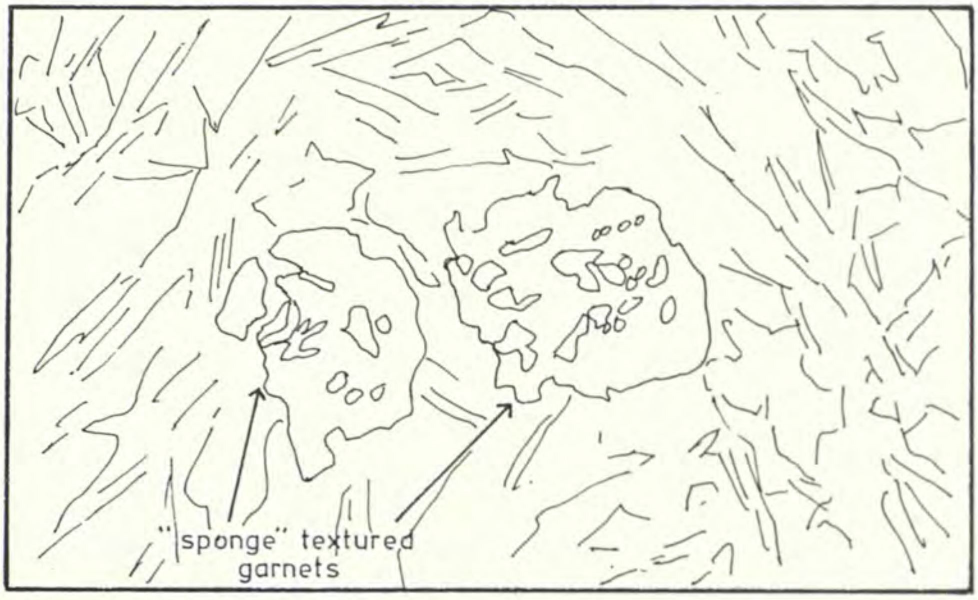


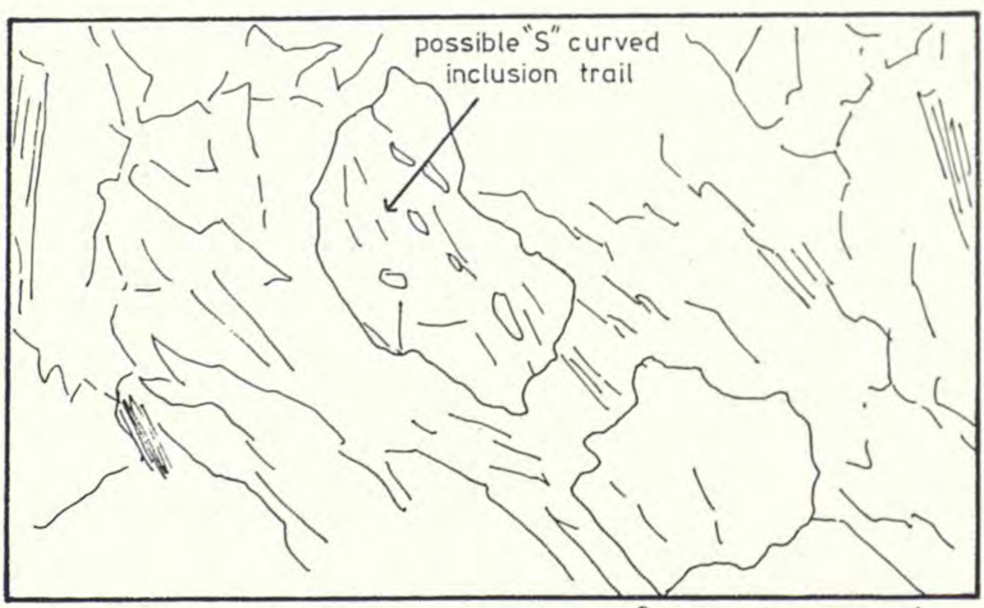
Figure 27 (cont). Photomicrographs of garnet textures in semi-pelitic and striped group lithologies.

(c)

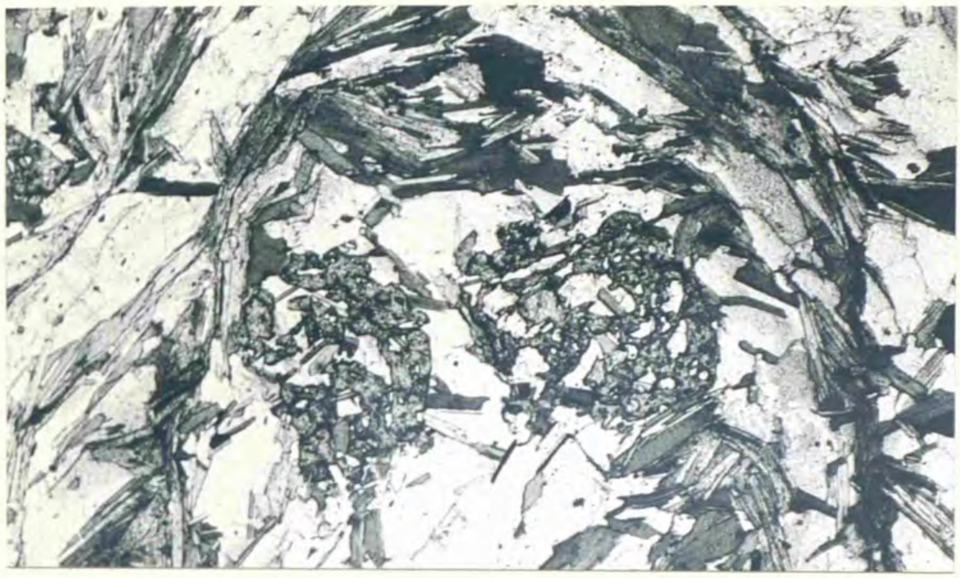


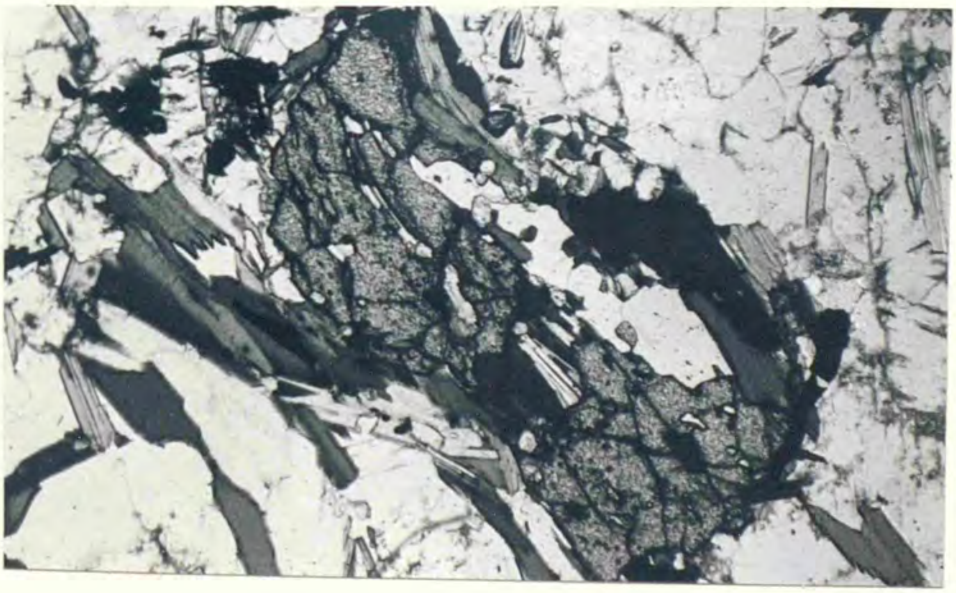
(Exp.129(a)/550).plane polars. 0 mm. 1

(d)



(Exp.8/60). plane polars. 0 mm. 1





and both of these minerals are more abundant than plagioclase,K-feldspar and quartz. The modal analyses of the striped lithologies and semi-pelites show that the semi-pelites are generally less quartz rich and more feldspathic than the striped group lithologies. The quantities of other minerals, chiefly micas, are relatively constant (see Fig.28).

Figure 28 can be used to compare these rocks with the psammites plotted similarly in Figure 19. These figures are combined to give Figure 29. In general the semi-pelites and striped group lithologies are less rich in quartz than the psammites, they have similar feldspar contents, or possibly slightly more feldspar and they have a considerably higher proportion of other minerals (essentially micas) than the psammites.

4.5) Pelitic rocks.

4.5a) Introduction.

Pelitic lithologies were not found in the Loch Eil Division. In the Morar Division a pelitic horizon, the Ranochan Pelite, crops out in the extreme west of the area mapped (Map 1) and although it was not mapped or sampled in detail it is compared to the pelites of the Glenfinnan Division.

4.5b) Lithology.

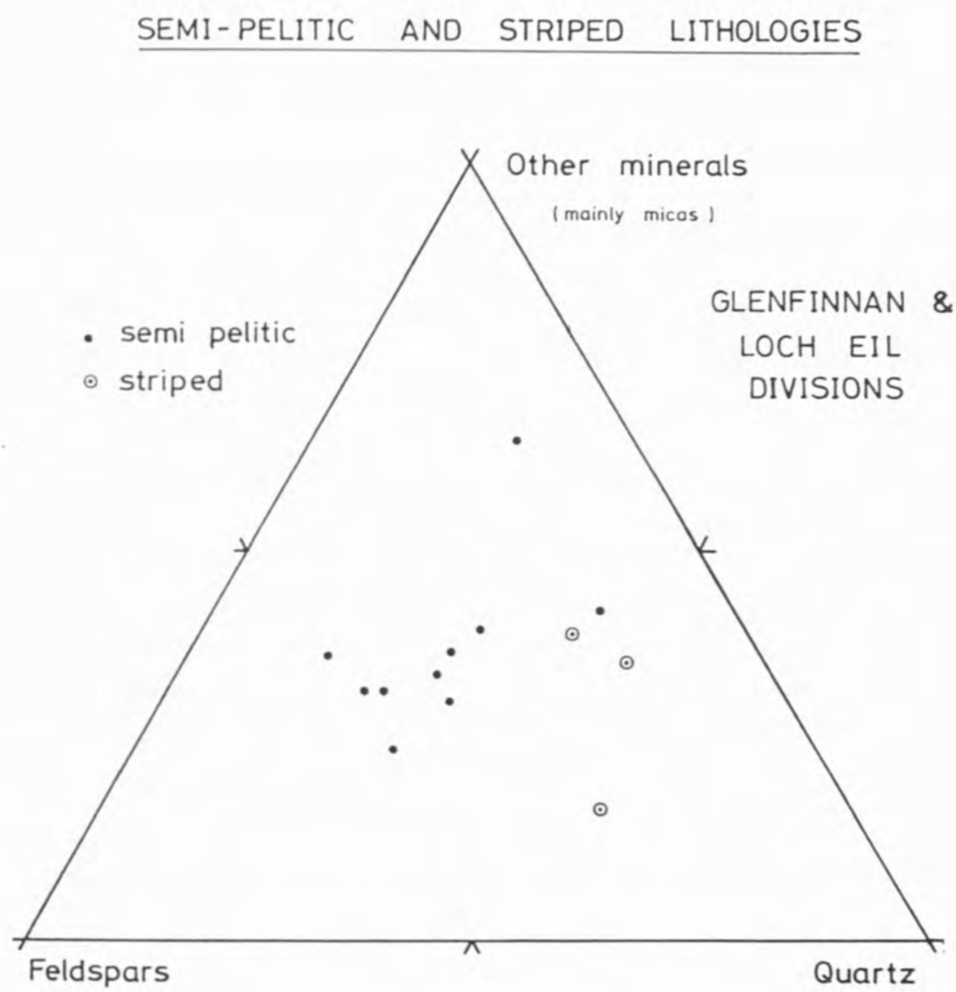
There are three mappable pelitic units in the Glenfinnan Division. the Glas Charn, Sgurr a Mhuidhe and Druim na Saille pelites (Figs.10 & 16).

Throughout the striped lithology there are thin micaceous and pelitic layers between thicker psammitic ribs.

Exposure in the west is good, especially on higher ground, however in the east exposure of the Drimsaille pelite is partly obscured by a mature coniferous forest.

In the field fresh exposures are medium to dark grey coloured while weathered surfaces are usually rusty brown coloured. Grain size is highly variable within individual samples and between samples. Pelitic rocks in the east of the Glenfinnan Division (eg. 12/70, 16/106, 17/107, 25/155, 30/209, 47/231) tend to be relatively fine grained and homogeneous (Fig.30a).

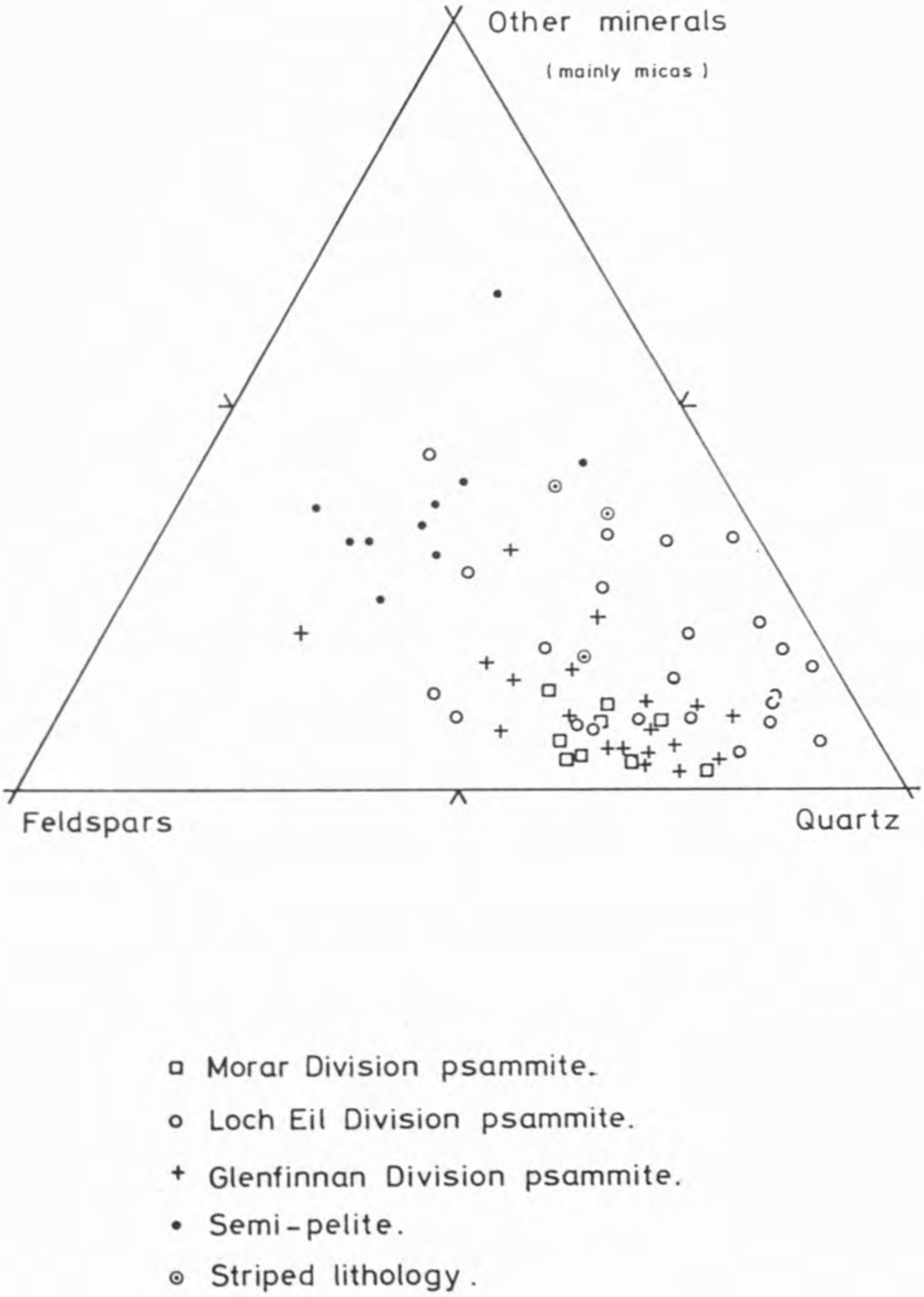
Figure 28.



Triangular diagram to show the composition of semi-pelitic and striped group lithologies.

Figure 29.

PSAMMITIC, SEMI-PELITIC AND STRIPED LITHOLOGIES



Triangular diagram to show the composition of psammitic, semi-pelitic and striped group lithologies.

(composite of Figures 19 & 28).

Felsic minerals have grain sizes of 1-2 mm, micas have lengths of 1-3 mm. Frequently the strong micaceous foliation is crenulated by microfolds (Fig. 30a). These pelites contain relatively few co-planar thin veins or pegmatites.

The Sgurr a Mhuidhe Pelite and the Glas Charn Pelite in the west of the Glenfinnan Division are much more coarse grained and gneissose in texture. Grain size of the felsic minerals is 2-3 mm. but the felsic minerals are found as elongate augen wrapped by micaceous foliae. The augen or lits have lengths from a few mm. up to a few cm.(Fig.30b) and grade upwards in size into larger co-planar pegmatites. Crenulation of the foliation is much less common than in pelites further east. Thin selvages of biotite are often found wrapping the larger of the quartzo-feldspathic augen. In the Glas Charn and Sgurr a Mhuidhe Pelites bedding is often difficult to trace. It can only be detected where there are more semi-pelitic layers within the pelite and then it is usually at a very low angle to the dominant planar mica foliation in the pelite.

4.5c) Mineralogy and petrography.

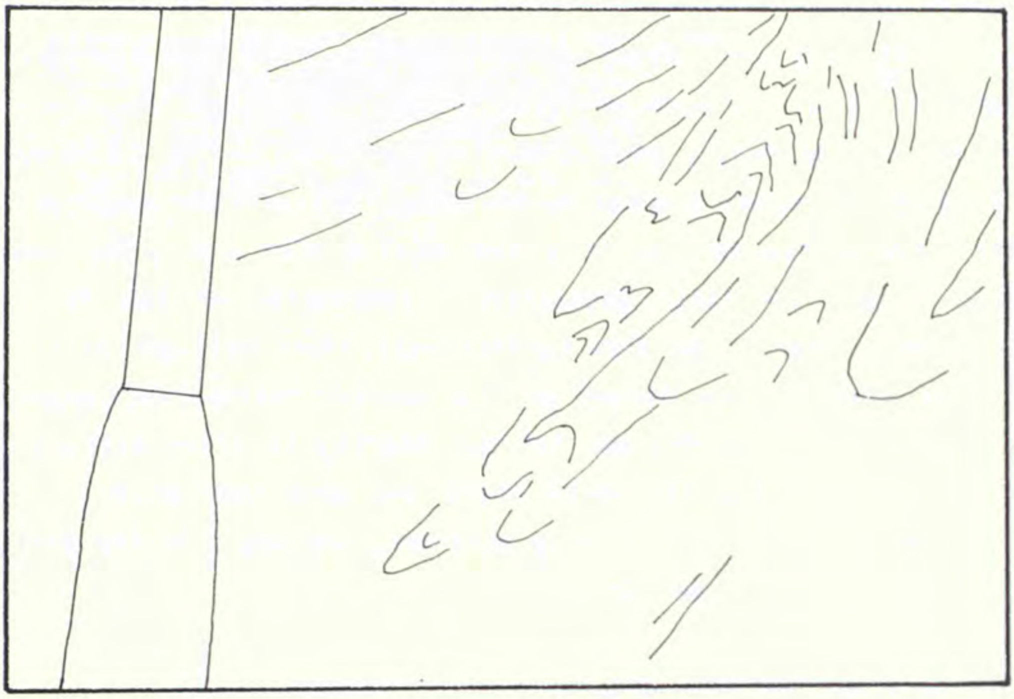
The pelites are composed predominantly of biotite, muscovite, feldspar and quartz with lesser quantities of garnet,chlorite, sillimanite, staurolite and other accessory minerals. Modal analyses of the samples collected are listed as Figure 31 (and in Appendix 3). A comparison of this figure with the modal analyses of striped and semi-pelitic rocks (Fig.26) shows that the pelites and semi-pelites contain similar mineral phases but the pelites generally have less quartz and feldspar and more mica and garnet plus small quantities of sillimanite and staurolite.

Because thin sections of Glenfinnan Division Division Pelites are very inhomogeneous, containing variable amounts of coarse pegmatites,augen etc. modal analyses are only presented as an indicator of mineral abundances and no statistical value is placed on the absolute values recorded in Figure 31.

In the mineral descriptions given below the following terms are used descriptively without any genetic implications. LEUCOSOME is used to describe the quartzo-feldspathic augen and co-planar pegmatitic portions of the pelites. PALAEOSOME is used to describe the remainder of the rock. Chapter 6 contains a more detailed discussion of descriptive and genetic terminology used here and by other authors.

Figure 30. Photographs to show the typical lithology of:

(a) eastern Glenfinnan Division pelite.



(b) western Glenfinnan Division migmatitic pelite.

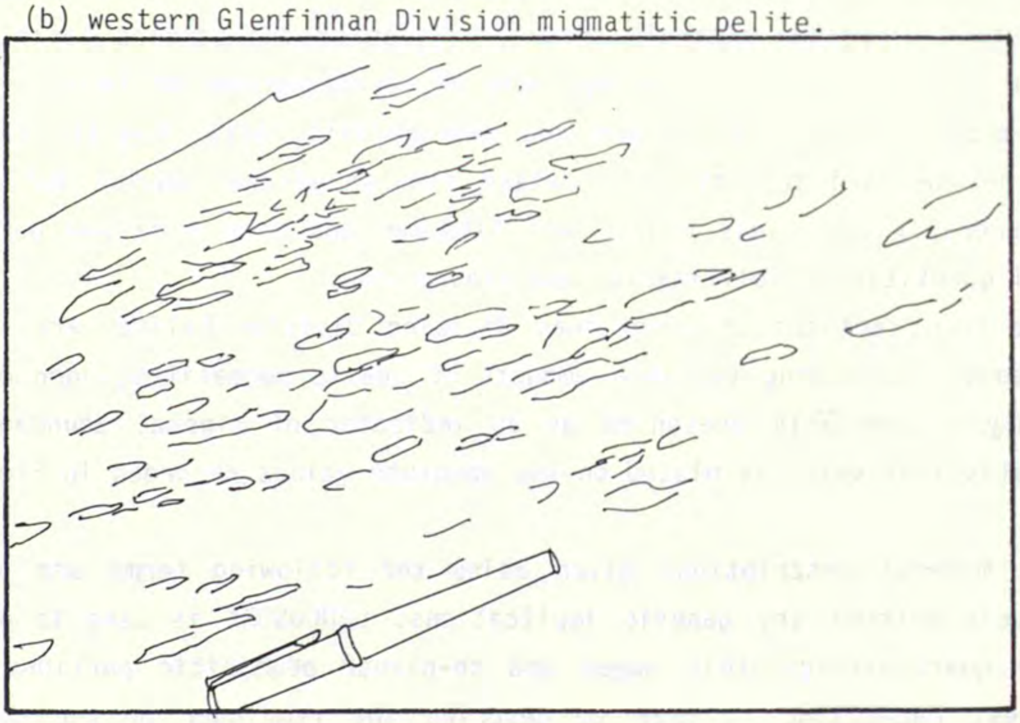


Figure 31.

Modal analyses of Glenfinnan Division Pelites.

	10/64	12/70	16/106	17/107	24/143	25/155	42/230	43/230	47/231
QUARTZ	15.6	23.8	20.8	23.4	20.8	18.6	40.8	20.6	21.6
FELDSPAR	26.4	31.4	52.4	39.8	46.2	47.6	35.6	34.8	18.2
BIOTITE	30.0	19.8	23.6	27.6	24.8	17.8	14.4	23.2	23.6
MUSCOVITE	23.2	16.6	1.4	7.6	5.8	10.0	8.8	18.0	34.8
GARNET	3.6	5.4	0.0	0.6	1.8	6.8	0.0	3.0	0.4
CHLORITE	0.6	1.2	0.2	0.0	0.0	0.2	0.2	0.2	0.0
ACCESSORIES	0.6	0.6	1.4	0.6	0.2	0.0	0.2	0.0	0.2
IRON ORES	0.0	1.2	0.2	0.4	0.2	1.0	0.0	0.2	1.2
SILLIMANITE	0.0	0.0	0.0	0.0	0.2	0.0	0.0	0.0	0.0

	135/583	82a/251	82b/251	Total 82/251	137/609	153/721	157/727	160/776	162/794
QUARTZ	29.0	5.8	17.4	11.6	44.8	24.8	31.2	7.8	30.2
FELDSPAR	0.2	28.0	48.0	38.0	26.0	37.4	11.8	23.6	18.6
BIOTITE	22.8	51.6	17.4	34.5	0.4	32.4	18.0	8.2	12.6
MUSCOVITE	17.0	0.2	0.6	0.	0.4	1.6	37.6	31.2	32.4
GARNET	0.6	2.6	1.6	2.2	15.8	3.0	0.2	8.4	3.8
CHLORITE	0.0	0.0	0.0	0.0	0.4	0.0	0.0	15.8	0.0
ACCESSORIES	0.2	0.2	0.6	0.4	0.4	0.4	0.4	0.0	0.4
IRON ORES	0.2	0.8	0.4	0.6	0.8	0.4	0.8	1.6	0.8
AMPHIBOLE	0.0	10.4	12.6	11.5	11.0	0.0	0.0	0.0	0.0
CLINOZOISITE	0.0	0.4	1.4	0.9	0.0	0.0	0.0	0.0	0.0
STAUROLITE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.4	0.0
SILLIMANITE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	1.0	1.2

	164/802	168/883	171/948	173/954	176/961	179/971	181/986	186/1012	189/1031
QUARTZ	31.2	30.0	28.4	45.4	36.8	11.8	20.6	19.0	31.0
FELDSPAR	27.2	16.2	14.6	20.4	27.4	63.0	17.8	27.2	28.6
BIOTITE	15.2	22.6	15.0	11.2	10.6	21.2	33.8	30.2	18.8
MUSCOVITE	18.4	30.8	35.6	19.0	24.2	1.6	25.4	21.6	15.6
GARNET	5.4	0.0	5.4	2.4	0.0	0.2	1.8	1.2	1.4
CHLORITE	0.0	0.0	0.0	0.2	0.2	2.0	0.0	0.0	0.6
ACCESSORIES	1.2	0.2	0.2	0.0	0.2	0.2	0.4	0.6	0.2
IRON ORES	1.4	0.2	0.8	0.4	0.6	0.0	0.0	0.2	1.8

500 Points per section

Figure 31.
(continued)

Modal analyses of Glenfinnan Division Pelites.

	199/1045	201/1050	203/1064	206/1065	211/1086	217/1089	223/1090	224/1090	232/1092	234/1093
QUARTZ	31.4	37.6	38.8	26.0	21.0	26.6	32.2	23.4	18.6	28.2
FELDSPAR	18.2	12.0	18.4	33.0	36.2	17.0	39.2	41.0	26.8	19.0
BIOTITE	15.6	19.6	15 2	28.2	20.0	13.2	22.8	17.6	31.4	14.6
MUSCOVITE	31.6	29.6	25.4	11.4	14.6	39.4	4.2	15.0	21.6	31.4
GARNET	0.6	0.0	0.0	0.2	5.0	0.0	0.4	1.6	0.8	2.6
CHLORITE	0.2	0.0	0.0	0.6	0.8	0.0	0.8	0.6	0.4	0.6
ACCESSORIES	0.4	0.2	0.0	0.2	0.2	0.2	0.2	0.2	0.2	0.0
IRON ORES	1.2	1.0	0.8	0.2	1.6	3.6	0.2	0.6	0.0	2.2
SILLIMANITE	0.8	0.0	0.4	0.2	0.6	0.0	0.0	0.0	0.2	1.4

	235/1093	237/1093	243/1096	245/1104	246/1104	247/1104	248/1104	249/1104	250/1104
QUARTZ	9.0	20.4	42.0	29.6	25.2	23.4	15.6	22.8	22.6
FELDSPAR	45.8	40.2	37.2	27.4	44.4	39.6	36.4	47.4	40.0
BIOTITE	39.0	25.4	6.6	22.6	24.6	31.4	33.8	21.2	24.0
MUSCOVITE	5.8	10.2	11.8	11.6	5.2	5.4	11.2	8.0	12.0
GARNET	0.0	2.0	0.2	5.4	0.2	0.2	1.4	0.0	0.8
CHLORITE	0.0	1.0	1.8	0.2	0.2	0.0	0.0	0.4	0.4
IRON ORES	0.2	0.6	0.0	1.8	0.0	0.0	1.0	0.0	0.2
ACCESSORIES	0.2	0.0	0.2	0.0	0.0	0.0	0.6	0.2	0.0
STAUROLITE	0.0	0.0	0.0	0.2	0.0	0.0	0.0	0.0	0.0
SILLIMANITE	0.0	0.2	0.2	1.2	0.2	0.0	0.0	0.0	0.0

500 Points per section

Quartz.

Quartz is found in all sections, ranging from about 5% to 40% in the sections examined. Grain size varies from 0.5 to 2.0 mm. in the palaeosomatic portions of the pelites. In the leucosomes quartz grain size ranges up to 4-5 mm. Crystal shapes are anhedral, mostly equidimensional with no obvious elongation within the strongly developed mica foliation although a few sections show shape elongations of up to 2:1 within the foliation. Quartz is generally not, or only very weakly strained, as evidenced by weakly developed broad deformation bands. Strained crystals are more common in the leucosomatic portions of some sections. Well developed crystallographic C-axis fabrics were not detected (using a gypsum plate). Crystal edges are either fairly straight or convex, often bulging slightly into adjacent feldspar crystals.

Feldspar.

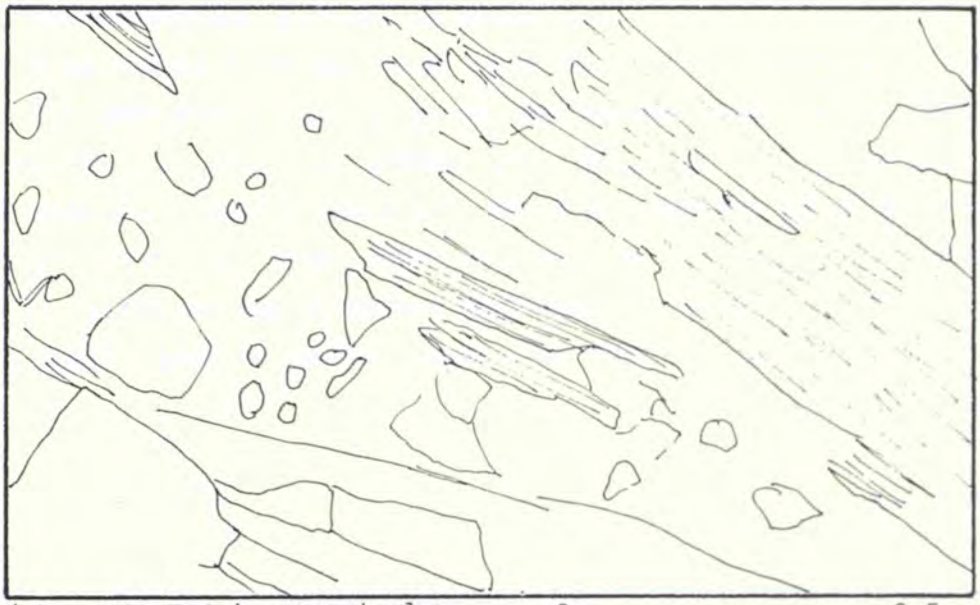
Feldspar is abundant in all sections, comprising about 10% to 60% of the modal totals (Fig.31). Twinned plagioclase can be seen in all sections, optical estimates of the anorthite content in different sections range from An.22% to An.54%. Virtually all the plagioclase crystals are un-zoned, rarely some show weak irregular patchy zoning. In contrast to psammitic and semi-pelitic lithologies, microcline was not found. Good examples of perthitic textures were not observed although some of the plagioclase crystals are altered along thin veins which may be vein anti-perthites. Likewise myrmekitic textures were not observed but in some sections a few plagioclase crystals contained small globular inclusions of quartz (Fig.32). Thus K-feldspar has not been positively optically identified though most sections contain some untwinned feldspar crystals which could be K-feldspar. Staining (using Sodium cobaltinitrite: Chayes,1952) has not revealed any K-feldspar. The grain size of feldspar crystals ranges from 0.5 to 3 mm. in the palaeosome and up to 10 mm. in the leucosomes. Crystal shapes are anhedral with no relationship between crystal edges and crystallography. The feldspars in most sections are fresh or only slightly saussuritised.

Biotite.

Biotite is abundant in all sections, ranging from 11% to 39% (Fig.31). It occurs as laths from 0.5 to 3 mm. long which are always elongate and orientated, usually very strongly, into a planar foliation which is either

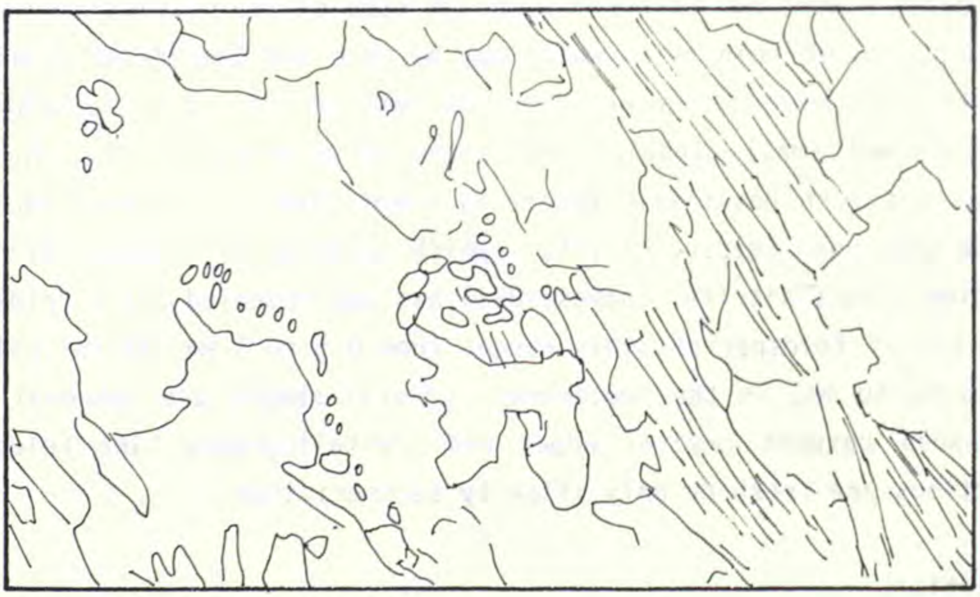
Figure 32. Photomicrographs to show quartz/feldspar textures in Glenfinnan Division pelites. (see text for discussion).

(a)

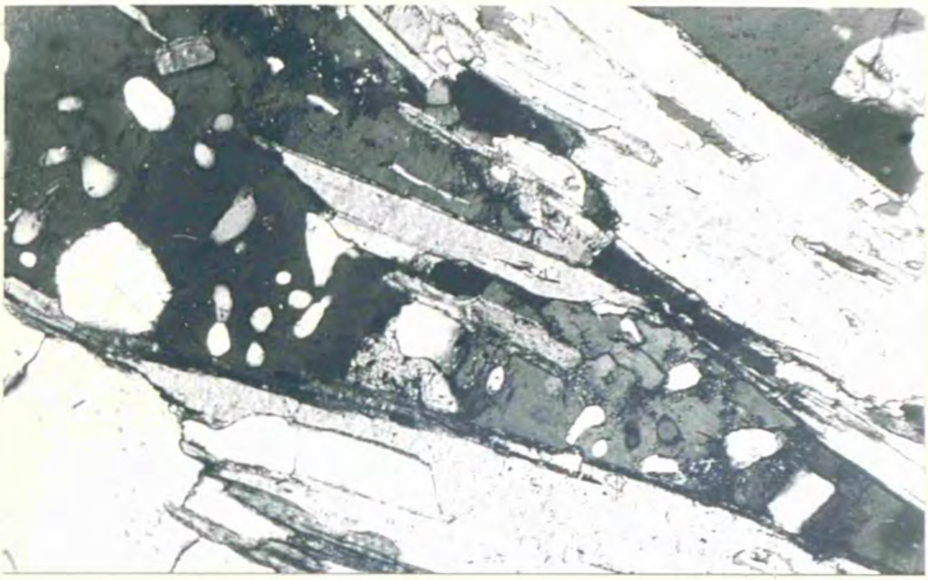


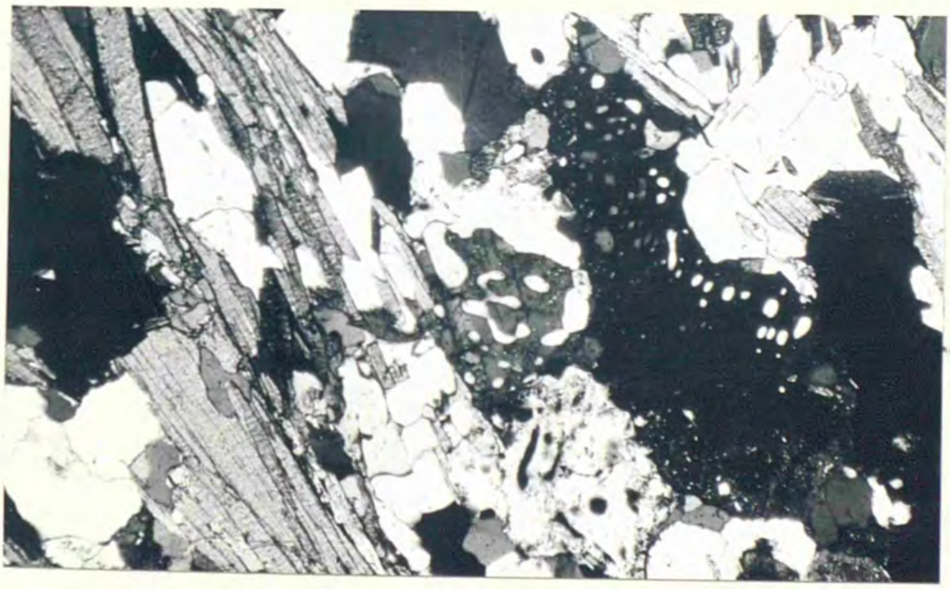
(Exp.162/794). crossed polars. 0 mm. 0.5

(b)



(Exp.162/794). crossed polars. 0 mm. 1





wrapped around quartzo-feldspathic augen or crenulated by a series of microfolds. The laths, although often folded around fairly tight crenulations, never show strong undulose extinction. In different specimens biotite shows a range of pleochroic schemes similar to biotites in the semi-pelitic and striped lithologies. Maximum pleochroic colours ($\beta = \gamma$) can be dark brown, blood red ar dark green.

Muscovite.

Muscovite is abundant in many of the thin sections, occuring in a variety of modes. Most crystals occur as fairly thin laths 0.1 to 3.0 mm. long, aligned co-planar with biotite laths. There are also some muscovite crystals which are up to 5 mm. long, non elongated, and showing no preferred orientation of the basal cleavage in relation to the ubiquitous micaceous foliation. These large porphyroblastic crystals occur in two modes, some are found enclosed within micaceous foliae and are often mantled by sub-grains of muscovite. Other large non-orientated porphyroblasts cross cut the dominant micaceous foliation. Of all the muscovites only the augened large porphyroblasts show any sign of deformational strain.

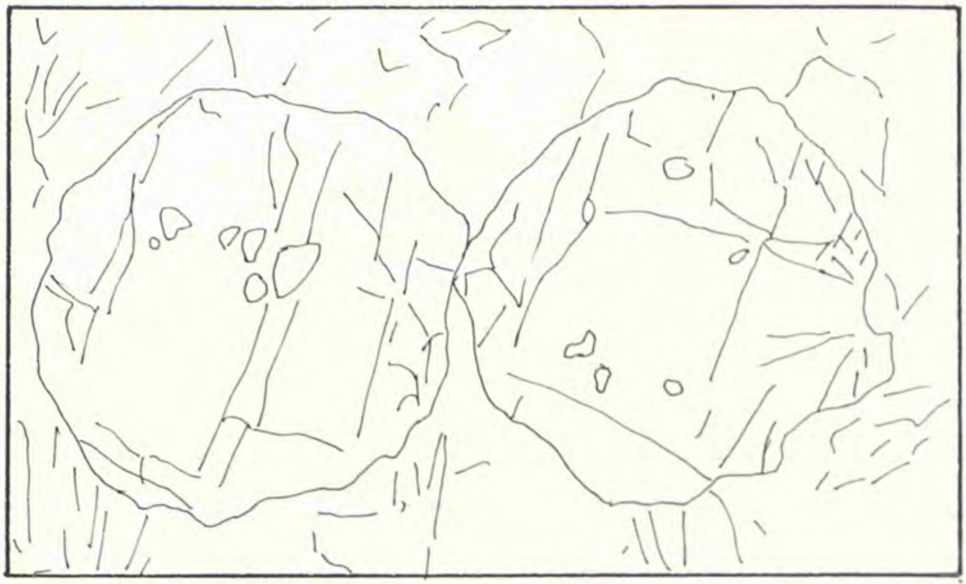
Garnet.

Burgundy coloured garnet is found in small quantities in most of the sections (Fig.31). Most garnets are small (1-2 mm.),fairly equidimensional and quite fresh. Crystallographic faces are not present. No abrupt zoning is present but many of the small garnets have small numbers of small round quartz inclusions in the central zone surrounded by an inclusion free zone which is occasionally rimmed by an outer zone containing slightly larger irregularly shaped quartz inclusions (Fig.33a). There are very few iron ore inclusions. No inclusion trails are found in any of the garnets. Occasionally partial replacement of garnet by aggregates of biotite, muscovite and feldspar has occured.

Larger irregularly shaped garnets (up to 6 mm.) occur sporadically in a few sections. They contain large quartz inclusions and a few iron ore inclusions. Frequently they are highly replaced by biotite, muscovite, feldspar, quartz and chlorite. These large garnets form augen around which the mica foliation is wrapped (Fig.33c). Probe analyses of garnets from the Lochailort-Glenfinnan area (Anderson & Olympio,1977) show them to be internally homogeneous, unlike those further west which often show strong internal chemical zoning.

Figure 33. Photomicrographs to show garnet textures in Glenfinnan Division pelites.

(a)

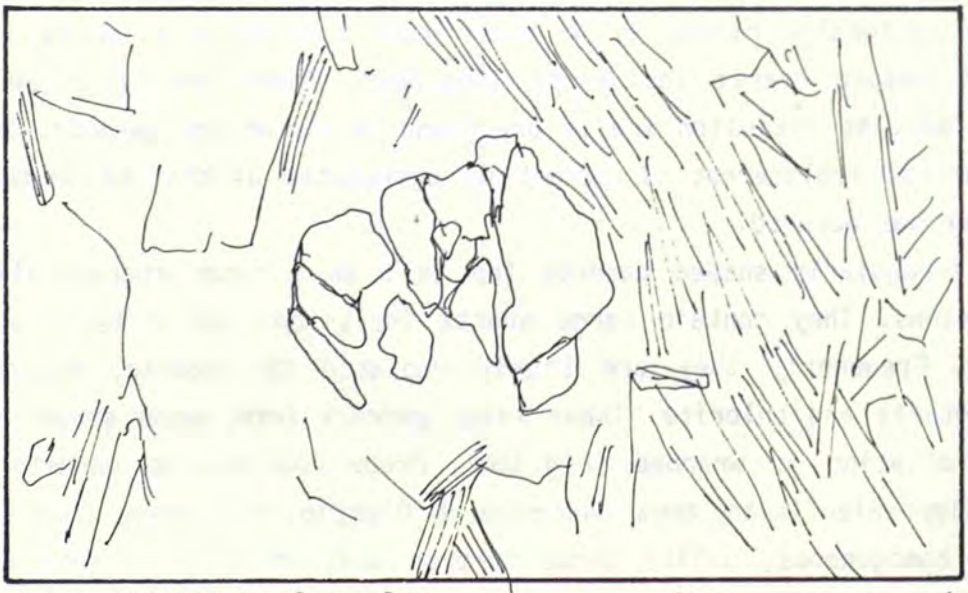


(Exp.25/155). plane polars. 0 mm. 1

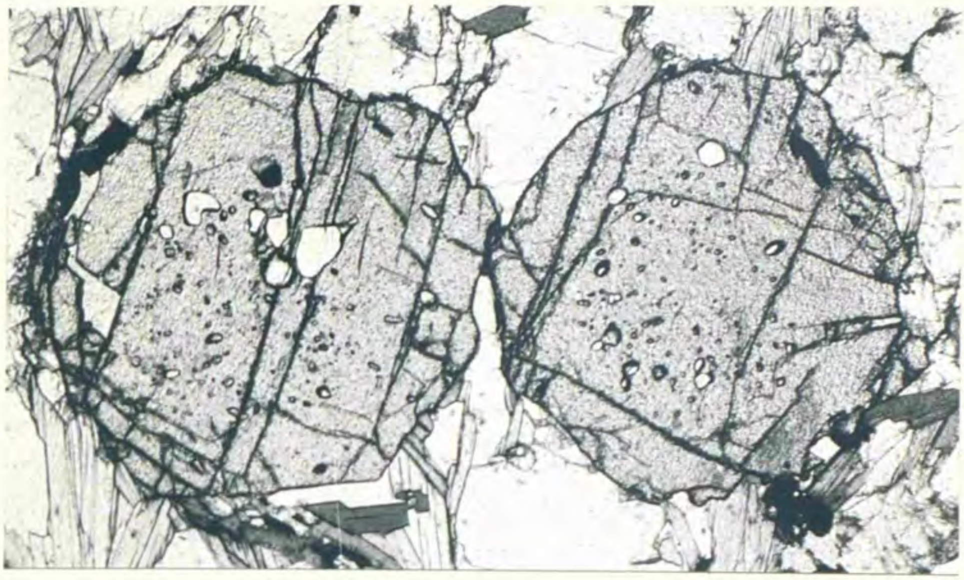
(b)

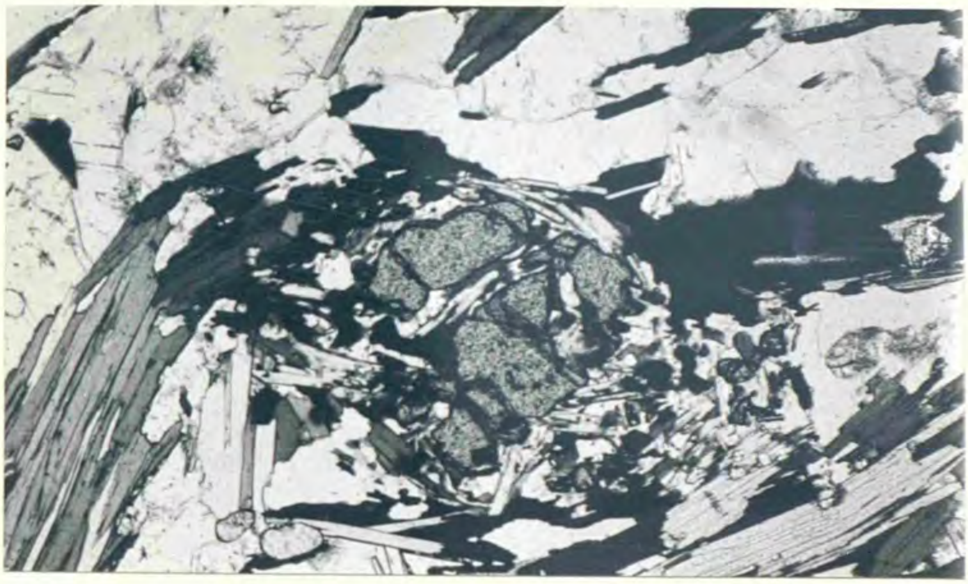
(Exp.135/583). plane polars. 0 mm. 1

(c)



(Exp.135/583). plane polars. 0 mm. 1





Sillimanite (Fibrolite).

Sillimanite is found in very small quantities in some pelites (Fig.34). It occurs as fine needles up to 0.2 mm. long, and always associated with muscovite. The mats of fibrolite needles anastomose across single muscovite crystals, showing no crystallographic relationship to the enclosing mica. Sillimanite has not been observed in any of the elongate laths which together with the biotite crystals produce the strong foliation in the pelites, it seems to be restricted to the larger, crystallographically unorientated porphyroblasts which frequently have marginal sub-grains and are augen wrapped by the strong micaceous foliation. Sillimanite has not been observed in any of the porphyroblasts which appear to cross cut the foliation, and it does not grow over the mantle of the early porphyroblasts it is concluded that the fibrolitic sillimanite grew prior to the deformation of the early porphyroblasts.

Other minerals.

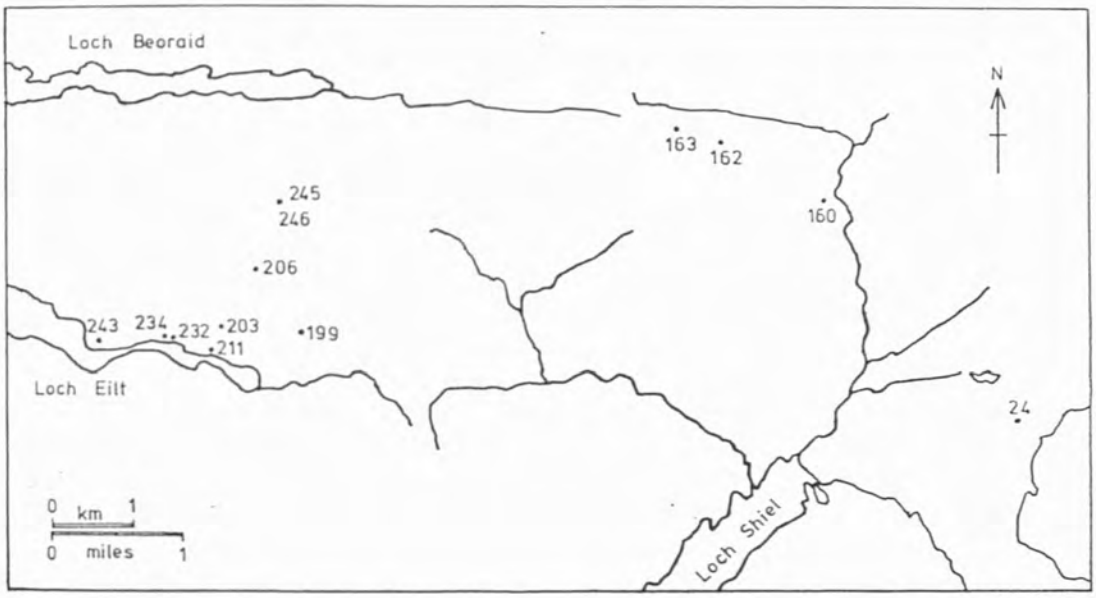
Small quantities of staurolite occur in sections 160/776 and 245/1104. Fibrous chlorite occurs, replacing biotite, in a few sections. Accessory amounts of iron ore, apatite and zircon occur in most sections.

4.5d) Morar Division pelite.

Although the Ranochan Pelite in the extreme west of the area has not been studied in detail, certain points can be made which show its petrographic similarity to the pelites of the Glenfinnan Division.

Twinned plagioclase is common, K-feldspars have not been observed. Quartzo-feldspathic fabrics are similar to those in the Glenfinnan Division pelites. Intergrown biotite and muscovite define a micaceous foliation. Sillimanite is found in large, partly deformed muscovite crystals. Garnets are 1-2 mm. long and texturally inseparable from the small garnets in the Glenfinnan Division pelites.

Figure 34.



Map to show sillimanite (fibrolite) localities.

(for Grid References see Appendix 4).

4.6) Ardgour granitic gneiss.

4.6a) Introduction.

A description of the lithology, mineralogy and petrography of the Ardgour granitic gneiss is included in this chapter on metasedimentary rocks even though the origin of the gneiss has been the subject of a great deal of research and debate. The structure of the gneiss and its relationship to the surrounding metasediments is discussed in Chapter 5, and its metamorphic history in Chapter 6.

The outcrop of the Ardgour granitic gneiss within the mapped area is shown on Map 1. East of Glenfinnan village, road cuttings on the A 830 expose many excellent large vertical rock faces up to 10-20 m. high and 100 m. long, and although the junction between the gneiss and the surrounding metasediments is not exposed, nearby exposures are lithologically gradational between striped lithologies and the Ardgour granitic gneiss. Exposure away from the road section is not so good, recently the western slopes of Glean Dubh have been afforested, while the eastern slopes of Glean Dubh are covered by a mature coniferous forest. Exposure is generally good on the western side of the outcrop of the gneiss.

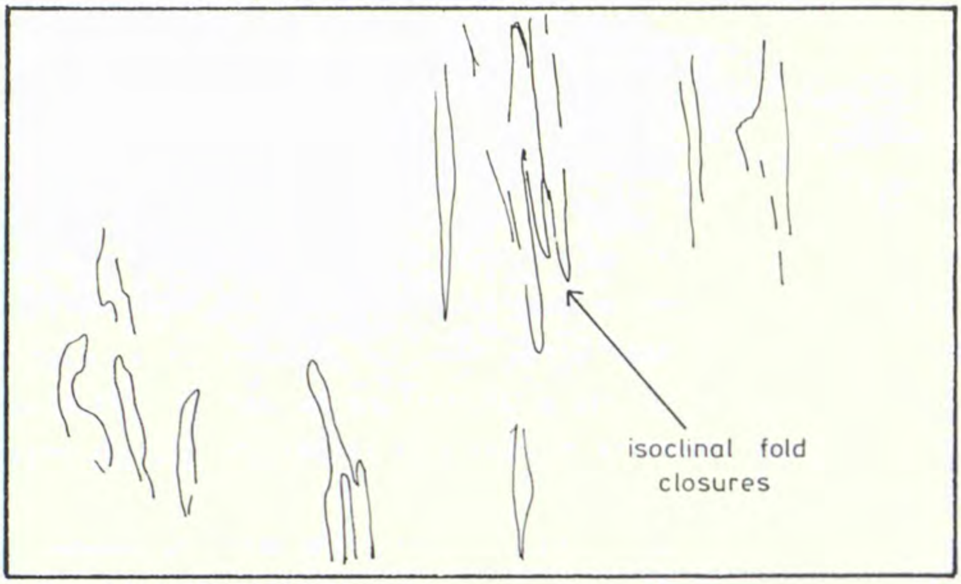
4.6b) Lithology.

The Ardgour granitic gneiss is strongly gneissose, often highly deformed, and does not resemble any of the surrounding striped and pelitic rocks. It is characterised by containing large numbers of pegmatites which have a variety of structural relationships to the gneissose foliation ranging from small deformed pegmatite pods, often with a biotite selvage, to large transgressive pegmatitic sheets (Fig.35).

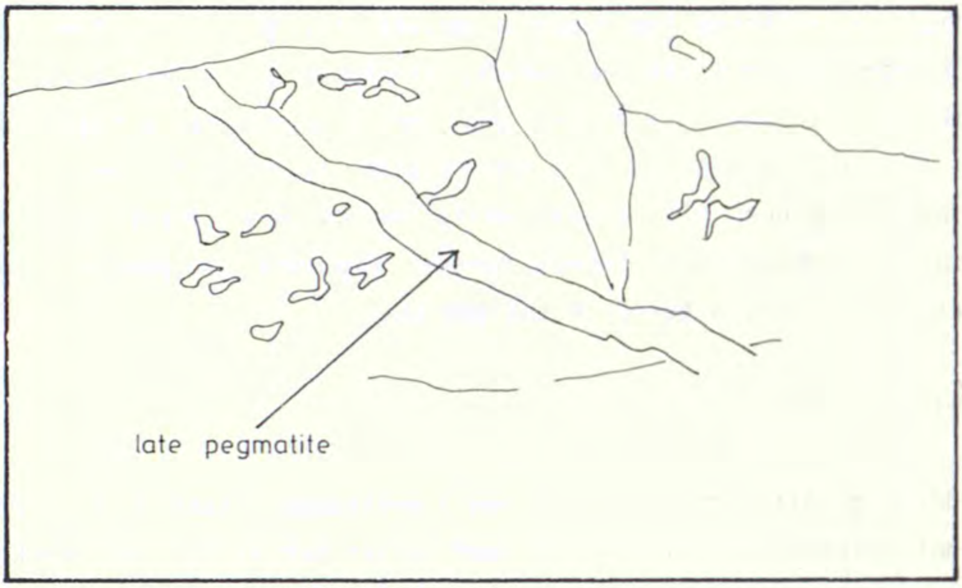
Excluding pegmatitic material, the gneiss has a strong gneissose foliation defined by planar, orientated biotite crystals 1-2 mm. long which are discrete and not found as well defined mica layers. Quartz and feldspar crystals (1-3 mm. long) are the other main components of the gneiss (Fig.36). Some of the feldspars weather to a cream or pink colour but most are fresh and white. Often micas are wrapped around small felsic augen 10-20 mm. long and 2-5 mm. wide. Felsic augen grade upwards into pegmatitic lenses (10-20 cm. long) which are generally concordant with the foliation. The gneiss, with its strong gneissose foliation and pegmatitic lenses is often folded by later folds.

On the southern slopes of Beinn an Tuim a large mass and a much smaller

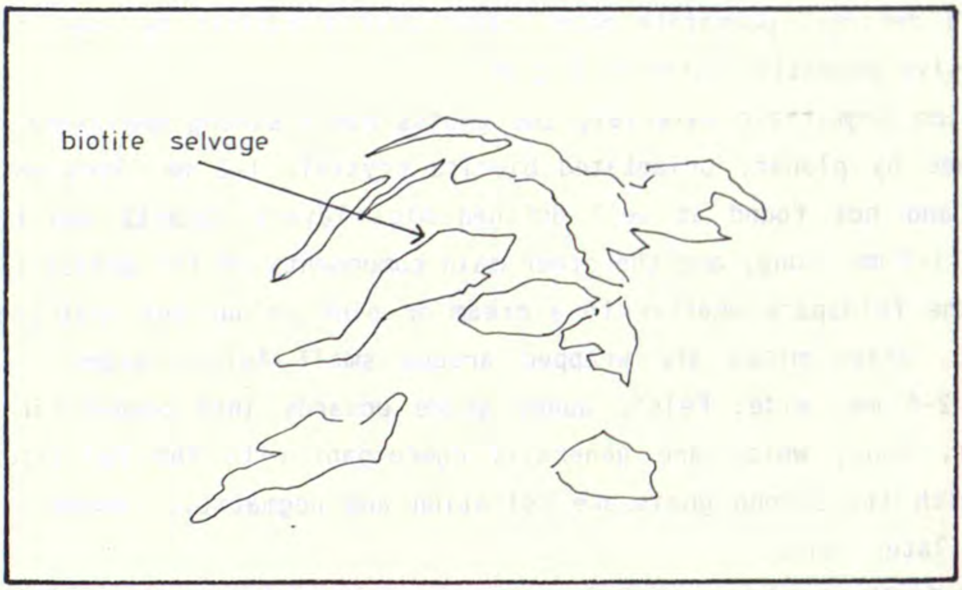
Figure 35. Photographs to show the field appearance of the Ardgour granitic gneiss.



(Exp.1233 NM 91458031).



(Exp.1235 NM 91778020).



(Exp.1235 NM 91778020).

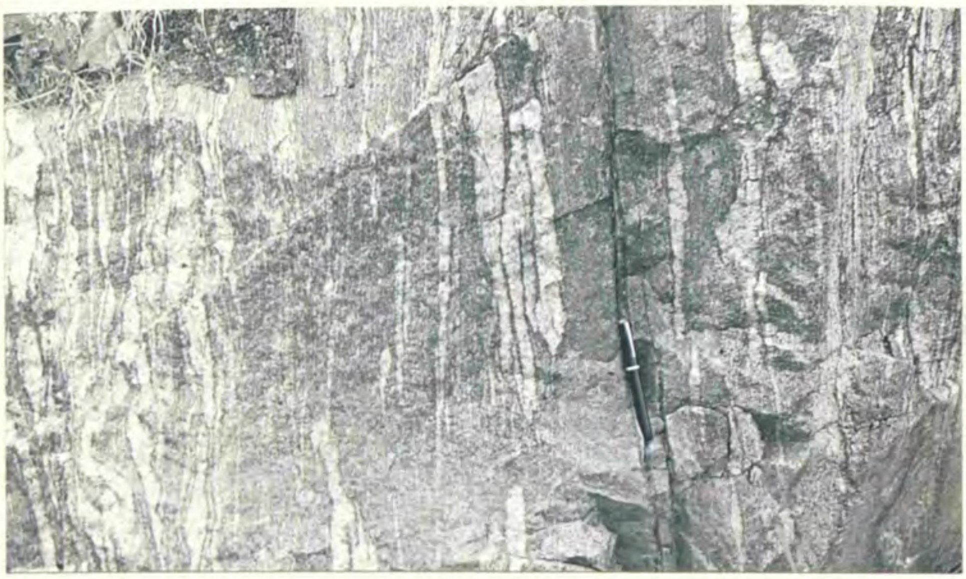


Figure 36.

Modal analyses of Ardgour granitic gneiss.

	6/60	7/60	14/92	15/93	27/164	53/232	55/232	58/235	59/235
QUARTZ	28.2	30.6	36.4	40.0	38.6	36.6	14.6	40.0	15.6
PLAGIOCLASE	25.0	29.8	46.6	36.4	15.8	39.4	41.8	42.6	31.2
K-FELDSPAR	30.2	21.6	12.4	16.2	19.2	10.6	5.0	1.4	53.2
MYRMEKITE	6.4	2.0	0.0	2.0	4.4	0.4	0.0	0.0	0.0
GARNET	0.0	0.0	0.0	0.0	0.0	0.6	0.0	1.8	0.0
BIOTITE	9.4	15.4	0.4	2.6	13.6	11.0	37.0	13.6	0.0
MUSCOVITE	0.0	0.4	2.4	1.4	8.0	0.0	0.0	0.0	0.0
CHLORITE	0.0	0.2	1.4	1.4	0.0	0.2	0.0	0.0	0.0
'EPIDOTE'	0.0	0.0	0.2	0.0	0.0	0.4	1.2	0.4	0.0
ACCESSORIES	0.8	0.0	0.2	0.0	0.4	0.8	0.4	0.2	0.0

	68/235	70/235	77/244	277/1235	278/1235	280/1235	281/1235	Peg
QUARTZ	35.4	28.0	30.6	24.6	29.6	26.6	32.8	23.8
PLAGIOCLASE	46.8	53.8	48.6	27.8	41.0	28.0	23.6	38.0
K-FELDSPAR	1.6	1.0	0.8	28.0	2.6	6.6	16.0	30.2
MYRMEKITE	0.0	0.0	0.0	3.8	2.6	0.6	1.4	5.2
GARNET	1.4	2.0	2.0	0.0	1.6	0.0	0.2	0.0
BIOTITE	14.0	15.2	17.4	14.8	24.6	29.2	26.0	2.0
MUSCOVITE	0.0	0.0	0.0	0.8	0.0	8.8	0.0	0.8
CHLORITE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
'EPIDOTE'	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
ACCESSORIES	0.6	0.0	0.6	0.2	0.0	0.2	0.0	0.0

500 Points per section

mass of metasediment (Map 1) are found within the gneiss, but generally the gneiss is remarkably homogeneous throughout its outcrop.

4.6c) Mineralogy and petrography.

Modal analyses of the gneiss samples (Fig.36), bearing in mind the heterogeneity of the gneiss at hand sample scale, indicate that the gneiss is composed of quartz,feldspars and biotite, with much smaller quantities of muscovite, garnet and other minerals. The modal analyses exclude any portions of discordant late pegmatite in the section and likewise mineral descriptions are of minerals within the gneiss exclusive of obvious discordant late pegmatitic material. Figure 36 contains an averaged modal analysis of concordant early pegmatitic from samples 280/1235 and 281/1235 (see Chapter 6.2c).

Quartz.

Quartz is abundant in all sections, its abundance ranging from 24% to 40%. Grain size ranges from 0.2 to 1.5 mm., crystals are anhedral with cuspate margins, convex towards other minerals. Quartz appears to overgrow plagioclase and K-feldspar (Fig.37b,bottom right). Strain within quartz crystals is usually weakly developed and can be seen as deformation bands and occasional sub-grains. Quartz occurs in minor quantities as the small "worms" in the myrmekites which often form between plagioclase and K-feldspar crystals and also as small globules within plagioclase crystals where myrmekite is not common.

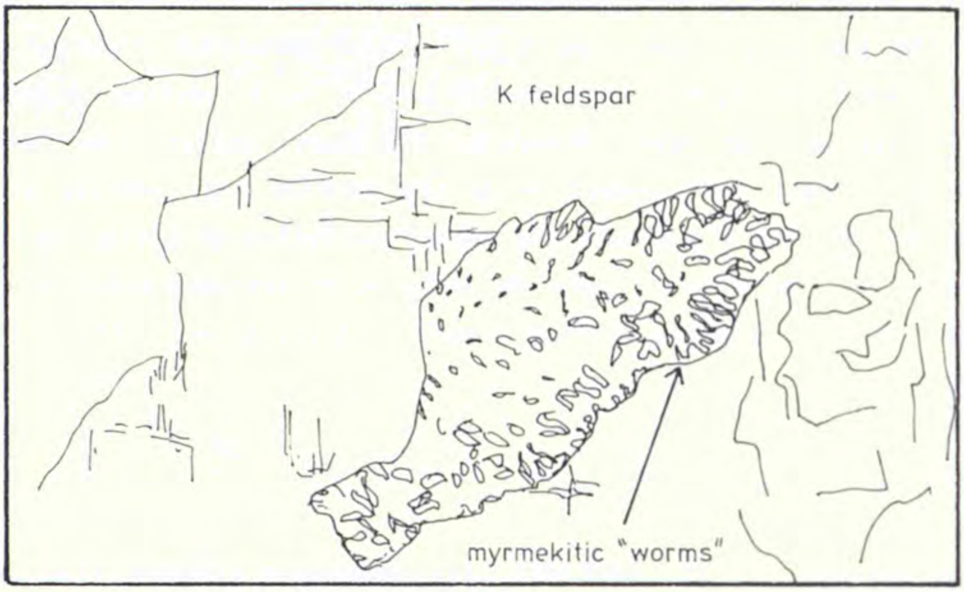
In the deformed concordant early pegmatites quartz crystals have grain sizes up to 2 mm.,anhedral shapes, often cuspate, bulging into feldspar crystals and occasionally quartz appears to be interstitial to much larger feldspar crystals.

Feldspar.

Feldspars make up a large proportion of all the sections examined. Plagioclase often with multiple albite twinning, can be identified in all sections. K-feldspar (microcline) can be identified optically in most but not all sections though it may be present as untwinned feldspar in all sections. K-feldspar content varies from 1% to 53% in different sections. Plagioclase comprises 16% to 54% in different sections, grain size ranges from 1-3 mm. and the anorthite content varies from An.17% to An.46% in different

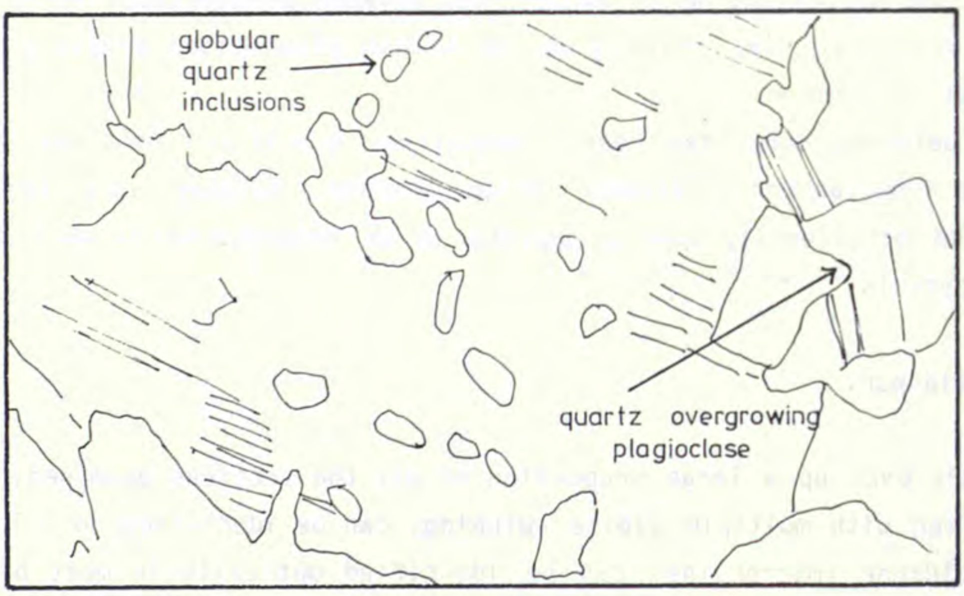
Figure 37. Photomicrographs to show feldspar textures in the Ardgour granitic gneiss.
(see text for discussion).

(a)



(Exp.6/60). crossed polars. 0 mm. 0.5

(b)



(Exp.77/244). crossed polars. 0 mm. 0.5

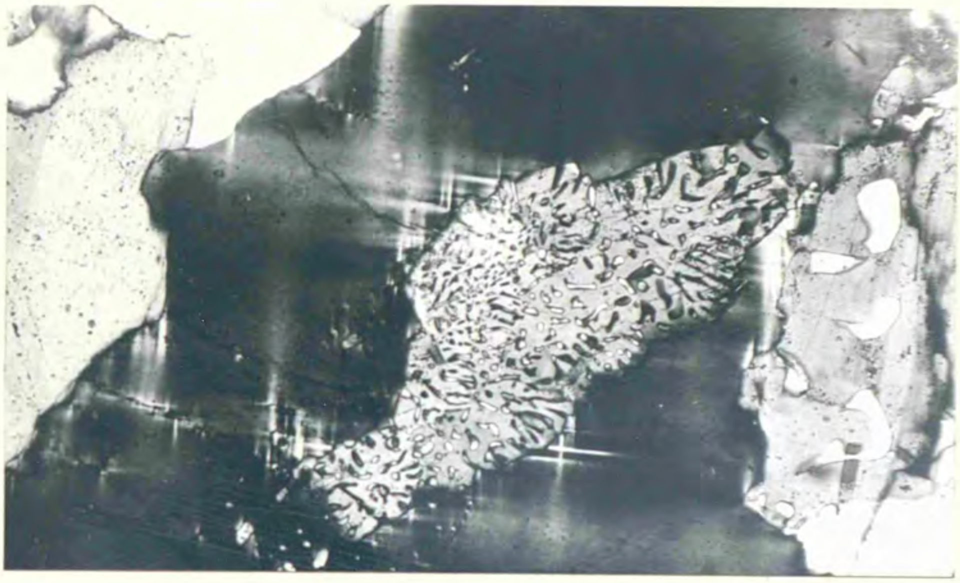
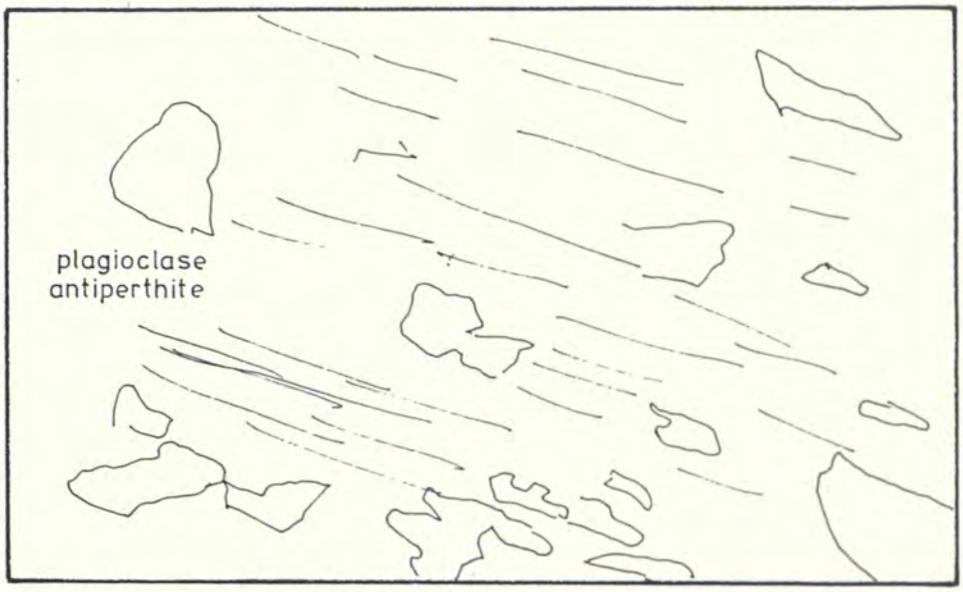


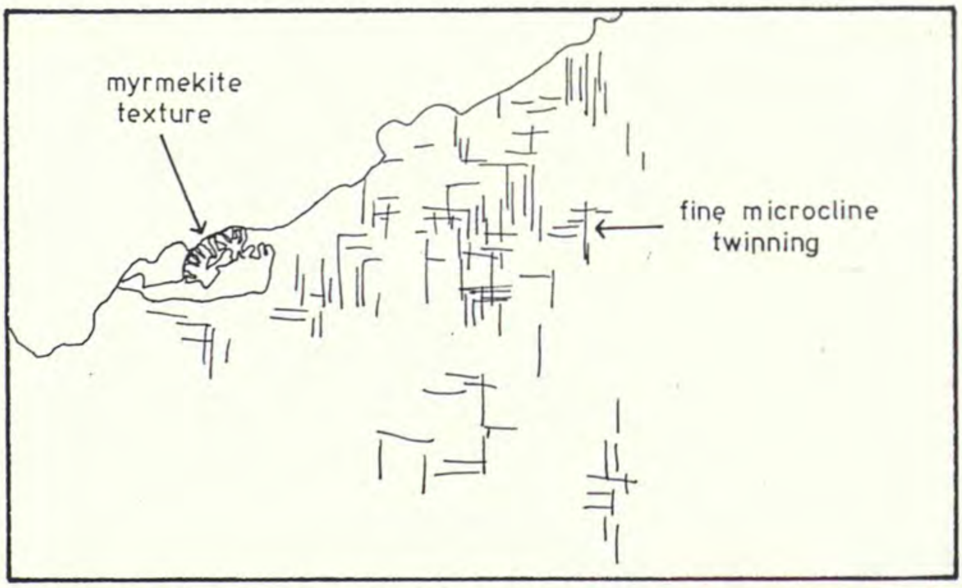
Figure 37 (cont). Photomicrographs to show feldspar textures in the Ardgour granitic gneiss. (see text for discussion).

(c)

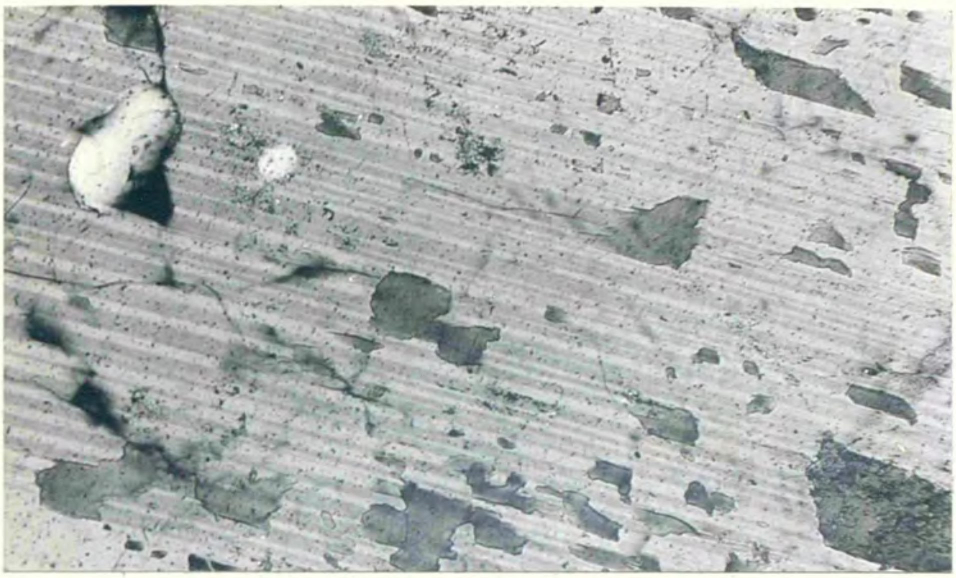


(Exp.280/1235). crossed polars. 0 mm. 0.5

(d)



(Exp.280/1235). crossed polars. 0 mm. 1



sections. Crystals are generally unzoned and fairly fresh with slight to moderate saussuritisation. crystal shapes are always anhedral, quartz tends to be convex towards plagioclase.

Plagioclase/K-feldspar interfaces are often myrmekitic, plagioclase being convex towards K-feldspar with quartz worms within the plagioclase (see Fig.37a). There are no large areas of replacement myrmekites of the type described by Ashworth (1972). Where there are no obvious K-feldspars of myrmekites in a section plagioclase often contains small globular quartz inclusions (Fig.37b). Plagioclase antiperthites, with a chequered appearance are fairly common (Fig.37c).

The k-feldspar does not contain well developed microcline cross-hatched twinning (in common with that in the psammites and semi-pelites) but it shows diffuse fine cross-hatching which is more pronounced adjacent to thin fractures and veins within the crystals (Fig.37d).

In some sections there are thin planar zones of fine grained quartzo-feldspathic material (0.1 mm. wide) which are thought to have formed as a result of deformation. Myrmekites, often found in the same sections, do not show any preferred orientation.

In the deformed pegmatites both plagioclase and K-feldspars are abundant, grain size is larger, up to 1 cm., and both myrmekites and antiperthites are common.

Biotite.

Biotite is abundant in nearly all samples (see Fig.36). The laths which are generally fresh, range in length from 1-2 mm. and define an undulose diffuse foliation. Only rarely are they segregated into discrete planar mica layers. The laths are often wrapped around small quartzo-feldspathic augen. In some exposures biotite crystals are seen which are co-planar with the early pegmatites and which (together with the pegmatites) have been folded and deformed into the dominant (S_2) gneissose foliation. In rare instances an early gneissic foliation can be seen which together with the early pegmatites is isoclinally folded and reworked during the formation of the dominant gneissic foliation (Fig.52). Thus the gneissic foliation is a composite foliation although more often than not this cannot be proved at an individual exposure.

Later folds of the gneissic foliation produce microfolds in which the biotite laths are rotated to lie nearly parallel to the microfold limbs but as there are very few micaceous layers the microfolds are rarely seen as crenulations, so typical of the adjacent pelites and semi-pelites. In all sections biotite crystals are un-strained.

Muscovite.

Muscovite, which is present in small quantities in some of the samples (Fig.36), forms up to 8.8% of the modal analyses and has two distinct modes of occurence. It occurs intergrown with biotite, enhancing the mica foliation of the gneiss and, like biotite, does not show undulose extinction. It also occurs growing as non-orientated, non-elongate crystals within plagioclase crystals.

Garnet.

Garnet is found in small quantities in about half of the sections examined (Fig.36). The crystals are small (1-2 mm.), round and contain small numbers of small round quartz inclusions with fewer iron ore inclusions. The inclusions tend to be concentrated towards the centre of the garnets. No strong zoning or inclusion trails have been noted. A few of the garnets are larger (3-4 mm.) and contain relatively large, irregularly shaped quartz inclusions. Frequently the garnets are partly replaced by aggregates of biotite and feldspar with or without muscovite and epidote.

Other minerals.

There are relatively small proportions of other minerals in the gneiss, these include chlorite,growing at the expense of biotite, epidote, iron ore, sphene, zircon and apatite.

C H A P T E R 5.

Structure.

5.1) General introduction.

A detailed structural analysis of the area, traversing all three of the major divisions of the Moine succession, has not previously been attempted. The area to the west has been the subject of study and controversy since the 1930's (see Chapter 2.1) and the area east of Glenfinnan village was the subject of a structural study of the Ardgour granitic gneiss, the study extending southwards from Glenfinnan towards Strontian (Dalziel,1966 and Dalziel & Johnson,1963).

Within this chapter the area is divided into a number of sub-areas and the structure of each sub-area discussed in detail. Chapter 7 is a discussion of the structural and metamorphic geology of the Sgurr Beag slide, Chapter 8 details the intrusive and deformational history of the suite of microdiorite sheet intrusions. Chapter 9 synthesises the complete structural history of the area and examines it in the light of proposed models of the structure of the NW Scottish Highlands.

The general attitude of the rocks within the area is indicated on Maps 2,3 & 4 which show that throughout the western part of the area the rocks strike NE-SW and have nearly vertical dips. Further east, in the Loch Eil Division, the amount of dip decreases rapidly and the dip direction varies from NE to SE in a regionally developed "flat belt".

5.2) Recognition of fold phases.

Detailed mapping within the area has revealed the presence of four fold phases, and the possibility of a fifth phase, whilst correlations with adjacent areas (Powell et al,1981:Baird,1982) reveal the existence of a fifth fold phase. Each of the mapped fold phases is characterised by several features and the recognition of structures as belonging to one of these fold phases is based on the following criteria;

1) Trend of major and minor fold axial planes and hinge lines.
2) Asymmetry of minor folds.
3) Structures associated with minor folds, eg. axial planar mica fabrics, axial planar migmatitic fabrics, crenulation of planar fabrics etc.
4) Geometrical style of minor folds.
5) The re-folding of structures of one phase by those of a later phase.
6) Relationship of folds to the Sgurr Beag slide.

None of the criteria listed above is diagnostic by itself, but a consideration of all of the criteria usually enables folds to be assigned to one of the episodes of folding.

The area contains rocks of all three major divisions within the Moine succession. Because of the variety of suggested and implied correlations between fold phases in the different divisions (see, for example, Johnstone et al.1969; Piasecki,1980; Lambert et al.1979) the structure is described in a number of sub-areas. The sub-areas lie either totally within a division or across the junction between adjacent divisions. Correlations are then made between the structural histories in each of the divisions and the significance of the junctions is discussed.

5.3) Loch Eil Division.

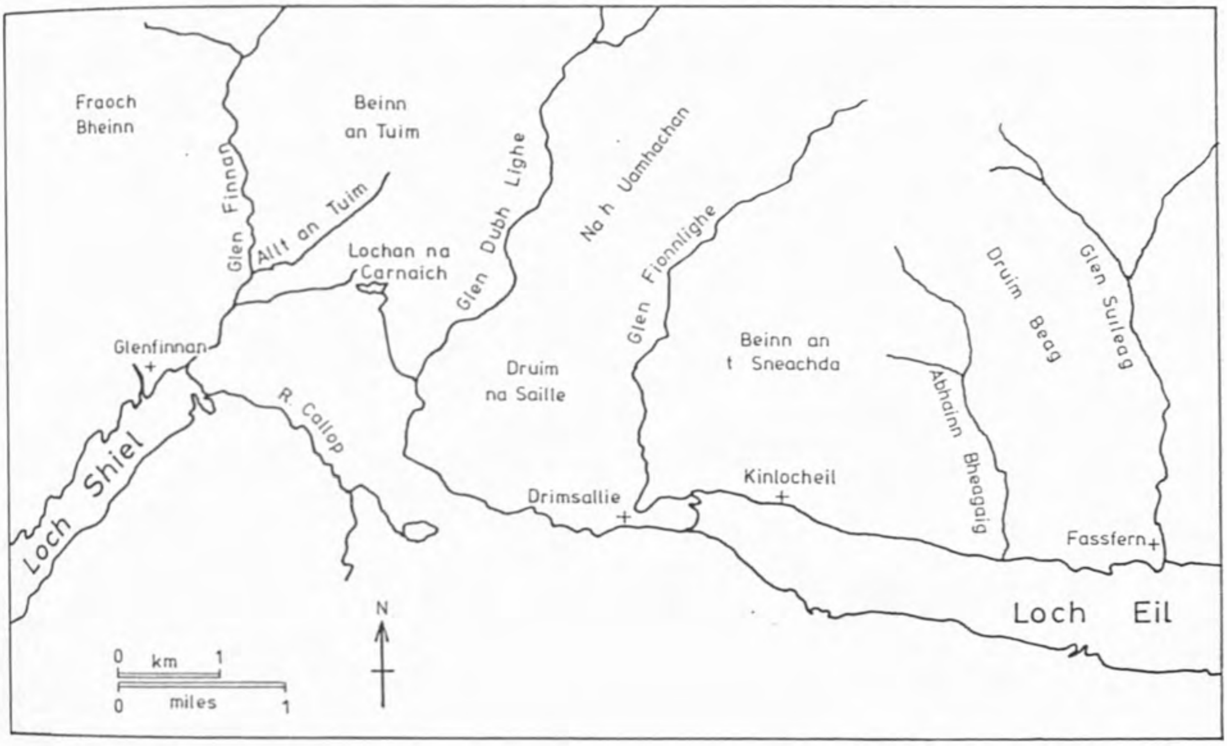
5.3a) Structural description.

The rocks of the Loch Eil Division form the most easterly sub-area and comprise banded psammites and quartzites with no mappable pelitic horizons. They are the least well exposed rocks in the area, exposure is reasonable on Beinn an t-Sneachda (Fig.38) but decreases eastwards. In Glean Suileag exposure is good in the stream section but is very poor elsewhere. Much of the eastern area is forested and the exposure from here eastwards to the Great Glen fault is generally very poor.

It is clear (Map 2) that the Loch Eil Division psammites structurally overlie the rocks of the Glenfinnan Division. The few younging directions obtained from the cross-bedded units within the banded psammites (Map 2) indicate that the psammites young eastwards and upwards away from the Glenfinnan Division. Most of the psammites are finely banded with a bedding parallel schistosity strongly developed within fine micaceous laminae separating the quartzo-feldspathic layers.

Bedding or structurally modified bedding dips generally eastwards, (Fig. 39a) but in detail shows a great circle distribution about an axis plunging

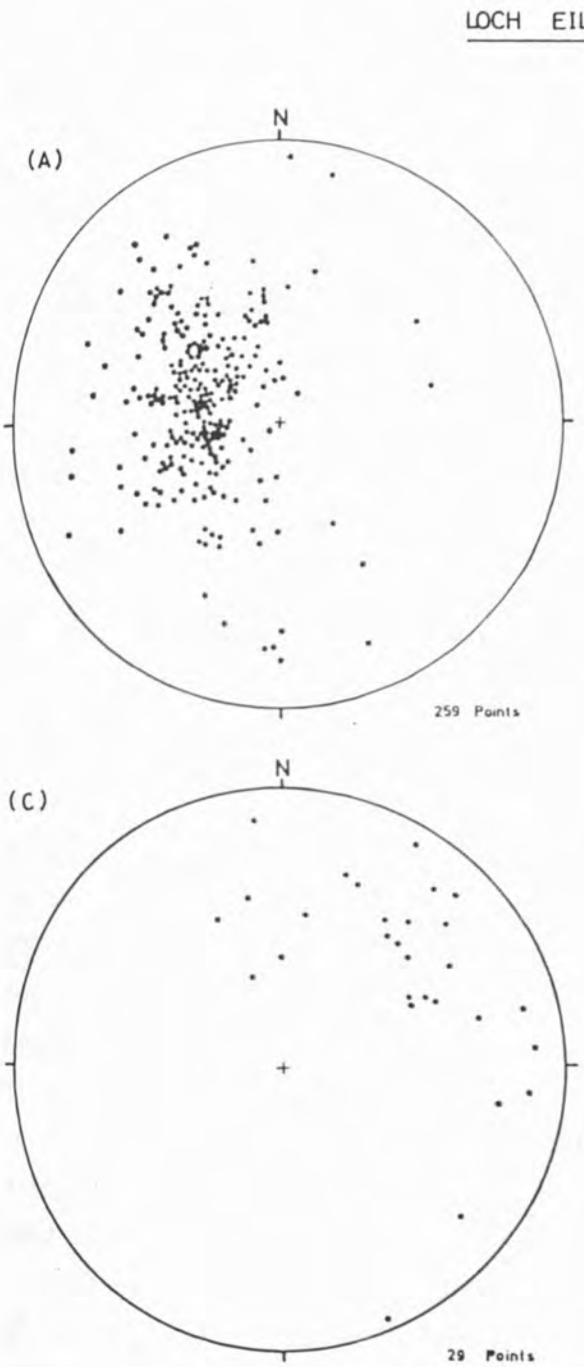
Figure 38.



Map of the main place names, rivers
and hills east of Glenfinnan.

Figure 39.

Stereonets drawn from data collected from the Loch Eil Division.



(A) Poles to bedding.

(B) Hinge lines of minor F_3 folds (Druim Beag Synform)

(C) Hinge lines of minor F_2 folds (Beinn an Tuim Synform)

(D) Hinge lines of minor F_4 folds (Glen Dubh Lighe Antiform) shallowly to 100°. The spread of orientation of bedding is related to the last phase of deformation which has produced a series of very open large scale warps which have wavelengths of up to 1 Km. The hinge lines plunge shallowly to the E or ESE. No minor structures were seen which could be related to these large scale open warps which are F_5 within the local deformation sequence discussed below.

Map 2 shows the axial plane of a major synform, the F_3 Drium Beag synform which crops out in the extreme east of the area mapped. The axial plane of the fold dips shallowly to the SE and the plunge of associated minor folds is very shallow towards the SW.(Fig.39b). The fold is virtually isoclinal (see Map 2). Minor F_3 folds related to the major synform are very common on the flat top of Druim Beag and show vergence clearly related to the major synformal axis. Where the rocks are semi-pelitic thin sections show that the minor folds have produced very tight crenulations of a pre-existing planar mica fabric. There are some strongly zoned garnets related to this planar fabric which have acted as rigid porphyroblasts during the crenulation event, deflecting the crenulation axial planes of the later (F_3) folds.

Psammites west of the axial plane trace of the F_3 Druim Beag synform younging upwards and eastwards indicate that the F_3 Druim Beag synform faces upwards. East of the axial plane trace in the valley of Glean Suileag in two places the rocks still young upwards and to the east, unfortunately the exposures do not contain minor F_3 folds and it is suspected that these two exposures may be on the short limbs of minor folds related to the major synform. Strachan (1985) mapping a larger area of the Loch Eil Division, mapped an F_3 antiform, the Stronchreggan antiform, immediately to the east of the F_3 Druim Beag synform. He mapped these folds as "upright" although Strachan (1983,Table 2) shows that the axial plane of the F_3 Druim Beag synform trending to 050° and dipping at 0-80° to the SE or NW!

Towards the west of the Loch Eil Division (Map 2) the axial plane traces of two sets of folds have been mapped. The earlier fold has (F_2) has been termed the Beinn an Tuim synform (Dalziel,1966) and a series of later crenulation folds (F_4) have been termed the Glen Dubh Lighe antiform (Dalziel, 1966).

The minor folds associated with the Beinn an Tuim synform plunge shallowly to moderately to the NE (Fig.39c) and have steep to vertical axial planes. The minor folds in more pelitic lithologies have axial planar penetrative mica fabrics. The fold limbs are relatively open in the psammites of the Loch Eil Division, but the trace of the F_2 axial plane continues WSW into the rocks of the Glenfinnan Division where the fold limbs rapidly tighten to become isoclinal. Within some of the pelitic lithologies of the Glenfinnan Division near to the Loch Eil Division psammites the rocks are gneissose with a migmatitic fabric which is axial planar to minor folds associated with the F_2 Beinn an Tuim synform. The migmatitic (or gneissose) pelites of the Glenfinnan Division outcrop as a thin strip just east of the stream in Glen Dubh Lighe (Map 2. extreme top left) and grade eastwards into non-migmatitic pelites which themselves become more psammitic as they pass eastwards, apparently conformably, into the psammites of the Loch Eil Division (but see Strachan,1982).

In the psammites of the Loch Eil Division the F_2 folds fold a composite bedding/schistosity fabric in which the micaceous laminae separating thicker quartzo-feldspathic laminae have strong bedding-parallel schistosity. Rarely there are minor isoclines within this bedding/schistosity foliation.

The axial plane traces of the later (?)F_4 folds in the Glen Dubh Lighe-Glen Fionn Lighe area (Map 2) are vertical and trend NNE-SSW. The plunge of associated minor crenulation folds is generally shallow to the NNE (Fig. 39d). There is a series of (?)F_4 folds in the area which,in total, produces an open "anticlinorium" here termed the Glen Dubh Lighe antiform. The shorter axial plane traces, indicated on Map 2 are inferred from minor fold vergences. The pelite/psammite junction is not well enough exposed to justify the drawing of a suitably crenulated lithological junction.

The minor folds produce well developed open crenulations of the earlier penetrative mica fabric (Fig.40). The interlimb angles of these crenulations are always open (interlimb angles approx.120-150°) even where they are crenulating the pelitic lithologies of the Glenfinnan Division. Type 1 re-fold patterns (Ramsay,1967) are common where (?)F_4 minor folds re-fold F_2 minor folds (Fig.41).

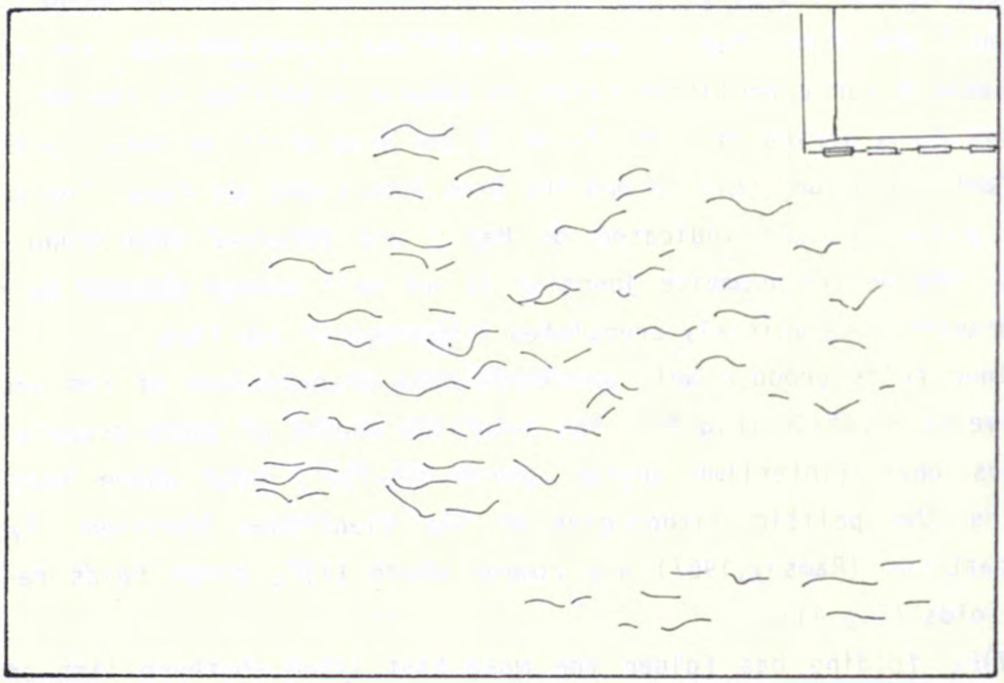
The (?)F_4 folding has folded the more flat lying southern limb of the F_2 Beinn an Tuim synform whereas the orientation of the near vertical northern limb of the F_2 Beinn an Tuim synform is much less affected. This refolding appears to be at least partly responsible for the opening up eastwards of the interlimb angles of the synform in the psammites of the Loch Eil Division.

5.3b) The relative ages of the folds in the Loch Eil Division.

In the preceding paragraph the Druim Beag synform and the Glen Dubh Lighe antiform have been described, arbitrarily, as of F_3 and (?)F_4 ages respectively without any consideration of their relative ages.

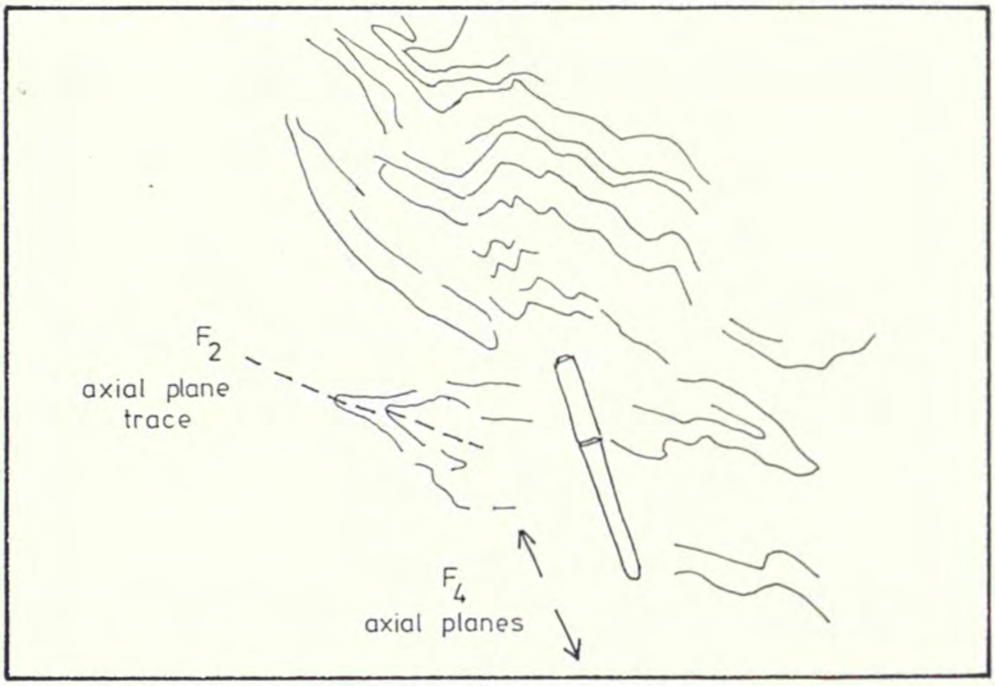
The Beinn an Tuim synform has an axial planar penetrative fabric in

Figure 40. Photograph of F_4 crenulations in the Druim na Saille pelite.



(Exp.208 NM 95078117).

Figure 41. Photograph of minor F_2 folds re-folded by F_4 folding.



(Exp.681 NM 92598295).

its core but on its limbs it is seen to fold a pre-existing bed-parallel schistosity (S_1). Therefore the Beinn an Tuim synform is considered to be at least F_2 in age.

The minor folds of the Glen Dubh Lighe antiform are frequently seen to fold minor F_2 folds and to crenulate a pre-existing planar mica fabric.

The Druim Beag synform is a very tight to isoclinal "crenulation" fold; it is seen to crenulate a pre-existing planar mica fabric.

Therefore it can be argued, most simply, that as both the Glen Dubh Lighe antiform and the Druim Beag synform crenulate a pre-existing planar fabric, they could be considered to be of the same age, ie. an F_3 crenulation of an F_2 fabric. Consequently, using this argument, the latest warps would become F_4 in this numerical sequence.

However, this simple correlation does not consider the geometries of both folds. The Druim Beag synform is a very tight to isoclinal fold with a shallow ESE dipping axial plane. The associated minor folds have hinges which plunge to the SW and develop a tight axial planar mica crenulation fabric. The adjacent Glen Dubh Lighe antiform is a very open series of folds with vertical axial planes and open interlimb angles. The minor folds plunge shallowly to the NE and associated micro-crenulations are relatively open, even in highly pelitic lithologies.

The obvious differences in the fold geometries leads one to suggest a difference in age between the two folds, with the nearly isoclinal Druim Beag synform being earlier than the much more open Glen Dubh Lighe antiform. Consequently the Druim Beag synform has been termed F_3 and the Glen Dubh Lighe antiform F_4 in the description above.

It is conceivable that if the Glen Dubh Lighe antiform had tightened up along its axial plane then it would have become geometrically indistinguishable from the Druim Beag synform. However in the area studied it must be assumed that if the crenulation folds are nucleating and developing as a series of folds within a single phase of deformation,then the Druim Beag synform being the more intensely developed, would be relatively older than the Glen Dubh Lighe antiform but possibly only by a short period of time.

Arguments are presented later concerning the regional development of fold geometry which also suggest that the Druim Beag synform is an F_3 fold and the Glen Dubh Lighe antiform is an F_4 fold (Chapter 7.5 & 9), and that there may be a considerable time interval between the two phases of deformation (Chapter 8).

Alternatively, from the preceding arguments it is possible to suggest that, since both the Druim Beag synform and the Beinn an Tuim synform are older than the Glen Dubh Lighe antiform, then both folds could be of the same age. However this is extremely difficult to visualise geometrically since the folds are adjacent and both are tight synforms which have very dissimilar geometries and open out towards each other. It can also be noted that the Beinn an Tuim synform locally develops an axial planar mica fabric whereas the Druim Beag synform always develops a tight crenulation mica fabric.

5.3c) Summary of the structural sequence within the Loch Eil Division.

1) Development of a bed-parallel schistosity with very rare intrafolial minor isoclines.
2) F_2 major and minor folds with locally developed axial planar penetrative mica fabrics (and axial planar migmatitic fabrics in some of the adjacent pelites).
3) F_3 tight to isoclinal major crenulation folds. Axial planes trend NE-SW and dip to the SE. Hinges plunge shallowly towards the SW.
4) F_4 relatively open upright crenulation folds with axial planes trending NNE-SSW to NE-SW. Minor fold hinges plunge towards the NNE.
5) Very open large scale F_5 warps which plunge shallowly eastwards.

This summary does not include faulting which is considered within the whole mapping area at the end of this chapter. The structural setting of a suite of deformed microdiorite sheets is considered in Chapter 8.

5.4) Glenfinnan Division east of the Beinn an Tuim fault.

5.4a) Introduction.

The rocks in this area show the greatest degree of lithological variation and are the most complexly deformed of the whole region mapped. They include the most northerly portion of the Ardgour granitic gneiss, pelitic and striped lithologies of the Glenfinnan Division and an infold of the Loch Eil Division psammites occurring west of the granitic gneiss, near Glenfinnan village (see Map 3).

The quality of exposure is extremely varied in the area, the eastern slopes of Glen Finnan, the southern slopes of Beinn an Tuim above Lochan na Carnaich and the roadside exposures for 3 Km. east of Glenfinnan village are very well exposed. Elsewhere exposure is not so good, especially in Glen Dubh Lighe which is heavily forested. Consequently the interpretation of fold axial plane traces and hinge line trends is partly based on regional correlations and extrapolations to the hills south of Glenfinnan,mapped by Dalziel in 1963.

5.4b) Lithological distribution.

Map 3 shows the distribution of lithologies in the area. Typical striped lithologies of the Glenfinnan Division crop out high on the southern slopes of Beinn an Tuim, these are succeeded southwards by a strip of much more mixed rock types. Outcrops are mainly pelitic with some more striped pelite and psammite together with at least three thin strips of granitic or highly migmatitic gneiss. A forth highly migmatitic or granitic gneiss outcrop is exposed in the stream bed of the river Dubh Lighe north of Wauchan bothy (NM 945820), it extends northeastwards out of the area mapped and appears to continue northwards to the granitic gneiss mapped at the western end of Loch Arkaig (I.G.S.1:50,000 Sheet 62W, Scotland).

The strips of granitic gneiss can each be traced a few hundred metres along strike. They do not appear to be connected to each other nor are they at precisely the same tectono-stratigraphic level.

Loch Eil Division psammites outcrop in the east of the area and occupy the core of the Beinn an Tuim synform, the axial plane trace of which can be traced into the banded pelites outcroping below and south of the striped lithology on Beinn an Tuim. The psammites are also folded by the later F_4 Glen Dubh Lighe antiform.

The Loch Eil Division psammites pass westwards into a thick pelitic unit, termed the Druim na Saille Pelite, which has a wide outcrop on the hillside between Glen Dubh Lighe and Glen Fionn Lighe (NM 9580). This relatively homogeneous pelite is continuous with and grades into the more varied pelites and psammites on the slopes of Beinn an Tuim (NM 9382).

The Ardgour granitic gneiss occupies most of the southern and western portion of the area. Within the granitic gneiss a large area of recognisable metasediments (mostly fine grained pelites and striped pelitic and psammitic lithologies) outcrop south of Lochan na Carnaich. A smaller outcrop of striped metasediments is found within the granitic gneiss on Mam Chreagain.

SW of the granitic gneiss there are thin strips of the striped lithology, pelite and then psammite which has been correlated by Dalziel (1966) with the psammites east of the granitic gneiss, ie. the Loch Eil Division psammites.

The distribution of lithologies is in a broad arcuate pattern in the southern part of the area, with the trend swinging from E-W in the west, through SE-NW to N-S in the south of the area. The eastern part of this arcuate outcrop has already been described as the Glen Dubh Lighe antiform (F_4) in the Loch Eil Division.

5.4c) F_4 folding.

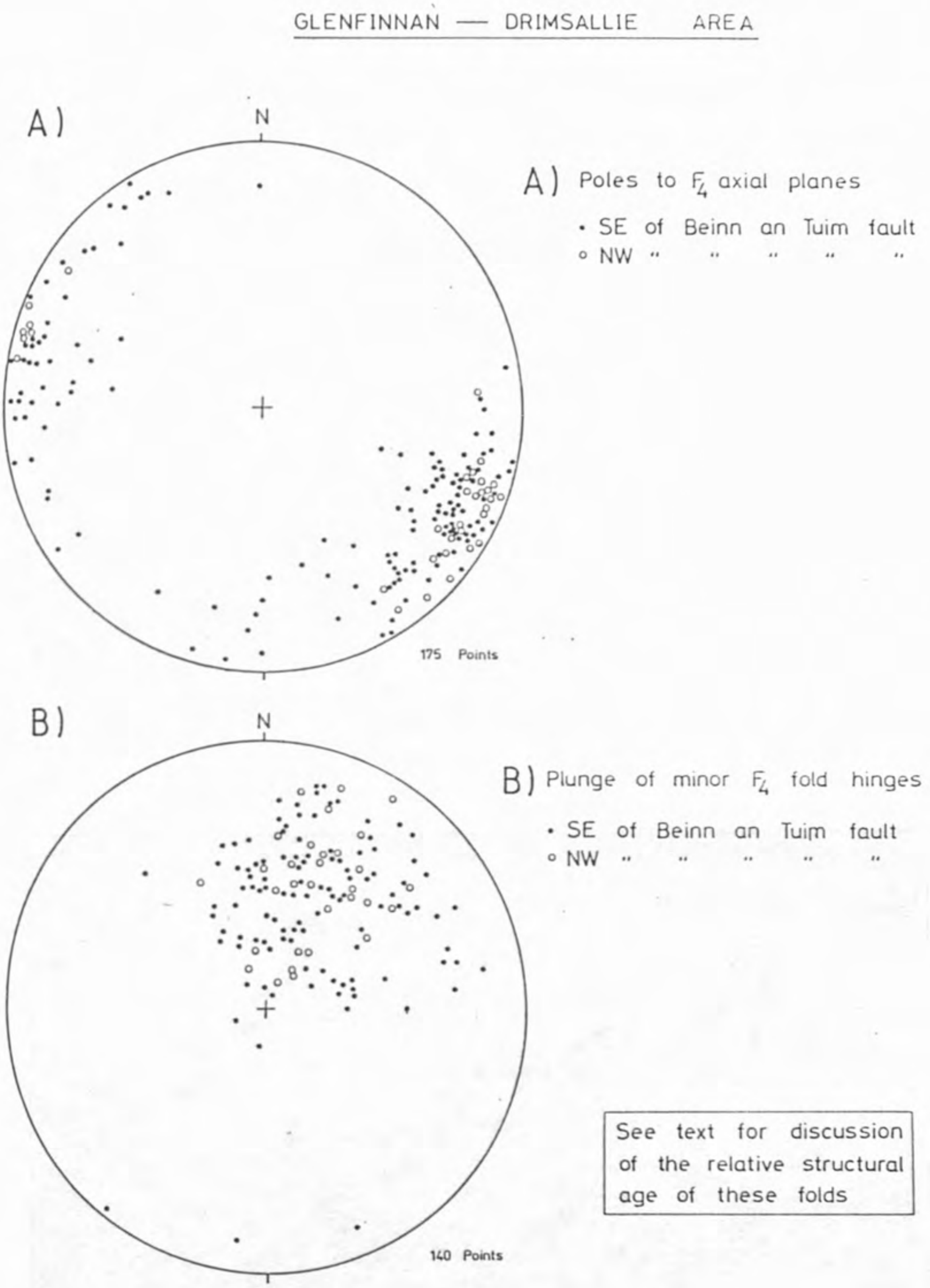
Map 3 shows the trend of bedding and S_2 foliation in the area. These planar elements are clearly seen to be deformed into two arcuate shapes (see also Fig.45). A series of F_4 minor crenulation folds have been mapped which appear to be related to these arcuate patterns.The axial planes of the minor folds strike approximately NE-SW and have near vertical dips (Fig.42a). The hinge lines of the F_4 minor folds plunge variably towards the NE (Fig.42b). The F_4 folds produce moderate to tight crenulations of the planar mica fabric (S_1/S_2) in the pelitic horizons. In the striped lithologies the psammitic bands are folded into moderately tight (interlimb angles 60-90°) generally asymmetric folds with wavelengths from a few tens of cms. up to 2-3 metres (Fig.43a). Within the granitic gneiss F_4 folds the dominant foliation (S_2) into moderately tight folds (Fig.43b). No development of an axial planar penetrative fabric has been observed in any of these folds.

Figure 44b shows that the F_4 minor fold axial planes have a fanning spread of orientation about a "common axis" which is vertical. The mean trend of these vertical planes is approximately NE-SW and may define the XY plane of the D_4 strain ellipsoid. One might assume that the "common axis" about which this fanning occurs is the Y axis of the D_4 strain ellipsoid, in which case the orientation of the D_4 strain ellipsoid would be constrained to; Y axis vertical; X axis horizontal,NE-SW; and Z axis horizontal, NW-SE.

By analogy with fanning cleavage around an individual fold (Fig.44a) where one observes that the "common axis" of the cleavage fan coincides with the hinge line of the fold, one would expect the "common axis" of the F_4 minor fold axial planes to coincide with the mean trend of the F_4 minor fold hinges. Examination of Figures 42a & b shows this not to be the case. However this analogy is considering the simplistic case where the initial planar surface is co-planar with the YZ plane of the imposed strain ellipsoid.

The spread of orientation of F_4 minor fold hinge lines may be due to

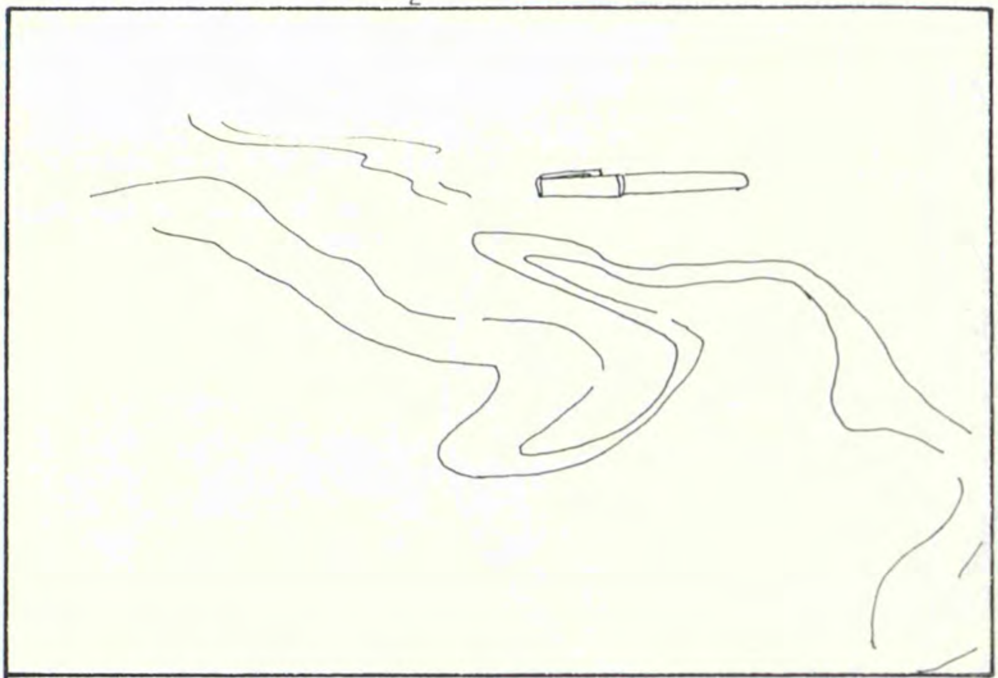
Figure 42.



Stereonets from the Glenfinnan-Drimsallie area.

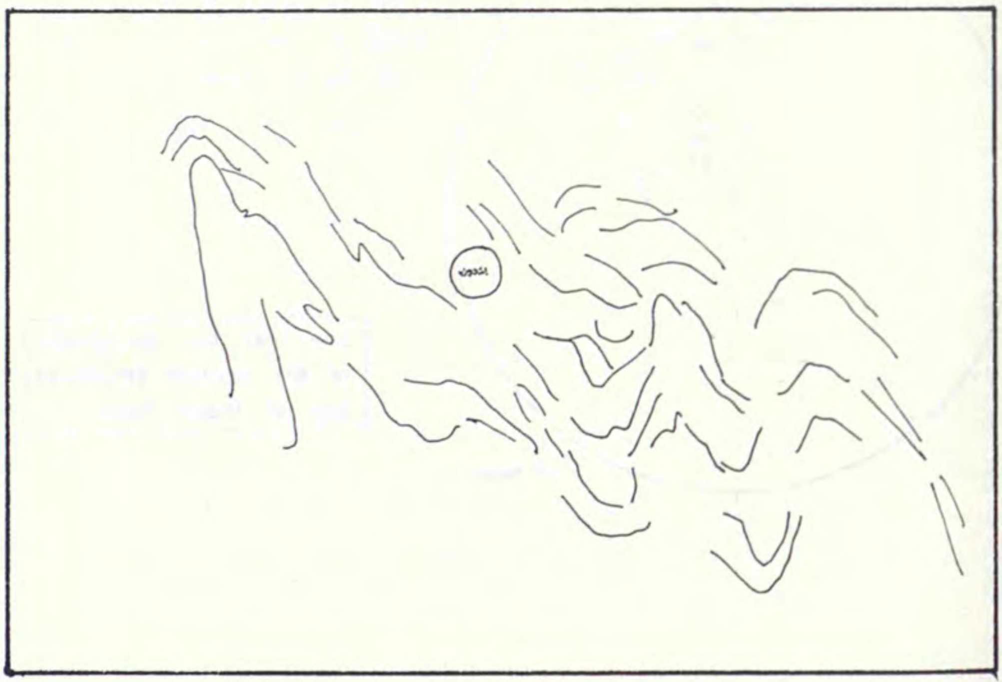
Figure 43. Photographs of minor F_4 folds.

(a) Re-fold of a minor F_2 fold.



(Exp.622 NM 93448257).

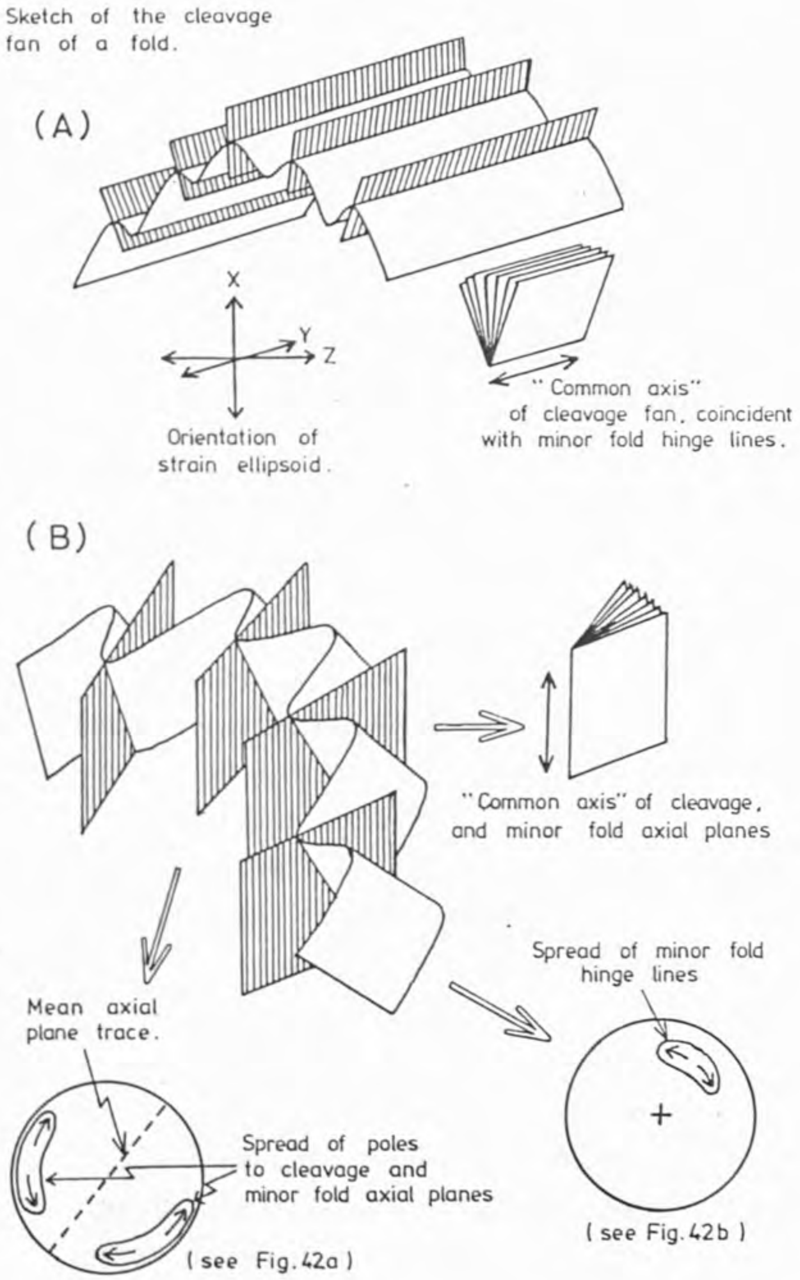
(b) Minor F_4 folds in the Ardgour granitic gneiss.



(Exp.1306 NM 91568030).

Figure 44.

Sketches to show the geometry of the F_4 Glen Dubh Lighe antiform.



a combination of three factors. Firstly, prior to D_4 deformation the surfaces may have had variable orientation so that the F_4 minor fold hinge lines produced were not co-linear, but lie within the XY plane of the D_4 strain ellipsoid. Secondly, heterogeneous progressive deformation may have rotated the F_4 minor fold hinge lines by varying amounts within the XY plane towards the X axis of the D_4 strain ellipsoid. Thirdly, since the F_4 minor fold axial planes have a fanning distribution about the XY plane of the D_4 strain ellipsoid, it follows that F_4 minor fold hinges which are on this fan of F_4 minor fold axial planes will have a spread of distribution which fans about the XY plane, trending towards the Z axis.

Thus these three factors, when considered, will produce a spread of orientations within the XY plane and towards the Z axis of the D_4 strain ellipsoid, possibly producing a distribution pattern similar to that of Figure 42b.

It would seem that the fanning spread of orientation of F_4 minor fold axial planes can be used to define the orientation of the imposed D_4 strain ellipsoid whereas the orientation of the F_4 minor fold hinge lines can be used, at most, to define a linear element within the XY plane of the ellipsoid.

An alternative explanation for the distribution of F_4 structural elements (Fig.42) is that the spread of distribution is due to subsequent (F_5) folding.Prior to the later folding the F_4 minor fold axial planes may have had a constant orientation (vertical, NE-SW) and the F_4 minor fold hinge lines a spread of orientation within the vertical NE-SW F_4 axial planes.

Although no folds obviously later than the F_4 minor folds have been recorded in the area, further east in the Loch Eil Division there is a weak phase of D_5 deformation, and in the extreme west of the area mapped, at the eastern end of Loch Eilt, there are numerous late minor folds which fold the local F_3 crenulation folds and the regionally developed F_4 major folds (see Chapter 5.6c and Chapter 7).

If one postulates that part of the spread of orientation of F_4 minor structures is due to D_5 deformation it becomes impossible to define the exact orientation of the D_4 strain ellipsoid since the orientation of the X and Y axes within the XY plane of the D_4 strain ellipsoid cannot be established.

The orientation of the regional D_4 strain ellipsoid is considered further in Chapter 8.1b where the D_4 deformation of a suite of post-D_3/pre-D_4 microdiorite sheet intrusions is discussed.

The F_4 Glen Dubh Lighe antiform is part of the large fold which produces the arcuate outcrop pattern of the area. The orientation of F_4 minor fold hinges and F_4 minor fold axial planes have been related to this major fold, however if the vergence of minor F_4 folds is studied (see Map 3) it is seen that the F_4 minor fold vergence shows very little correspondence to the geometry of the major fold. However it can be argued that the minor folds are parasitically disposed about larger folds which themselves parasitic to the major Glen Dubh Lighe antiform. It is noted that the outcrop pattern of the gneiss and its contained metasediments around Lochan na Carnaich appears to show folding on a scale intermediate between the F_4 minor folds and the regional arcuate Glen Dubh Lighe antiform.

It is interesting to examine the vergence of F_4 minor folds in the east of the area on either side of the trace of the F_2 Beinn an Tuim synform (see Fig.45). South of the F_2 axial plane the F_4 minor folds plunge NNE and verge to indicate the presence of a synform to the east. North of the F_2 axial plane the F_4 minor folds again plunge to the NNE but verge to indicate the presence of an F_4 antiform to the east. This apparent anomaly can be explained by envisaging the two limbs of the F_2 Beinn an Tuim synform being geometrically on either side of the XY plane of the D_4 strain ellipsoid and consequently, during D_4 deformation, rotating in opposite directions towards the XY plane of the ellipsoid.

Thus a combination of fold generation on different limbs of the F_2 Beinn an Tuim synform and rotation of opposite limbs in opposite directions towards the D_4 XY plane would produce F_4 minor folds with the opposite senses of vergence on each limb of the F_2 fold.

Further, the opening up of the F_2 Beinn an Tuim synform during D_4 deformation requires that prior to the onset of D_4 deformation the F_2 fold was not isoclinal in this particular area.

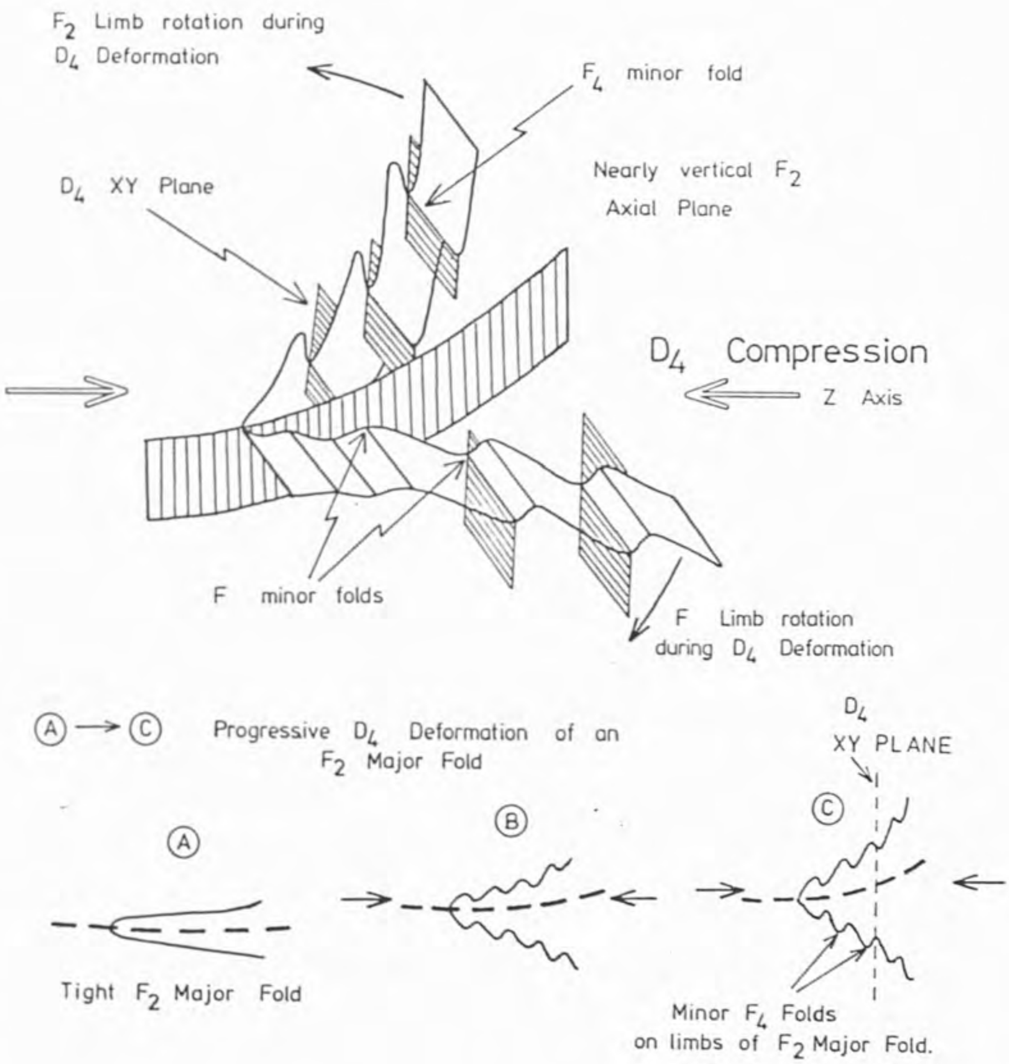
5.4d) D_2 deformation.

5.4d1) Pattern of S_2 foliation.

Map 3 shows the pattern of S_2 foliation planes in the area. The F_2 Beinn an Tuim synform and F_2 minor folds are isoclinal or nearly so within the area and the S_2 foliation pattern is very similar to the pattern of lithological distribution. Only in the extreme NE of the area and eastwards into the psammites of the Loch Eil Division is this not the case. Here the regional F_2 Beinn an Tuim synform is a relatively open fold. The planar mica fabric in the Loch Eil Division psammites which is folded by the Beinn an Tuim synform, though initially mapped as S_2 is now considered to be S_1 foliation.

Figure 45.

Diagrammatic sketch to show the F_2 Beinn an Tuim synform "opening" during D_4 deformation.



The pattern of lithological distribution and S_2 foliation (see Fig.46) is controlled by the subsequent F_4 folding which has produced the arcuate patterns in the area.

5.4d2) Nature of the S_2 foliation.

The striped and banded psammites and pelites exposed on the slopes of Beinn an Tuim (Map 3 and Fig.38) contain a planar mica fabric in the pelitic units which, except in the cores of F_2 folds, is generally virtually co-planar with the lithological layering. The fabric is axial planar to minor tight to isoclinal folds (F_2) and is often crenulated by later folds related to the major F_4 Glen Dubh Lighe antiform (Fig.47a). In a few places the F_2 folds in the striped group are seen to fold not only bedding but also an earlier set of minor folds which themselves possess an axial planar penetrative mica fabric in the pelitic layers. At exposure 231 (NM 91278033), where both F_1 and F_2 minor folds can be distinguished, the fabric in the hinges of the minor F_2 folds appear penetrative, there are no signs of it being a very tight crenulation of the pre-existing S_1 planar fabric (Fig.47b).

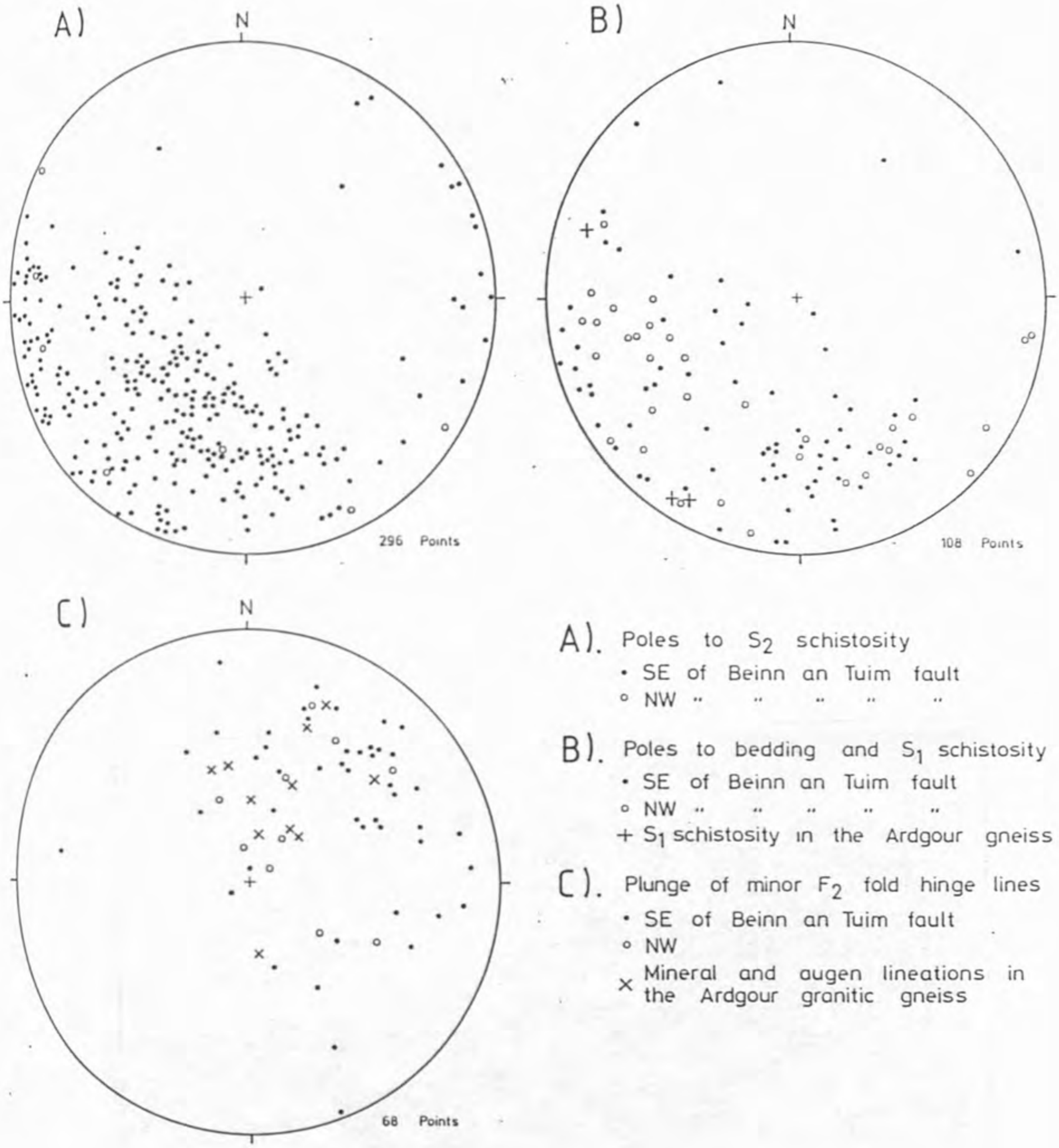
Locally the S_2 penetrative mica fabric contains lobate lensoid felds-pathic segregations, such rocks are termed migmatitic pelites. There are no indications that the segregations are very tightly developed crenulations of a pre-existing S_1 migmatitic fabric, although this could be the case if crenulation development was extremely intense. Thus in many areas S_2 is a true penetrative fabric which has totally re-worked and obliterated the earlier penetrative S_1 fabric.

In the homogeneous Druim na Saille pelite, the strong migmatitic fabric, comprising aligned micas and co-planar lensoid feldspathic segregations was initially mapped as the S_2 fabric even though bedding was frequently undefined within individual exposures. However as the fabric is folded by the F_2 Beinn an Tuim synform it is now re-interpreted as an S_1 fabric. Thus on this basis there appears to be evidence for two phases of migmatite development.

The Ardgour granitic gneiss, the structure of which is more fully con-sidered in Chapter 5.4e contains a very strongly developed foliation which is composed of micas co-planar with small lensoid quartzo-feldspathic seg-regations. In addition the gneiss contains abundant deformed lensoid pegmat-ites which are nearly co-planar with this dominant foliation. Frequently the pegmatites, which have biotite selvages, can be seen to be folded into this foliation. Very occasionally an earlier migmatitic foliation can be

Figure 46.

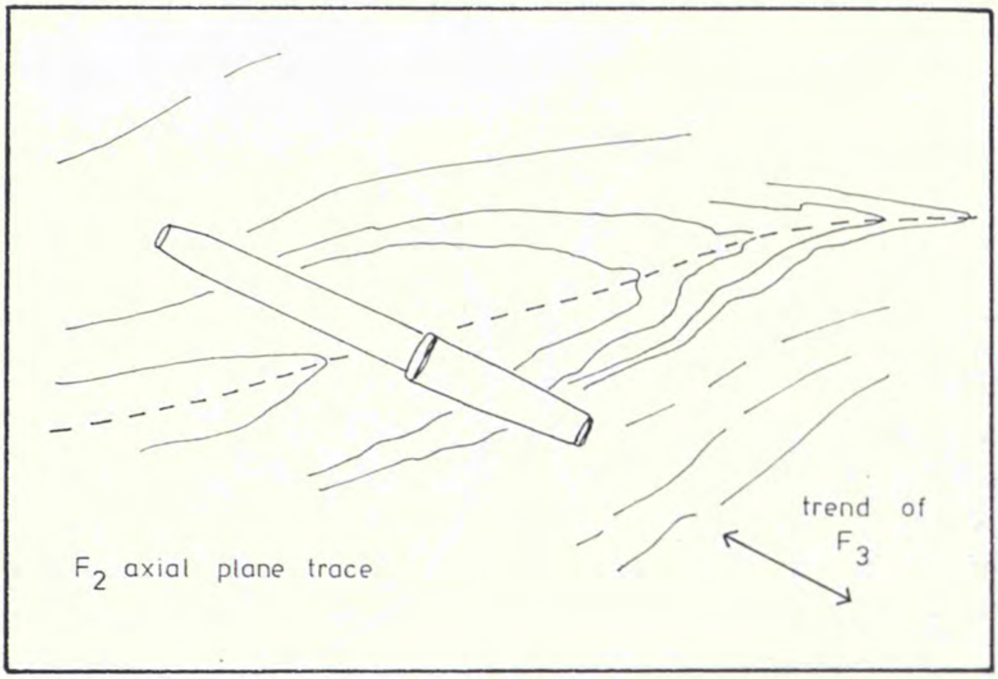
GLENFINNAN DRIMSALLIE AREA



Stereonets from the Glenfinnan-Drimsallie area.

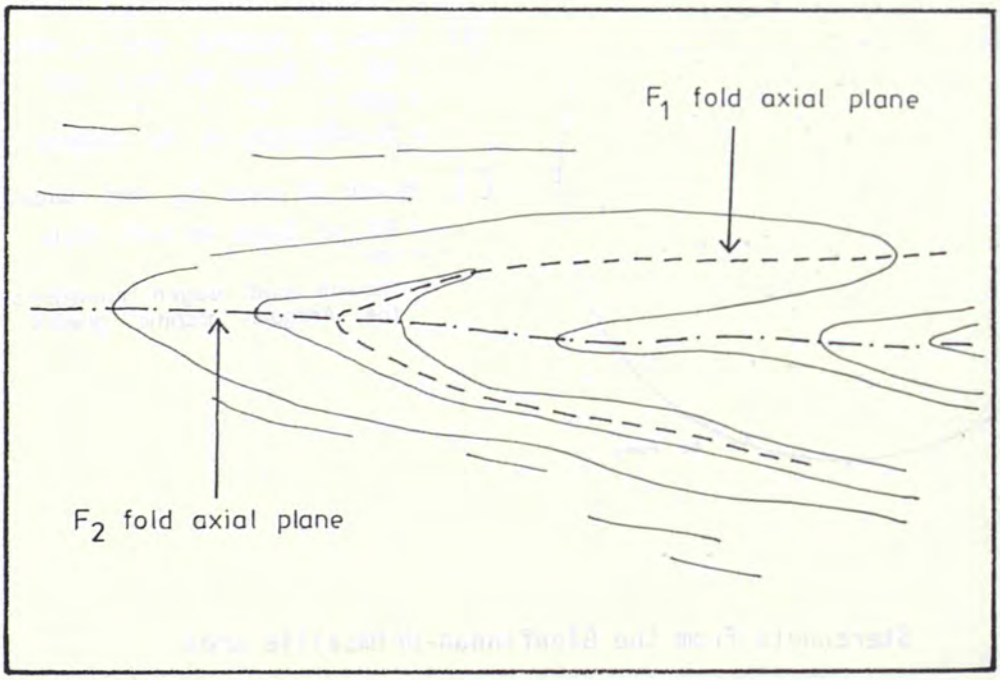
Figure 47. Photographs of F_2 minor folds.

(a) F_2 fold re-folded by F_3 crenulation fold.



(Exp.230 NM 91068047).

(b) F_2 re-fold of F_1 minor fold.



(Exp.231 NM 91278033).

seen which is highly deformed into near parallelism with the dominant (S_2) foliation (Fig.52).

Both the axial planar S_2 fabric of the Glenfinnan Division metasediments and the dominant foliation (S_2) of the Ardgour granitic gneiss have been deformed in the same manner by the folds of the F_4 Glen Dubh Lighe antiform. Throughout most of the area, especially within the exposure of the Ardgour granitic gneiss, the S_2 foliation is so strongly developed as to transpose most of the earlier structural elements and because minor fold vergences are often difficult to discern, it is very difficult to position accurately the traces of F_2 fold axial planes. In conclusion it is seen that the migmatisation is associated with both S_1 and S_2 foliations. D_2 deformation has re-worked and obliterated much of the evidence of the early S_1 foliation.

5.4d3) Geometry of F_2 folds.

In the typical striped lithologies of Beinn an Tuim and the psammites of the Loch Eil Division F_2 minor folds are relatively common. Within the granitic gneiss and associated metasediments around Lochan na Carnaich the S_2 foliation is well developed but minor folds are much less frequent. Much of the structural interpretation which follows is based on extrapolation from unfortunately little information. These inferences are then extended southwards by comparisons with the work of Dalziel (1963a,1966).

Beinn an Tuim synform.

The eastern end of this structure has already been described where it occurs in the Loch Eil Division. Here and in the metasediments in the upper Glen Dubh Lighe (Map 3) the fold is a relatively open synform which opens upwards and to the NE. Minor folds in all the non migmatitic or gneissose rocks have associated axial planar penetrative mica fabrics. In the migmatitic pelite and granitic gneiss on the eastern slopes of upper Glen Dubh Lighe (which is believed to be the southern end of the Loch Arkaig granitic gneiss) the fabric associated with the F_2 minor folds is an axial planar migmatitic fabric of co-planar mica foliae and oblate quartzo-feldspathic segregations.

The axial plane of the Beinn an Tuim synform can be traced westwards into the mixed pelites and striped lithologies which crop out to the north of Lochan na Carnaich (Map 3). The bedding planes in this area are co-planar with the S_2 foliation and the lithological outcrop pattern indicates that the fold is isoclinal. The fold axial plane can be traced and extrapolated westwards westwards until it is cut by the Beinn an Tuim fault.

Figure 48 is a sketch map to show the geometry of observed F_2 minor folds together with the orientations of inferred F_2 minor folds and fold axial planes (Fig.46c shows the orientation of F_2 minor fold hinge lines). The Beinn an Tuim synform over most of its outcrop has minor folds whose hinges plunge to the E and NE (eg.Fig.48 folds(1) and (2)) but when traced westwards along its axial plane the major fold has minor folds which have hinge lines which have passed through the vertical (Fig.48 fold (3)). The Beinn an Tuim "synform" still opens to the east however since the minor folds plunge westwards the fold is now geometrically an antiform(Fig.49)

In the area west of Lochan na Carnaich (Map 3,NM 9282) the few F_2 minor folds which have been observed, when taken together with the geometry of S_1/S_2 intersections and rodding within the granitic gneiss, indicate that there are a series of isoclines in the area rather than just the cores of two major folds, the Beinn an Tuim "synform" and the Meall nan Damh fold (Fig.48,folds(3),(4),(5),(6) etc.).

Dalziel (1963a and 1966) has described a fold, the Beinn an Tuim antiform outcropping to the north of the Beinn an Tuim synform, SE of the Beinn an Tuim fault and trending ENE-WSW. This fold closure was not found in the area, although it is possible that its axial plane outcrops further to the north.

Meall nan Damh fold.

This fold takes its name from a hill 6 Km. south of Glenfinnan in western Ardgour (Dalziel,1963a). In the area under consideration the western limb is well exposed but the eastern limb is very poorly exposed.

The F_2 minor folds in the Ardgour granitic gneiss on the road section east of Glenfinnan village and on Mam Chreagain are on the western limb of a major fold which locally plunges steeply northwards and opens out to the north (see Fig. 48,folds(7) and (8)). In the exposures further north-eastwards the relationship of the dominant foliation (S_2) to the earlier migmatitic fabric (S_0/S_1) indicates that the fold axial plane has been crossed.

Minor folds and rodding within the granitic gneiss indicate that the structure generally plunges steeply northwards with only occasional F_2 minor folds plunging steeply southwards (Fig.48,Folds(4),(5),(6)). Occasionally F_2 minor folds (Fig.48,folds(4),(5),(6)) to the northeast of the major axial plane of the Meall nan Damh fold have geometries which indicate the presence of more than one F_2 axial plane in the area.

Figure 48.

Diagrammatic sketch of the Beinn an Tuim area to show the relationship between major and minor F_2 folds.

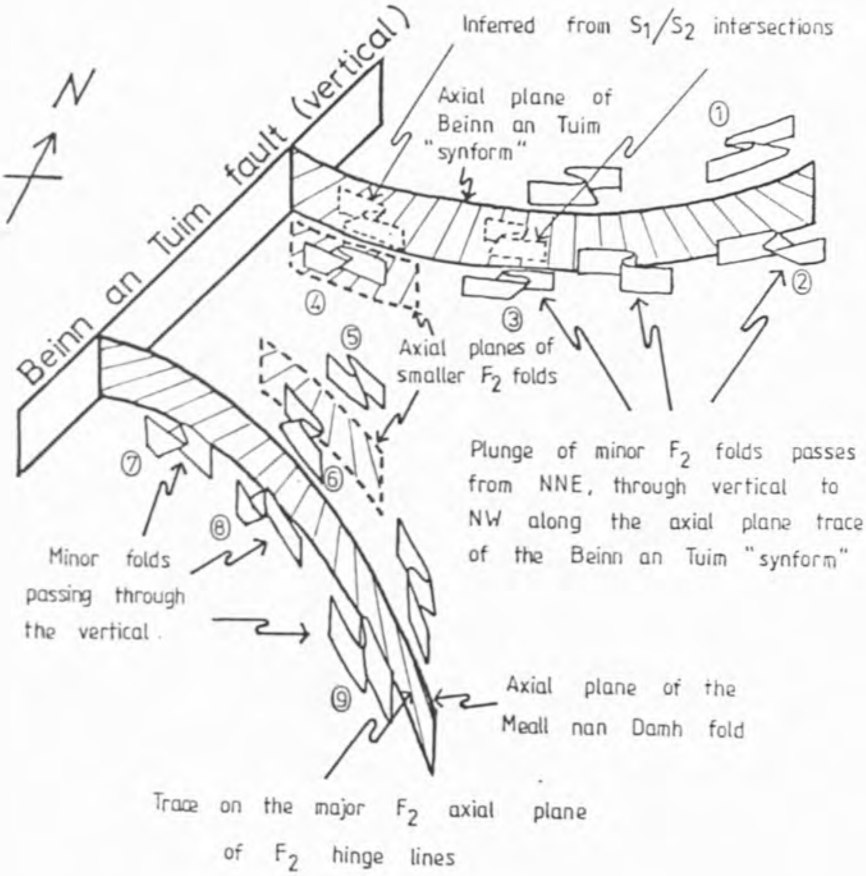
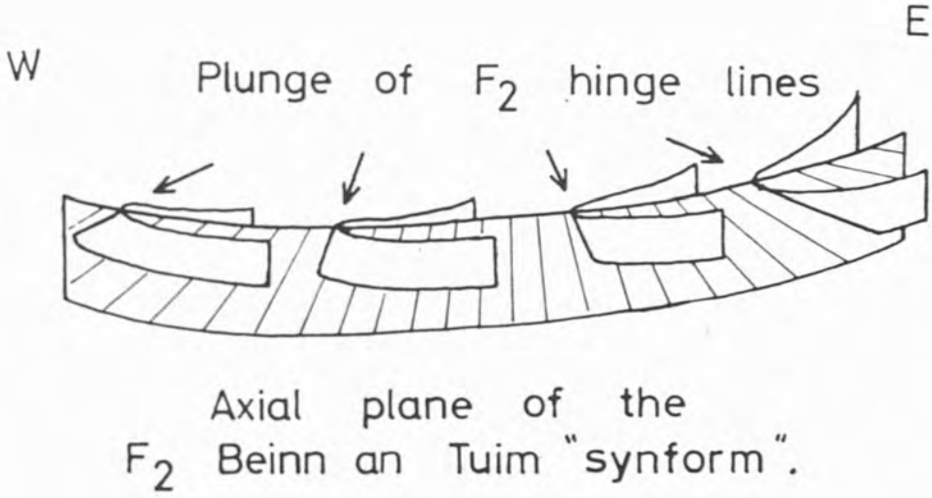


Figure 49.

Sketch to show the Beinn an Tuim "synform" passing through the vertical.



F_2 minor folds become tighter towards the West

Bending of the F_2 axial plane is due, at least in part, to later folding.

Over most of the area under consideration the Meall nan Damh fold is a northerly plunging, nearly isoclinal synform, however Dalziel (1963a,1966) has mapped a much larger area which is a southerly extension of this fold and over most of this area its axis is inferred to plunge steeply southwards and its shape is therefore inferred to be antiformal, the fold opening northwards. Moving northwards along its axial plane trace the hinge line is inferred to pass through the vertical so that in the extreme north of Dalziel's area (north of the A 830 road and Callop river, Maps 3 & 5) the fold is a northwards opening synform (see Fig.50, taken from Dalziel,1966). Dalziel has also mapped an F_2 major fold, the Meall a Bhainne synform (see Fig.50) the axial plane trace of which is extended northwards through the northern end of Loch Shiel, however no evidence for the existence of this fold has been found in the rocks north of the river Callop.

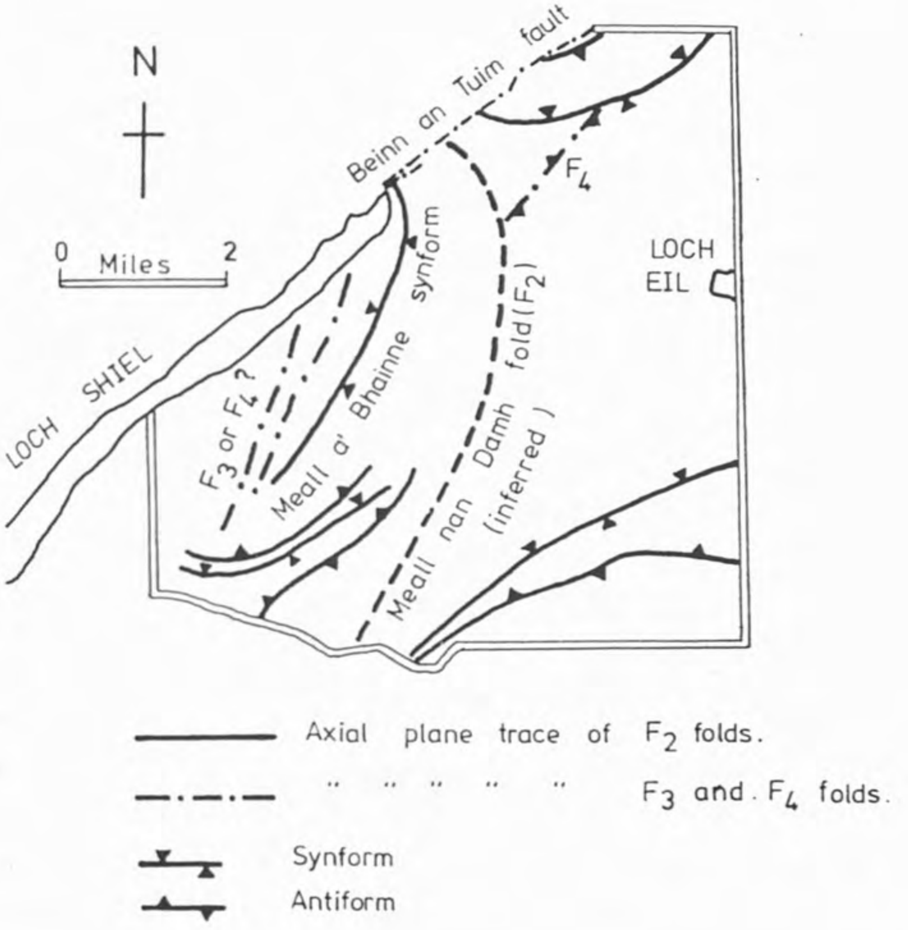
5.4d4) Variations in strain.

The area east of the Beinn an Tuim fault (Map 3) is clearly one of heterogeneous deformation, demonstrated by the variations in plunge of F_2 minor fold hinges and variations of interlimb angle of F_2 major folds. The heterogeneity could have been produced during D_2 deformation, during D_4 deformation, or by a combination of both. The variation in states of deformation across the Glenfinnan/Loch Eil Division boundary is considered in more detail later (Chapter 5.5). However even within the rocks of the Glenfinnan Division there are obvious strain variations. It is possible to consider that the variation in the plunge of F_2 minor fold hinges along the length of the Beinn an Tuim synform is a result of folding F_2 minor fold hinges during the D_4 deformation (ie. by the Glen Dubh Lighe antiform). However an examination of Map 3 shows the western half of the axial plane of the Beinn an Tuim "synform" is essentially planar yet F_2 minor fold hinges vary in plunge from 30-50° towards the east in the east, through reclined northerly plunging, eventually plunging at 25-60° towards the west in the west. Thus the plunge variation cannot be explained by variations in D_4 strain and consequently must be due to strain variations during D_2.

The interlimb angle of the major F_2 Beinn an Tuim synform tightens rapidly to isoclinal moving westwards from Glen Dubh Lighe. This seems to correspond to the change of plunge of the F_2 minor fold hinges from synformal in the east, through reclined to antiformal in the west. Therefore it is simplest to visualise the Beinn an Tuim synform forming as a near horizontal or easterly plunging synform, cored by Loch Eil Division psammites, which has been progressively more deformed by D_2 deformation in the west. The

Figure 50.

Map to show the trend of major fold axial plane traces southeast of the Beinn an Tuim fault.



Modified after Dalziel 1966, fig 10.

progressive deformation has presumably rotated the hinge lines of minor folds in the XY plane of the strain ellipsoid towards the X direction of the imposed strain ellipsoid.

The western section of the Beinn an Tuim fold axial plane dips approximately northwards at 50-60° (Map 3). Rotation of F_2 minor hinge lines from approximately 30-50° towards the ENE through the reclined position, to end up plunging at 25-60° towards the WNW in the west, requires a very large amount of rotation, possibly through an angle of more than 90° and if this is the case then a rotation of the principal axes of stress is required.

It is noteworthy that only in the extreme east of the area where D_2 strain is relatively low is there good evidence for the existence of a well developed pre-S_2 schistosity (a schistosity which initially was presumed to be S_2 in the psammites and pelites on Druim na Saille).

Virtually everywhere else in the area the dominant S_2 fabric appears to be the earliest fabric. Nowhere in the area does the S_2 fabric appear to be a crenulation of a pre-existing fabric, it always seems to have completely transposed or enhanced the early fabric so that the S_2 fabric always looks axial planar and penetrative.

The Meall nan Damh fold is a northwesterly opening, steeply plunging synform in the area mapped. It changes southeastwards along its axial plane into a northwesterly opening antiform. By analogy with the Beinn an Tuim synform, this fold is probably at a lower state of strain in the south and more highly deformed in the north.

In the northwest of the area, near the Beinn an Tuim fault, the axial planes of the two F_2 major folds are in close proximity and the D_2 strain is high. Moving eastwards along the trace of the Beinn an Tuim synform and southeastwards along the trace of the Meall nan Damh fold the D_2 strain decreases, the width between the fold axial planes increases and the interlimb angle of one of the folds increases. This decrease of D_2 strain together with the folding by the F_4 Glen Dubh Lighe antiform and the effect of topography on Druim na Saille all combine to produce the dramatic change in width of the outcrop of the Ardgour granitic gneiss in the area (see Map 3).

5.4e) Structural setting of the Ardgour granitic gneiss.

The term "Ardgour granitic gneiss" here is not used to imply that the rock had an igneous origin, rather it is used to indicate that the rock has the composition of a granite, ie. quartz, two feldspars and biotite. A summary of the views of earlier workers has been given in Chapter 2.2.

5.4e1) Appearance of the Ardgour granitic gneiss.

The Ardgour granitic gneiss consists of suites of segregations and pegmatites intruded into, or sweated out of, a "host rock" which is composed of quartz,feldspar and biotite,in the form of a fairly homogeneous gneiss.

The "host rock" is coarsely foliated, gneissose and has a mottled black and white colour when fresh. The black colour is due to foliae of biotite which anastomose around white lenses of quartzo-feldspathic material (see Figs.51a & b). In some places the biotite defines more pronounced bands and the host rock then more closely resembles some of the adjacent metasediments.

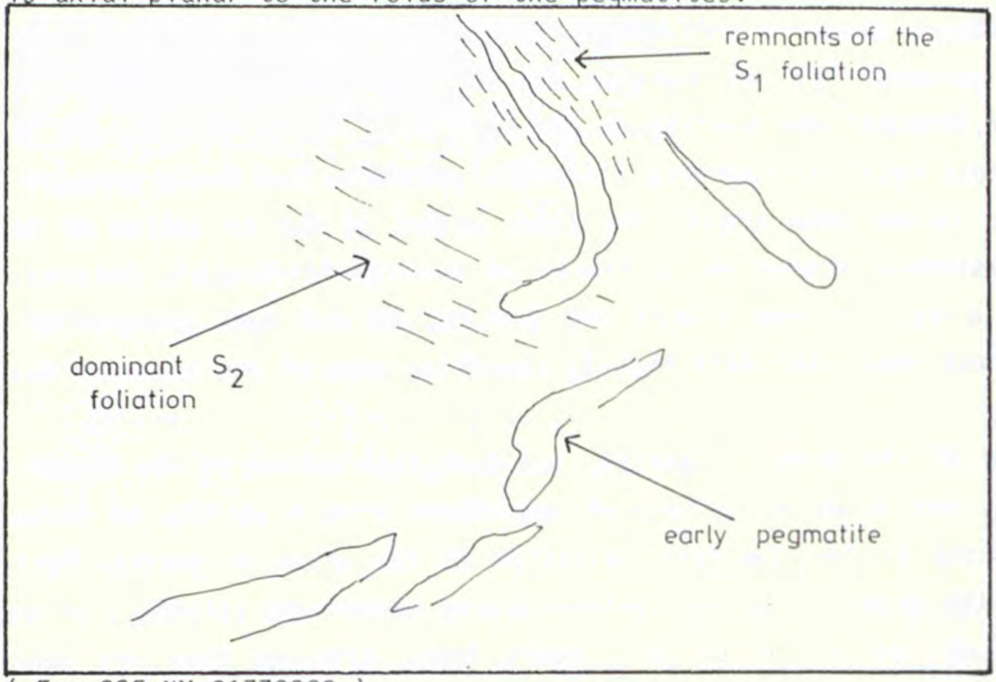
Typical of the granitic gneiss, and much less common in the mapped metasediments, are a whole variety of pegmatites with a variety of structural relationships to the dominant foliation of the granitic gneiss. Most outcrops of the granitic gneiss contain nearly concordant stringers of pegmatite 1-10 cm. thick and up to 1 metre long, although they are generally shorter. They are frequently lensoid and always have a thin selvage of biotite. The nearly concordant pegmatite lenses are often hook shaped (Fig. 51b) and appear to mark the remnants of fold closures indicating that the pegmatites have been deformed towards the strong S_2 foliation in the gneiss. In some instances, near to the end of the hook pegmatites, there is a suggestion of a migmatitic fabric (S_1) co-planar with the pegmatites and slightly oblique to the dominant S_2 fabric. In a few exposures the pegmatites are strongly discordant to the dominant S_2 foliation (Fig.51a), the biotite laths comprising the selvages are co-planar to the pegmatites and crenulated into the dominant (S_2) foliation (Fig.51a).

In many exposures there is a suggestion that the dominant schistosity (S_2) is composite, re-working an earlier fabric which was probably migmatitic. At exp.1323 on Mam Chreagain (NM 91988078) the gneiss is dominated by the S_2 foliation and superficially the lensiod pegmatites appear virtually co-planar with this foliation (Fig.52a). However, it can be seen that the lensoid pegmatites have been tightly folded to lie within the dominant foliation. In addition, an earlier, apparently migmatitic,fabric is also isoclinally folded to lie within the dominant (S_2) foliation (Fig.52b). This exposure also contains two types of later "pegmatites". A 2-3 cm. thick vein of aplite which cuts obliquely across the S_2 foliation and appears to be later than the D_2 deformation is truncated by a large "late pegmatite", one of many found in the area (see Chapter 5.4e4).

Figure 51. Photographs to show the gneissic foliation(s) in the Ardgour granitic gneiss.

S_1 foliation is co-planar to the pegmatites. S_2 foliation is axial planar to the folds of the pegmatites.

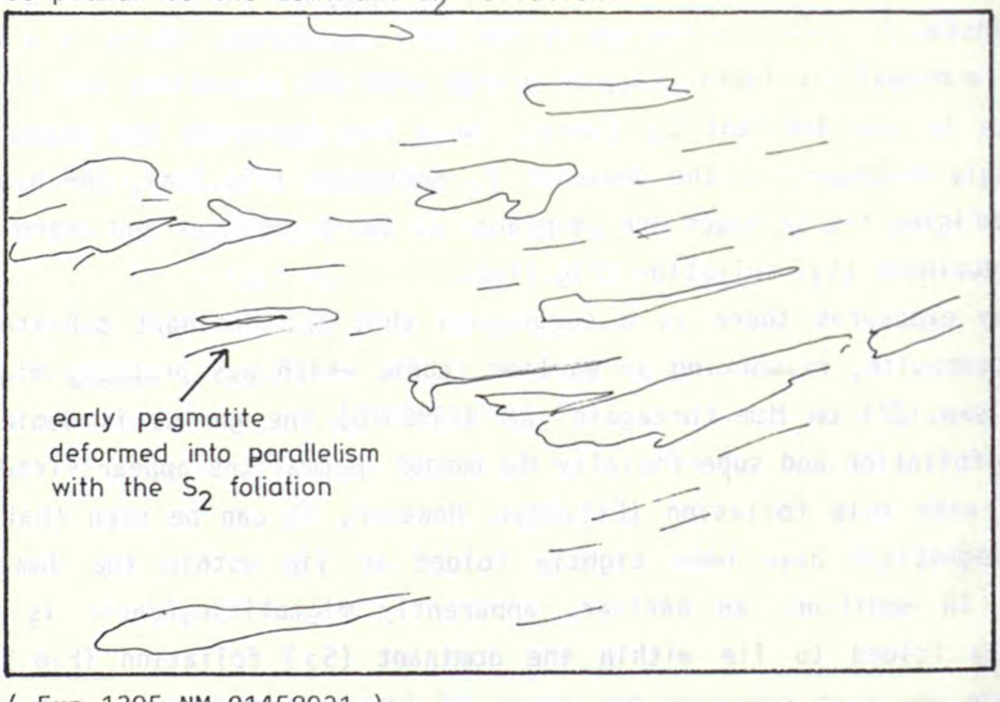
(a)



(Exp.235 NM 91778020).

"Hook" pegmatites tightly folded to lie almost co-planar to the dominant S_2 foliation.

(b)



(Exp.1305 NM 91458031).

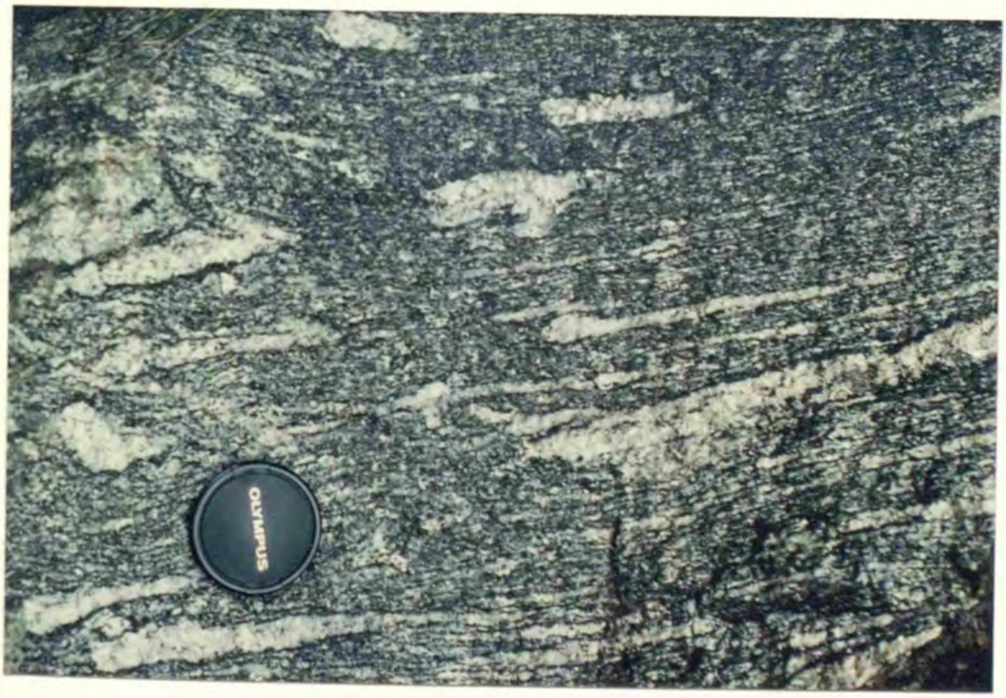
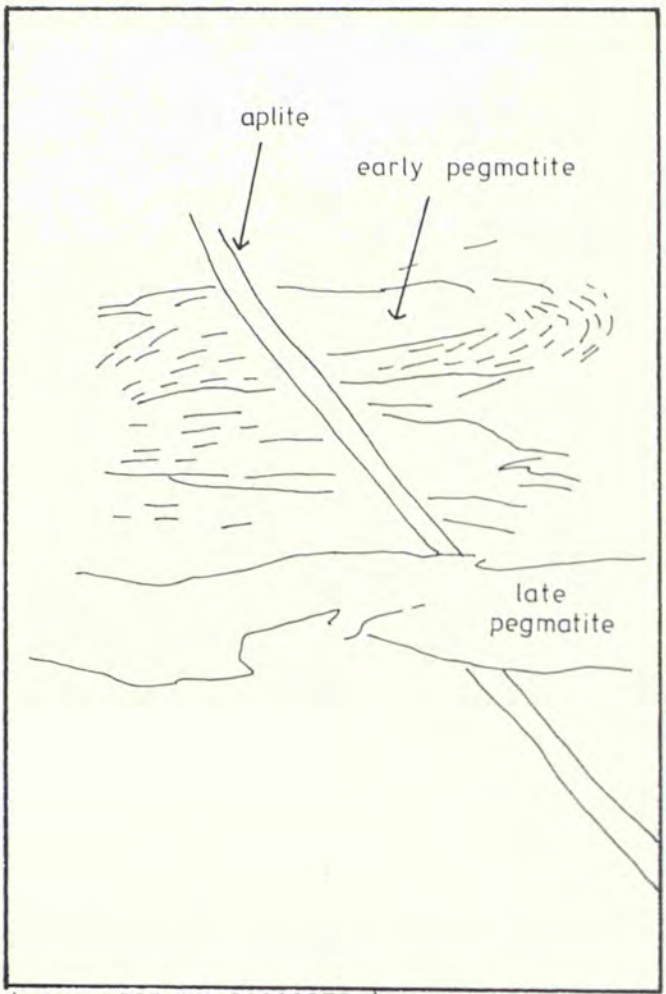


Figure 52. Photographs of the Ardgour granitic gneiss.

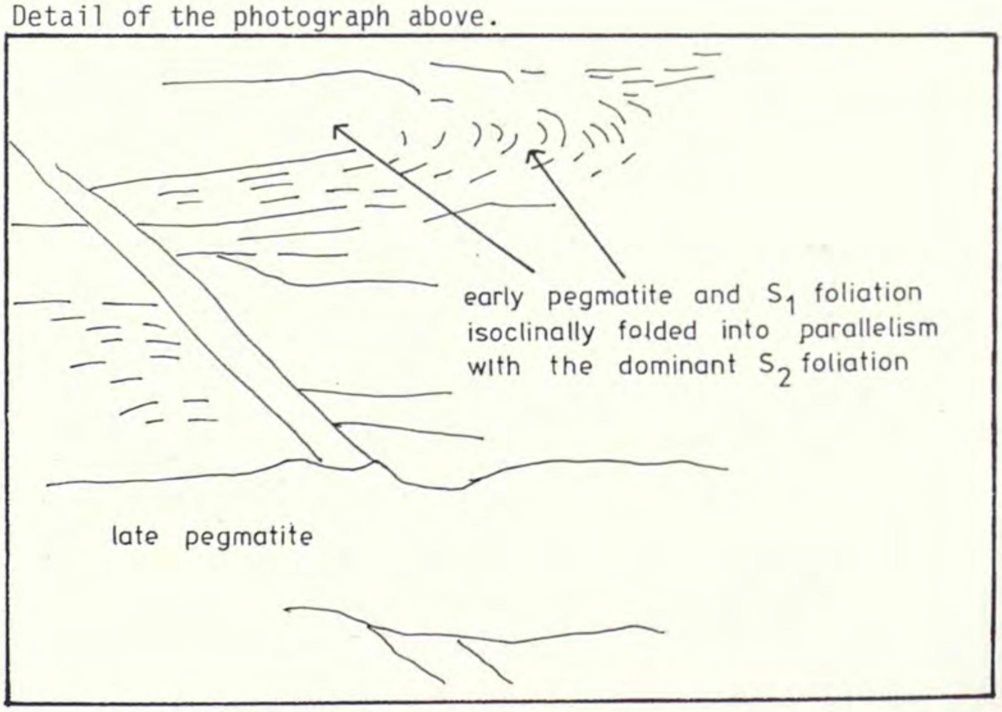
(a)



(Exp.1323 NM 91988078).

Detail of the photograph above.

(b)



5.4e2) Nature of the enclosed sediments.

There are two masses of metasediment enclosed within the granitic gneiss. The larger one outcrops around Lochan na Carnaich (Map 3). It is composed of metasedimentary types ranging from psammites,through stripes lithologies to garnetiferous pelites. The junction between the metasediments and the enclosing granitic gneiss is gradational over a distance of a few metres as the lithological banding becomes obvious in the metasediments. The S_2 foliation in the metasediments is co-planar with the dominant S_2 foliation in the granitic gneiss (Map 3). Both rocks seem to have undergone exactly the same sequence of deformation. There is no evidence of cross-cutting relationships, xenoliths, a metamorphic aureole etc. at any of the granitic gneiss/metasediment contacts which would clearly indicate a high level intrusive origin for the granitic gneiss.

The smaller metasedimentary mass exposed on Mam Chreagan (NM 918810) shows similar structural relationships to the granitic gneiss as does the larger mass described above. The metasediment is distinguishable from the granitic gneiss because it is more obviously laminated or striped than the homogeneous "host rock" granitic gneiss. Gradation from metasediment into the granitic gneiss occurs in directions both normal and parallel to the composite foliation in the metasediment.

There are many fewer early pegmatites in the metasediments than in the granitic gneiss and they are generally concordant with S_2 and less highly deformed in the metasediments. It is suggested elsewhere (Chapter 6.2c) that the metasediment is a "palaeosome" within the "neosome" of the granitic gneiss.

5.4e3) Formation and deformation of the early pegmatites.

The early pegmatites are petrographically similar to the "host rock" granitic gneiss, containing quartz, two feldspars (microcline and plagioclase togather with perthites, antiperthites and myrmekites) biotite and muscovite. They do not contain an internal foliation. They occur in a variety of states of deformation; as fairly continuous concordant veins co-planar with the dominant S_2 foliation and as tectonically isolated lenses slightly discordant to the dominant S_2 foliation. The lenses are often isolated hooks or fold noses which have been progressively rotated towards the dominant S_2 foliation. Less frequently the pegmatites are markedly discordant to the dominant S_2 foliation, but they too have undergone D_2 deformation.

From the foregoing description it appears that the early pegmatites plus their biotite selvages have been, with increasing strain, rotated towards near parallelism with the dominant S_2 foliation. Where the early pegmatites are strongly oblique to the dominant S_2 foliation there are frequent signs of an earlier (S_1) migmatitic fabric which is co-planar to the pegmatites (Fig.51a). Figure 52b shows an early pegmatite co-planar to the early migmatitic foliation. Both the pegmatite and the early fabric are isoclinally folded to lie within the dominant (S_2) foliation.

In summary the early pegmatites are apparently co-planar to an early (S_1) migmatitic fabric in the gneiss and they are variably rotated into the dominant S_2 foliation by deformation. Since they seem co-planar to the S_1 migmatitic fabric and they have an S_1 biotite selvage, the pegmatites must be either the product of high grade metamorphism and partial anatexis during an early (pre- or syn-F_1) metamorphic event or pre-F_1 intrusions subsequently metamorphosed during D_1. No F_1 folds were observed so it is not known whether the pegmatites are pre-F_1 or syn-F_1 segregations. Since the pegmatites are co-planar with the S_1 foliation and contain a co-planar S_1 selvage they cannot be post-F_1/pre-F_2 in age.

Examples of pegmatites which have unequivocally formed during D_2 have not been observed. The very well developed S_2 migmatitic fabric need not necessarily signify migmatitic or partial melt conditions during D_2 as the fabric may be the result of intense solid state recrystallisation of the S_1 migmatite.

The pegmatites are mineralogically similar to the granitic gneiss and are thus probably closely related to the formation of the granitic gneiss. Both the early pegmatites and the granitic gneiss are rich in K-feldspar which is generally lacking in the surrounding metasediments. Dalziel (1963a) noted a suite of early pegmatites in both the granitic gneiss and surrounding metasediments which is K-feldspar free and he observed that these pegmatites mineralogically resemble the quartzo-feldspathic foliae and groundmass of the regionally injected "oligoclase gneisses" (ie. the more gneissose of the regional metasediments). The formation of the granitic gneiss and its metamorphism is considered in Chapter 6.

5.4e4) Late veins and aplites.

An aplite vein has already been described which is later than the formation of the dominant (S_2) foliation and which is cross-cut by a member of the suite of "late pegmatites".

The thick late pegmatites always post-date D_2 deformation , some of them appear to post-date F_4 folds in the Ardgour granitic gneiss although in the Glenfinnan Division further west there are late pegmatites deformed by the D_3 deformation which produced the Sgurr a Mhuidhe synform.

Thus there may be more than one phase of late pegmatite intrusion (cf. van Breemen et al.1974), alternatively the intrusion of late pegmatites may have commenced before D_3 and continued until post-D_4 times. The latter alternative is viewed as unlikely since arguments are presented elsewhere (Chapter 8.1b) that a post-D_3/pre-D_4 suite of microdiorite sheet evidences a significant change in the crustal stress field during this period.

The late pegmatites do not contain a biotite selvage but internal, layer parallel compositional zoning is common. Books of mica up to 2-3 cm. long and 1-2 cm. thick are quite common. Where pegmatites are folded by F_3 folds in the west the books of mica take up a "crenulated" appearance, that is, the books of mica previously with random orientations have rotated towards the XY plane of the D_3 deformation strain ellipsoid producing the overall appearance of a crenulated fabric.

5.5) The Loch Quoich line.

The Loch Quoich line has already been described in all but name in the earlier parts of this chapter, however such a variety of interpretations have been placed upon it by different authors that it is worth summarising previous descriptions and interpretations and then commenting on the relevant observed field evidence from the Glenfinnan-Loch Eil area.

Leedal (1952) has described an eastern flat belt composed of "the Upper Psammitic Group" (ie. the Loch Eil Division psammites) and going westwards he took the first main outcrops of his "Pelitic Group" to define the start of the "highly inclined" belt. He emphasised that this outcrop did not mark a continuous tectonic structure.

Clifford (1957) working farther north in the Loch Arkaig area, postulated the existence of a klippen of Glenfinnan Division rocks resting on top of the Morar Succession in Kintail. Looking eastwards for a root zone to this nappe he noted that the junction between the "highly inclined" belt and the "flat" belt was also the eastern limit of intense regional injection in the Moine. He termed this junction the Loch Quoich line.

Dalziel (1966) noted that the Loch Quoich line is broadly parallel to the axial plane trends of F_1, F_2 and F_3 structures but F_4 folds (which in this description are F_5 in the Loch Eil Division, see Chapter 5.3b) have a more E-W trend and are too poorly developed to be closely connected with the origin of this feature. He noted that his F_3 Glen Dubh Lighe antiform axial plane trace crosses the Loch Quoich line and therefore considered that the line originated prior to the F_3 fold movements. He further noted that the lower and more mobile structural levels were raised to the west of the line which could therefore mark the outcrop of a zone of Decollement between un-migmatised, relatively rigid cover and a highly mobile, migmatised infrastructure. He postulated that formation of this line may have commenced during D_1 deformation which he believed to have been low grade (cf. Chapter 5.4e3). However he favoured the idea that the Loch Quoich line became a "significant tectonic boundary throughout the F_2 deformation".

Lambert et al.(1979) suggested that the Loch Eil Division appeared to be structurally and metamorphically simpler and geochemically distinct from the Morar and Glenfinnan Divisions and in their view the simplest explanation of these "facts" is that the Loch Eil Division rests unconformably on the metamorphosed Glenfinnan Division.

Piasecki & van Breemen (1979) described a basement (Central Highland Division) and cover assemblage (Grampian Group) separated by a zone of sliding in the Central Highlands east of the Great Glen fault. They suggested that their Grenvillian basement assemblage is analogous to the Glenfinnan Division and that their (Morarian age) cover succession is analogous to both the Morar and Loch Eil Divisions west of the Great Glen fault. They concluded that the Loch Quoich line is a tectonic slide zone between the Loch Eil Division cover and the Glenfinnan Division basement.

Strachan (1982), mapping an intricate set of subtly different psammitic rocks at the base of the Loch Eil Division, claimed that the thickness variations in these units could not be due to sedimentary facies variations. He therefore invoked the existence of a series of tectonic slides to explain their exposure pattern. However he cited no evidence for the existence of the slides other than his map of lithological pattern. Strachan further claimed the sequence of complex slides to have been formed prior to D_2 deformation and he considered the deformation sequences in the Glenfinnan and Loch Eil Divisions to have been the same.

In contrast to these views, Roberts & Harris (1983), in mapping across the type area of the Loch Quoich line concluded that the line marks the eastern limit of D_3 re-working of an earlier metamorphic complex. They mapped pre-D_3 structures which are flat lying in the Loch Eil Division, whereas west of the Loch Quoich line all the pre-D_3 structures have been rotated towards the vertical during the D_3 deformation. West of the line these authors demonstrate the existence of major F_3 folds with vertical axial planes and sub-horizontal hinge lines.

Roberts et al.(1984) claimed that the intensity of D_3 deformation progressively increases westwards and is reflected by the rotation of F_3 hinge lines towards the vertical and the development of curvilinear fold hinge lines indicative of vertical extension within the vertical F_3 axial planes.

Obviously many of these observations already made in this chapter have a bearing on these previously stated views. Of foremost significance is the observation herein that the structural history of the Glenfinnan Division rocks is the same as that of the Loch Eil Division. In both Divisions the Beinn an Tuim synform is an F_2 fold which folds rocks containing bedding and an earlier (S_1) fabric. Consequently the Loch Quoich line cannot be considered to be a junction where the Loch Eil Division psammites rest unconformably on previously deformed and metamorphosed rocks of the Glenfinnan Division (cf. Lambert et al.1979). Likewise by the same evidence it cannot be considered to be a slide zone between cover and basement sequences (cf. Piasecki & van Breemen,1979).

The trace of the Glen Dubh Lighe antiform locally crosses the Loch Quoich line, a fact which Dalziel (1966) used to suggest that the line originated prior to this phase of deformation. However by this argument he should have considered the formation of the line to pre-date the F_2 Beinn an Tuim synform which also crosses the Loch Quoich line.

The pattern of lithological variations in the area can be explained by lateral facies changes without the need to invoke the presence of tectonic slides. Strachan (1985) uses facies changes to explain some of the largest lithological variations and cites no other evidence for the existence of tectonic slide contacts.

D_3 deformation in the Loch Eil Division has produced the flat lying, nearly isoclinal Druim Beag synform. Thus the Loch Quoich line cannot be an eastern limit of D_3 re-working (cf. Roberts & Harris,1983).

Within the Glenfinnan Division (see Chapter 5.4 & 5.6) the major F_3 folds have vertical axial planes and show variations of hinge line plunge which are related to the intensity of D_3 deformation, but the vertical orientation of planar elements is due to subsequent D_4 deformation. Prior to the D_4 deformation the variation of F_3 hinge line orientation occurred within F_3 fold axial planes which dipped shallowly to the east (see Chapter 5.6e, 7.5 and 9.1).

The Loch Quoich line represents a front of intense D_4 deformation: west of the line pre-existing relatively flat lying structures have been rotated and folded into a sub-vertical orientation. East of the line the pre-D_4 structures remain relatively flat lying and little affected by D_4 deformation.

5.6) West of the Beinn an Tuim fault.

5.6a) Distribution of lithologies.

The rocks which outcrop west of the Beinn an Tuim fault are dominantly metasediments of the Glenfinnan Division. Morar Division metasediments only outcrop in the extreme west ot the area, west of the Sgurr Beag slide (Maps 1 & 4). The detailed structural and metamorphic relationships across the Sgurr Beag slide at the eastern end of Loch Eilt are considered in Chapter 7.

Over large areas the rocks of the Glenfinnan Division are mixed psammites and thin pelites which have proved impossible to sub-divide into mappable units, however near the summit of Beinn an Tuim and from Sgurr a Mhuidhe westwards lithological sub-divisions have been mapped (Map 4).

5.6b) Chronology of folding and deformation.

The dominant phase of deformation in the area (D_3) has produced major steeply plunging F_3 folds and associated minor folds which have a well developed S_3 crenulation fabric. Locally an intensification of D_3 strain has caused F_3 folds to tighten into a D_3 slide zone (Hutton,1979) termed the Sgurr Beag slide.

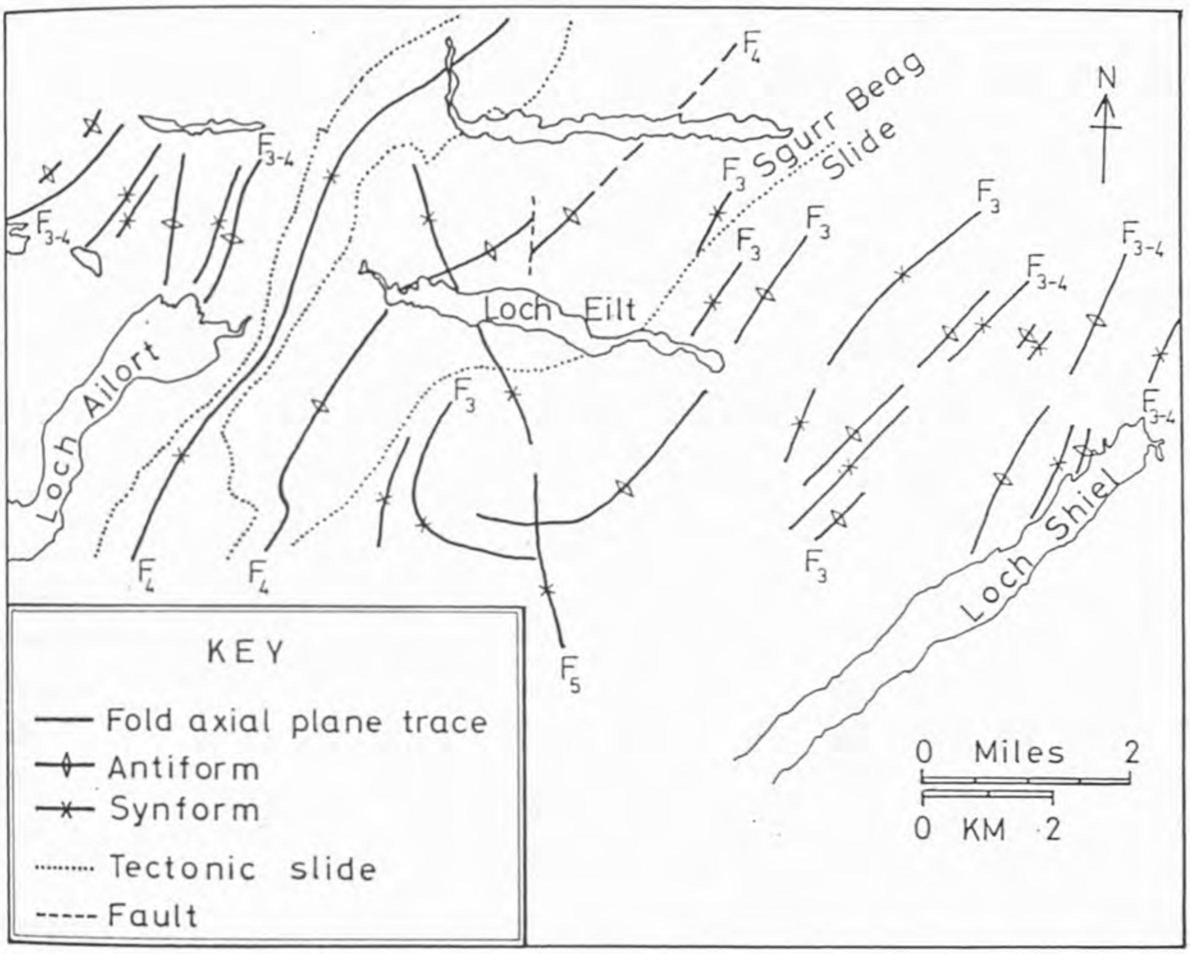
D_3 deformation deforms a pre-existing planar mica fabric (?S_2). Minor folds coeval with this planar penetrative mica fabric were mapped as F_2 folds, but they could equally well be of an earlier (F_1) generation. It is not at all clear at individual outcrops whether pre-D_3 folds and fabrics are of D_1 or D_2 age. None of the exposures west of the Beinn an Tuim fault was seen to contain direct evidence for two fold phases prior to D_3 deformation.

Major folds which fold the D_3 Sgurr Beag slide and F_3 folds in the Loch Eilt area are regional F_4 folds. A set of relatively open crenulation folds which has a constant sense of asymmetry and plunges steeply to the SE appear to be minor folds related to the major open fold seen south of Loch Eilt (Fig.53 and I.G.S 1:63,360 Sheet 61.Scotland) and are F_5 folds.

5.6c) D_5 deformation.

The episode of deformation which has produced the least intense folding and the most easily understood geometry is here termed D_5. It has produced a set of open minor crenulation folds which are found only in the extreme

Figure 53.



Fold axial plane traces between Loch Shiel (Glenfinnan) and Loch Ailort (partly after Brown et al. 1970 and Powell 1974). DIAGRAM TAKEN FROM BAIRD 1982.

west of this area around Loch Eilt. Fold interlimb angles in the psammitic layers are usually greater than 90°. Open mica crenulations are produced in the pelitic layers.

No exposures have been seen where the open crenulation folds fold the more common tight crenulation folds or vice versa so that their relative structural ages, based on evidence from individual exposures, has not been proven conclusively. However, the hinge lines of the open minor folds and their axial planes are remarkably constant in their plunge and trend (Fig.54 a & b) and they show no marked change across the axial planes of the F_3 major folds in the area. Where the F_3 major folds are isoclinal however, no change in amount or direction of the plunge of the F_5 minor hinge lines would be expected. Where the F_3 major folds are not isoclinal and have been deformed by F_5 crenulations, the hinge line orientation of F_5 folds is dependent on the orientation of the planar elements being deformed (see Fig.55a). In nearly isoclinal folds (eg. the F_3 Ranochan synform and the F_3 Chreag Bhan antiform, Map 4) the variation of F_5 hinge line orientation would be relatively small and this small variation could cause the small spread of orientation visible in Figure 54a.

In considering the possibility that the F_5 open crenulation folds are older than the F_3 crenulation folds it is noted that the hinge lines of the F_5 minor folds lie approximately normal to those of F_3 and within the F_3 axial plane. If the F_3 isoclinal and nearly isoclinal folds had re-folded the F_5 open crenulations then the F_5 open crenulations on both limbs of the F_3 major folds would have similar orientations, but the vergence of the F_5 minor folds would change across the F_3 axial plane (see Fig.55b). All the F_5 crenulation folds mapped in the area have the same sense of vergence: a sense which does not change across the traces of the F_3 major folds, therefore the F_5 folds cannot be earlier than the F_3 folds and re-folded by them.

When considered regionally (I.G.S 1:63,360 Sheet 61.Scotland and Chapter 7) the F_5 minor folds seem to be related to a large open fold seen south of Loch Eilt (Fig.53).

The poles to F_3 minor fold axial planes show a very well developed girdle reflecting a gentle regional warp of F_3 axial planes (and earlier structures) produced by the D_5 deformation (Fig.56a).

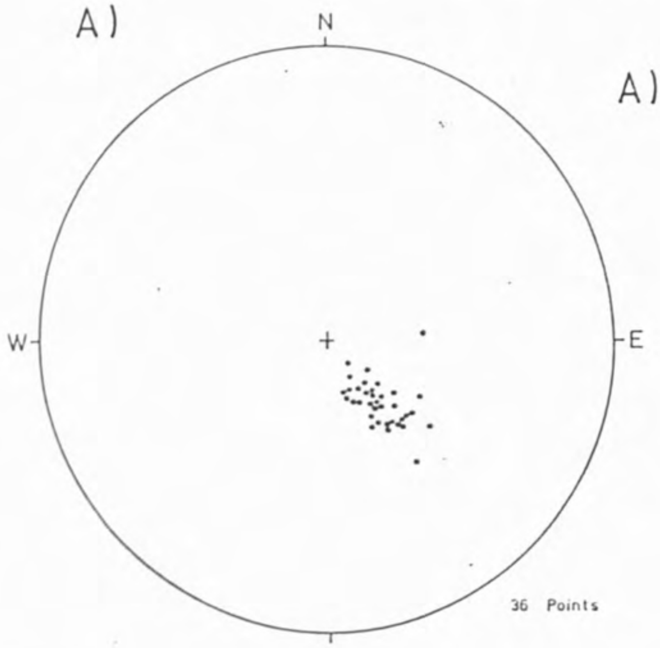
5.6d) D_3 deformation.

This, the dominant phase of deformation, has produced major folds, which when mapped across the area show progressive changes in geometry

Figure 54.

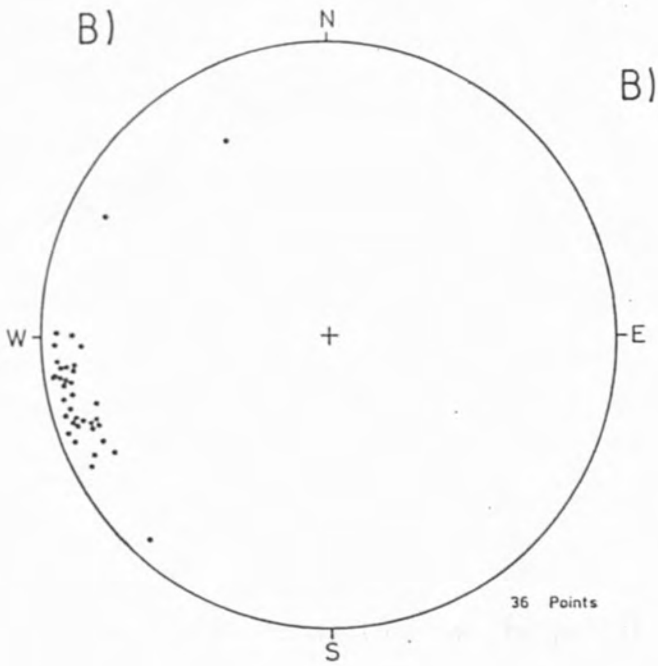
WEST OF THE BEINN AN TUIM FAULT.

A)



A). Plunge of F_5 minor fold hinge lines.

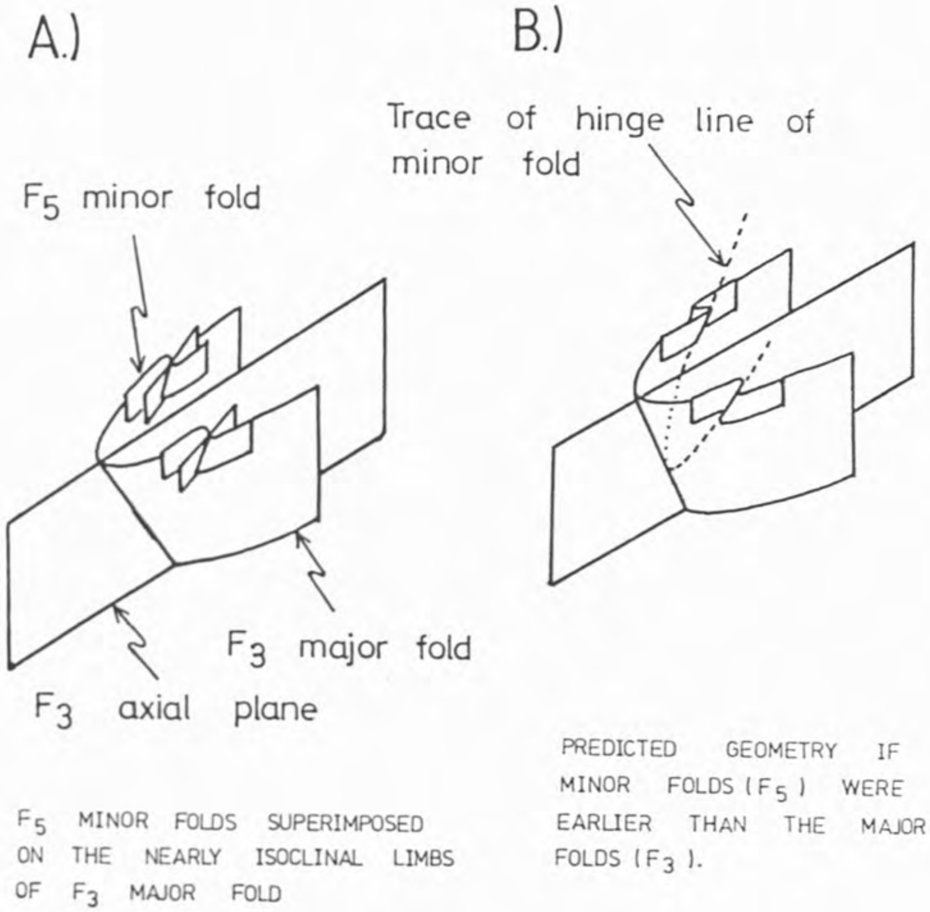
B)



B). Poles to axial planes of F_5 minor folds.

Stereonets from the area west of the Beinn an Tuim fault.

Figure 55.

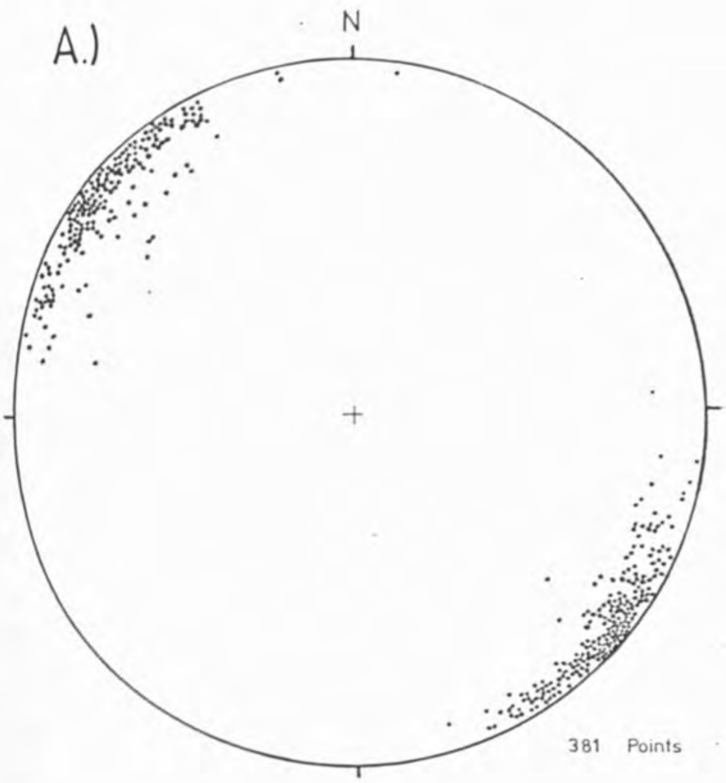


Diagrammatic sketches to show the possible relationships between F_3 and F_5 minor folds (see text for discussion).

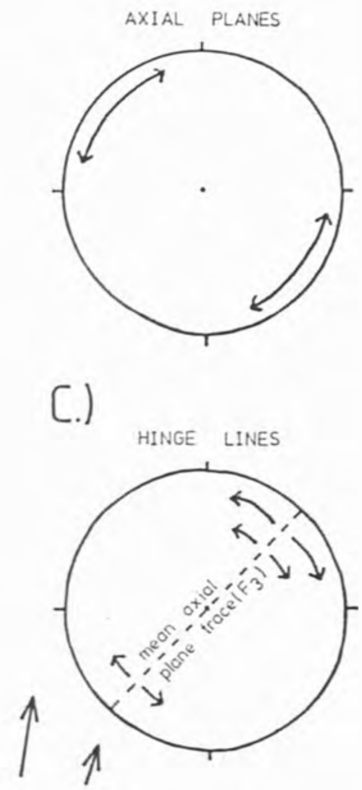
Figure 56.

GLENFINNAN DIVISION — WEST OF THE BEINN AN TUIM FAULT.

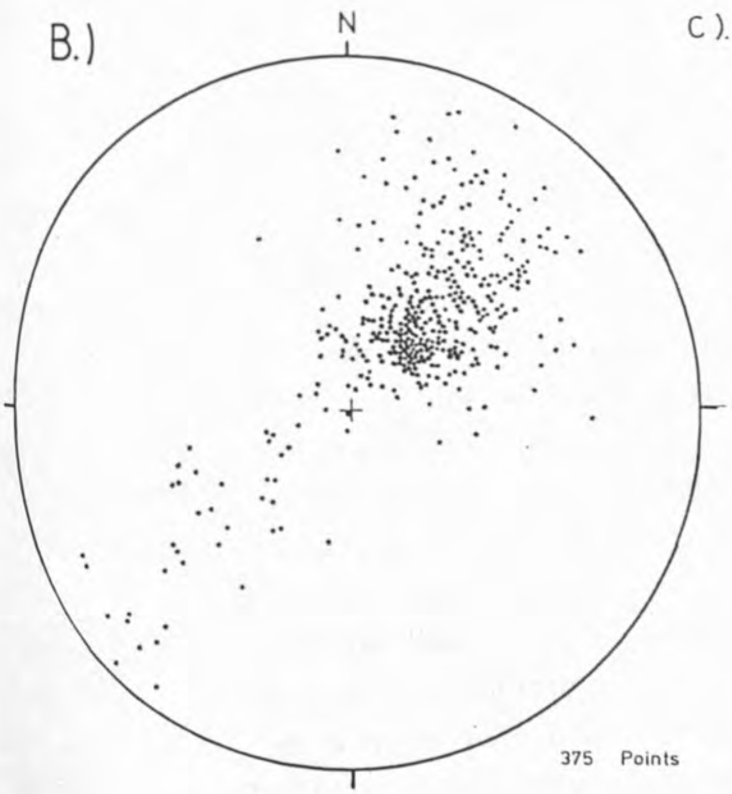
A.)



C.)



B.)



C). Sketches to show rotation of axial planes and hinge lines of minor F_3 folds during F_5 deformation. Arrows indicate the dispersion of data during F_5 deformation.

A). POLES TO AXIAL PLANES OF MINOR F_3 FOLDS.

B). HINGE LINES OF MINOR F_3 FOLDS.

related to the intensity of deformation.

In pelitic layers the associated minor folds are tight crenulations of a pre-existing planar mica fabric. In the east the crenulations are tight, further west they become extremely tight. In psammitic layers the minor folds are tight with interlimb angles as low as 30° (Fig.57a).

In some exposures the planar mica fabric deformed by the F_3 crenulations is seen to be axial planar to tight to isoclinal folds (Fig.57b) which were mapped as F_2 folds with an S_2 fabric (but see discussion of early deformation, Chapter5.4d2).

The traces of the F_3 major folds are indicated on Map 5. The Allt an Tuim synform is a relatively shallowly plunging F_3 fold. This is followed westwards by a complex fold termed the Tom na h Aire antiform. The fold can be defined by the change of orientation of bedding around the hinge zone (see Map 4). Changes in minor fold vergence show that there are a number of folds of intermediate scale related to the major antiform, some of which are indicated on Map 5 . The Tom na h Aire antiform is much more steeply plunging than the adjacent Allt an Tuim synform, a situation which could be due to higher amounts of strain causing progressive rotation of F_3 hinge lines away from the Y, and towards the X direction of the imposed D_3 strain ellipsoid.

The Tom na h Aire antiform is succeeded westwards by the Sgurr a Mhuidhe synform, a nearly isoclinal fold, the hinge line of which plunges steeply towards the NE, this fold is succeeded westwards by the isoclinal Creag Bhan antiform and in turn by the isoclinal Coille Chreag synform.

The Coille Chreag synform is separated from the Ranochan synform to the west by the Sgurr Beag slide (Map 4), a zone of ductile shearing across which there has been large scale displacement (Baird,1982 (included as Appendix 6) and Kelley & Powell,1985). The structural development of the area between the Ranochan synform and the Sgurr a Mhuidhe synform is discussed in more detail in Chapter 7.

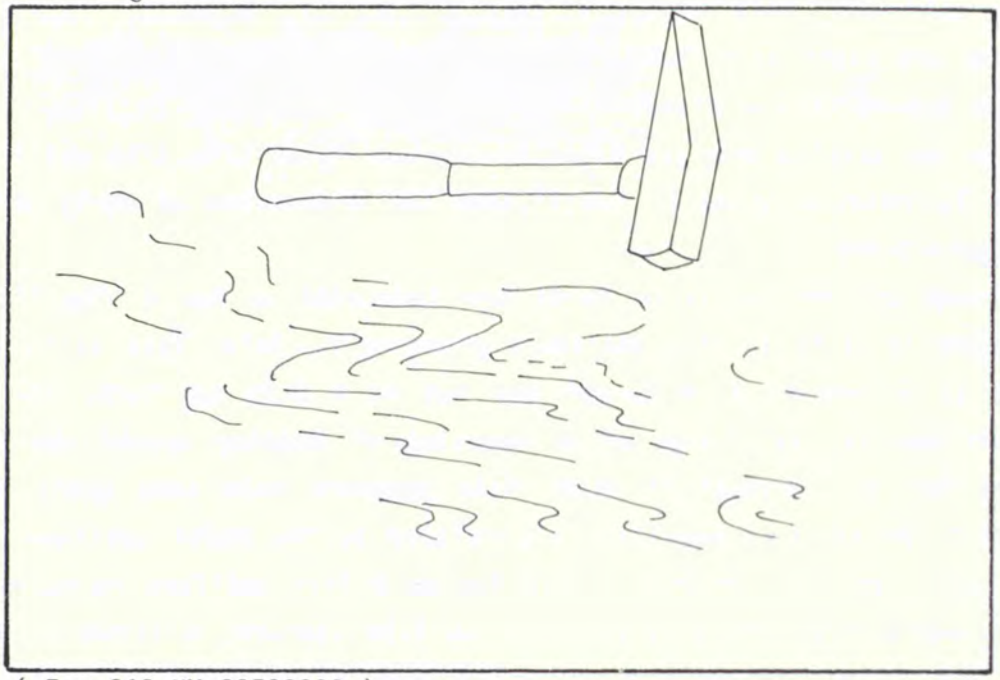
Figures 56a & b show the orientation of F_3 minor fold axial planes and hinge lines respectively. The plot of poles to axial planes shows that they are all approximately vertical and strike from NNE-SSW to ENE-WSW. The spread of strike directions is due to the very open regional F_5 folding and can be seen clearly on Map 4 and Figure 53. Moving westwards the strike of F_3 fold axial planes and bedding planes rotate gradually from NNE-SSW through NE-SW towards ENE-WSW.

Figure 56b shows the orientations of F_3 minor fold hinge lines. The spread of data can only be partly explained by later D_5 deformation. The removal of the effects of D_5 deformation would result in the spread of

Figure 57. Photographs of F_3 minor folds.

Tight F_3 fold in psammitic layers.

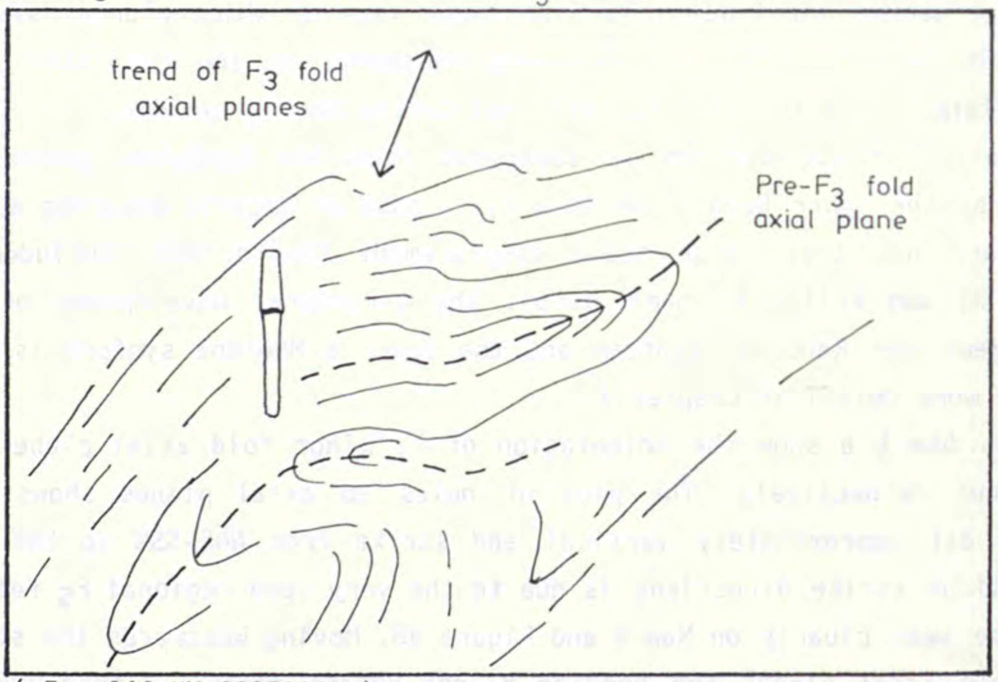
(a)



(Exp.819 NM 89508239).

Pre-F_3 minor folds re-folded by F_3 folds.

(b)



(Exp.846 NM 89058348).

F_3 hinge lines being contained within a vertical plane striking approx. NE-SW (see Fig.56c). The F_3 hinge lines which have low plunges, within the vertical F_3 axial planes will have been displaced more by the removal of D_5 deformation which has involved rotation about a near vertical axis (rotation of near vertical pre-existing linear elements such as F_3 hinge lines about a near vertical axis produces little change in orientation of the linear element). After the removal of the effects of F_5 folding, there is still a large plunge variation within the vertical F_3 axial plane with a concentration of hinge lines plunging towards the NE at 50-70°. The variation in the amount of plunge of F_3 hinge lines may be a function of the amount of D_3 strain, alternatively it may be related to the imposition of D_4 strain.

In general it is seen that in the east the F_3 major folds are relatively open and F_3 minor folds plunge shallowly to the NE. Moving westwards the F_3 major folds tighten to isoclinal and the F_3 minor fold hinges plunge more steeply to the NE. The isoclinal major folds are succeeded westwards by a D_3 ductile high strain zone, the Sgurr Beag slide (see Chapter 7) and it is argued that as the D_3 strain increases progressively westwards into the slide zone the F_3 fold hinges rotate progressively within the XY plane of the D_3 strain ellipsoid towards its X axis.

Regional analysis (Powell et al,1981: Baird,1982) shows that the D_3 Sgurr Beag slide is highly folded by the F_4 Loch Eilt antiform to the west of the area under consideration and by the F_4 Glenshian synform yet further west. Both of these folds have vertical axial planes and hinge lines which plunge shallowly to the SW.

It is necessary to consider the possibility that the variation of plunge of F_3 fold hinges is due to the imposition of varying amounts of D_4 strain. In such an analysis one might assume that D_4 strain is at its lowest level in the core of the F_4 Loch Eilt antiform and increases eastwards into the area under consideration. In this particular geometric situation one would expect the F_3 major folds to be tighter in areas of high D_4 strain, ie. away from the core of the F_4 Loch Eilt antiform and towards the east of the area under consideration. However, this is the exact opposite of what has been observed. There seems to be no sensible way to correlate the observed strain variations in D_3 structures with a plausible D_4 strain profile across the eastern limb of the F_4 Loch Eilt antiform (the mechanism of deformation during D_4 is considered more fully in Chapter 8.1b4).

5.6e) D_4 deformation.

Two F_4 major folds have been described above, the Loch Eilt antiform and the Glenshian synform, both of which have vertical NE-SW trending axial planes and shallowly SW plunging hinge lines. Since this geometry is similar to that of the F_3 major folds previously described, the only unequivocal way of establishing the relative F_3 or F_4 age of an individual major fold is by its relationship to the D_3 Sgurr Beag slide. The Sgurr Beag slide being tightly folded by F_4 major folds.

Since it is impossible to relate individual minor crenulation folds with vertical NE-SW trending axial planes to the D_3 Sgurr Beag slide , they could be of either F_3 of F_4 age. Map 4 shows the vergence and amount of plunge of tight crenulation folds mapped as F_3 minor folds. Over the vast majority of the area they plunge to the NE and relate simply to the F_3 major folds already described. F_3 minor fold plunge varies from relatively shallow (20-40°) to the NE, steepening up towards vertical and occasionally through vertical to plunge steeply towards the SW. Where the plunge of minor folds passes through vertical on the limb of a major fold then all the minor folds still possess the same sense of vergence ("s" or "z"), when viewed down-plunge.

There is one area on Druim na Brein Choille (NM 8882) to the SE of the axial plane of the F_3 Sgurr a Mhuidhe synform (Map 4) where a number of minor crenulation folds have very low plunges towards the SW and "s" vergence, whereas the majority of minor folds on the same limb of the major fold have "z" vergence and plunge steeply to the NE. The "s" vergence of the SW plunging folds precludes the possibility that they have been rotated from NE plunging, through vertical to plunge to the SW. It is possible to interpret the SW plunging minor folds as F_4 minor folds related to the F_4 Loch Eilt antiform to the west. However, if this were the case, one would expect "z" vergence of the minor folds. Alternatively, if D_3 deformation initially produced F_3 minor folds with sub-horizontal hinge lines, mostly plunging shallowly to the NE, but with some plunging to the SW, then progressive D_3 deformation (where the Y axis of the D_3 strain ellipsoid is horizontal trending NE-SW and the X axis is vertical) would cause progressive rotation of the minor fold hinge lines, within the XY plane of the ellipsoid, towards the X direction. In such circumstances F_3 minor folds to the east of an F_3 synform (eg. the Sgurr a Mhuidhe synform) would have "z" vergence when they plunge to the NE and "s" vergence when they plunge to the SW, the more nearly horizontal hinge lines indicating either areas of low D_3 strain, or hinge lines which are co-linear to the Y axis of the D_3 strain ellipsoid, and which have not rotated. There is, however, no geographical relationship between the amount of plunge of the minor fold hinges (ie. strain) and their positions in relation to the F_3 Sgurr a Mhuidhe synform.

If this explanation of varying F_3 minor fold plunge and vergence is accepted then it enables one crudely to estimate the present orientation of the X and Y axes within the XY plane of the D_3 strain ellipsoid. If the SW plunging folds are considered to be of a different age to the dominant NE plunging F_3 minor fold then it is impossible to estimate the orientation of the X and Y axes within the XY plane of the D_3 strain ellipsoid.

It is very important to note that the analysis above of D_3 strain is made using the present orientation of the D_3 strain ellipsoid and does not take into account the possibility that this ellipsoid has been rotated into its present orientation during subsequent deformation.

The significance of D_4 deformation in re-orientating the D_3 strain ellipsoid, and its orientation prior to D_4 deformation, is discussed later (Chapter 7.5 & 9.1).

5.6f) Early deformation (D_1 and D_2).

It is only by comparison of fold geometries and fabrics developed in the Glenfinnan Division metasediments on both sides of the Beinn an Tuim fault that the dominant folds west of the fault have been designated F_3 folds. Nowhere to the west of the Beinn an Tuim fault have F_3 folds been seen to deform two sets of earlier isoclinal minor folds, consequently it is not possible to state unequivocally the structural age of the pre-D_3 deformation events.

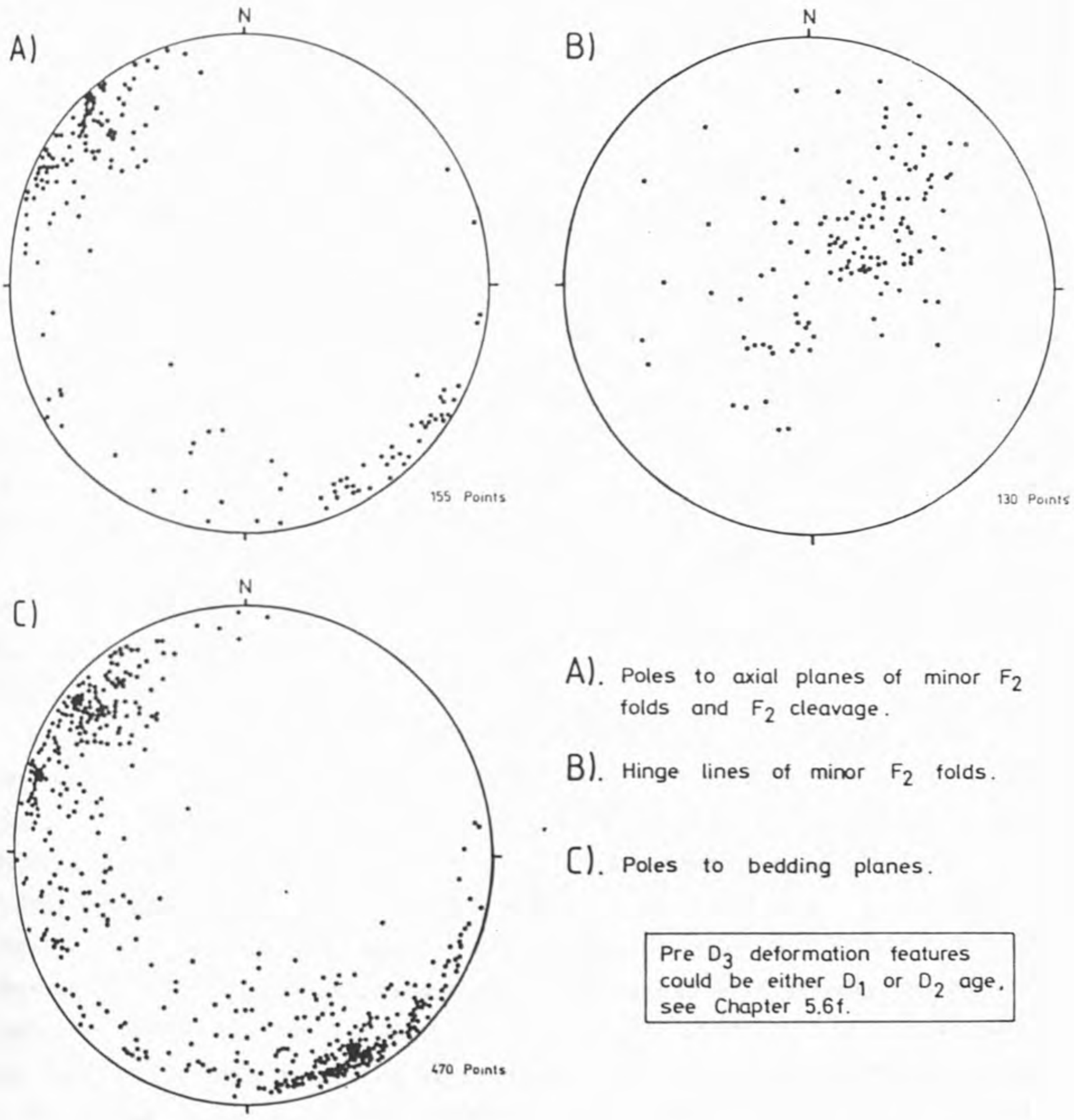
F_3 folds deform metasediments which contain modified bedding and a strongly developed nearly co-planar mica fabric. In many places the planar mica fabric is axial planar to tight to isoclinal minor folds. Figures 58a,b and c plot the orientation of such early fold axial planes, early minor fold hinges and bedding planes respectively.

It is frequently difficult to determine the vergence of the early minor folds as they are often very long limbed symmetrical isoclines. Away from the cores of major F_3 folds the early fold hinges are frequently co-linear to the F_3 minor fold hinges, but they are not co-linear in the vicinity of F_3 fold cores, eg, on the summit of Beinn an Tuim and near the summit of Sgurr a Mhuidhe. The vergence of the minor early folds does not define the closure of any major early isocline in the area: nor is the sense of

Figure 58.

Stereographic projections.

GLENFINNAN DIVISION WEST OF THE BEINN AN TUIM FAULT.



A). Poles to axial planes of minor F_2 folds and F_2 cleavage.

B). Hinge lines of minor F_2 folds.

C). Poles to bedding planes.

Pre D_3 deformation features could be either D_1 or D_2 age, see Chapter 5.6f.

vergence constant so that the rocks would be on one limb of a major early isocline or fold nappe. However any interpretation of early fold vergence is made all the more complex when the possibility is considered that the early minor folds may be of two generations.

5.7) Brittle deformation.

This section discusses the Beinn an Tuim fault, the other minor faults and the prominent easterly dipping joint set possibly associated with easterly dipping cataclastic thrusts.

5.7a) The Beinn an Tuim fault.

This fault (Maps 3 & 4) has been used to define structural sub-areas in the region. The fault is a very obvious major geological and topographical feature; it produces a fault valley up to 100 metres deep with slopes of over 45°. The deep fault valley rises to a height of 550 metres on the SW side of Beinn an Tuim where it splays into a set of less well formed valleys. Further NE the fault is no longer a topographic feature.

There is no exposure in the floor of the fault valley, but at the top of the valley (NM 928827) fault breccias are exposed and are intruded by later basalt dykes. In view of topographic relief and the straight trace of the fault, the fault plane must be nearly vertical.

Across the fault there are marked changes of lithology and structure, the most obvious being the restriction of the Ardgour granitic gneiss to the SE. The F_2 structures of the Beinn an Tuim synform and Meall nan Damh fold cannot be traced NW across the fault (see Maps 3 & 5), and since the folds are a synform/antiform pair with vertical axial plane traces and locally only moderate plunge, their axial plane traces should continue west of the fault plane assuming that the folds had a relatively cylindrical geometry and fault displacement was vertical. If the fault displacement was lateral then the displaced axial plane trace should be traceable.

To the NW there is no sign of the displacement of the Loch Arkaig granitic gneiss (I.G.S Geol.Map 1:50,000 Sheet 62W, Scotland) where the fault may reasonably be expected to cut across it, suggesting either that the fault dies out in this direction within a distance of 5-10 Km. or that the displacement direction of the fault at Loch Arkaig is co-linear with the intersection of the granitic gneiss and the fault plane so that fault movement has produced no apparent displacement on the map.

The trace of the fault plane is perfectly straight and shows no indication of having been folded during the formation of the F_4 Glen Dubh Lighe antiform and is presumably post-D_4 in age. The general NE-SW Caledonian trend may indicate that the fault is possibly of late-Caledonian age.

No other major faults have been identified in the area.

5.7b) Joints and thrusts.

The exposures on the hillsides of the area, especially west of Glenfinnan village are bounded by an extremely well developed set of joint planes which dip eastwards at 30-40°. These planes have no obvious geometrical relationship to any of the major fold phases of the area; they appear to be later than all of the ductile deformation. However, some of the joints are occupied by sheets of post-D_3/pre-D_4 microdiorite (see Chapter 8.1) which are frequently highly schistose and metamorphosed. The age of the microdiorite deformation, with its bearing on the age of the jointing is discussed in Chapter 8.

In the Loch Eil Division part of the course of the Abhain Bheagaig (Fig. 38) follows the outcrop of a small thrust plane which dips to the east at 32°. The thickness of the thrust zone varies from 1 metre to 10 cm. along its length. Within the zone blocks of country rock psammite up to 1 metre long and 40 cm. thick are contained within a green rock flour matrix. Smaller blocks are partly streaked out into a crude schistosity.

The rock flour is highly chloritic and veined by calcite which is altering and replacing the brecciated material. The thrust plane cuts across large late transgressive pegmatites. A similar small thrust plane has been mapped in the stream bed of the river Fionn Lighe. These thrusts have a similar orientation to the major joint set mentioned above and could be of the same age, however the low grade (greenschist) metamorphic mineral assemblage of the rock flour in the thrust zone suggests a relatively late (Caledonian ?) age.

C H A P T E R 6.

Metamorphism.

6.1) Introduction.

The rocks of the area are predominantly a sequence of mixed psammites and pelites which have undergone a complex sequence of metamorphism and deformation. The petrography of the various rock types and their history of deformation have already been discussed (chapters 4 & 5). It is the object of this chapter to examine the metamorphic history of the area in terms of its grade, its relationship to the phases of deformation and its timing.

Although pelitic and semi-pelitic schists and gneisses are fairly common throughout the area they generally do not contain mineral assemblages which can be used to establish a detailed pattern of metamorphic zones across the area. In particular the polymorphs of Al_2SiO_5, Kyanite and sillimanite, although found occasionally, are generally absent from the pelites of the area. However their sporadic occurrence and the much more widespread occurrence of migmatitic pelites indicate that medium to high grade metamorphism has occurred throughout the area.

The general absence of Al_2SiO_5 has been ascribed to a whole rock chemical compositional control (Winchester,1974b).

As is commonly the case elsewhere, siliceous and feldspathic sandstones tend not to show complex mineralogical changes in relation to variations in temperature. Kennedy (1949), acknowledging these difficulties in establishing the metamorphic zonation of the western Highlands, proposed a zonal succession based on the mineral assemblages found within very subordinate but widespread calc-silicates. His map of metamorphic zones in western Inverness-shire and northwestern Argyllshire is reproduced below (Fig.59). Any interpretation of this and other maps of metamorphic zones in the area must be made with a knowledge of the history of deformation of the area. Consideration must be given to the possibility of folding and re-folding of isograds by subsequent phases of deformation and the possibility of polyphase metamorphism related to the polyphase sequence of deformation.

6.2) Mineral assemblages and metamorphic grade.

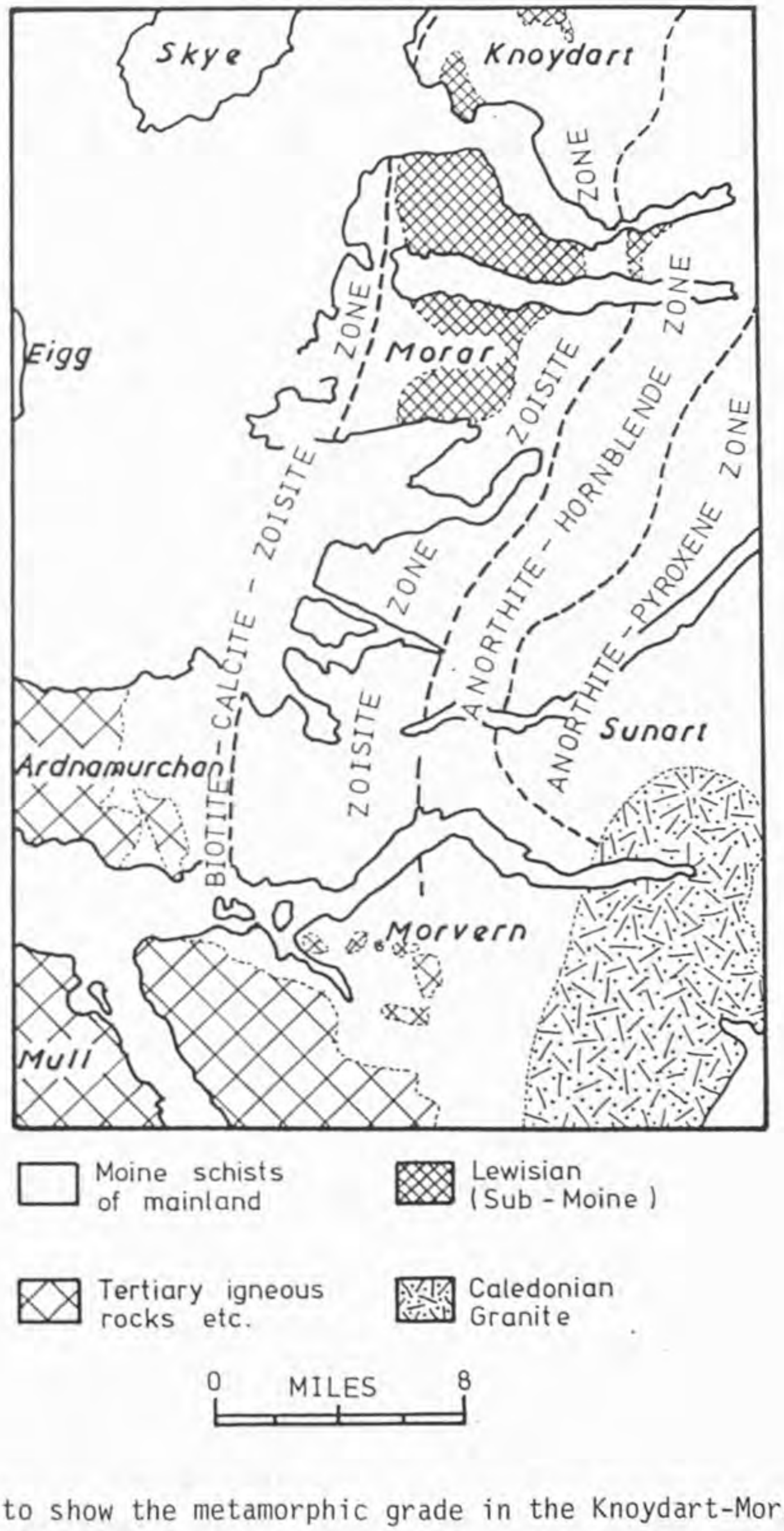
6.2a) Pelitic rocks.

Pelitic and semi-pelitic rocks are common in the Glenfinnan and Morar Divisions, but only a few thin semi-pelites have been recovered from the Loch Eil Division.

The usual mineral assemblage found is biotite-muscovite-plagioclase-quartz-garnet with accessory amounts of iron ore, apatite and zircon. Sillimanite, staurolite and chlorite have also been found in some sections. No obvious K-feldspar is present (for modal analyses of pelites and semi-pelites see Figs.31 & 26 respectively).

The widespread presence of garnet places the pelites in the "garnet zone" (Barrow,1912). Winkler (1979) correlates the first appearance of almandine rich garnet with the higher temperature part of low grade metamorphism and he uses reactions which produce staurolite as reaction-isograds to define the onset of medium grade metamorphism. However, these staurolite reactions are of limited use as the formation of staurolite appears to

Figure 59.



Map to show the metamorphic grade in the Knoydart-Morar area based on calc-silicate metamorphic mineral assemblages.
(Re-drawn after Kennedy,1949).

be governed by the bulk composition of the rocks and unfortunately the chemical composition of most pelitic sediments does not allow for its formation in medium grade metamorphism (Hoschek,1969).

Sillimanite, in the form of fibrolite, occurs in very small quantities in a number of sections (see Fig.31). Winkler (1979) notes a number of reactions which produce sillimanite and biotite at the expense of chlorite+ muscovite±quartz±staurolite within the range of medium grade metamorphism. However all the fibrolite observed in thin section appears to have grown within large muscovite crystals and is apparently not related to any of the other reactants mentioned by Winkler.

The general absence of Al_2SiO_5 polymorphs in high grade Moine pelites has been discussed by Winchester (1974b). Whole rock analyses of these pelites when compared to others indicate that Al_2SiO_5 polymorphs do not occur in pelites which are relatively rich in CaO and deficient in Al_2O_3.

The common occurrence of quartz-feldspar segregations within the pelites of the area lead naturally to considerations of the pelites as migmatites which have undergone partial anatexis. Winkler (1979,p.320) discusses the anatexis of rocks containing Plagioclase + Quartz + Biotite + Muscovite which at P=5Kb. and approximately 680° begin to melt. Muscovite disappears and the melt contains K-feldspar togather with Quartz and Albitic Plagioclase. The reaction involved is:

Musc. + Qtz. + Plag.(+ Biotite)→K-feld. + Plag. + Qtz.(in the melt)
+ An richer Plag. + Sill. + H O (+ Biotite).

At higher temperatures the biotite reacts with sillimanite to produce more K-feldspar plus either almandine or cordierite.

Thus theoretically, pelites containing no K-feldspar produce an anatectic melt of granitic composition and should contain K-Feldspar.

In pelites containing Plagioclase + Quartz + Biotite, but lacking muscovite, only a small amount of biotite is removed at the onset of anatexis. K-feldspar is produced and frequently hornblende and sphene are also produced during anatexis.

In summary, anatexis of both of the assemblages discussed above should produce a melt containing K-feldspar together with Plagioclase and Quartz.

None of the pelites of the study area contain K-feldspar, either in the quartz-feldspar segregations or their mica rich melanosomes. It is assumed that these segregations are not, therefore, the product of partial anatexis, as considered above. Thus they would appear to be the result of solid state metamorphic diffusion and segregation, or some process of metasomatism.

In conclusion, on the basis of pelite parageneses, the sporadic occurence of fibrolite and staurolite indicates that the metamorphism reached at least medium grade. The widespread development of garnet as an index mineral and the general absence of staurolite and Al_2SiO_5 polymorphs precludes the accurate establishment of "Barrow zones" of metamorphism within the area.

Barr (1983) using techniques of statistical analysis of geochemical data and normative calculation reached similar conclusions, namely that Moine migmatites have formed by sub-solidus segregation rather than partial melting. He noted that rocks of greywacke-like or arkosic composition are more readily migmatised than rocks of pelitic, quartzose semi-pelitic, and quartzose psammitic composition.

6.2b) Calc-silicates.

The petrography and modal analyses of all of the calc-silicates collected in the area are given in Chapter 4,Figure 20. The dominant assemblage is Quartz + Plagioclase + Amphibole + "Epidote".

"Epidote" is used here to indicate all mineral compositions in the solid solution series between the end members epidote and clinozoisite. Zoisite was not found in any of the sections. Occasionally small quantities of pyroxene are found together with amphibole. Other minerals which occur in small quantities are biotite, muscovite, chlorite, calcite, iron ore sphene, apatite and zircon. Plagioclase feldspar occurs in all sections and varies in composition from An.42% to An.97%.

Kennedy (1949) noted progressive changes in the mineralogy of calc-silicates collected between the west coast and the Great Glen fault. Various parageneses were recognised and were used to define four metamorphic zones, increasing in grade eastwards (Fig.59, redrawn from Kennedy,1949).

Kennedy's four zones are:

1) Zoisite - (calcite) - biotite zone. (West)
2) Zoisite zone.
3) Anorthite - Hornblende zone.
4) Anorthite - Pyroxene zone. (East)

He noted that an eastward increase in grade is accompanied by changes in plagioclase composition which he claimed was not gradual but changed abruptly from albite to bytownite (Kennedy,1949,p.48).

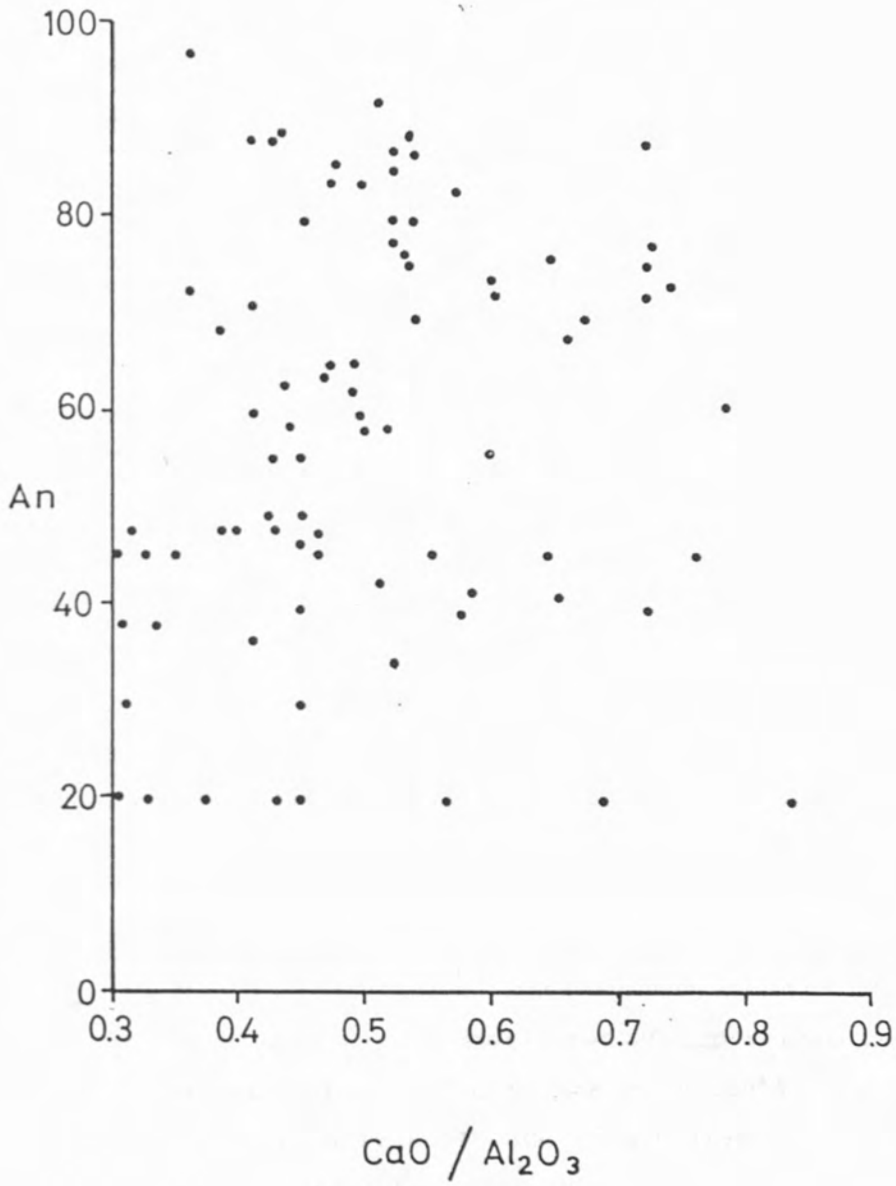
The calc-silicates collected from the western side of the area occupy Kennedy's Anorthite - Hornblende zone and Anorthite - Pyroxene zone, considered by him to correlate with the Kyanite and Sillimanite zones respectively, of "normal" pelitic schists.

Regional studies after the work of Kennedy (1949) reduced the Albite-Bytownite gap in plagioclase composition to a much narrower Andesine-Bytownite gap (see Tanner,1976) and Tanner suggested that more extensive collecting might prove that a gap in plagioclase composition does not exist.

Powell et al.(1981) using calc-silicates collected within the area studied by Kennedy have shown, on the basis of optical and microprobe analysis, that a compositional gap does not occur (Fig.60, taken from Powell et al: text fig.3). The preliminary work of Powell et al.(1981) strongly suggests that the anorthite content of plagioclase, apparently almost irrespective of calc-silicate whole rock chemistry, provides an indication of metamorphic grade (see Chapter 6.3e). A very simplistic view is that plagioclase has recrystallised during solid state metamorphism with a composition which is in equilibrium with the metamorphic grade and the other minerals (pyroxene, hornblende, biotite etc.) are formed from what "remains" after plagioclase has established equilibrium composition.

Attempts to study mineral assemblages and the chemical reactions responsible for the variety of assemblages (eg. Winchester,1972: Tanner,1976) indicate the complexity of the thermodynamic systems involved. In different areas of the Moine rocks of NW Scotland there are reportedly different prograde and retrograde metamorphic sequences of mineral assemblages and chemical reactions (compare Winchester,1972: Tanner,1976:Tanner & Miller, 1980 and Powell et al,1981). Accepting these reportedly different sequences of events, it may be that they have proceeded under different conditions of P, T, p_{H_2O}, p_{CO_2} with the result, for example, that the relative stabilities of epidote group minerals may have varied.

Figure 60.



Graph of Anorthite content of plagioclase versus whole rock CaO/Al_2O_3 ratio in calc-silicates.
(see text for discussion).

6.2c) The Ardgour granitic gneiss.

The petrography of the Ardgour granitic gneiss and some indication of its modal mineral composition has been given in Chapter 4. The dominant mineral paragenesis is Quartz + Plagioclase + K-feldspar + Biotite with much lesser amounts of garnet, muscovite, chlorite and epidote plus accessory iron ore, zircon and apatite (Fig.61 shows typical textures of the Ardgour granitic gneiss).

It is worthy of note that sillimanite has not been observed in the gneiss, cordierite is also absent and garnet is present only in very small quantities (up to 2%, see Fig.36). Muscovite is present in small quantities, either along the cleavage of biotite or as irregular patches within the felsic minerals. It is considered, on textural grounds, to be later than the Qtz.+ Plag.+ K-feld.+ Biot. migmatite assemblage (Fig.62).

The gneiss contains Qtz.+ Plag.+ k-feld. segregations varying in size from small augen a few mm. long up to "early" concordant pegmatites, usually with thin selvages of biotite. There are no striking mineralogical differences between the leucosomes and palaeosomes, only differences in the ratios of major mineral components.

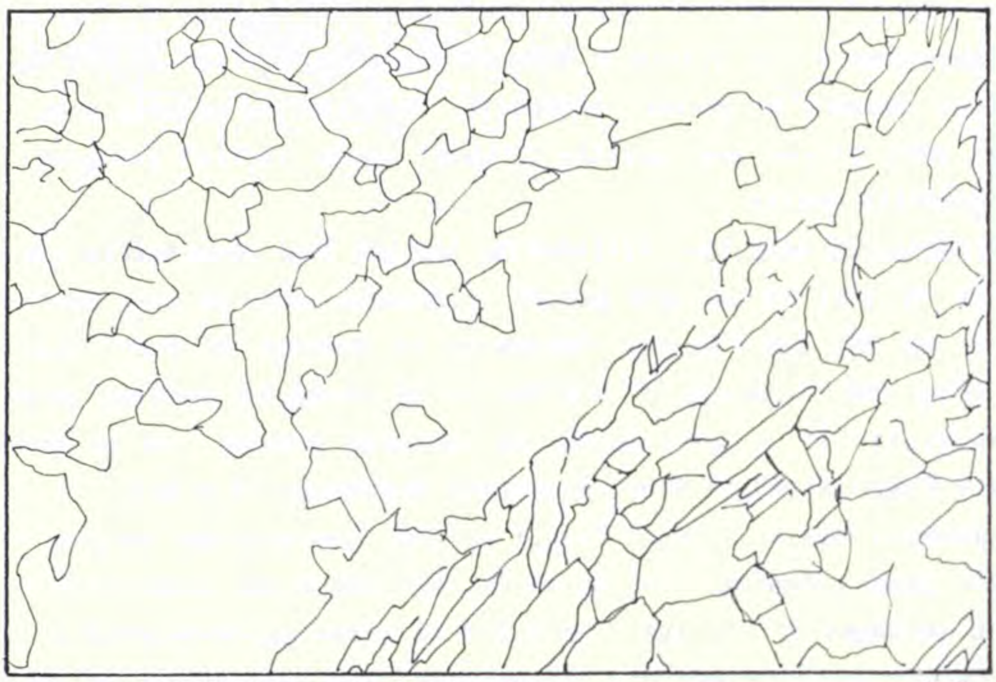
If the interpretation of muscovite as a product of secondary retrogression unrelated to the main migmatite forming event is correct, then the absence of primary muscovite in this assemblage would indicate that the migmatite has undergone high grade metamorphism (Winkler,1979. p,83-88).

The paragenesis Qtz.+ Plag.+ K-feld.+ Biotite has been the subject of a number of experimental studies (see Winkler,1979. Chapter 18). The results of these studies are discussed below in relation to the Ardgour granitic gneiss and the adjacent Glenfinnan Division "migmatitic" pelites. In the presence of water vapour at temperatures of approximately 650-670°c. the paragenesis begins to melt. Initially the melt contains approximately equal proportions of Quartz, K-feldspar and Plagioclase. Melt is generated rapidly within 10-20°c. temperature range until the least abundant of the solid phases Qtz.- K-feld.- Plag. is totally melted. A further rise in temperature is then required to continue melting the remaining two solid phases. With a further rise in temperature the less abundant of the two remaining solid phases may melt completely leaving only one of the quartzofeldspathic phases as a solid, together with biotite. Large temperature rises in the order of another 100°c. may be necessary to melt the remaining solid phases.

Only a very small part of the biotite dissolves incongruently into the melt, increasing the amount of K-feldspar in the melt as temperature rises

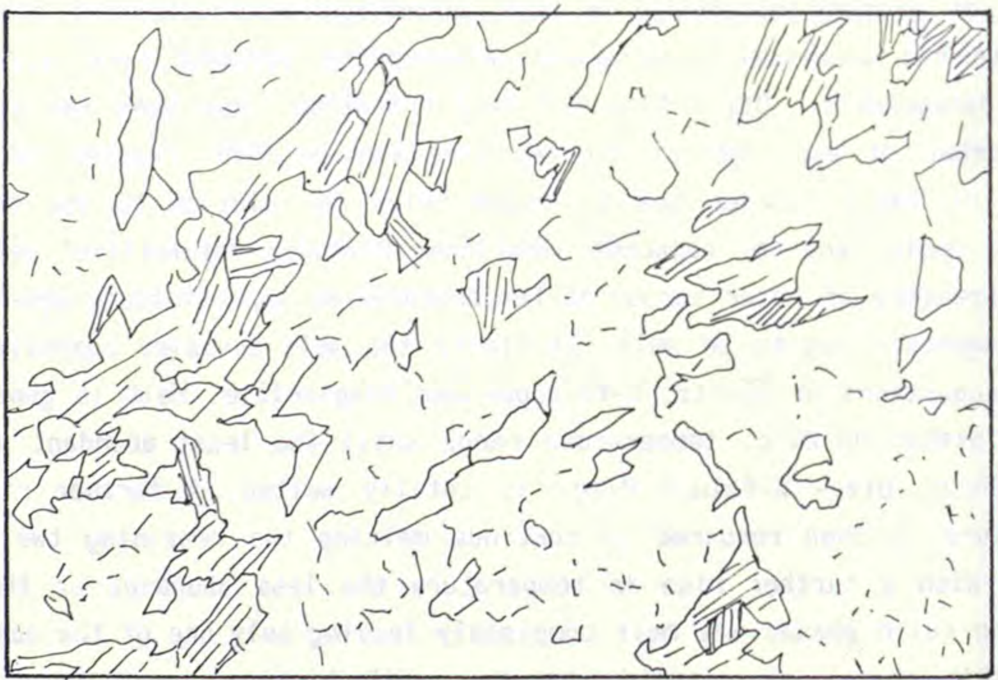
Figure 61. Photomicrographs of textures which are typical of the Ardgour granitic gneiss.

(a)

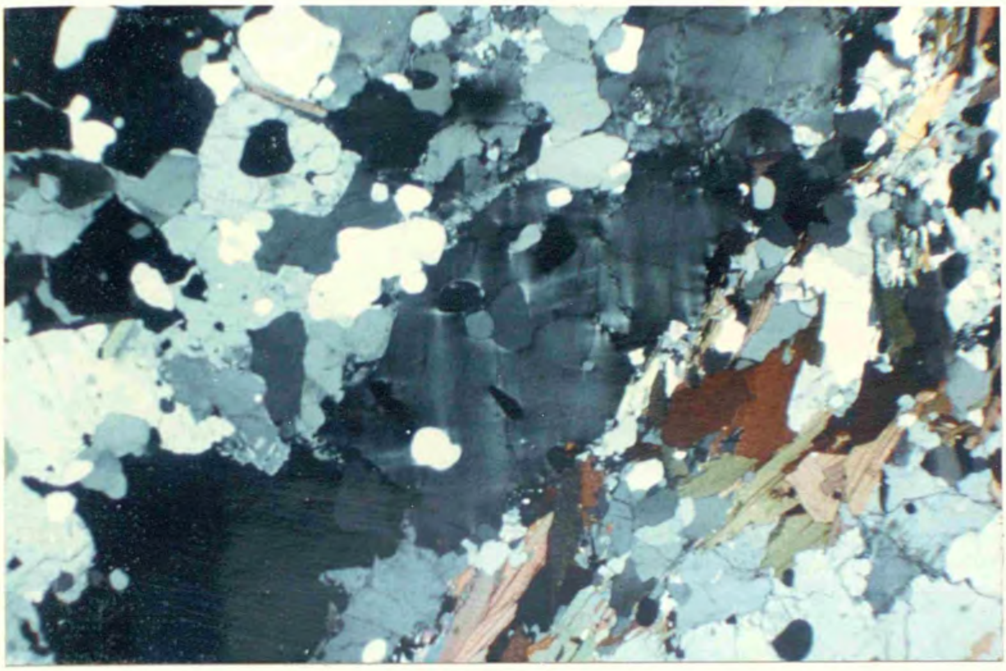


(Exp.277/1235).crossed polars 0 mm. 0.5

(b)



(Exp.278/1235). plane polars. 0 mm. 0.5



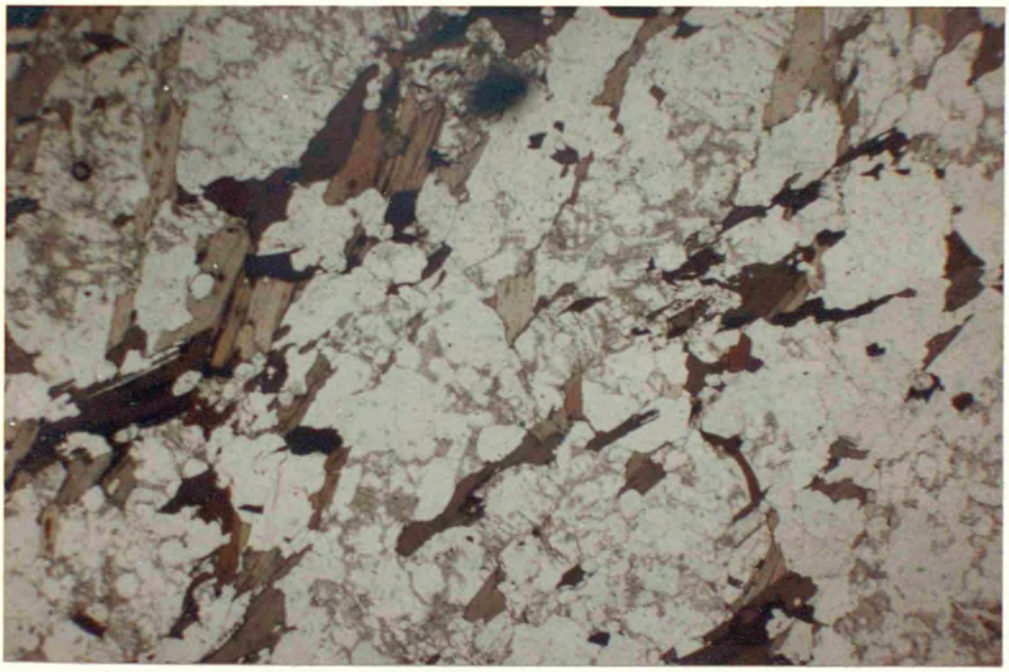
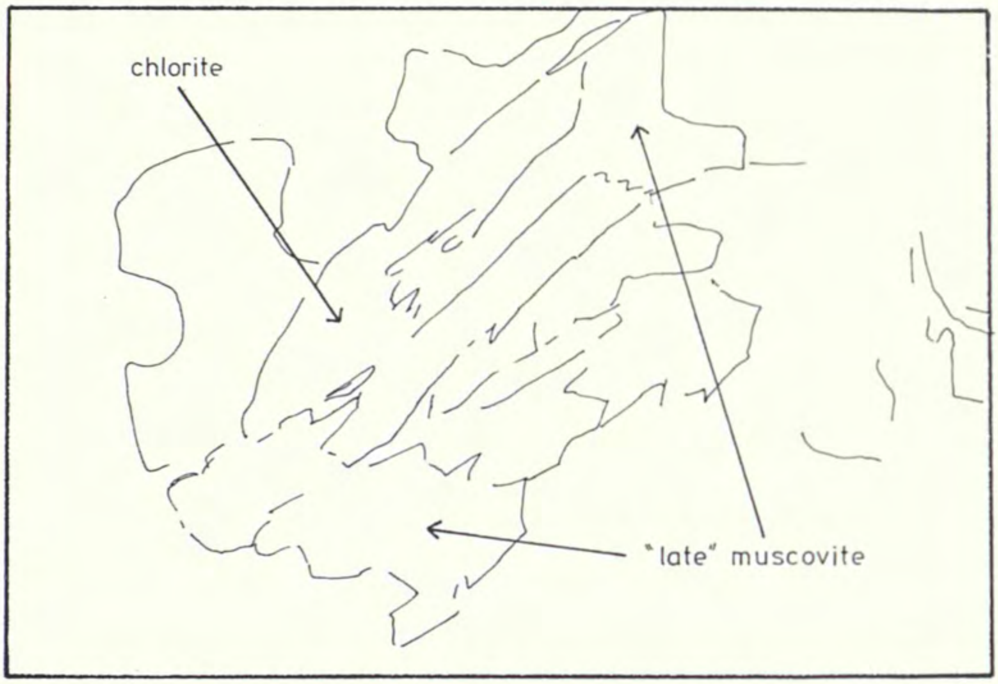
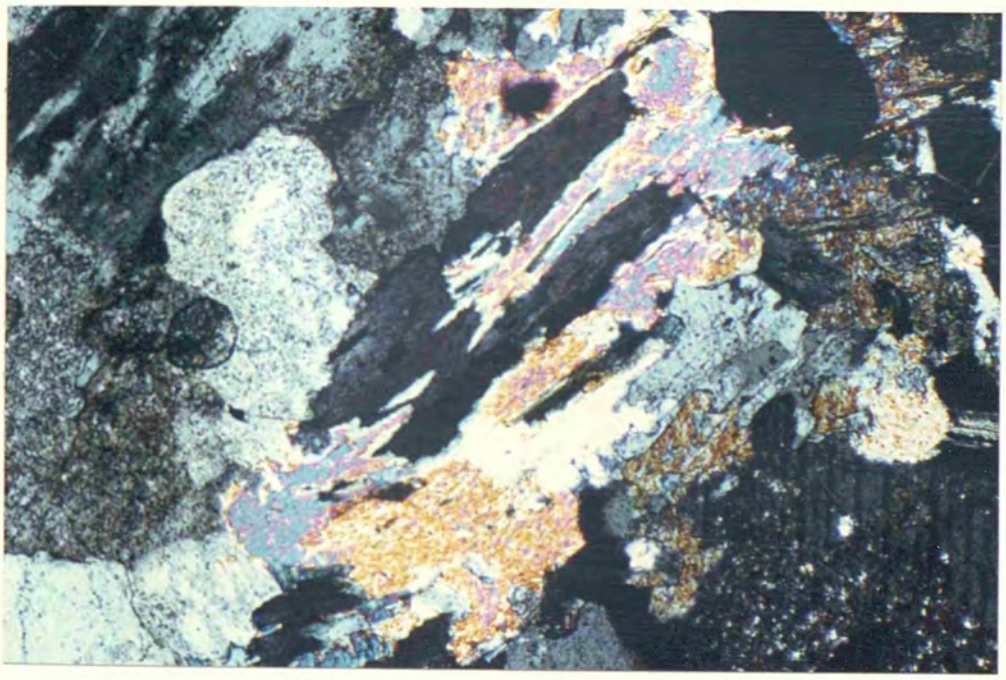


Figure 62. Photomicrograph to show muscovite textural relationships within the Ardgour granitic gneiss.



(Exp.14/92). crossed polars.

0 mm. 1



by 50-60°c. from the minimum temperature of melt.

Thus in these cases with sufficient H_2O available, anatexis of the paragenesis Qtz.+ Plag.+ K-feld.+ Biotite produces granitic, granodioritic or trondhjemitic leucosomes within a few tens of degrees centigrade of initial melting, leaving a melanosome (unmelted residue) of biotite plus residual feldspar and / or quartz.

With insufficient water available to saturate the system completely, melting occurs rapidly within a limited temperature range until there is no longer enough water to saturate the melt. Continued melting then requires much higher temperatures. The result of this is the production of limited amounts of leucosome and melanosome together with areas of palaeosome which have lacked sufficient H_2O to undergo anatexis.

If the composition of the plagioclase in the paragenesis is more anorthitic then slightly higher temperatures (10-20°c.) are required to initiate melting.

The experimental results outlined above assume an initial assemblage of Qtz.+ Plag.+ K-feld.+ Biotite with subsequent anatexis producing granitic (sensuo lato) leucosomes.

If the Ardgour granitic gneiss had formed from the adjacent Glenfinnan Division pelites or semi-pelites then a comparison of the two assemblages shows that the granitic gneiss contains K-feldspar and generally lacks primary muscovite. Consequently it is necessary to consider possible reactions involving Quartz, Plagioclase, Biotite, Muscovite and K-feldspar which may be responsible for the production of K-feldspar in the granitic gneiss, possibly at the expense of muscovite. Experimental work on this system has concentrated on the equilibrium.

(1) Musc.+ Qtz. $\rightleftharpoons$ K-feld.+ Sillimanite + H_2O (Althaus et al.1970)

or when plagioclase is included the reaction is.

(2) Musc.+ Qtz.+ Plag. $\rightarrow$ K-feld.+ Plag.+ Qtz. (in the anatectic melt)
+ $An._{richer}$ Plag. + Sillimanite + H_2O
(Winkler,1966)

These prograde metamorphic reactions which could explain the conversion of pelite to "granitic gneiss", would be responsible for the production of new K-feldspar and the loss of muscovite, however sillimanite, which would also be produced by the reactions, does not occur in the granitic gneiss. If biotite is present during reaction (2) then at higher temperature

Biot.+ Sill.+ Qtz. $\rightarrow$ K-feld.+ Cordierite + H_2O
$\searrow$ (or at Higher Pressure)
K-feld.+ Almandine + H_2O (von Platen & Holler,1967)

This higher temperature reaction could remove any sillimanite produced by reactions (1) or (2) but would produce either cordierite and/or garnet. Cordierite has not been observed in the Ardgour granitic gneiss and garnet occurs only in very small quantities, not enough to be related to the large quantities of K-feldspar which are present in all sections.

It is concluded that while biotite as well as muscovite can constitute a source for K-feldspar, the other products of such reactions, sillimanite, cordierite and/or garnet are not present in sufficiently large quantities for such postulated reactions to have been the source of all the k-feldspar in the Ardgour granitic gneiss. Thus, arguing from the theoretical grounds outlined above, it would appear that the Ardgour granitic gneiss is not the product of high grade metamorphism of local pelitic rocks.

The possibility that the Ardgour granitic gneiss was generated from the adjacent pelites has been examined geochemically by Gould (1966). Figure 63 (modified after Gould,1966, Table 27) shows an averaged composition of the granitic gneiss together with various analyses of "Ardgour pelites" and "Ardgour psammites".

It is obvious that the Ardgour granitic gneiss is highly potassic in comparison to local pelites, psammites and combinations of pelite and psammite. It has been argued above from a consideration of possible metamorphic reactions that the granitic gneiss is not a product of partial anatexis of adjacent pelites. However the granitic gneiss could be a partial melt of sediments orriginally rich in K_2O in marked contrast to the adjacent pelites. If the line of granitic gneiss bodies, reportedly similar to the Ardgour granitic gneiss, is traced northwards it is seen to transgress eastwards from the mixed pelites and psammites of the Glenfinnan Division into the psammites of the Loch Eil Division (see Fig.6 and I.G.S. 1:50,000 Sheet 62W,Scotland) which seemingly precludes the possibility that these granitic gneiss bodies are the direct anatectic products of a potassic sediment unless there was a narrow discontinuous zone of potassic sediments sub-parallel, but crossing, the Glenfinnan - Loch Eil Division boundary. Such sedimentological variation is extremely hard to envisage.

It is possible that the precursors to the granitic gneiss may have become potassic as a result of some form of metasomatism or "granitisation", so that the line of granitic gneissbodies marks an elongate zone of metasomatism of Moine sediments.

There remains the final possibility of the granitic gneiss being the partial anatectic melt of a granitic intrusion, intruded prior to the complex sequence of deformation and metamorphism which has obscured any field evidence of an intrusive origin such as sharp igneous contacts or a meta-

Figure 63.

Geochemical and Modal analyses from the Ardgour granitic gneiss.
(tabulated from Gould,1966).

	(A) Ardgour pelitic rock	(B) Ardgour psammitic rock	(C) Ardgour granitic gneiss
Si^{4+}	59.73	81.90	68.12
Ti^{4+}	0.68	0.22	0.41
Al^{3+}	18.66	9.16	14.67
Fe^{3+}	0.67	0.28	0.70
Fe^{2+}	3.86	0.70	2.26
Mn^{2+}	0.09	0.02	0.06
Mg^{2+}	2.82	0.47	0.82
Ca^{2+}	2.61	0.94	1.50
Na^{+}	6.92	4.02	5.78
K^{+}	3.80	2.24	5.55
P^{5+}	0.17	0.04	0.13

Eskola Molecular norms, recalculated to high - grade metamorphic minerals.

Quartz	23.33	61.43	31.64
K-feldspar	4.65	10.04	27.94
Plagioclase	45.42	22.42	29.05
Muscovite	2.18	1.66	—
Biotite	2369	3.42	9.02
Hornblende	—	—	1.00
Magnetite	0.32	0.32	0.73
Ilmenite	0.02	0.24	0.27
Apatite	0.43	0.10	0.33

A). Average of 15 new analyses of Ardgour pelitic rock (from Gould 1966, Table 27)

B). Average of 10 new analyses of Moine psammites, one from the Druim na Saille pelitic group and 9 from the Beinn an Tuim striped group (Gould, table 27).

C). Average of all analyses of Ardgour granitic gneiss (Gould , 1966 , table 14) morphic aureole. The regional discordance of the granitic gneiss bodies traced northwards is easily explained in terms of an original, slightly discordant to stratigraphy, suite of intrusions.

If the precursor to the Ardgour granitic gneiss was a granite then the lack of a metamorphic aureole would suggest either that it has not been observed, or that it has been destroyed by later metamorphism and deformation, or that the host rocks were hot at the time of intrusion and an aureole was not formed.

Comparisons of the metamorphic mineralogy and chemistry of the granitic gneiss and the Glenfinnan Division pelites in the light of experimental observations indicate that the pelite has not undergone partial anatexis during the production of its quartz-plagioclase segregations (=migmatitic lits). Anatexis would have produced K-feldspar in a granitic melt along with sillimanite, a reaction indicative of high grade metamorphism.

The granitic gneiss has, however, undergone partial anatexis under conditions with sufficient H_2O present to produce granitic and related melts from a Qtz.+ Plag.+ K-feld.+ Biotite assemblage. This anatexis occured under conditions of high grade metamorphism.

The Ardgour granitic gneiss contains two mappable masses of recognisable metasediment which appear to be gradational into the typical granitic gneiss (see Chapter 5). There are a number of reasons which could explain why these masses have not undergone partial anatexis. Assuming that the granitic gneiss was originally a granite then the sedimentary masses must have been xenolithic. The parallelism between the xenolithic modified bedding and the modified bedding in the adjacent metasediments would suggest that the granite intrusion was passive and not very far travelled from its source. If the granitic gneiss precursors were potassic sediments then it is possible that the metasedimentary masses within the granitic gneiss were either less potassic or contained plagioclase of a higher An. content, raising its anatectic melting temperature by some tens of degrees centigrade, or were in an area of low pH_2O with insufficient water to produce anatectic melts, ie. they are palaeosomes within a mixed leucosome/melanosome granitic gneiss.

There are no indicators that the Ardgour granitic gneiss and host metasediments have undergone different tectono-metamorphic histories, therefore the process which produced lit-par-lit segregations in the Glenfinnan Division by solid state processes at temperatures lower than those required for partial anatexis must also be presumed to have been operative in the granitic gneiss before a continued rise in temperature produced anatectic melts within the granitic gneiss. This would apply whether the precursor to the granitic gneiss was a potassic sediment or an intrusive granite.

The petrological and experimental work outlined above suggests that the Ardgour granitic gneiss has not formed by partial anatexis of the pelitic metasediments of the Glenfinnan Division. The granitic gneiss may have originated as an elongate zone of potassic sediments which was slightly oblique to the overall sedimentary facies boundary between the Glenfinnan and Loch Eil Divisions, but this seems sedimentologically improbable. Alternatively the potassium enrichment may be due to some form of metasomatism. In the light of the discussion above the author considers an intrusive origin for the Ardgour granitic gneiss to be most likely.

It is of course realised that further experimental work may reveal other reactions which can explain the presence or absence of K-feldspar, muscovite, Al_2SiO_5 etc. in the metasediments and granitic gneiss.

6.2d) Amphibolites.

This section concerns the mineral parageneses and metamorphic grade of the amphibolites in the area, the geochemistry and structural setting of the basic and metabasic rocks in the area is discussed in Chapter 8.

The most common amphibolite paragenesis is Amphibole + Plagioclase + Quartz ± Biotite ± Garnet + accessory minerals. The amphibole, which occurs in all sections and comprises from 40-80% of the rock, is hornblende with a pale coloured pleochroic scheme.

α = Straw yellow.
β = Mid-dark green.
γ = Dark green.

Plagioclase is found in all sections, comprising 20-40% of each section. The anorthite content ranges from An.28% to An.69%. Frequently the plagioclase is at least partly saussuritised. Quartz is found in virtually all sections, comprising about 5-20% of most sections. Biotite is present in approximately half of the sections, as is garnet, but there is no relationship between the presence or absence of either mineral. Other minerals found in relatively small quantities include chlorite, calcite, clinozoisite and epidote. Accessory amounts of sphene and iron ore are present in nearly all sections.

Comparisons of amphibolites from the Glenfinnan and Loch Eil Divisions show that the anorthite content of plagioclase is more varied within the Glenfinnan Division (An.28% to An.69%) whereas the anorthite content in the Loch Eil Division (An.44% to An.47%) is restricted to the central portion of the range exhibited within the Glenfinnan Division (Map 6). Garnet and biotite are less common and epidote/clinozoisite is more common in the amphibolites of the Loch Eil Division.

All the amphibolites have at least a moderate linear shape and crystallographic fabric defined by orientated, elongate hornblende crystals. There is often some suggestion that the linear fabric is slightly schistose,ie. planar. This L-S fabric is coeval with the penetrative planar fabric in the surrounding metasediments, which may be of S_1 or S_2 age.

The paragenesis Hornblende + Plagioclase + Quartz ± Biotite ± Garnet derived from the metamorphism of mafic rocks (see geochemistry, Chapter 8.3) is indicative of medium to high grade metamorphism (Winkler,1979). Plagioclase composition of Andesine/Labraborite indicates relatively high grades of metamorphism within this range (Wenk & Keller,1969). However, within the mapping area, the anorthite content of plagioclase in amphibolites is not sufficiently sensitive to variations of metamorphic grade to enable it to be used as a reliable indicator of metamorphic grade (Map 6).

The textural and structural relationships of the minerals chlorite, epidote, clinozoisite and calcite to the main paragenesis described above indicate the occurrence of either a subsequent lower grade metamorphism or a phase of retrogression separated from the main metamorphism by at least one phase of deformation (see Chapter 6.3d).

6.3) The relationship of metamorphism to deformation.

6.3a) Micas.

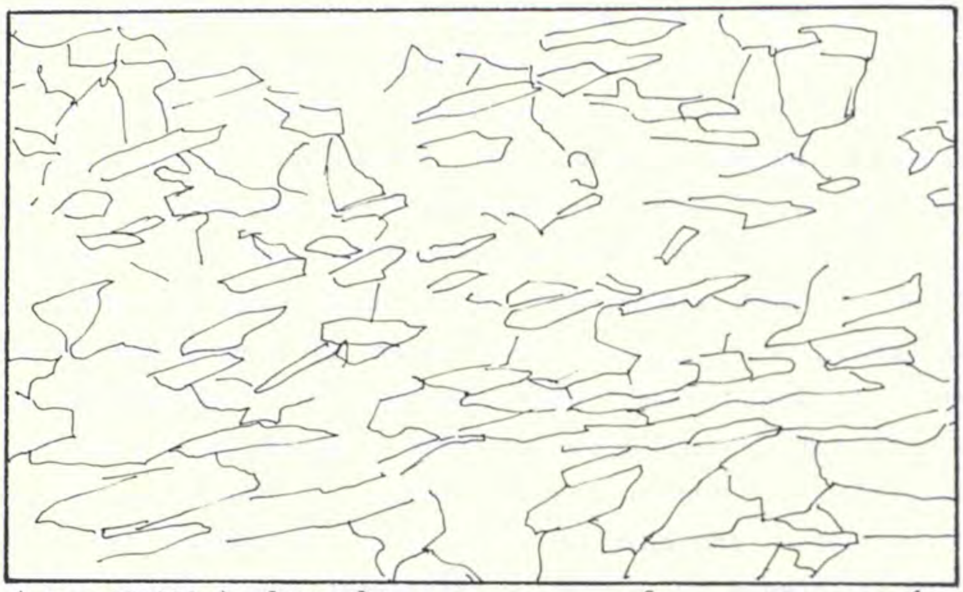
Both biotite and muscovite are abundant in the pelitic and semi-pelitic rocks of the area. In the more psammitic rocks both minerals are found in smaller quantities. Within the Ardgour granitic gneiss biotite is abundant but muscovite is either absent or present in only small quantities (see modal analyses Figs.12, 15, 17, 26, 31 and 36).

In all cases biotite laths define a very strong planar fabric. In rocks which show a compositional banding the fabric is generally parallel to the banding (foliation). In the hinges of minor tight to isoclinal F_2 intrafolial folds the fabric is axial planar and penetrative (Fig.64). Where two phases of early minor isoclinal folds can be seen (F_1 and F_2) the biotite fabric appears to be axial planar to both F_1 and F_2 folds (Fig.47b).

In the core of the F_2 Beinn an Tuim synform(Maps 2 & 3) the biotite fabric is axial planar to the tight minor F_2 folds in the core of the major

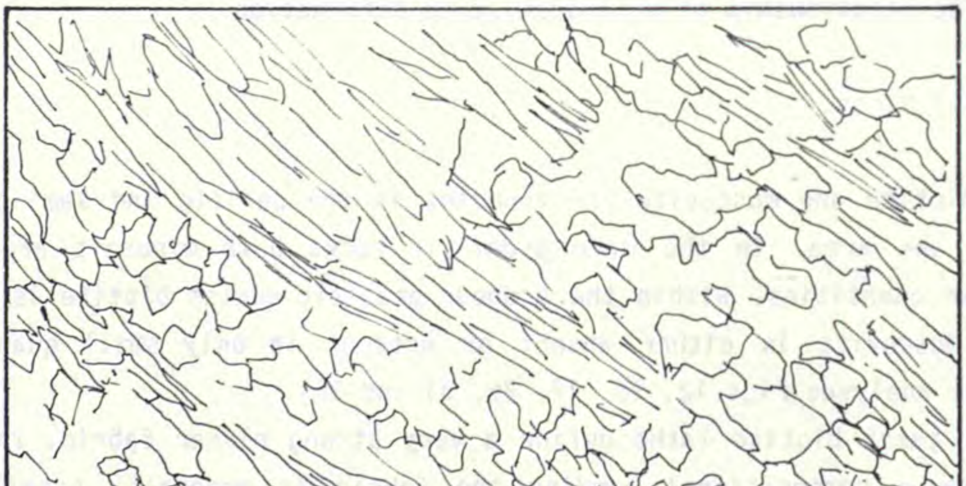
Figure 64. Photomicrographs of the planar S_2 biotite fabric in psammitic rocks.

(a)

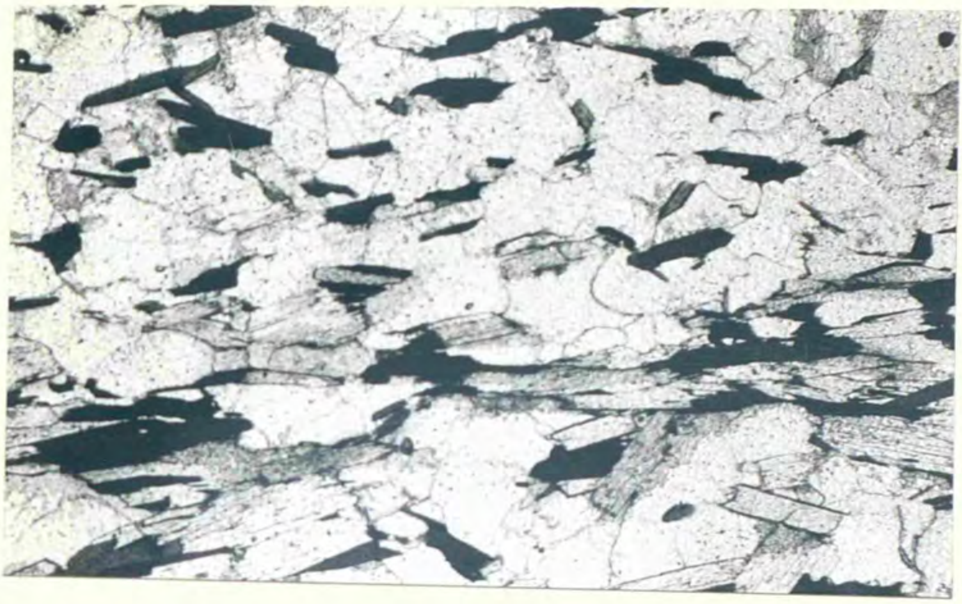


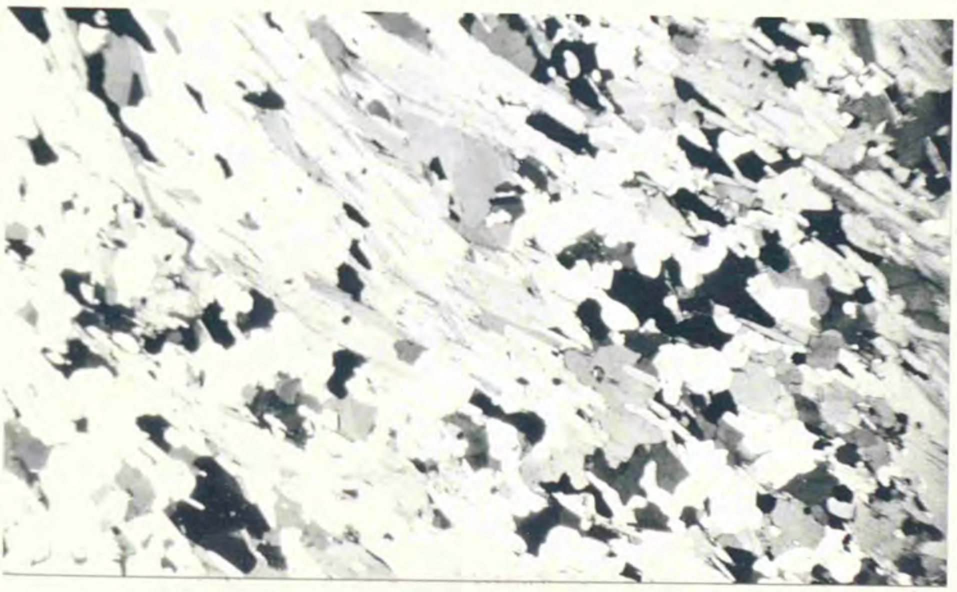
(Exp.38/229). plane polars. 0 mm. 1

(b)



(Exp.45(a)/230). crossed polars. 0 mm. 1





fold but in the more open limbs of the major fold, the minor F_2 folds fold a bedding parallel foliation but do not develop their own penetrative axial planar biotite fabric.

Throughout the area the strongly developed planar biotite fabric is highly crenulated by minor local F_3 folds. The crenulations are often very tight but generally the biotite crystals are completely unstrained (Fig.65a). Only very occasionally do some biotite crystals show slight straining of the lattice. Minor folds of F_4 generation were not observed (or recognised as such). F_5 minor folds crenulate the planar biotite fabric and again no strain can be seen in biotite crystals. In a few sections biotite has been extensively replaced by aggregates of chlorite.

Muscovite crystals most frequently have the same mode of occurrence as biotite and they reinforce the biotite planar fabric and are folded by F_3 crenulation folds. Like biotite, the muscovite laths generally do not show straining related to the F_3 or F_5 crenulations (Fig.65b).

Where biotite and muscovite are intergrown and form an intense planar mica fabric, the two minerals are in textural equilibrium with each other. There are no textural signs of any replacement of one mineral by the other, although sometimes it could be suggested that muscovite seems to be growing along the cleavage of biotite laths (Fig.65a,centre-right). Where F_3 crenulations are intense, occasionally a few muscovite crystals seem to have recrystallised parallel to the axial planes of the F_3 minor folds (Fig.65b centre-right).

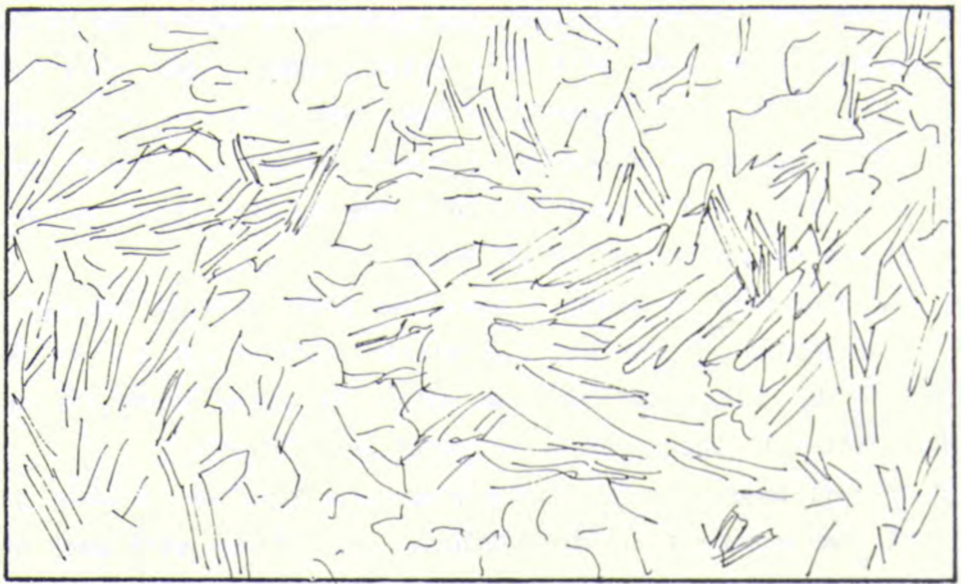
In much smaller quantities muscovite adopts two other modes of occurrence; Firstly, large porphyroblasts (up to 5 mm. long) are seen which have no preferred crystallographic or shape orientation and cross cut the dominant muscovite-biotite planar fabric (Fig.66a). Secondly, muscovite crystals up to 5 mm. long occur as porphyroblasts which have no preferred crystallographic orientation, they have a mantle of muscovite sub-grains and are wrapped by the muscovite-biotite planar fabric. These porphyroblasts tend to show the most internal lattice distortion (Fig.66b) and may be the sporadic remnants of a suite of early porphyroblasts which, subsequently, have been strained, and for the most part totally recrystalised.

In the granitic gneiss the small quantities of muscovite which occur in some sections occur as co-planar intergrowths with the planar biotite fabric and also as a late replacement of (?K-) feldspar.

Throughout the region major folds have been mapped and their structural age relative to other structures established (see Chapters 5 & 7). Along the axial plane trace of the F_2 Beinn an Tuim synform the minor structures range from tight to isoclinal (F_2) folds with an extremely intense axial

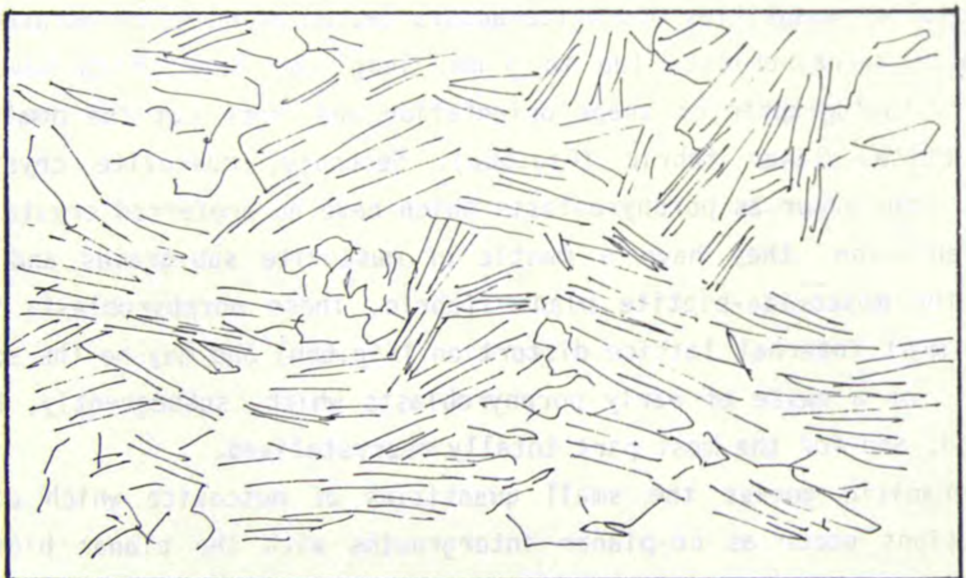
Figure 65. Photomicrographs showing the S_3 crenulation of the planar (S_2) fabric.

(a)

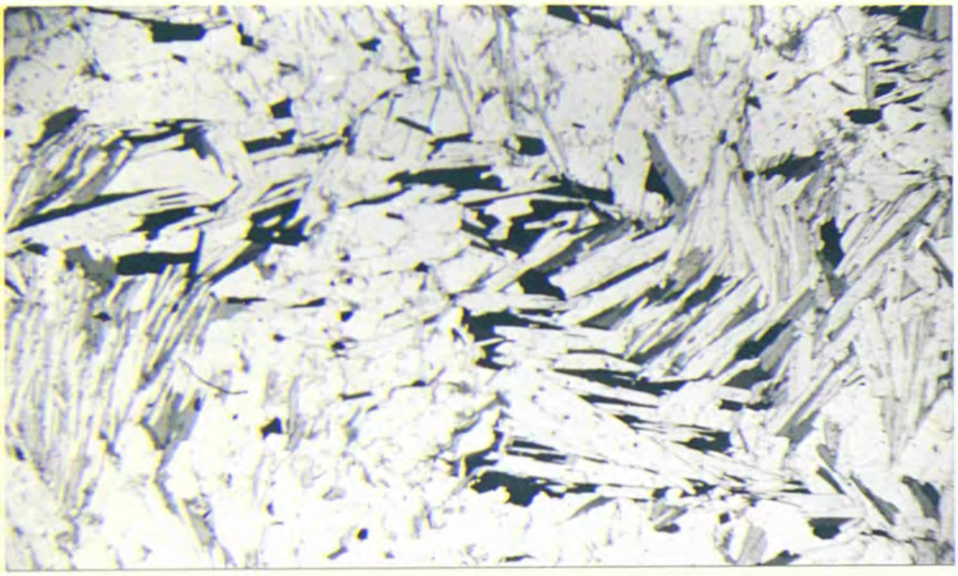


(Exp.47/231). plane polars. 0 mm. 1

(b)



(Exp.217/1089). plane polars. 0 mm. 1



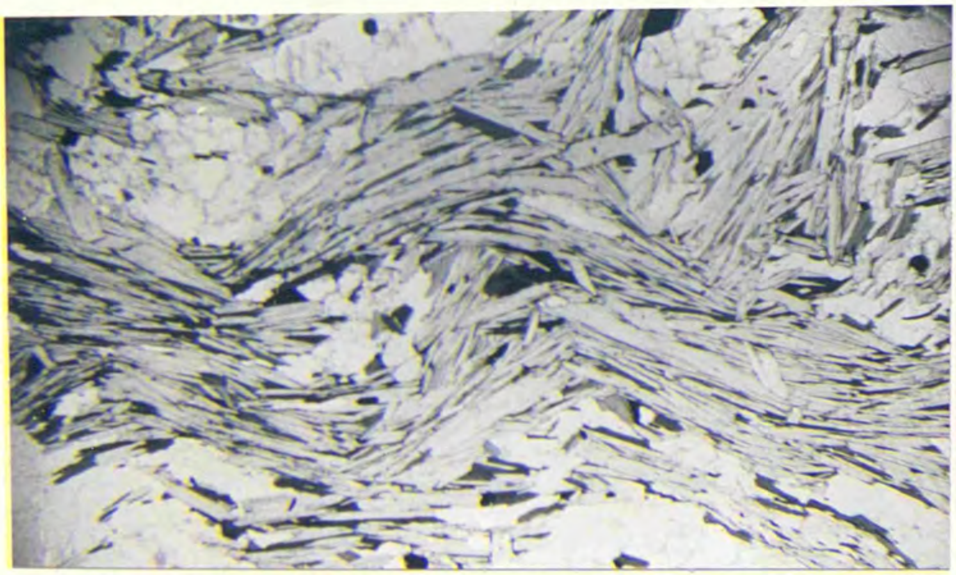
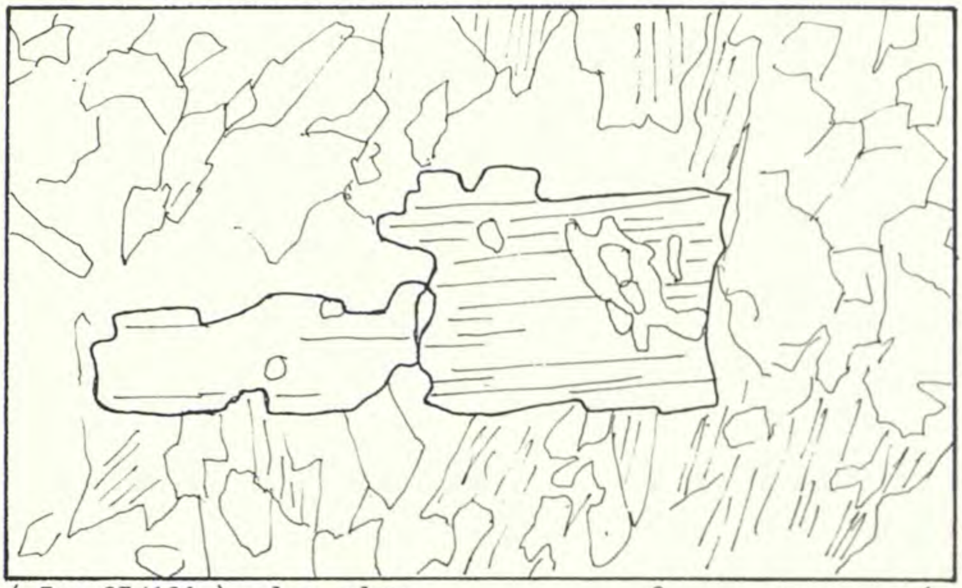


Figure 66. Photomicrographs to show muscovite textures.

Large non-orientated muscovite porphyroblasts.

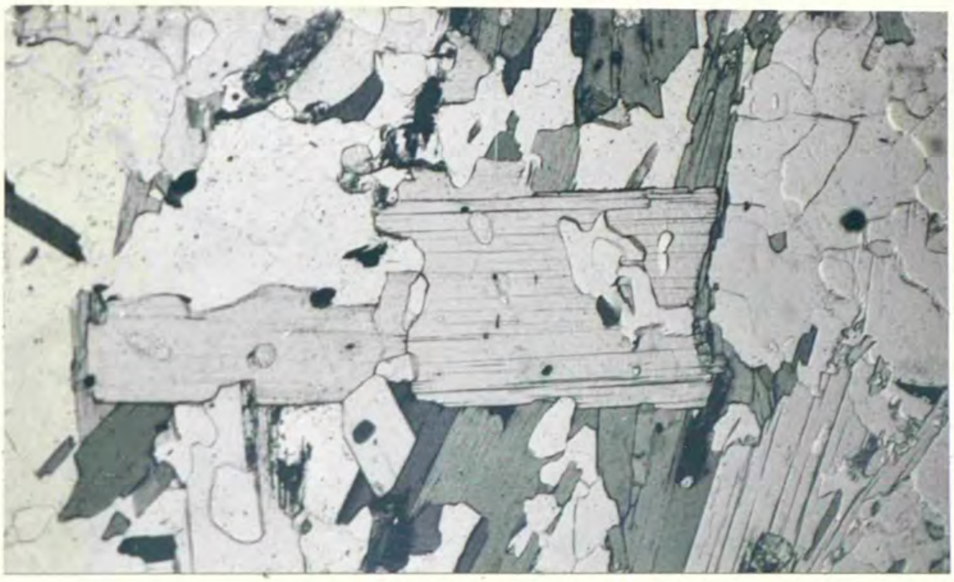


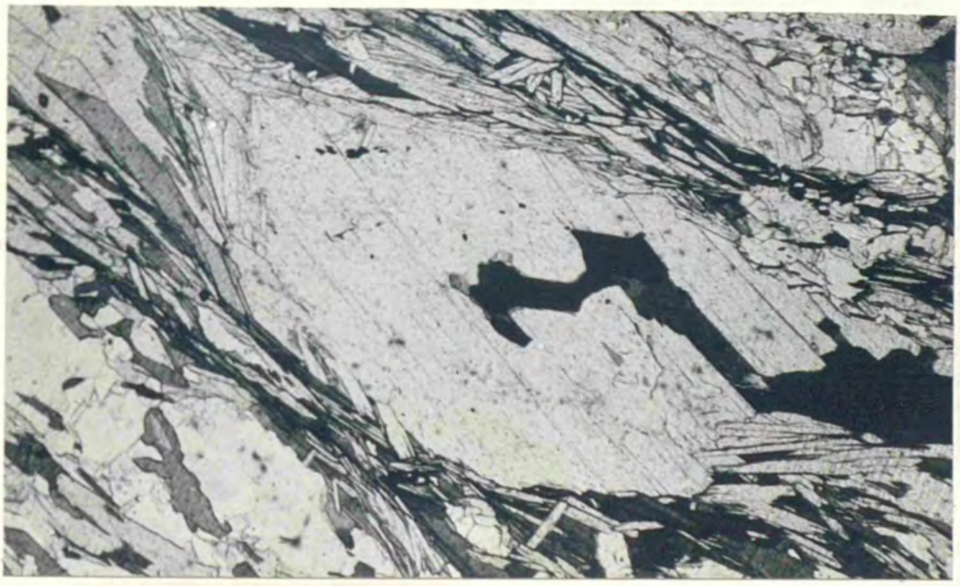
(Exp.27/164). plane polars. 0 mm. 1

(see text for discussion).

Deformed, early muscovite porphyroblasts.

(Exp.201/1050). plane polars. 0 mm. 1





planar gneissose fabric (S_2) which almost totally transposes the earlier gneissose fabric (S_1) in the pelites, to more open minor (F_2) folds which fold a pre-existing planar mica fabric (S_1) but do not contain a penetrative fabric axial planar to the F_2 minor folds.

Thus locally there is evidence of two planar penetrative mica fabrics which are pre-crenulation in age. There are very rare exposures which contain two phases of minor isoclinal folds, both of which appear to contain truly penetrative axial planar fabrics, and are deformed by crenulation folds.

The situation where the planar penetrative mica fabric may be the product of either D_1 and/or D_2 deformation is seen to be even more equivocal when it is realised that minor crenulation folds may be a product of not one but two phases of major folding (the earlier phase related to the production of the Sgurr Beag slide and the later one responsible for the folding of it: see Chapters 5 & 7).

Although not unequivocal, the following table is an attempt to summarise the history of mica growth and recrystallisation related to deformation.

D_1 Growth of "early" muscovite porphyroblasts, possibly after the production of a planar mica fabric.

D_2 Enhancement and syn-tectonic development of the dominant planar fabric with the deformation of "early" muscovite porphyroblasts.*

D_3–D_4 Strong crenulation of the planar fabric; followed by selective recrystallisation to produce unstrained micas (remnant mantled "early" muscovite porphyroblasts retain their internal strain)

D_5 Open crenulation and recrystallisation.

Post-D_5 Post tectonic static growth of randomly orientated muscovite porphyroblasts.

* Note: It is possible that the "early" muscovite porphyroblasts have been deformed by flattening during D_3 (or D_4) deformation. The porphyroblasts could then be late- or post- D_2 porphyroblasts.

6.3b) Al_2SiO_5 polymorphs.

Sillimanite is found sporadically in small quantities in some of the pelites in the area. Kyanite has not been observed in any rocks in the area but it occurs very sporadically in adjacent areas (Powell et al,1981. text Fig.1). Andalusite has not been recorded in the area or adjacent to to it but it occurs in the inner aureole of the Strontian granite 20 Km. to the south (Tyler & Ashworth,1982).

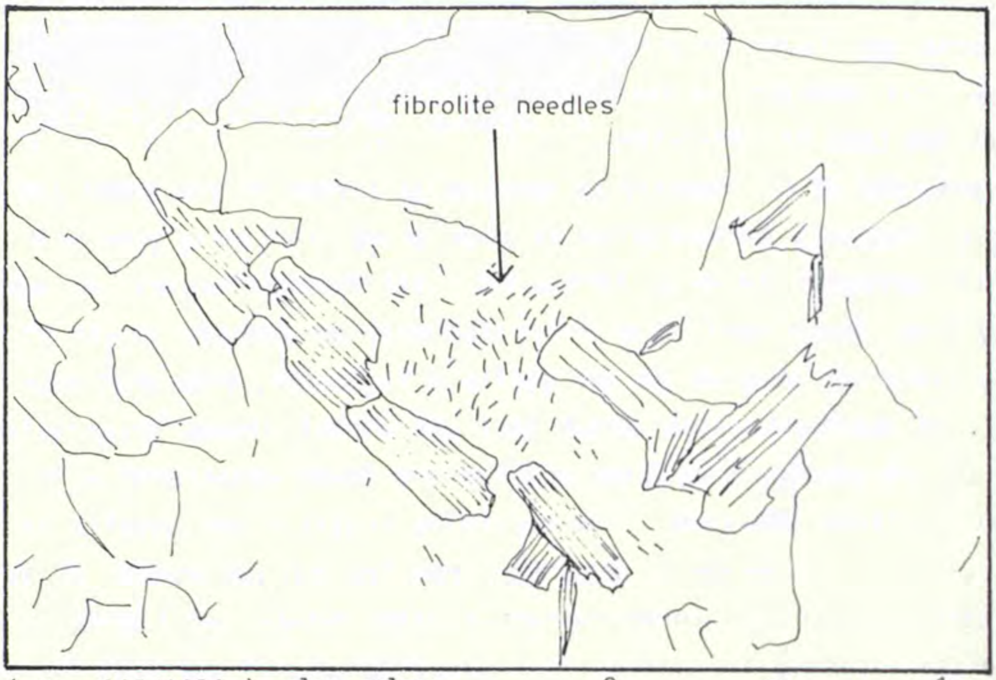
Sillimanite was observed in one pelitic rock collected from the Morar Division just west of the Sgurr Beag slide. In the Glenfinnan Division sillimanite was found only in pelitic and semi-pelitic lithologies, figure 34 shows its sporadic distribution. Sillimanite was not observed in the rocks ot the Loch Eil Division.

The presence of sillimanite as opposed to either of the other two Al_2SiO_5 polymorphs suggests metamorphism to high grade. Confining pressure is difficult to estimate (Winkler,1979.p,92). Sillimanite occurs as aggregates of very fine small needles of fibrolite with an average length of 0.1-0.2 mm. The fibrolite needles do not define a lineation, they anastomose within crystals of muscovite with which they are almost always invariably associated. In one section fibrolite needles are found associated with (? plag) feldspar (Section 246/1104). The muscovite crystals are usually relatively large, 2-3 mm., frequently deformed, mantled by sub-grains of muscovite (ie. "early" muscovite porphyroblasts). The "waves" of fibrolite needles show no relationship to either the crystallography of the muscovite crystals or to planar or crenulation fabrics of the pelitic rocks in which they were found. There are no signs of deformation of individual needles which constitute the "waves" of fibrolite (Fig.67a & b). It seems reasonable to assume that the fibrolite texture has been produced by fibrolite ovegrowing muscovite, rather than muscovite nucleating on fibrolite.

Fibrolite is found overgrowing relatively large micas which are strained, deformed into sub-grains and often wrapped by the dominant muscovite-biotite foliation. The flattening of these augen crystals seems to have occurred either during the production of S_2 or during D_3 deformation which also produced the strong crenulation of the planar fabric. Fibrolite was not observed growing in any muscovite in the dominant planar fabric, nor was it seen to grow in the large muscovite porphyroblasts which are undeformed and grow across the strong mica crenulation fabric, ie. fibrolite is found only in the "early" muscovite porphyroblasts. This could be interpreted as indicating fibrolite growth at an early stage of metamorphism in "early" muscovite porphyroblasts before any of the muscovite had been deformed and recrystallised. Deformation of the "early" porphyroblasts with subgrain production and presumably some total recrystallisation may have destroyed more widespread fibrolite growth. However, it is also possible to argue that fibrolite grew after the growth and recrystallisation of all the muscovite crystals, but that it only grew on the "early" porphyroblasts because they had retained strained lattices and were thus ideal sites for

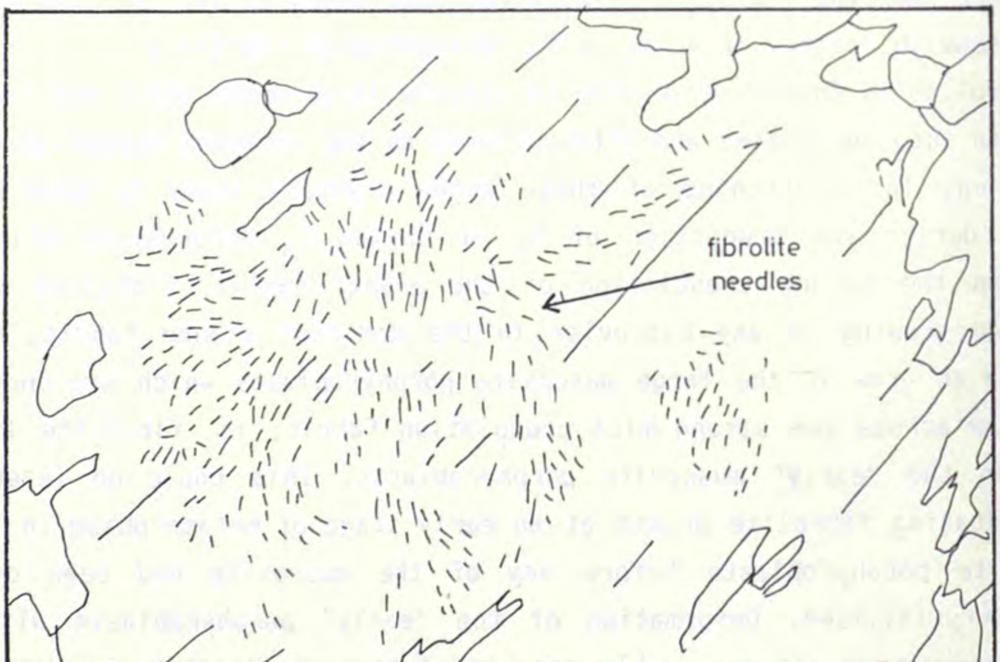
Figure 67. Photomicrographs to show fibrolite "needles" in "early" muscovite porphyroblasts.

(a)

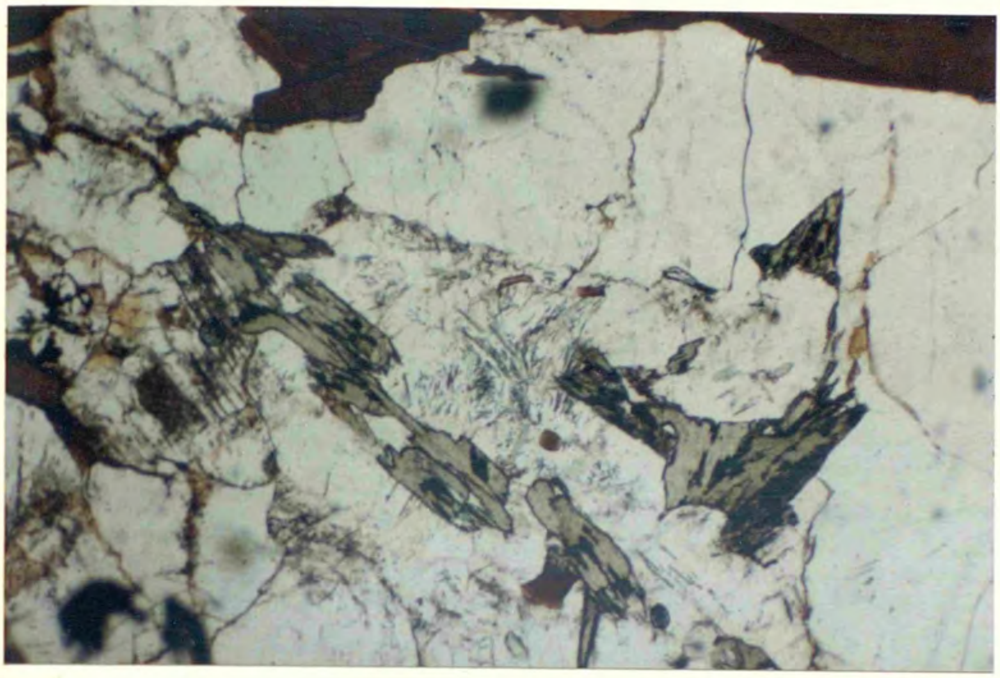


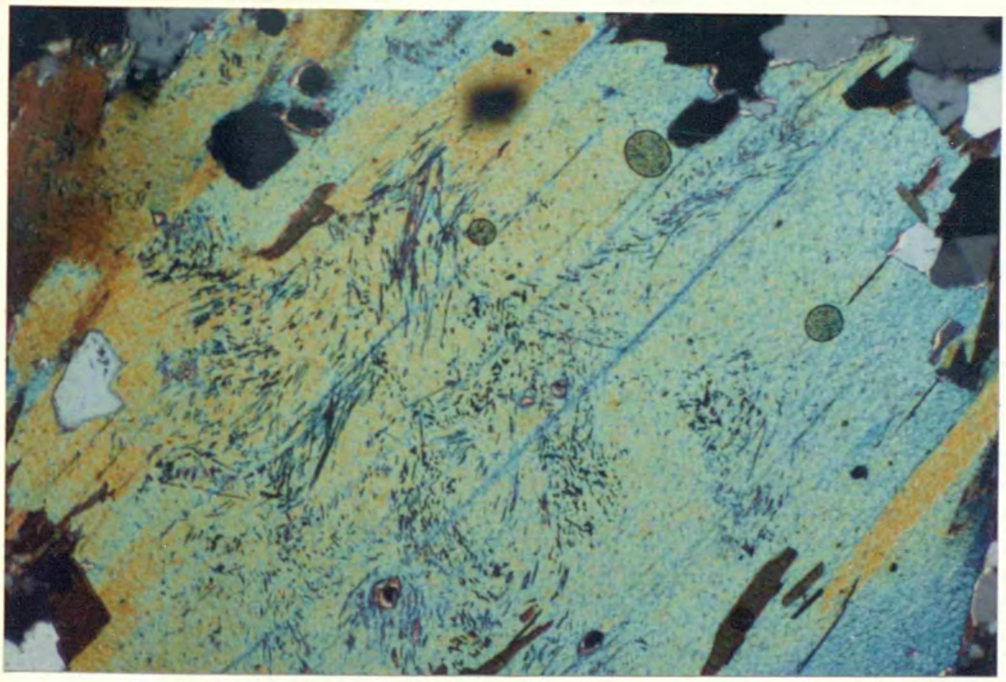
(Exp.246/1104). plane polars. 0 mm. 1

(b)



(Exp.199/1045). crossed polars. 0 mm. 1





fibrolite nucleation. The former possibility is the one favoured by the author.

The sporadic distribution of fibrolite throughout the Morar and Glenfinnan Division rocks of the area suggests that sillimanite grade (high grade) metamorphism was widespread.

Winchester (1974b) ascribes the lack of sillimanite in high grade pelitic rocks to unfavourable whole rock chemistry, however the absence of fibrolite could be due to the absence of a suitable host mineral ("early" muscovite porphyroblasts) which in turn could reflect variations in the intensity of early (?D_2) deformation, the latter being responsible for the destruction of the "early" porphyroblasts.

Dalziel & Brown (1965) describe sillimanite occurrences in the Moine rocks of Ardgour and Moidart which are "closely associated with biotite (p.306)", they also note that "fibrous felts of sillimanite overgrow biotite laths, frequently so strongly that the biotite almost disappears".

6.3c) Garnet.

Garnet crystals occur in small quantities in most of the pelitic and many of the semi-pelitic rocks of the area. They are mostly small (1-2 mm.) and fairly equidimensional. Well formed crystal faces are not found. Occasionally larger irregular shaped crystals (up to 6 mm.) occur. All the garnets contain small round inclusions of quartz which are usually most dense in the middle of the crystals. Neither inclusion trails nor any link between the inclusion and matrix fabric were observed. Well developed internal zonal patterns of inclusions are absent. Figures 27 & 33 show some of the variety of garnet shapes and textures.

By themselves these features do not shed much light on the relationships between the structural and metamorphic histories of the area, but when considered in conjunction with work further west (eg.MacQueen & Powell,1977; Powell & MacQueen,1976; Anderson & Olympio,1977 and Olympio & Anderson,1978) they may be more informative. In the west, from the Sleat of Skye eastwards through the rocks of the Morar Division to the Sgurr Beag slide, garnets which are frequently subhedral to euhedral, overgrow bedding and an early (S_1) low grade metamorphic fabric. They have grown synchronously with the local second phase of deformation (D_2) and have, whilst growing, keyed F_2 microfolds (MacQueen & Powell,1977). Moving east through the Morar Division the garnets have a pre-tectonic core with a syn-tectonic margin implying either or both garnet growth and F_2 folding was slightly diachronous across the Morar Division. Powell & MacQueen (1976) have shown that the garnet crystals in the west of the area (on the Sleat of Skye) have acted as passive rigid porphyroblasts which were bodily rotated during local F_3 folding. Here the garnet inclusion trails are continuous out into the modified bedding fabric (S_0/S_1) in the matrix. The fabric is disrupted by the S_3 crenulation.

If one attempts to extrapolate further east, across the Sgurr Beag slide into the Glenfinnan Division then one might expect to find garnets which are completely or almost completely pre-tectonic in relation to F_2. As post-D_2 deformation appears to be more intense in the Glenfinnan Division than further west, one might expect the pre- to syn-F_2 garnets to occur as rigid porphyroblasts within a strongly developed S_3 schistosity. Such intense D_3 deformation would be expected to disrupt and partly transpose earlier foliations so that the internal inclusion fabrics would not be traceable into the surrounding matrix.

To an extent the observed features fit the extrapolation: there is no connection between the inclusion fabric and the surrounding matrix fabric and the garnets have acted as rigid bodies during the deformation which produced the S_3 crenulation schistosity. However the garnets do not contain good inclusion trails;neither totally pre-tectonic straight trails nor partly syn-tectonic curved inclusion trails are seen. The garnet shapes are not euhedral and there is no obvious internal zoning. The garnets appear to have recrystallised after their postulated pre- to syn-F_2 growth. This could imply that on a regional scale there is an overall increase eastwards in metamorphic grade of a post-F_2 metamorphic event.

Garnets in the few pelitic samples collected from the Morar Division at the eastern end of Loch Eilt are texturally similar to the garnets in the nearby rocks of the Glenfinnan Division and dissimilar to those described by MacQueen & Powell (1977) from the Morar Division further west. This would imply that the change in garnet "morphology" is not due to the juxtaposition of the two assemblages of rocks characterised by different types of garnet across the Sgurr Beag slide but is more likely to be the result of modification of originally similar garnets by differing grades of post-F_2 metamorphism.

Anderson & Olympio (1977) reporting chemically homogeneous garnets in the Lochailort-Glenfinnan area, unlike the chemically zoned ones further west, dismissed the possibility that the homogeneous garnets had grown after the zoned ones and they offered two possible explanations of their observations. Either the eastern garnets grew at higher metamorphic grade (during D_2) enabling internal homogenisation to occur during growth, or the garnets homogenised during a later metamorphic event.

6.3d) Amphibolites.

The amphibolites of the area can be divided into a group of coarse garnetiferous amphibolites (where sometimes the garnet is partly or totally retrogressed) and a group of fine grained hornblende schists which are non-garnetiferous (see Chapter 8 and Smith,1979,p.694). The garnetiferous amphibolites, which chiefly occur as isolated boudin pods, are lineated and contain a schistosity co-planar with the penetrative planar fabric in the surrounding metasediments. Within the Ardgour granitic gneiss (exp.235, NM 91778020) a thin tightly folded sheet was observed. The dominant gneissose fabric (S_2) of the Ardgour granitic gneiss is co-planar with the fold axial plane (F_2) of the amphibolite, whilst the amphibolite contains an internal fabric (S_1) which is folded around the tight (F_2) fold suggesting that the garnetiferous amphibolites were intruded (or deposited ?) prior to D_1.

The hornblende schists are found as isolated boudin pods or thin sheets which, in the Ardgour granitic gneiss, are discordant to the dominant S_2 gneissose fabric. It has proved impossible to establish the time of intrusion (or ? deposition) of the hornblende schists relative to the earliest (D_1) phase of deformation in the area.

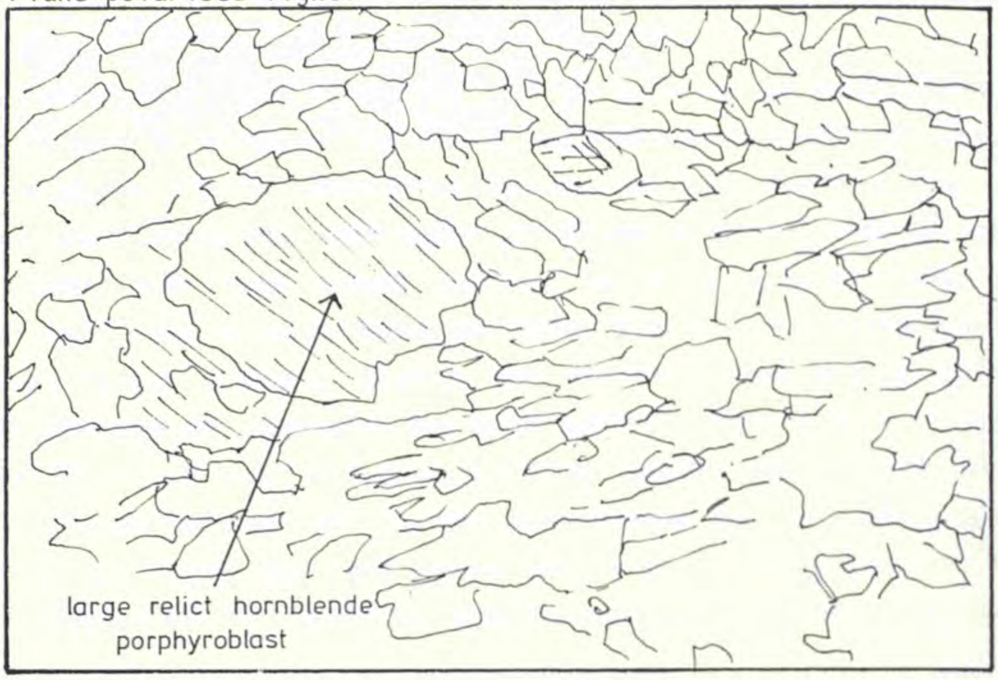
Whilst Johnstone et al.(1969) could find no evidence for intrusion and suggested a sedimentary origin, Winchester (1976) on geochemical grounds favoured a tholeiitic intrusive origin for the amphibolites. If, as was favoured in Chapter 6.2c, the precursor to the Ardgour granitic gneiss was a granitic intrusive body then both the garnetiferous amphibolites and hornblende schists must have been igneous intrusions.

Hornblende schists occur in the Loch Eil and Glenfinnan Division rocks, that is they occur across the loch Quoich line and their textural development and modification can be related to the later stages of the deformation history of the area. In the Loch Eil Division the hornblende schists frequently contain some relatively large porphyroblasts of hornblende in a finer hornblende matrix (Fig.68a & b). The porphyroblasts contain many small quartz inclusions parallel to the amphibole cleavage. Usually the porphyroblasts are mantled by sub-grains and show evidence of internal straining and sub-grain production.

Samples collected west of the Loch Quioch line are nearly all uniformly

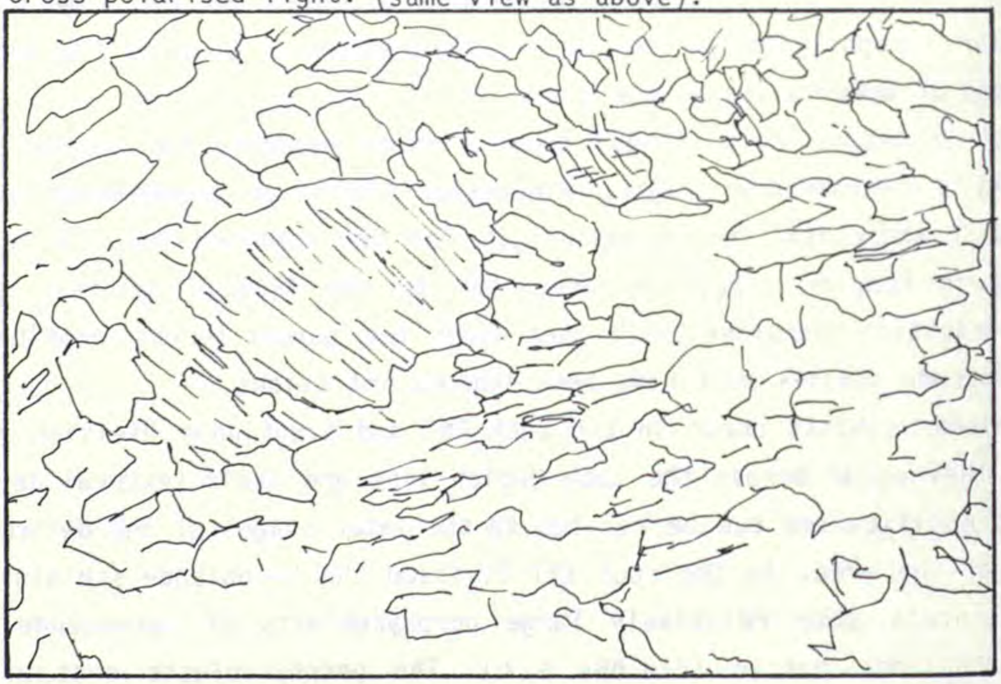
Figure 68. Photomicrographs of Loch Eil Division porphyroblastic hornblende schist. (see text for discussion).

Plane polarised light.

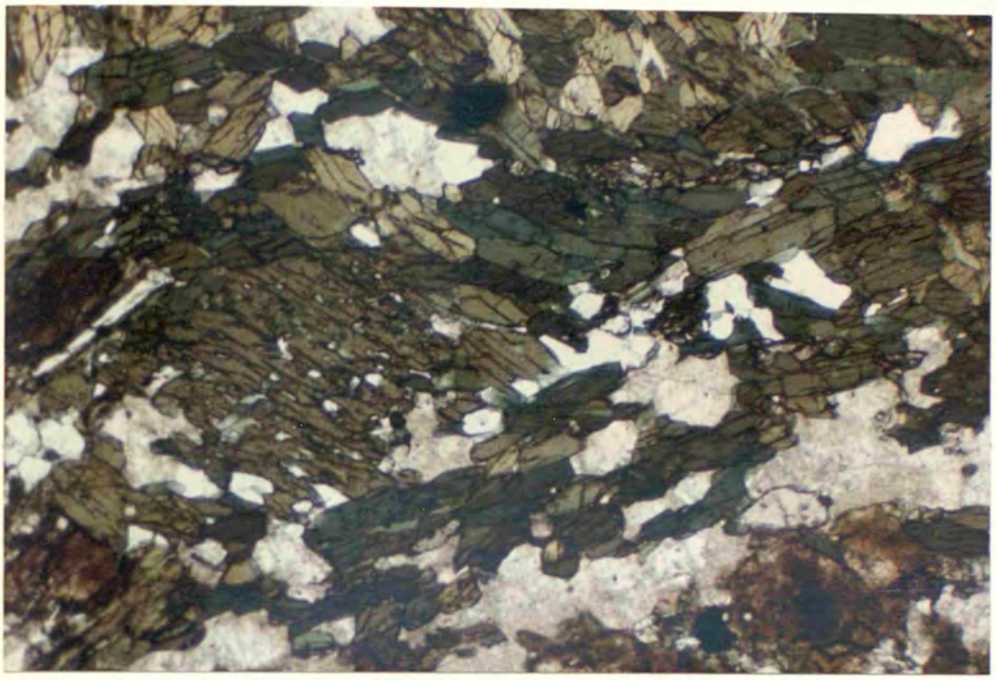


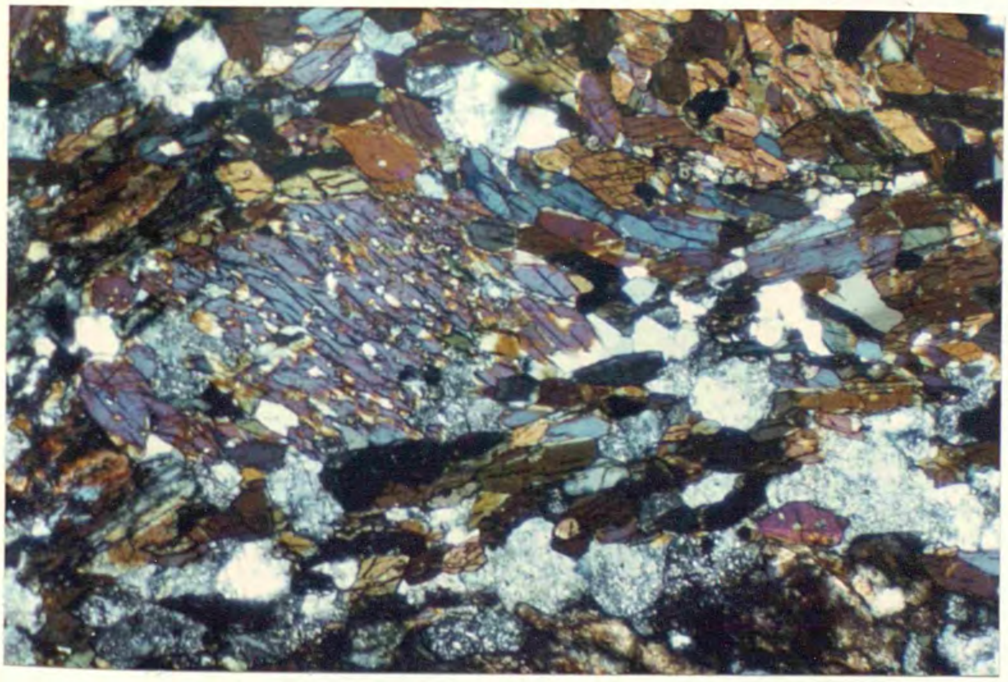
(Exp.88/281). plane polars. 0 mm. 1

Cross polarised light. (same view as above).



(Exp.88/281). crossed polars. 0 mm. 1





fine grained, containing very few, if any, hornblende porphyroblasts. The F_2 Beinn an Tuim synform, the axial plane of which crosses the Loch Quoich line, shows a considerable variation in intensity of D_2 deformation along its trace. There is no relationship between the D_2 deformation state and the presence and/or deformation of hornblende porphyroblasts, suggesting that the Hornblende porphyroblasts may have grown as syn- or post-D_2 crystals. Within the Loch Eil Division near the F_3 Druim Beag synform the porphyroblasts are very heavily deformed and recrystallised, suggesting pre-D_3 growth and D_3 deformation of the porphyroblasts. The porphyroblasts may also deform and recrystallise as a result of D_4 deformation, and similarly during D_5.

Consequently the presence of relict large hornblende porphyroblasts may be indicative of areas of relatively low D_3 (and later) deformation. This postulate is strengthened by consideration of the regional variations during D_3 and D_4. In the Loch Eil Division D_3 strain is at its lowest levels and increases progressively westwards towards the Sgurr Beag slide. The level of D_4 strain in the Loch Eil Division is also relatively low, it too increases westwards where it is responsible for the re-orientation and folding of the sub-recumbent F_3 major folds and D_3 Sgurr Beag slide to produce the D_4 "steep belt" of the Glenfinnan Division (see Chapters 7.5 & 9.1).

6.3e) Calc-silicates.

Kennedy's map of calc-silicate zones (Fig.59) shows a series of N-S trending metamorphic zones corresponding to a progressive easterly increase in metamorphic grade. In general zone boundaries conform to the regional strike of the rocks but the zones are not symmetrically disposed about major folds. Kennedy (1949) envisaged that the zones owe their development to the heat transmitted outwards from a thermal focus — a heat source which he also saw as being responsible for the production of the regional migmatites. Since the metamorphic zones are not related to major folds, one has to assume that the regional metamorphism occurred after all the major phases of folding. A problem arises when one recalls that Kennedy believed that this metamorphism produced the regional migmatites. These migmatites are,however, deformed by the D_3 deformation associated with the formation of the Sgurr Beag slide (see Chapter 7) which itself has been folded on a regional scale.

Subsequent work has shown that the mineral parageneses in calc-silicates vary with calc-silicate whole rock chemistry as well as with metamorphic grade (Winchester,1974a: Charnley,1976) so that mineral parageneses alone cannot be used to define metamorphic grade at any one locality.

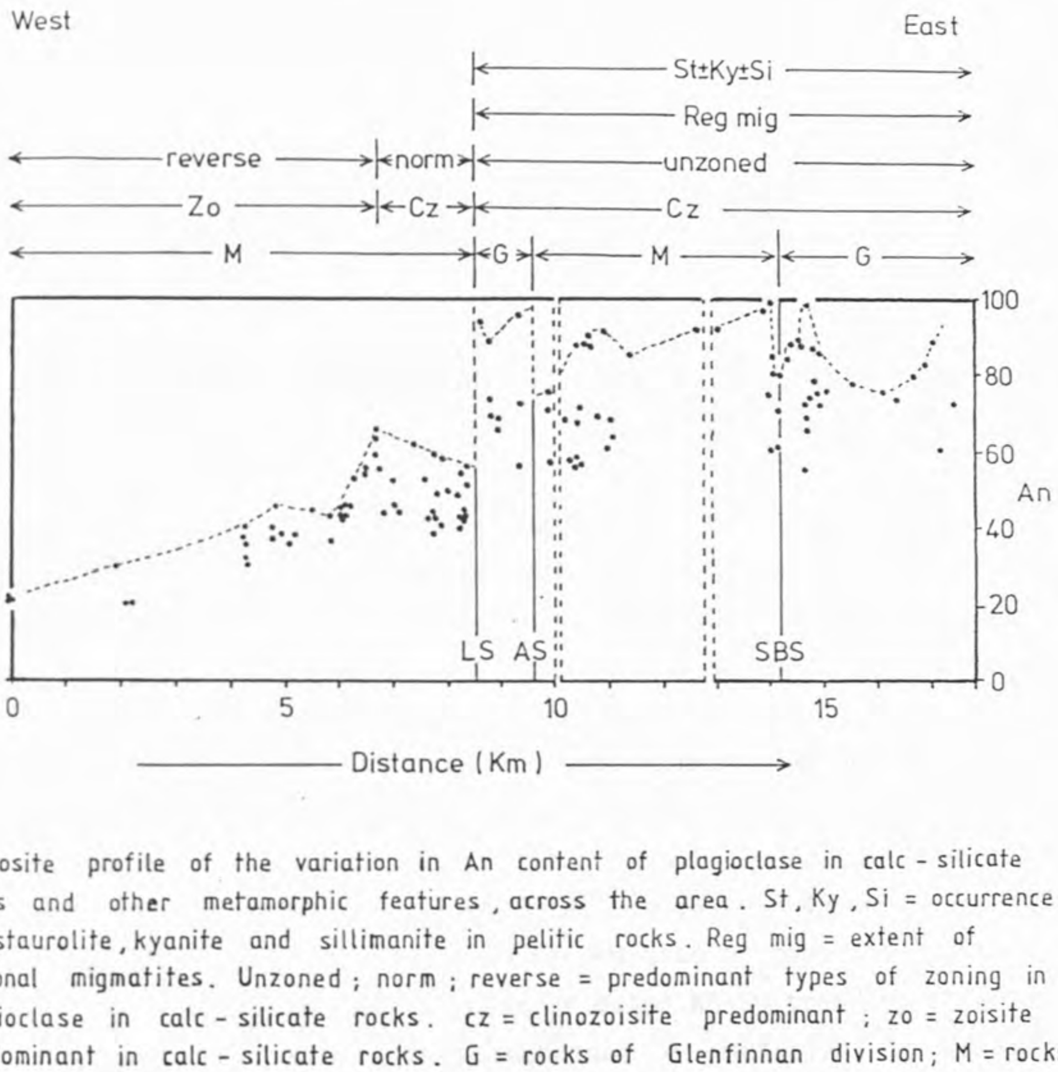
Kennedy (1949) noted that eastwards across his metamorphic zones there was an increase in the An. content of plagioclase in the calc-silicates, the transition is abrupt with a large gap between albite and bytownite. Subsequently, detailed work by Tanner (1976) reduced this gap to An.55-70% and more recent work (Powell et al.1981.text fig.3(see Appendix 5)) has shown that the An. content of plagioclase is independent of whole rock chemistry (as expressed by the ratio CaO/Al_2O_3) but apparently related to metamorphic grade.

Powell et al.(1981) noted that the variation in plagioclase composition across the area is not random; there is a general, and in places gradual, increase in An. content of the plagioclase from W to E ie. with overall increase in metamorphic grade (Fig.69, taken from Powell et al.1981). Lowest values of An. content come from the calc-silicates at western localities which have highly variable whole rock chemistry, as expressed by the ratio CaO/Al_2O_3. Conversely highest values of An. content come from the eastern localities where nearby pelitic rocks contain staurolite and/or kyanite, and/or fibrolite. Again the calc-silicates show highly variable whole rock chemistry.

As a consequence of the apparent relationship between An. content of plagioclase in calc-silicats and metamorphic grade it has been possible to draw a profile across much of the SW Northern Highlands showing variations of An. content in calc-silicates (Fig.69) and to interpret it as showing variations of metamorphic grade. From this diagram it can be seen that, as Kennedy (1949) observed, metamorphic grade increases, for the most part, gradually eastwards. However the diagram also shows that this gradual increase is interrupted in places which, from structural considerations, have been held to be the outcrop of the Sgurr Beag slide repeated by folding along the line of section.

To the east of the Sgurr Beag slide the "isograd" pattern shows a distinct trough (see Fig.69) which corresponds to the position of the F_3 Sgurr a Mhuidhe synform, suggesting that the "isograds" are folded by F_3 folds. On this evidence alone the "isograd" pattern is pre-F_3; confirmation is provided by the disruption of the metamorphic pattern across the Sgurr Beag slide which itself is a D_3 structure (see Chapter 7). It can further be argued that within the slide zone, where strain exceeds a critical value (Powell et al.1981 text figs. 2 & 10) there is a downgrading of the pre-existing metamorphic grade during D_3.

Figure 69.



Composite profile of the variation in An content of plagioclase in calc - silicate rocks and other metamorphic features, across the area. St, Ky, Si = occurrence of staurolite, kyanite and sillimanite in pelitic rocks. Reg mig = extent of regional migmatites. Unzoned ; norm ; reverse = predominant types of zoning in plagioclase in calc - silicate rocks. cz = clinozoisite predominant ; zo = zoisite predominant in calc - silicate rocks. G = rocks of Glenfinnan division; M = rocks of Morar division. SBS = Sgurr Beag Slide ; AS = Arieniskill Slide ; LS = Lochailort Slide.

(Taken from Powell et al, 1981)

Comparisons of figures 69 and 70 (taken from Baird,1982. text fig.3) leads to the suggestion that on the E-W line of section at the eastern end of Figure 69, the "isograd" surfaces are broadly coincident with the composite bedding/schistosity fabric, both of which are folded by the F_3 Sgurr a Mhuidhe synform.

At the eastern end of Loch Eilt an ill-fated attempt was made to establish the three dimensional geometry of the "isograd" surface, unfortunately calc-silicate samples giving a sufficient N-S spread of data points could not be found.

While it appears that the "isograd" and composite bedding/schistosity surfaces produce coincident traces on the vertical E-W section, the two surfaces need not be co-planar. The composite bedding/schistosity surface is folded to produce the F_3 Sgurr a Mhuidhe synform with its NE-SW trending vertical axial plane and steep NE plunge. D_3 deformation of the "isograd" surface would produce a synformal fold, the plunge of which would have to be somewhere within a vertical NE-SW trending plane, but only co-linear with the plunge of the F_3 Sgurr a Mhuidhe synform if the two surfaces were co-planar prior to D_3.

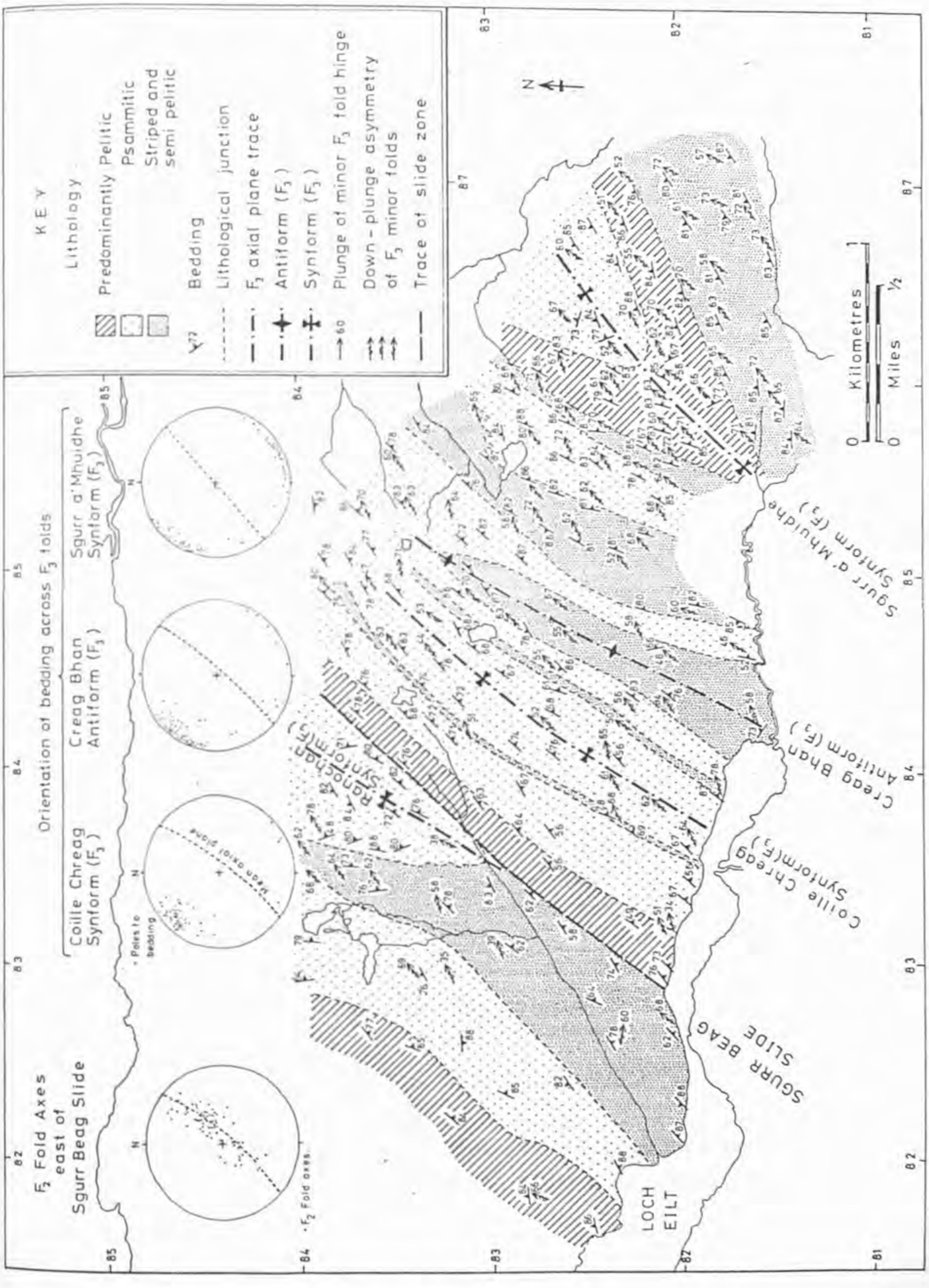
The possibility that the folded "isograd" surface plunges in a NE or SW direction may have significance in explaining the "band" of An, contents seen in the E-W line of section in Figure 69. Not only does the maximum An. content of plagioclase in calc-silicates vary progressively along the line of section, but the minimum An. content also seems to reflect this progressive variation, producing a band of An. contents at any given distance along the line of section. The "band" may be due partly to the method of construction of the diagram. Calc-silicates were collected from a thin strip of land approximately 20 Km. long (E-W) and up to 5 Km. wide (N-S). The position of the localities were transferred onto a composite E-W section by measuring their distance east or west of major NE-SW trending structures such as the Sgurr Beag slide. This technique is equivalent to combining a series of sections which are progressively further north (or south) of a reference section line. With the possibility that the folded isograd surfaces plunge to the NE then the more northerly sections would expose lower metamorphic grade rocks at higher structural levels (see Fig.71).

If this geometrical construction is solely resonsible for the "band" of An. contents seen in Figure 69 then it is possible, using very simplistic arguments, to obtain some idea of the plunge of the folded isograd surfaces.

The difference between the maximum and minimum values of An. content at any distance along Figure 69 is approximately 25% An. If it were possible to calibrate the variation of anorthite content of plagioclase in calc-

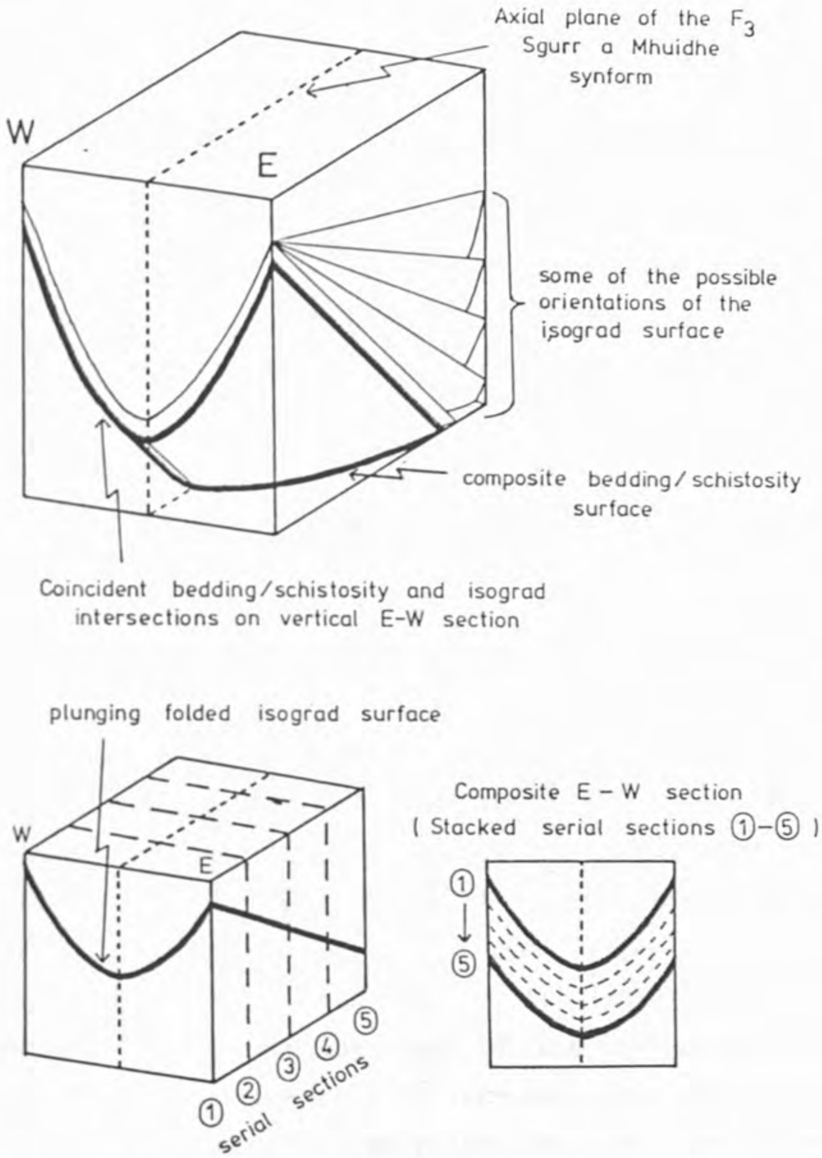
Figure 70.

Detailed map of the Moine rocks at eastern Loch Eilt.



(taken from Baird,1982).

Figure 71.



See text for discussion.

Diagrammatic sketches to show possible relationships between the metamorphic isograds and the F_3 Sgurr a Mhuidhe synform.

silicates versus temperature and depth of burial during metamorphism, then it might be possible to equate, for example, an increase of 25% An. content with an increase in depth of burial of 4-5 Km. This hypothetical 4-5 Km. increase in depth of burial would be the result of the fact that the isograd surfaces dip. If the N-S width of the area is taken as 5 Km., then the width, measured in the "plunge" direction (NE-SW) is approximately 7 Km. ($5 \times \sqrt{2}$ Km.). Therefore, theoretically, the hinge line of the folded isograd surface drops 4-5 Km. vertically over a horizontal distance of approximately 7 Km., indicating a plunge in the order of 30°.

If one assumes that the isograd surface is approximately co-planar with the composite bedding/schistosity fabric as well as being co-linear on the E-W line of section then one can go further and postulate that if the isograd surfaces formed as relatively flat lying surfaces or as a gentle dome around a focus of metamorphism, then the pre-D_3 geometry of the composite bedding/schistosity fabric is similarly flat lying or gently domed prior to D_3. Consequently any major pre-D_3 tight to isoclinal folds would have been sub-recumbent.

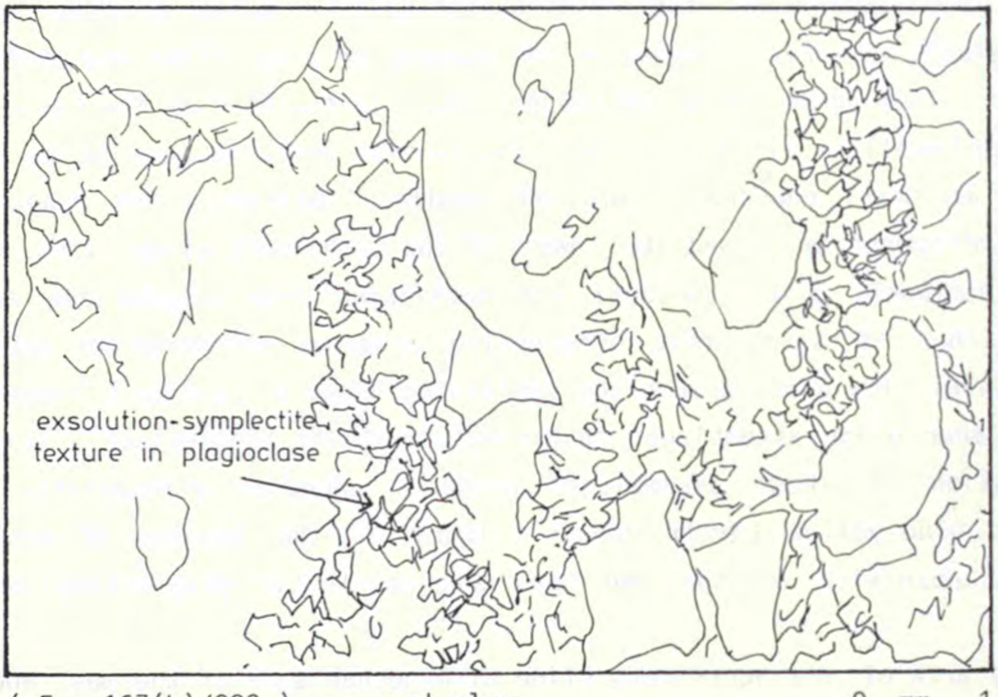
Calc-silicates occur so infrequently east of the axial plane trace of the Sgurr a Mhuidhe synform, that changes in mineralogy, texture and An. content of plagioclase crystals are difficult to relate with any certainty to the structural and metamorphic history of the area. The plagioclase crystals are usually highly or totally saussuritised. Where the plagioclase feldspars are not totally saussuritised the An. contents are lower than those further west, dropping from An.70-90% in the vicinity of the Sgurr Beag slide at Ranochan to a range of An.30-70% (Map 6). If the An. content is a reflection of the metamorphic grade during a pre-D_3 metamorphic event then it would indicate that the grade has dropped fairly considerably moving eastwards from the slide. However some of the calc-silicates collected from east of Glenfinnan village contain very well developed pyroxene crystals, indicative of high grade metamorphism regardless of whole rock chemistry (Winchester,1974: Charnley,1976). It is thus possible that the eastern area (ie, the eastern Glenfinnan Division and Loch Eil Division) has undergone an early high grade metamorphic event producing highly anorthitic feldspars togather with well developed pyroxenes, followed by intense retrogression or later lower grade metamorphism, the latter being responsible for the intense saussuritisation and recrystallisation of less anorthitic plagioclase. Possible support for this speculation is given by Johnstone et al.(1969) who drew a line representing an easterly limit of bytownite in calc-silicates (shown on Map 6) which seems to be unrelated to any of the structural trends in the area and may represent a western limit of intense retrogression.

Calc-silicate mineral textures visible in thin section are very difficult to interpret because of the complex relationship between the metamorphic and structural histories of the area and the geochemical variation which occurs between samples. Powell et al.(1981,p.663) have defined the main mineral assemblages occuring in the area under consideration and westwards to the coast. A detailed study of the metamorphic petrology of the calc-silicates requires study of samples from beyond the area so that the variations of metamorphic grade and whole rock chemistry can be taken into consideration. Such a study is beyond the scope of this work but a number of comments can be made which are relevant both to work on this area and in the region as a whole.

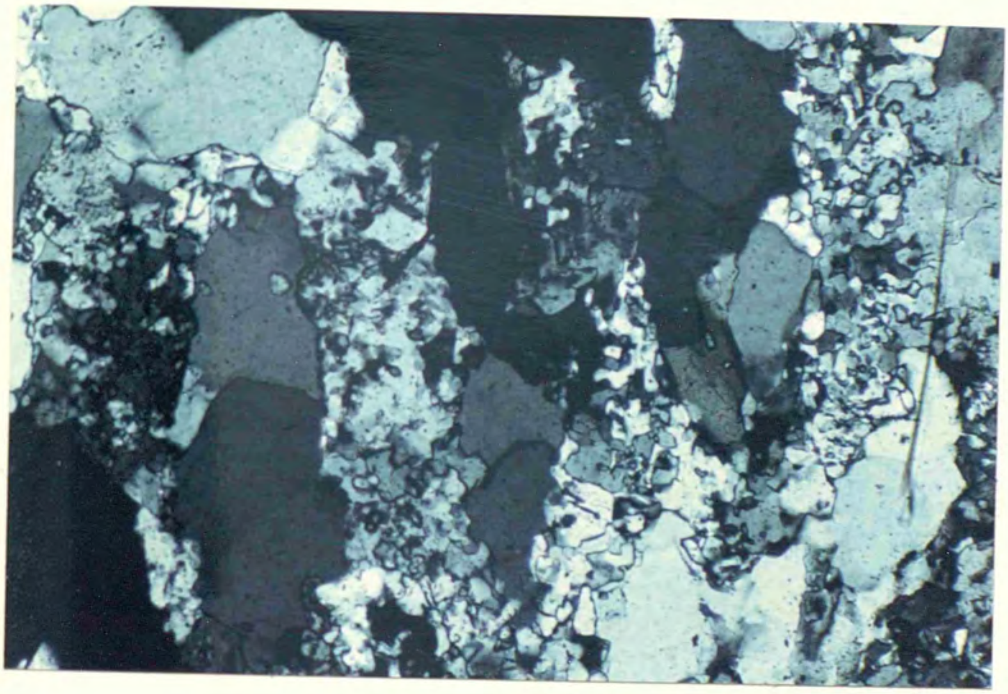
Zoisite has not been found in the area, all of the epidote group minerals present are members of the clinozoisite-epidote solid solution series and are always anhedral, growing at the expense of plagioclase and sometimes amphibole (? retrogressive metamorphism). Well to the west of the Lochailort slide (see Powell et al.1981) co-existing zoisite and clinozoisite are found , occuring as small euhedral laths and needles, apparently the product of prograde metamorphism. Immediately west of the Lochailort slide clinozoisite becomes dominant, often anhedral and developing from plagioclase and/or zoisite. Thus it seems that progressive prograde metamorphism converts the relatively low grade assemblage containing euhedral zoisite-clinozoisite into a higher grade assemblage containing anorthitic plagioclase + zoisite. Retrogression of these assemblages produces anhedral clinozoisite with anomalous blue-yellow birefringence colours at the expense of prograde zoisite-clinozoisite on the one hand and anorthitic plagioclase on the other.

In the area of the Sgurr Beag slide at Ranochan some of the most anorthitic plagioclase crystals contain small exsolution blobs of lower An. plagioclase (Fig.72). There is however, no geographical link between the calc-silicates near the Sgurr Beag slide which contain "exsolved" anorthitic plagioclase and the more widely distributed calc-silicates in which the plagioclase is highly saussuritised. The restriction of "exsolution" texture to calc-silicates with highly anorthitic plagioclase collected near to the Sgurr Beag slide suggests two things, firstly that the "exsolution" reaction is activated by high D_3 strain, or high D_3 strain rate and secondly that "exsolution" is a re-equilibrium reaction, retrogressing highly anorthitic plagioclase to less anorthitic plagioclase during D_3. Plagioclase which, prior to D_3, was less anorthitic may have remained stable or totally recrystallised without the production of exsolution texture during D_3.

Figure 72. Photomicrograph to show plagioclase "exsolution-symplectite" texture in calc-silicate.



(Exp.167(b)/883). crossed polars.



In several sections there are two generations of amphibole. Large hornblende crystals predominate but they are sometimes replaced by small acicular hornblendic amphiboles; again this is presumed to be a result of retrogressive metamorphism (c.f. Winchester,1972).

6.3f) The Ardgour granitic gneiss.

The Ardgour granitic gneiss contains early pegmatites of granitic composition which grade downwards in size into migmatitic segregations. Both the pegmatites and the migmatitic segregations have selvages of biotite crystals which are co-planar to their margins. The pegmatites are frequently very tightly folded so that they lie very close to the dominant gneissose schistosity (S_2) seen in the gneiss. Close observation of the biotite selvages in the hinges of F_2 folds reveals that they are frequently very tightly crenulated so that the axial planes of the crenulations are co-planar with the dominant S_2 foliation, an indication that the formation of the pegmatites and their associated biotite selvages occurred prior to D_2 deformation. The F_2 folds which fold the pegmatites also fold a gneissic (S_1) foliation,but more usually the early (S_1) fabric is almost totally transposed. Theoretical considerations of the mineral assemblages indicate that the apparent origin of the migmatitic segregations and "early" pegmatites was by partial anatexis, however the age of partial anatexis in relation to D_1 deformation is difficult to ascertain.

A pre-D_1 age of partial anatexis is possible, assuming that D_1 deformation rotated and deformed the segregations to produce the S_1 gneissose fabric. Under these circumstances one might expect to observe deformation and perhaps crenulation of the biotite selvage to produce the S_1 fabric, but given the intensity of re-working of the gneiss during D_2, this may not be observed.

A syn-D_1 age of anatexis is possible, under these circumstances segregation formation and growth of biotite crystals would have been co-planar with the S_1 fabric.

A post-D_1 age is possible only if the segregations migrated into pre-existing S_1 planes. However in a static post-D_1 environment it is difficult to envisage the growth of biotite crystals in the selvages being co-planar to the pre-existing S_1 gneissose fabric unless the biotite crystals nucleated epitaxially on some structural element within the pre-existing S_1 fabric.

It is important to consider whether during D_2 deformation the S_2 gneissose fabric was produced by partial anatexis or by a very intense re-working or transposition of the earlier S_1 gneissose fabric. It has already been noted that D_2 deformation has deformed the early pegmatites so that they now occur in two modes, firstly as elongate, isolated boudin pods which are nearly co-planar with the dominant S_2 gneissose fabric and secondly as isolated tight to isoclinal "fold hooks" which are intrafolial to the S_2 fabric. In the hinges of the "fold hooks" the biotite selvage fabric is tightly crenulated by D_2. Occasionally around the "fold hooks" remnants of the S_1 fabric are seen which are co-planar with the early pegmatites and folded by D_2. Away from the vicinity of the pegmatites the S_1 fabric in the gneiss is very quickly transposed by the dominant S_2 fabric.

The evidence outlined above is consistent with a suite of pre-D_2 pegmatites, co-planar with the S_1 gneissose fabric, which have been deformed during D_2. Some of the pegmatites have been folded to produce isolated F_2 fold hooks, while others have been rotated towards the plane of the S_2 gneissose fabric.

There is no field evidence, such as examples of folded pre-D_2 pegmatites cross cut by later pegmatites which are axial planar to the dominant S_2 foliation, which would conclusively prove the occurrence of partial anatexis during D_2. Although the widespread intense transposition of the S_1 fabric to produce the S_2 fabric suggests high grade metamorphic conditions during D_2.

The two high grade metamorphic events outlined above were followed by the intrusion of sporadic small dilational aplite veins which presumably indicate a period of brittle fracture perhaps suggesting their intrusion into relatively cooler host rocks.

The sporadic aplite veins were followed by a suite of large cross-cutting coarse grained pegmatites (= "late pegmatites") which are sometimes folded by F_3 folds and sometimes appear to post-date F_3 folds. The phase of pegmatite intrusion may have commenced before D_3 and outlasted it. However the irregular and random orientation of the pegmatites would suggest that they were not intruded during a period of compressional deformation and folding. In view of the difficulties in positively distinguishing between F_3 and F_4 folds, the intrusive age of the pegmatites in relation to D_3 and D_4 deformation is even more equivocal.

Where the late pegmatites are folded, large books of mica crystals have a pseudo-crenulated appearance because individual, previously randomly orientated books of mica have been rotated towards the XY plane of the D_3 (or D_4) strain ellipsoid. Presumably such ductile deformation must have occured under conditions of medium or high grade metamorphism during D_3 (or D_4).

6.4) Summary of conclusions.

The mineralogy of the pelitic rocks of the area is unfavourable for the delineation of the classical "Barrow zones" of metamorphic grade. Sillimanite, kyanite and staurolite occur very sporadically, possibly reflecting unfavourable whole rock chemistry since their occasional occurrence together with the widespread occurrence of migmatitic segregations of quartz and feldspar indicate the occurrence of medium to high grade metamorphism. The absence of K-feldspar from the migmatitic segregations suggests that their formation was not as a result of partial anatexis.

The Ardgour granitic gneiss; a K-feldspar + Plagioclase + Quartz + Biotite gneiss contains migmatitic segregations of K-feldspar + Plagioclase + Quartz which are the products of partial anatexis. The precursor to the Ardgour granitic gneiss could not have been the adjacent Moine sediments since partial anatexis of them would have produced large quantities of sillimanite and/or garnet or cordierite. The precursor must have been highly potassic, probably a granite. The "early" pegmatites and migmatitic segregations in the Ardgour granitic gneiss are a product of partial anatexis prior to D_2, and most probably synchronous with D_1 deformation. Intense D_2 re-working has produced a second, dominant gneissose foliation. There is no unequivocal evidence that this re-working during D_2 was accompanied by any partial anatexis.

The anorthite content of plagioclase in calc-silicates is used to establish variations of metamorphic grade across the area. The pattern of metamorphism is seen to be folded by major F_3 folds and disrupted by the D_3 Sgurr Beag slide. The calc-silicate "isograd" surfaces, which are probably co-planar with the composite bedding/schistosity surfaces, were probably flat lying prior to D_3 deformation, consequently any major pre-F_3 folds must have been sub-recumbent. There is evidence that at the highest levels of D_3 strain within the Sgurr Beag slide zone there was partial re-equilibration and downgrading of anorthite in calc-silicate plagioclase crystals. A widespread drop in anorthite content in calc-silicate plagioclase crystals in the east of the area, in rocks which frequently retain high grade pyroxene crystals may evidence an eastern zone of late retrogression or a zone of later lower grade metamorphism during which anorthitic plagioclase was able to equilibrate to more albitic compositions.

Garnet crystals in pelitic rocks behaved as rigid bodies during D_3 deformation. The lack of D_2 inclusion patterns similar to those found in garnets further west may indicate that during garnet growth inclusion fabrics were not formed or alternatively D_2 inclusion fabrics have been destroyed as a result of late- or post-D_2 recrystallisation.

Biotite and muscovite laths form a planar penetrative mica fabric which is folded by F_3 and later phases of folding. In rare exposures where two phases of isoclinal folding prior to F_3 folding can be seen, both fold phases contain an axial planar penetrative mica fabric.

Minor folds of F_3 and F_5 age crenulate the planar mica fabric, but individual mica crystals are unstrained indicating that recrystallisation has occured after F_5 folding and probably also after F_3 folding. Minor folds related to the major F_4 folds were not observed.

Muscovite crystals have two other less frequent modes of occurrence. Early porphyroblasts showing no preferred crystallographic orientation are marginally deformed and wrapped by the planar penetrative mica fabric. Late porphyroblasts without a preferred orientation or shape orientation cross cut the penetrative planar mica fabric.

Sillimanite, in the form of fibrolite needles, is found as irregularly curved trails growing through the early muscovite porphyroblasts, but not in any other muscovite crystals, leading to the conclusion that fibrolite growth occurred after the growth of early muscovite porphyroblasts but before the development of the planar penetrative mica fabric.

CHAPTER 7.

The Sgurr Beag slide at eastern Loch Eilt.

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7.4) Metamorphic features.
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- 7.4d) Status of the "early" metamorphic events.

7.5) The relationship of the Sgurr Beag slide to the "steep belt".

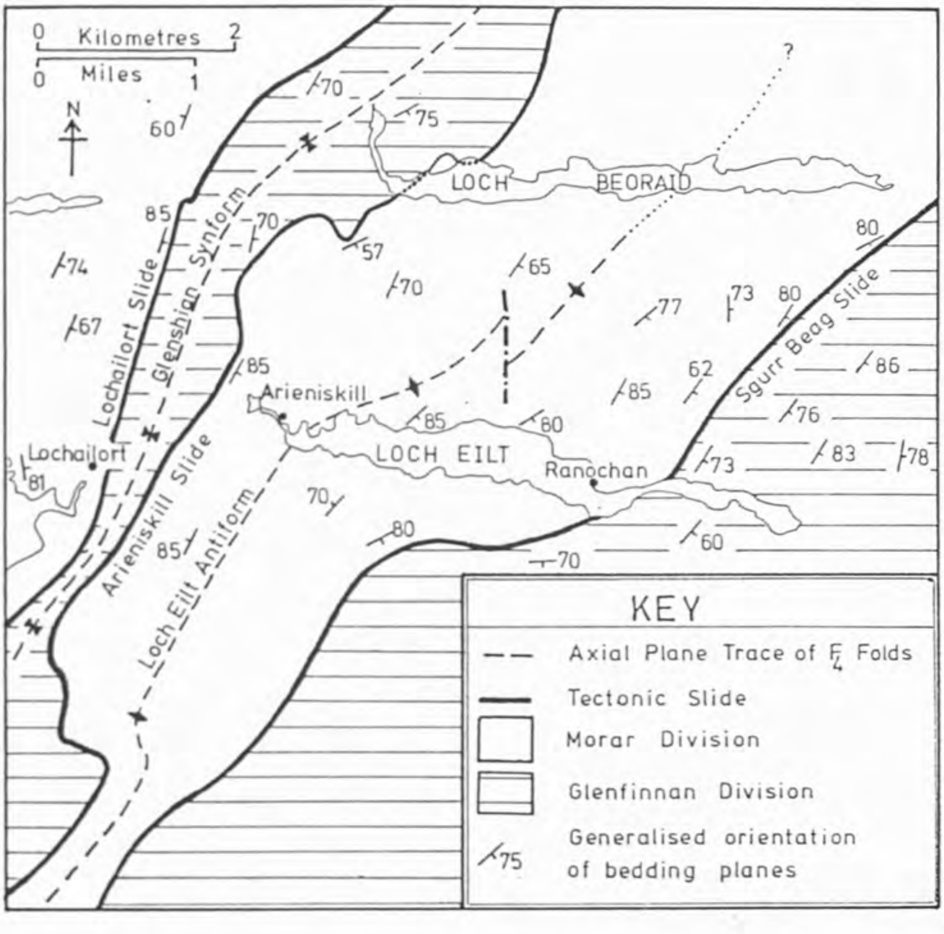
7.1) Introduction and regional significance.

A consideration of the regional litho-stratigraphy and structure indicates that the area at the eastern end of Loch Eilt has to contain the junction between the rocks of the Morar and Glenfinnan Divisions (Powell,1964, 1974: Johnstone et al.1969: Tanner et al.1970). Farther north this junction, termed the Sgurr Beag slide at Kinloch Hourn by Tanner (1971) is a major break within the Moine succession. This chapter describes the structural age and development of the slide within the local structural sequence and examines the metamorphic changes which occur across and in relation to the slide. In the later parts of this chapter the regional implications of the local structure and its relationship to the origin of the Glenfinnan Division "steep belt" are considered. Much of the detailed structure of the area has already been described by Baird (1982, included as Appendix 6).

The existence of a major tectonic break within the Moine Succession at Kinloch Hourn 30 Km. north of Loch Eilt was first noted by Tanner (1971) and termed the Sgurr Beag slide. The slide separates rocks grouped as the Morar and Glenfinnan Divisions (Johnstone et al.1969) which were thought to be in continuous stratigraphical succession (Brown et al.1970: Powell, 1964). Locally and only to the north of Kinloch Hourn, small tectonic slices of Lewisian basement rocks occur along and near the junction of the Morar and Glenfinnan Divisions. The presence of these slices at high stratigraphical levels within the Morar Succession, not only provides the evidence of the tectonic origin of the slide but also shows it to be a major feature (Tanner et al.1970: Rathbone et al.1983).

In the Lochailort-Loch Eilt area (Fig.73, taken from Baird,1982,text fig.2) the Lochailort pelite (Powell,1964) has been correlated with the Glenfinnan Division and the rocks east of the Lochailort pelite with the Morar Division (Powell,1964,1974). Rathbone & Harris (1979) indicated that there are large strain variations across the junction at Lochailort and considered it to be a southerly extension of the Sgurr Beag slide. The Morar-Glenfinnan Division junction is found on both sided of a major fold, the Glenshian synform (Fig.73) and therefore it is inferred that this fold post-dates the formation of the slide. Extrapolation of this slide to the Sgurr Beag slide at Kinloch Hourn requires that it has been folded by the Loch Eilt antiform (Powell,1974), its trace being repeated in the eastern Loch Eilt area (Fig.73) and continuing northeastwards to Kinloch Hourn (Powell et al.1981. and I.G.S. 1:63,360 & 1:50,000 Geol.Maps 61 & 62W, Scotland).

Figure 73.



Map of the Loch Eilt area to show the outcrop of the trace of the D_3 Sgurr Beag slide repeated by F_4 folding.

(taken from Baird,1982).

7.2) Local structure.

A sequence of events can be erected locally in keeping with the general tectonic sequence established in Chapter 5. Certain parts of the sequence of tectonic events however, are only weakly developed or can be surmised only from regional correlations and extrapolations. Other parts of the sequence are best developed in this locality and are used to enlarge the record of the regional deformation sequence. Recognition of the Sgurr Beag slide allows an insight into the structural history of the area in terms of the relative chronology of fold phases developed locally and permits correlations and extrapolations to adjacent areas.

7.2a) F_5 folds.

These open crenulation folds have near vertical axial planes and hinges which plunge steeply to the SE (see Fig.54). They have not been identified much further east than the eastern end of Loch Eilt. The minor folds appear to be related to major folds which have axial planes trending approximately NNW-SSE (see Fig.53). F_5 folding is responsible for the gentle change of trend of the axial planes of the major F_3 folds which occur between Glenfinnan village and Loch Eilt (see Chapter 5.6c). The major F_5 folds are the folds described by Powell (1974) as F_4.

7.2b) F_4 folds.

Minor folds of this generation were not recognised in the area, but on a regional scale the D_3 Sgurr Beag slide is folded by tight to isoclinal folds, the F_4 Glenshian synform and the F_4 Loch Eilt antiform (Powell et al. 1981, text-fig 5 and Baird,1982, text-figs.2 & 4), both of which plunge shallowly to the SW. A suite of post-D_3/pre-D_4 microdiorite sheet intrusions has undergone widespread D_4 deformation, the mechanism of deformation of these sheets is discussed in Chapter 8.1b4.

7.2c) F_3 folds.

F_3 folds, both on major and minor scales, are the dominant folds in the area, and eastwards throughout the Glenfinnan Division. Minor F_3 fold hinges generally plunge steeply towards the NE within near vertical axial planes (see Fig.56). Major F_3 folds become progressively tighter and their hinge lines become steeper westwards from Glenfinnan village to the area under consideration. These and other features described later in this chapter are taken as indicators of increasing intensity of D_3 deformation moving westwards. At the eastern end of Loch Eilt the extremely intense D_3 deformation has produced a syn- to late-D_3 tectonic slide zone which truncates both lithological units and a major F_3 synform (the Ranochan synform). The structural development of the slide zone and its relationship to the metamorphism of the area are discussed later in the chapter.

7.2d) Pre-F_3 folds.

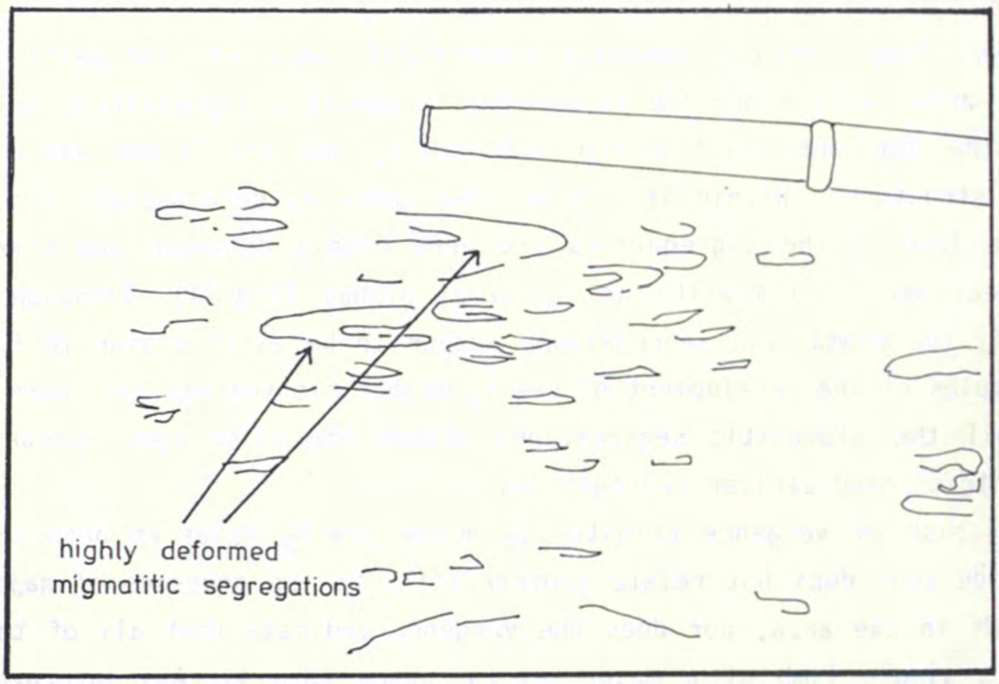
From Glenfinnan village westwards to the area under consideration major folds of pre-F_3 age have not been found. F_3 folds deform a strongly developed foliation which is nearly always modified bedding with a bed-parallel schistosity, but occasionally the schistosity is seen to be axial planar to minor, long limbed, frequently symmetrical isoclines. The pelitic rocks of the area contain quartzo-feldspathic migmatitic segregations which enhance the dominant axial planar schistosity and are folded and deformed by D_3 structures. Within the slide zone where D_3 deformation reaches its highest levels, the segregations are very highly deformed and crenulated into near parallelism with the F_3 axial planes (Fig.74). Although superficially the migmatitic segregations appear to be axial planar to F_3 folds no examples of the development of new F_3 migmatitic segregations were observed. All the migmatitic segregations within the slide zone appear to be highly transposed earlier segregations.

The sense of vergence exibited by minor pre-F_3 folds on both sides of the slide zone does not relate geometrically to the presence of major pre-F_3 folds in the area, nor does the vergence indicate that all of the area is on a single limb of a major pre-F_3 nappe (Map 4). Any interpretation of the vergence pattern of pre-F_3 minor folds is made more equivocal when it is realised that individual pre-F_3 isoclines could be either F_1 or F_2 folds.

7.3) D_3 structural development.

Figure 70 shows that the slide zone truncates lithological units. The slide zone does not show any cataclastic or mylonitic deformation textures or widespread hydrothermal, retrogressive features, either in the field or in thin section, such as might relate to a high level thrust. The slide zone pelites contain intensely crenulated pre-D_3 migmatitic segregations

Figure 74. Photograph of migmatitic pelite from the east of the Sgurr Beag slide at Ranochan.



(Exp.1259 NM 84128347).

which are so tightly crenulated as to appear superficially to be axial planar to the sporadic F_3 minor folds (see Fig.74).

Mapping towards the slide zone, either eastwards or westwards, reveals that D_3 structures become increasingly intensely developed as the slide zone is approached. Therefore the Sgurr Beag slide zone does not separate two major rock units with different structural histories; it is not a basement/cover contact (cf.Piasecki & van Breemen,1979.p.143). Since the truncation of lithological units produced when the axial plane of a major F_3 fold is cut out against the slide zone, the Sgurr Beag slide cannot be an unconformity. Thus it follows that the slide, apparent because of its mapped geometrical relationships and the intensification of D_3 strain towards it, was a result of ductile deformation which was followed by post-sliding recrystallisation which overprinted and destroyed cataclastic and/or mylonitic textures, or which proceeded under conditions of dynamic recrystallisation which prevented grain size reduction (Tanner,1971). Quantitive estimates of strain associated with the slide zone are impossible to obtain because no suitable strain markers are present (cf. Rathbone & Harris,1979).

7.3a) East of the slide.

In the area from Loch Eilt to Glenfinnan village it has been noted earlier in the chapter that the predominant folds in the area are F_3 in age. Large scale tight to isoclinal folds with NE-SW trending nearly vertical axial planes and steep NE plunges are present, togather with abundant associated minor folds (see Map 4). These D_3 structures deform tight to isoclinal minor F_1 or F_2 folds.

A progressive change in structural geometry, especially in the structures of D_3 age, can be traced westwards into the slide zone. The Sgurr a Mhuidhe synform (Fig.70) is a relatively open F_3 synform which plunges steeply to the NE. The complimentary F_3 Creag Bhan antiform to the west is tighter, and farther west the Coille Chreag synform is isoclinal. The westward tightening of the F_3 interlimb angles is indicated on the stereographic projections included in figure 70. From Glenfinnan village westwards, the plunge of major F_3 folds, derived from consideration of the plunge of the associated minor F_3 folds, progressively steepens within the near vertical fold axial planes. F_3 minor folds also tighten westwards towards the slide and the number of minor F_3 folds decreases rapidly in this direction so that the rocks become markedly planar. Moving west from the Sgurr a Mhuidhe synform (Fig.70) boudinage of psammitic beds becomes progressively more common. Boudin pods containing isoclinal F_1 or F_2 folds can be found near the slide and occasionally F_3 minor folds are boudinaged and thinned. These features are taken to indicate that boudinage occured during or after D_3 deformation.

The tightening up of F_3 fold interlimb angles, the increase in the amount of boudinage and the reduction in the number of minor F_3 folds approaching the slide can all be attributed to an increase of D_3 strain, with rotation of planar elements into the extension field of the strain ellipsoid (Flinn, 1962). The change of plunge of major F_3 fold hinges may be attributed to rotation within the XY plane of the strain ellipsoid towards the X direction (Transport direction is considered in Chapter 7.3c). Minor F_2 fold hinges (stereographic projection,Fig.70) are co-axial with the intersection of bedding and the migmatitic fabric and lie near the mean F_3 fold axial plane. There are however, no progressive changes in orientation of F_2 hinge directions within this mean plane as the slide is approached (see Map 4).

7.3b) West of the slide.

A major F_3 synform, the Ranochan synform, lies immediately to the west of the slide zone in the north of the area (Fig.70). The western limb of this major fold contains numerous open to tight F_3 folds which plunge moderately to the NE (Fig.70) and fold the migmatitic fabric related to occasional tight pre-F_3 minor folds. Moving eastwards across the axial plane trace of the Ranochan synform, the F_3 minor folds have the opposite sense of vergence (Map 4) and have tighter interlimb angles. Slightly farther east the F_3 folds become even more tight and much less frequent over a distance of 10-15 metres. There is, however, no progressive re-orientation of minor F_3 fold hinge directions or axial plane trends as the slide is approached (Fig.70). Within the pelitic unit adjacent to the slide (Fig.70 and Map 4), migmatitic segregations are tightly folded and are associated with tight mica crenulations and strong boudinage of pre-F_3 isoclines. These features are held to indicate that D_3 strain increases progressively eastwards toward the slide. As the vergence of minor F_3 folds is spatially related to the Ranochan synform and D_3 strain increases towards the slide it appears most probable that both the Ranochan synform and the slide formed during this phase of deformation. The evidence presented above, given that there is a progressive increase in strain towards the slide zone seems incompatible with the Ranochan synform being cut out by a later, unrelated tectonic break. Such ductile re-working, unrelated to and later than D_3 deformation would be expected, progressively, to re-orientate minor F_3 folds. Nevertheless, the axial plane trace of the Ranochan synform is clearly oblique to the trend of the lithological units mapped (Fig.70) and can be traced southwards to the pelitic unit adjacent to the slide but not through it. The Sgurr Beag slide is therefore regarded as a D_3 structure.

7.3c) Current orientation of the D_3 strain ellipsoid.

The Sgurr Beag slide at the eastern end of Loch Eilt is a vertical D_3 structure. Consideration of the rotation of F_3 fold hinge lines within the vertical NE-SW trending XY plane of the D_3 strain ellipsoid leads to the conclusion that the X axis plunges more steeply to the NE than any of the F_3 fold hinge lines, while the Y axis plunges less steeply to the NE than any of the F_3 fold hinge lines. If the minor folds (see Chapter 5.6e) which plunge to the SW then, following the geometrical argument developed in Chapter 5.6e, the orientations of the X and Y axes of the D_3 strain ellipsoid are more tightly constrained within the vertical NE-SW trending XY plane. The X axis must be approximately vertical and the Y axis approximately horizontal. Thus the extension direction of the D_3 strain ellipsoid (in its present orientation) is vertical.

It is possible to consider the geometry of the Sgurr Beag slide in terms of a thrust belt. The Ranochan synform can be considered to be a ramp, whose "branch line" intersection with the slide is co-linear with the F_3 fold hinge lines. However as the ramp could be frontal, lateral.or oblique (see Butler,1982.text-fig.3) the transport direction of the material in the hanging wall above the ramp (ie. Glenfinnan Division rocks) cannot be constrained.

If the Sgurr Beag slide had frequently cut up section to produce a large number of frontal, lateral and oblique ramps then it might be possible, by considering the spread of orientations of the ramp surfaces, to define more accurately the transport direction. Unfortunately the Sgurr Beag slide is typified by being bed-parallel over virtually all of its known outcrop (Tanner,1971: Baird,1982). The deduced D_3 extension direction is coincident with one of the possible transport directions and it seems highly likely that this sub-vertical direction within the NE-SW trending slide zone defines the direction of transport of the Glenfinnan Division rocks over those of the Morar Division.

The orientation of the extension and transport direction at their time of development is considered in Chapter 7.5.

7.4) Metamorphic features.

7.4a) Pelitic rocks.

Pelitic rocks from both sides of the Sgurr Beag slide contain quartzo-feldspathic migmatitic segregations and sillimanite overgrowths of the early muscovite porphyroblasts. Thus it could be assumed that metamorphism reached the same grade on both sides of the slide zone, though not necessarily at the same time. Evidence has, however, already been advanced showing that the tectono-metamorphic histories of the rock assemblages to both east and west of the slide within the area are essentially the same.

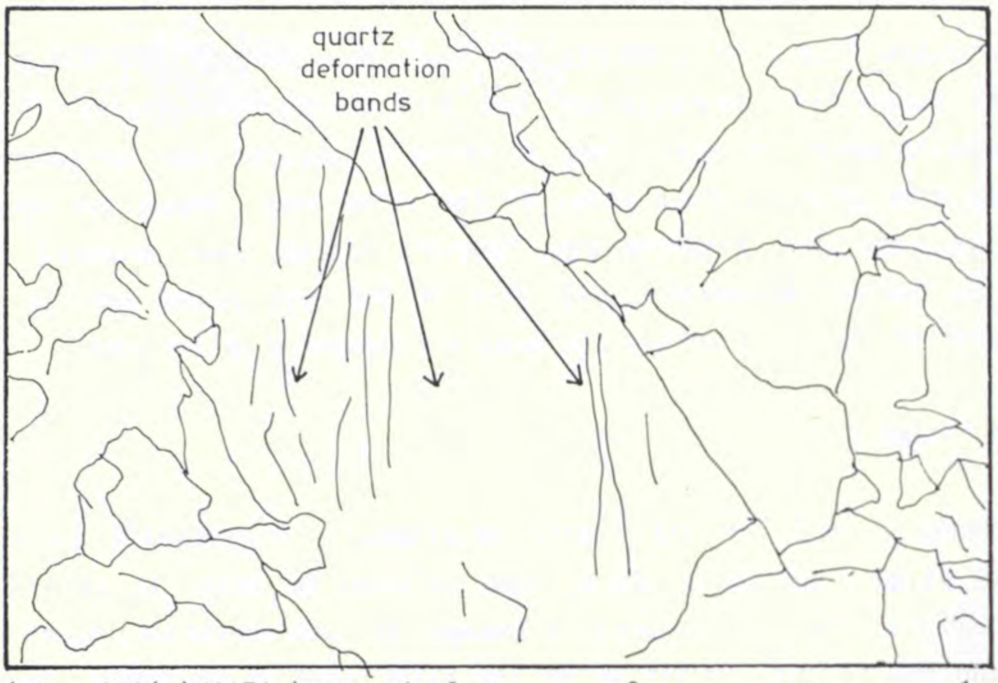
7.4b) Psammitic rocks.

A preliminary study of the fabric development in psammites collected across the slide zone has produced some information which is related to the metamorphic history of the area, although its interpretation, especially its relationship to the formation of the slide zone is equivocal. Samples 260/1163 to 273/1176 inclusive were collected at approximately 100 metre intervals in a WNW-ESE traverse across the slide zone. The samples collected farthest away from the slide zone (approx. 600-700 metres) are the ones which appear to have suffered the lowest levels of D_3 strain. Their textures are thus presumed to be closest to those of a pre-existing fabric with relatively little D_3 modification. In the psammites of the Morar Division, as the slide zone is approached, quartz crystals progressively contain more deformation bands and the pre-existing weak C-axis fabric either becomes more strongly developed or a new one develops as the slide is approached (Fig.75a). Plagioclase and K-feldspar crystals in Morar Division psammites farthest away from the slide zone do not contain myrmekitic textures. Approaching the slide zone small discrete co-planar zones of fine grained crystals occur which show myrmekitic textures (Fig.75b). Near the slide zone the planar zones become more common and randomly distributed throughout the rock.

Psammites collected from the east of the slide zone do not show changes of fabric which can be related to distance from the slide. Myrmekitic textures in feldspars are developed in a few sections and quartz deformation textures such as deformation bands and undulose extinction are sporadically and weakly developed. These textures in the Glenfinnan Division psammites are similar to those seen in psammites of the Morar Division distant from the slide, possibly implying that the rocks of the Glenfinnan Division

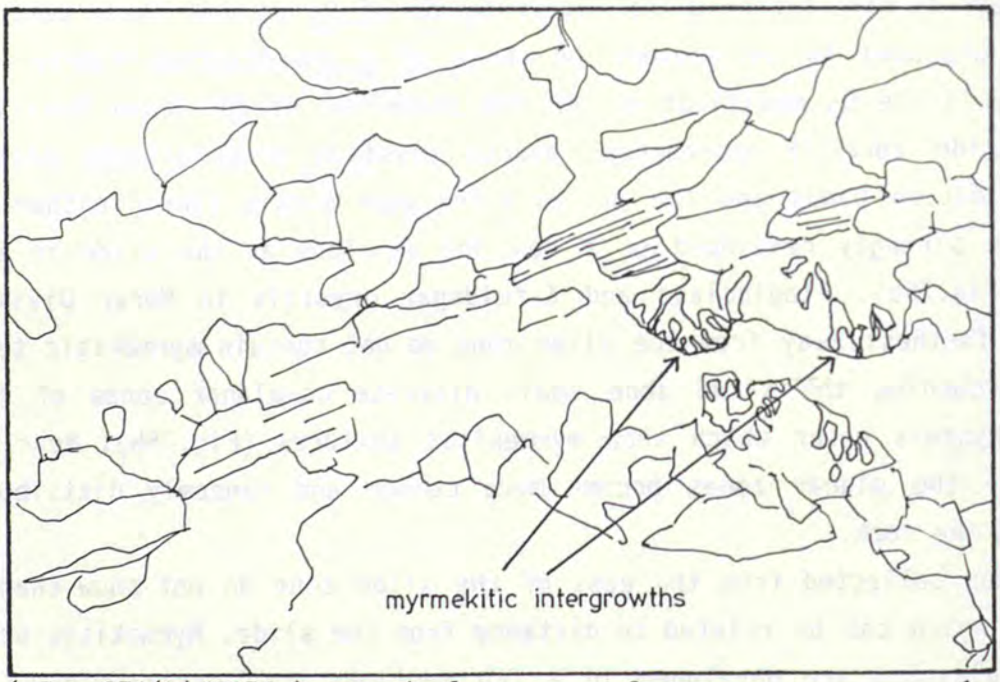
Figure 75. Photomicrographs to show deformation textures in (Morar Division) psammitic rocks near to the Sgurr Beag slide.

(a)

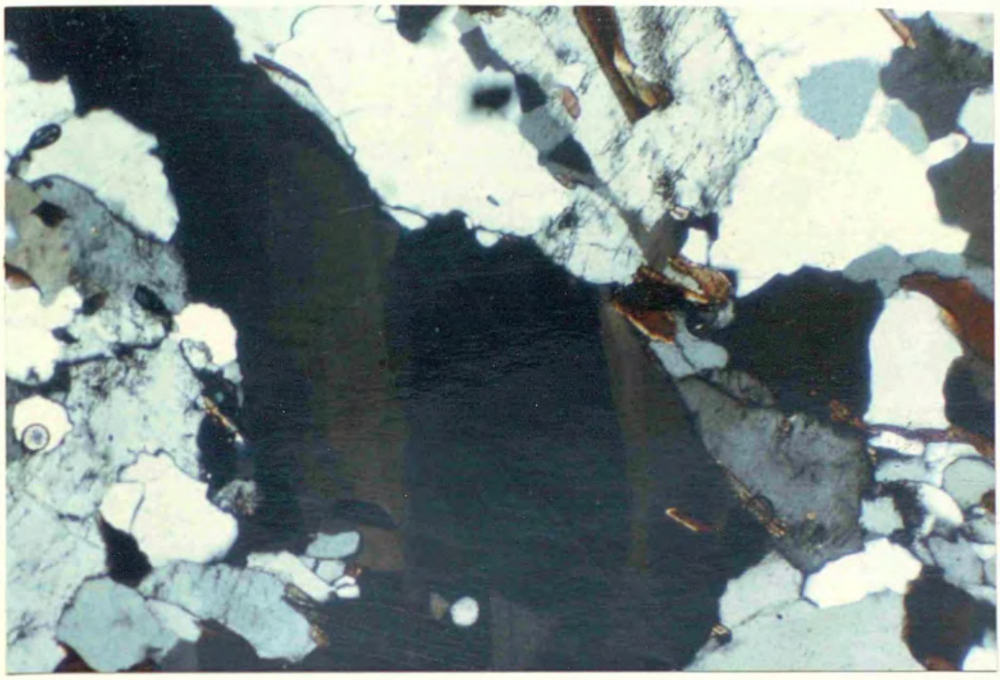


(Exp.268(c)/1171). crossed polars. 0 mm. 1

(b)



(Exp.271(b)/1174). crossed polars. 0 mm. 1



are at a generally lower state of D_3 strain than those of the Morar Division. It is possible to argue that the variations of strain may be a result of F_4 folding of the slide zone, heterogeneous layer parallel slip producing the strain variations. If D_4 deformation was at least partly by a mechanism of flexural slip and was responsible for the fabric variation near the slide zone at eastern Loch Eilt, then one would expect that such fabrics would not be produced in the large area of the major F_4 fold. However "slide zone" fabrics have been observed where the Sgurr Beag slide is folded around the hinge of the F_4 Loch Eilt antiform farther north in NE Morar (Dr.A.L. Harris, pers.comm.) so that it appears that the fabrics of the slide zone are probably D_3 fabrics rather than fabrics produced during the formation of the F_4 Loch Eilt antiform.

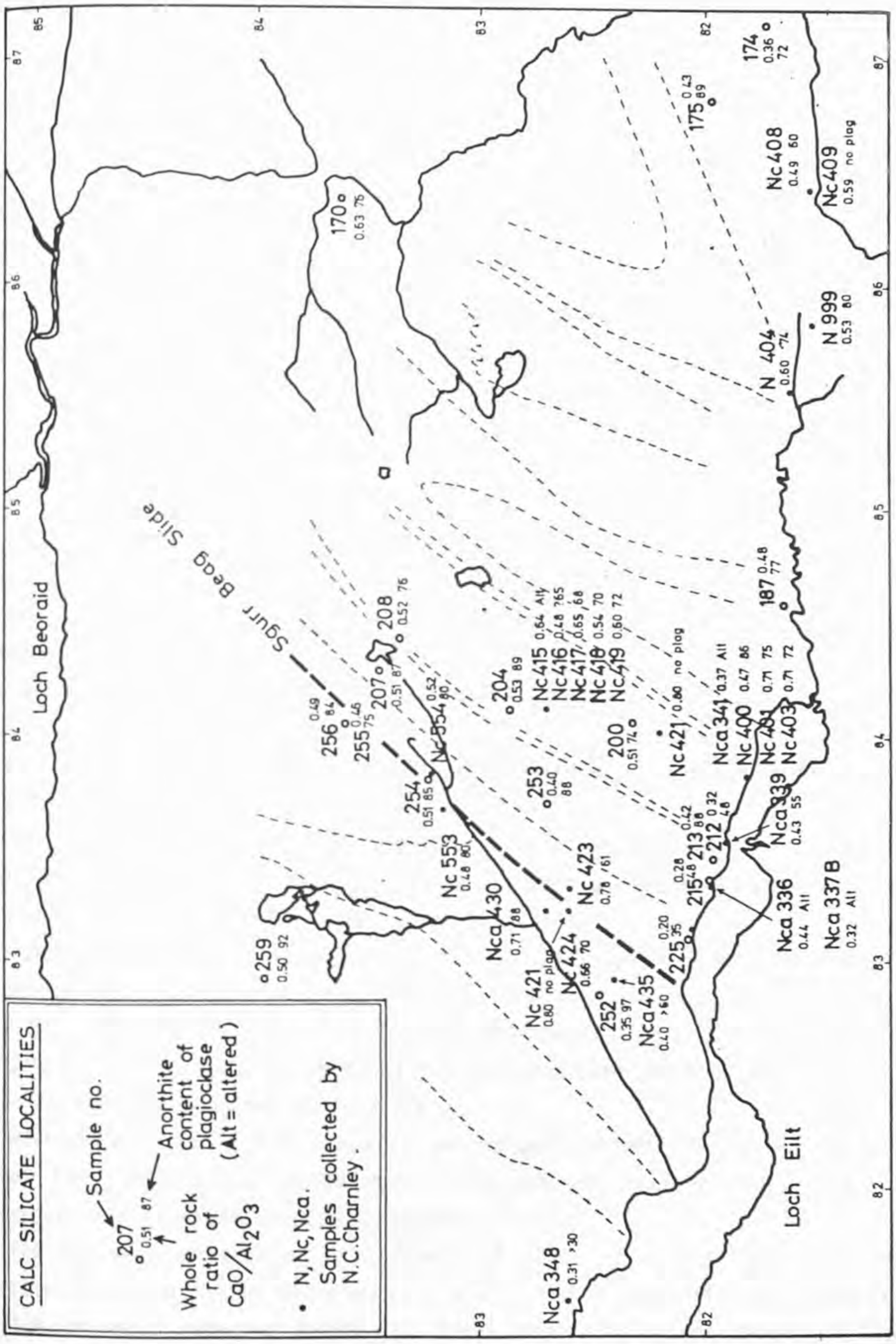
The postulate that the slide zone fabrics and strain variations are D_3 deformation features, with higher D_3 strain in the Morar Division is in agreement with the basic assumptions used by Powell et al.(1981) when producing a model to explain the production of the observed metamorphic pattern in a section from the west coast eastwards almost to Glenfinnan village. One of the major assumptions of this model is that D_3 strain increases asymmetrically, rising gradually in the Morar Division below the slide, reaching a peak within the slide zone and declining very rapidly in the rocks of the Glenfinnan Division. Rathbone & Harris (1979) noted similar high strains across the Sgurr Beag slide at Glen Shiel, 45 Km. further to the north.

7.4c) Calc-silicate rocks.

Calc-silicate mineral textures in the region of the Sgurr Beag slide at eastern Loch Eilt are described from samples collected and analysed by the author (see Appendix 2). This data together with data taken from Charnley (1976) has been used to examine the changes of metamorphic grade across the slide zone and, in conjunction with other regional data, has been used by Powell et al.(1981) to establish some idea of metamorphic grade patterns in the region.

The mineralogy of the calc-silicates has been described in Chapter 4. The most common assemblage of minerals found is Quartz + Plagioclase Garnet ± Biotite ± Pyroxene ± Clinozoisite plus various accessory minerals (see modal analyses. Fig.20 samples 167/883 to 259/1145). Figure 76 shows the location of the samples in relation to the slide zone togather with the An. content of the plagioclase and whole rock CaO/Al_2O_3 ratio. Included in the figure is similar information taken from samples listed by Charnley

Figure 76.



Map of calc-silicate localities in eastern Loch Eilt showing sample Nos. An. content of plagioclase and whole rock CaO/Al_2O_3.

(1976).

Petrographic study reveals that all calc-silicates sectioned contain abundant quartz plus plagioclase and nearly always some amphibole. Plagioclase crystals across the area in question show a range of textures which appear to be related to geographical position. The more easterly sample (167/883) contains plagioclase which has strongly developed exsolution or symplectite texture such that relatively high An. plagioclase crystals contain vermicular growths of lower An. plagioclase content. Virtually every plagioclase crystal shows this texture (see Fig.72), which has been interpreted by Spear (1977) as resulting from retrogression. Moving west-eards this texture becomes much less common in sections 170/948, 174/956, 175/961 and 187/1031, and in sections west of these the texture is very rare or absent.

Clinozoisite is a common phase in many rocks, it usually grows at the expense of plagioclase. In many sections it could be related to, and formed during the production of "exsolution-symplectite" texture in plagioclase: but in section 167/883 it appears to overgrow this texture suggesting that clinozoisite growth is the later, but often preferentially on the more anorthitic portions of the "exsolution-symplectite" textured plagioclase. Saussuritisation of plagioclase is common and occasionally very intense. It is spatially related to plagioclase/clinozoisite aggregates and seems to be a feature of a late stage retrogression which, when very intense, has been accompanied by the growth of some relatively large muscovite crystals within the saussurite.

Clinozoisite/epidote is also found growing at the expense of both amphibole and garnet. Very iron-rich epidotes are seen overgrowing mats of chlorite which themselves replace amphiboles (Fig,77a).

Pyroxene is found in the very small quantities in some sections and seems to be growing from large pale coloured hornblende. Pyroxene growth may be the culmination of a prograde metamorphic event. Acicular small pale hornblende crystals seem to have grown at a later period, also possibly as a result of retrogression (Fig.77b).

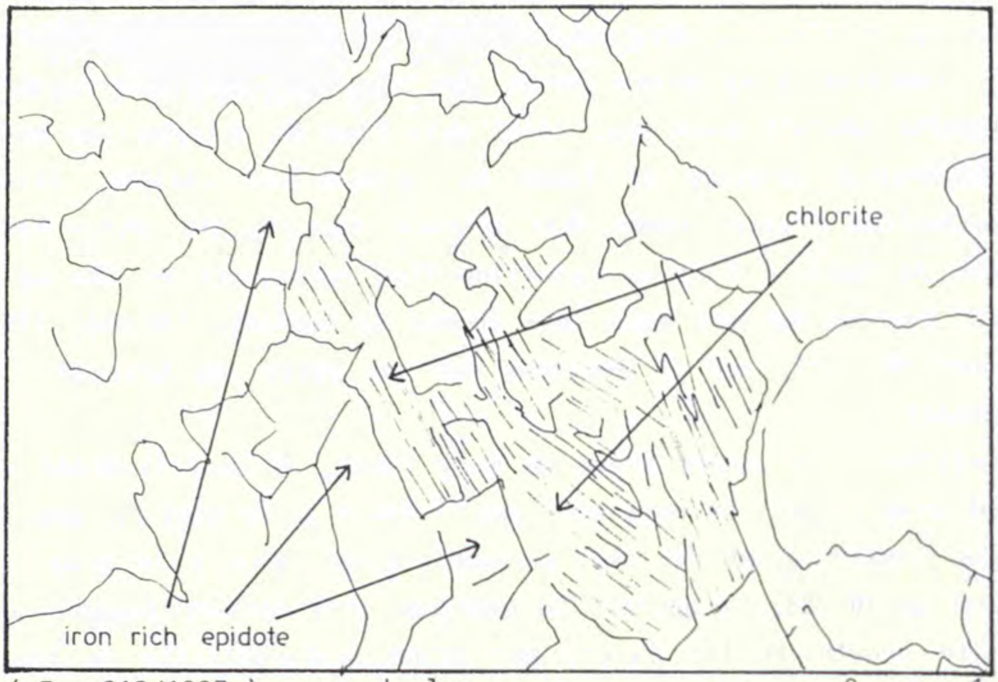
With such a small data base of petrographic observations, conclusions about the direction of metamorphic reactions and whether they represent prograde or retrograde events is always equivocal.

The presence or absence of the mineral phases pyroxene, amphibole and biotite seems partly to be related to whole rock geochemistry as expressed by the chemical ratio of CaO/Al_2O_3. Rocks with a relatively high CaO/Al_2O_3 ratio contain pyroxene (plus amphibole and biotite) whereas those with progressively lower ratios contain amphibole and biotite and then only

Figure 77. Photomicrographs of calc-silicates to show:

Epidote/chlorite textural relationships.

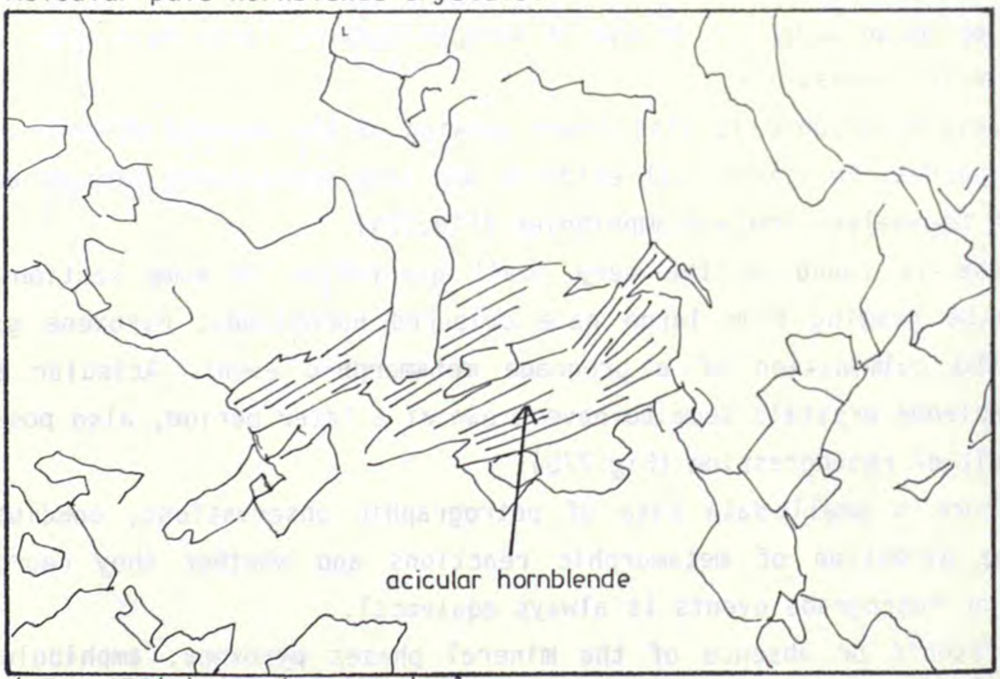
(a)



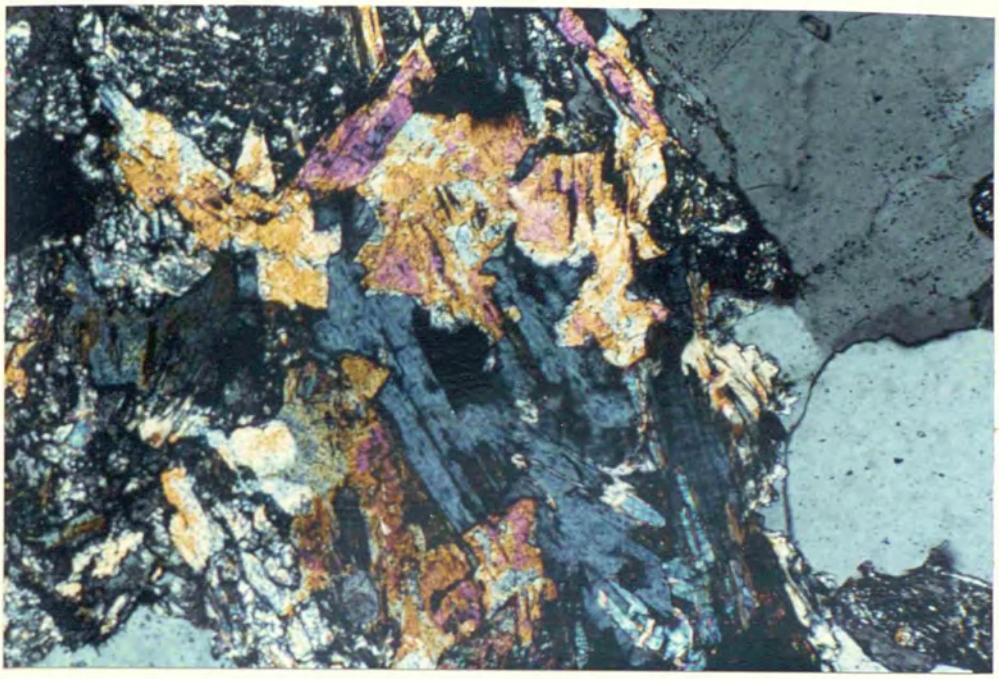
(Exp.213/1087). crossed polars.

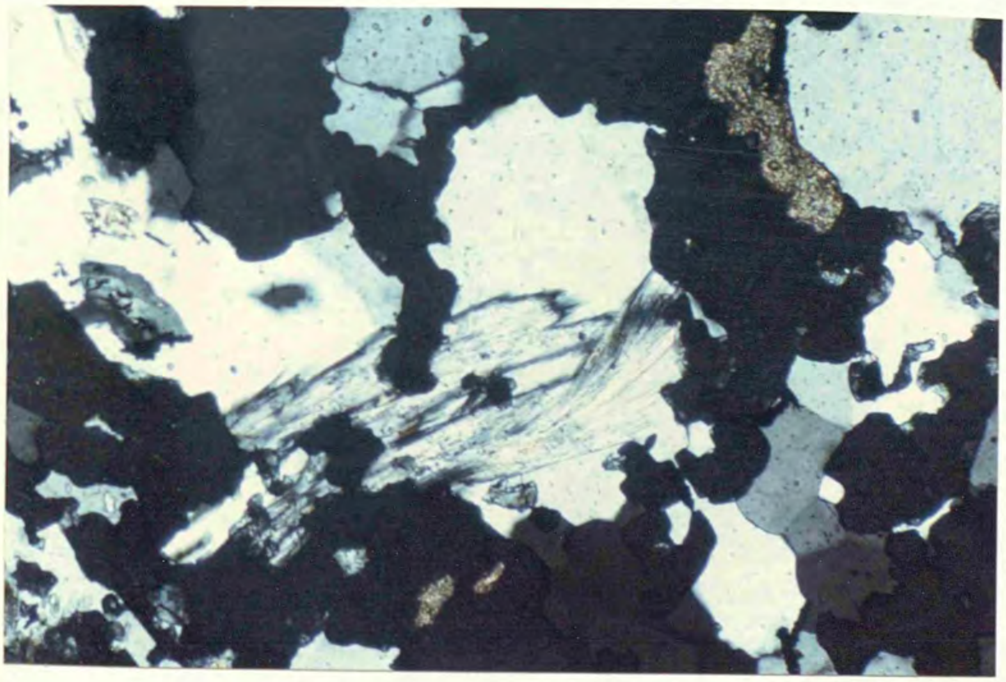
Acicular pale hornblende crystals.

(b)



(Exp.174(b)/956). crossed polars.





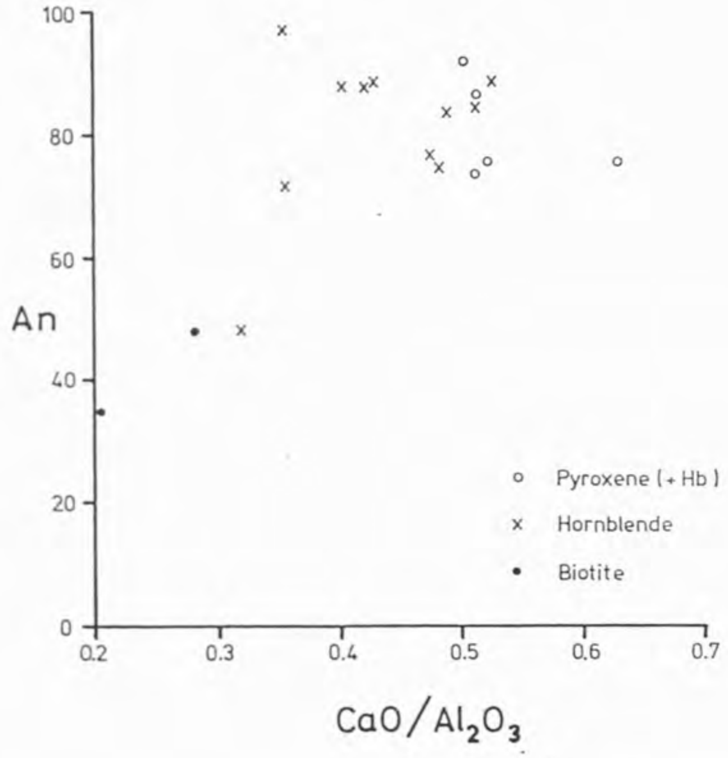
biotite (see Fig.78). In areas of much lower grade metamorphism (to the west) the changes of mineral phases biotite→amphibole (+ biotite)→Pyroxene (+ amphibole + biotite) occur at higher levels of CaO/Al_2O_3 ratio, so that the mineralogy of any particular sample is dependent on both metamorphic grade and whole rock chemistry.

An indication of the variation of metamorphic grade is given by the An. content of plagioclase in the rock (Powell et al.1981). An. content has been measured optically using the Michel Levy method and where sufficient readings have been obtainable from a section these results are in close agreement with electron probe data (Calc-silicate sample 204/1065: Optically An.89%: probe analyses across one crystal, An. 88.7%, An. 91.3%, An. 92.2%, 93.4%, 91.5%, 90.6% and 95.7% (see also Powell et al.1981 and Glazner,1980). Where saussuritisation is intense and albite twins are difficult to see, estimates of the An. content are less acurate and always lower than the actual composition. Figure 79 shows the An. content of plagioclase in calc-silicates plotted against distance from the trace of the Sgurr Beag slide (a similar diagram including this information is given by Powell et al.1981. text-fig.2). Locally the variations of maximum An. content coincide with major structures. The slide zone is clearly seen, as is the relatively open Sgurr a Mhuidhe synform to the east. The position of the Creag Bhan antiform approximately coincides with a rise in the maximum An. content but the adjacent Coille Chreag synform is not apparent on the diagram. If it accepted that the An. variation reflects metamorphic grade (see Powell et al.1981) then it is obvious that the slide zone and the major F_3 folds are deforming and displacing an earlier metamorphic complex.

Theoretical thermal modelling has been used by Powell et al.(1981) to suggest that the overall regional metamorphic pattern is explained best as an early metamorphic pattern which is deformed and modified by a syn-shearing metamorphic event. Only in the highest strain zones within the slide zone has the earlier metamorphic assemblage been re-worked and reset by the syn-shearing metamorphism. The theoretical modelling suggests that the syn-shearing metamorphic effects may be more clearly seen in areas of lower metamorphic grade (which occur to the west of the region) where both the An. content and the mineral assemblages, especially the zoisite-clinozisite relationships, appear to be related to syn-shearing metamorphism. However within the area of Figure 76 the pattern is most simply explained as an early metamorphic pattern folded by F_3 folds and truncated by sliding.

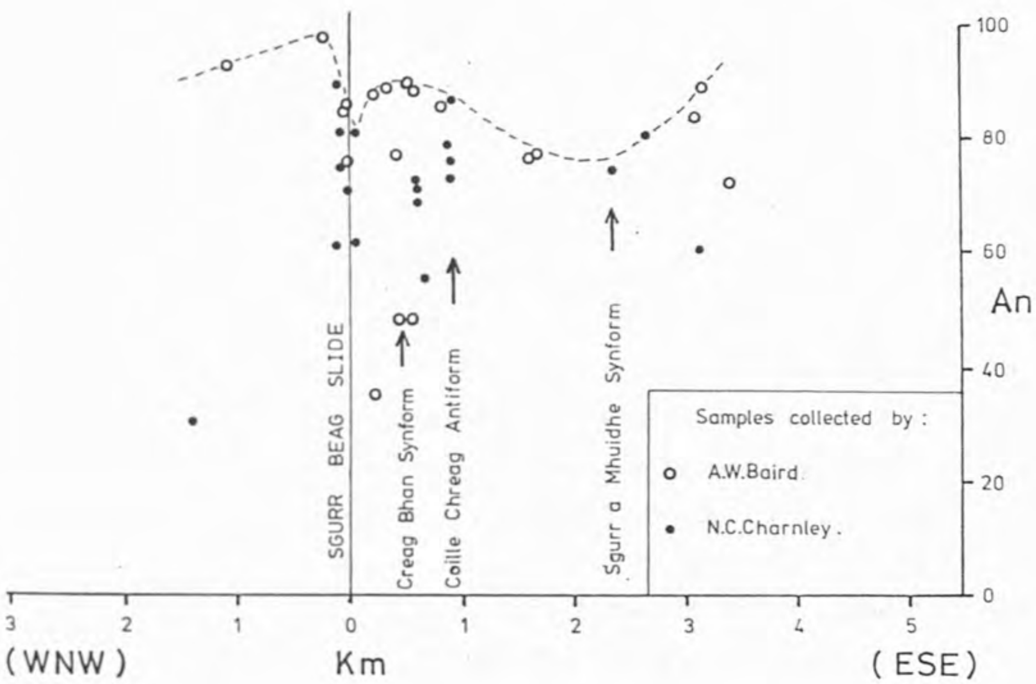
On a regional scale, the incomming of anhedral clinozoisite coincides with the onset of intense D_3 deformation and the production of the Sgurr

Figure 78.



Graph of An. content versus whole rock CaO/Al_2O_3 ratio in calc-silicates, also indicating the mineral assemblages present in thin section.

Figure 79.



Composite profile of variation of An. content in calc-silicates in eastern Loch Eilt.

Beag slide. Thus it can be argued that anhedral clinozoisite growth is related to syn-shearing metamorphism during D_3. However it is still not clear whether the early metamorphic assemblage in the calc-silicates is related to metamorphism associated with D_1 or D_2 deformation. The widespread development of exsolution-symplectite textures in the extreme east which preceeds the growth of clinozoisite (during D_3) may indicate that D_2 metamorphism is re-working or ? retrogressing the earlier D_1 metamorphic event.

7.4d) Status of the "early" metamorphic events.

The grade of an early (D_1) metamorphism in the region is not known in any detail although MacQueen & Powell (1977) report a low grade fine S_1 schistosity along the western seaboard. Within the Ardgour granitic gneiss migmatisation occurred during D_1 deformation and subsequently the gneiss has been re-worked and possibly re-migmatised during D_2.This may suggest that the grade of the earliest metamorphism rose progressively from west to east. The D_2 deformation and metamorphism (extremely intense to the east, in the Ardgour granitic gneiss) may be responsible for the production of the exsolution-symplectite textures in the plagioclase of calc-silicates. The intensity of this texture in the east could be explained either by the D_2 metamorphism which was high enough to re-work the early plagioclase crystals only in the east or by the possibility that further west the earliest metamorphic assemblage was at a grade too low to produce highly anorthitic plagioclase crystals which would re-equilibrate or retrogress by a process of exsolution-symplectite production. This latter suggestion is unlikely since some of the highest An. values have been recorded just west of the Sgurr Beag slide (see Fig.76d) and these rocks do not show widespread symplectite development. However both suggestions imply that the early, pre-D_3 sliding metamorphism is an event associated with D_1 deformation and in this is the case then somewhere presumably near the western seaboard, D_1 metamorphic grade must rise rapidly to exceed D_2 metamorphic grade and this relationship is retained eastwards to Glenfinnan and the Ardgour granitic gneiss. A third alternative is that only in the extreme east (where exsolution-symplectite texture occurs) has D_1 metamorphism exceeded the grade of D_2 metamorphism, so that in the east D_2 is a retrogressive event on a higher grade D_1 assemblage, whereas further west D_2 is progressive and has obliterated D_1 metamorphic patterns.

7.5) The relationship of the Sgurr Beag slide to the "steep belt".

Presently the Sgurr Beag slide at eastern Loch Eilt is a near vertical structure trending NE-SW. The major F_3 folds also have axial planes which are nearly vertical and trend NE-SW with hinge lines near the slide zone plunging steeply to the NE. Eastwards from the slide zone the interlimb angles of the major F_3 folds increase and the plunge of the hinge line decreases. The major F_4 folds, the Loch Eilt antiform to the west of the slide zone and the Glenshian synform further west (Fig.73) are tight to nearly isoclinal folds with near axial planes trending NE-SW and with shallowly SW plunging hinge lines.

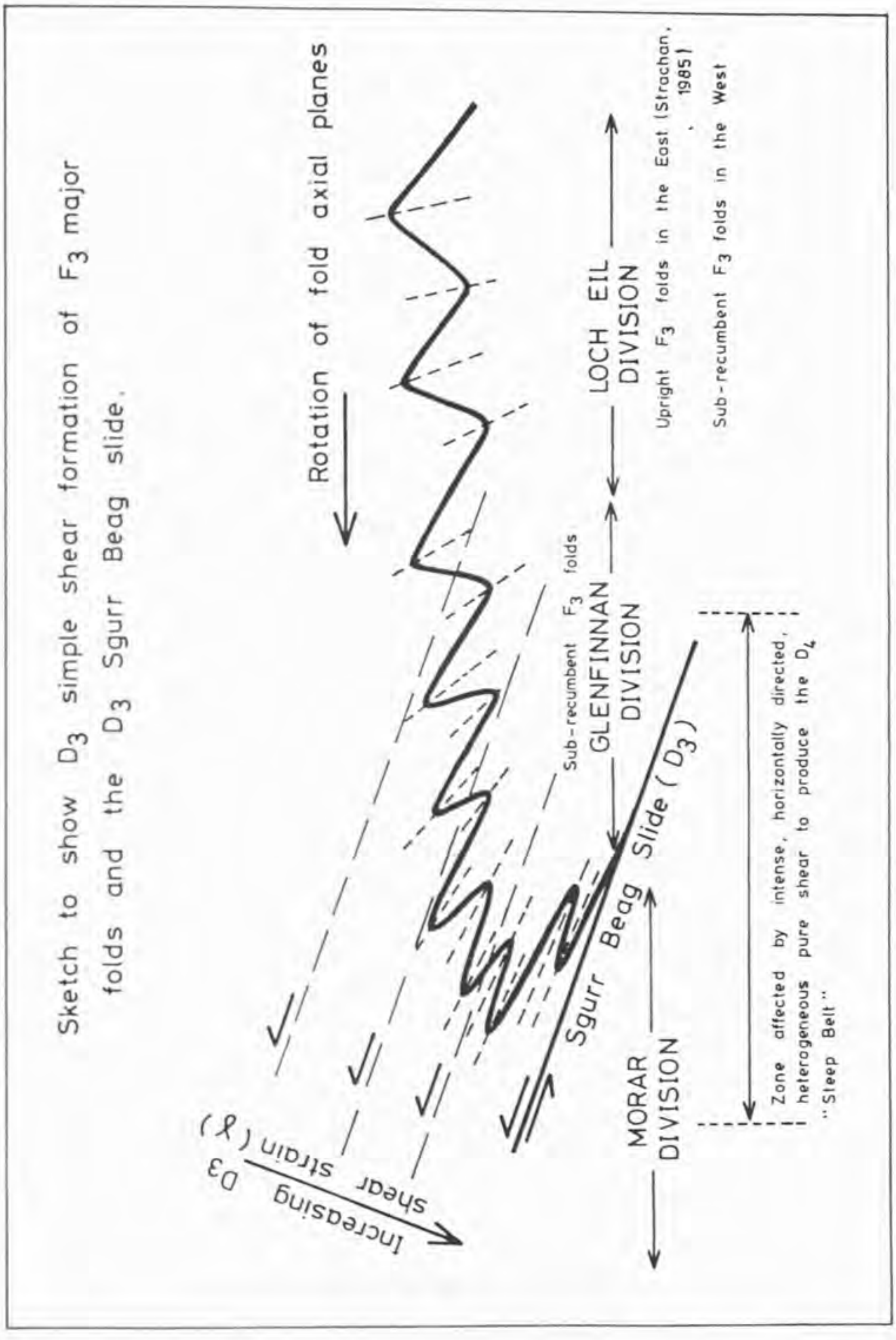
If a simplistic attempt is made to remove some of the effects of F_4 folding by envisaging rotation and unfolding of the fold limbs of the F_4 folds about an axis of rotation coincident with the plunge of the F_4 hinge lines and by ignoring the effects of D_4 bulk shortening by pure shear (see Chapter 8.1b4) which may have re-orientated pre-D_4 planar and linear structures, then the Sgurr Beag slide in post-D_3, pre-D_4 times would have been a relatively flat lying structure dipping shallowly to the east (this is accounting for the vergence of the F_4 folds (see Fig.81a)). The complete stack of major F_3 folds in the Glenfinnan Division would have rested on top of, and to the east of, this slide zone. Lowest in the stack and nearest the slide zone is the Coille Chreag synform which is succeeded upwards and eastwards by the F_3 Creag Bhan antiform followed by the F_3 Sgurr a Mhuidhe synform and the sequence of major folds between this and Glenfinnan village. A sense of F_3 vergence cannot be ascertained as the lengths of the fold limbs are not known and there are no indications of high and low strain limbs. Moving westwards and down through this stack of major F_3 folds in the Glenfinnan Division towards the slide zone one can observe that the folds progressively become tighter and have shorter wavelengths, boudinage increases dramatically and the number of minor F_3 folds decreases, all features which have been related to the progressive increase in the amount of D_3 strain towards the slide zone (see the early part of this Chapter). In such a post-D_3/pre-D_4 reconstruction a progressive change in the orientation of major F_3 fold hinge lines can be seen. In the extreme east (in the Loch Eil Division) the hinge lines would have been horizontal and trending NE-SW, mowing westwards the Sgurr Beag slide the trend of the hinge lines would have rotated clockwise within the easterly dipping fold axial planes and would have gradually plunged more to the E-NE so that just above the slide zone the F_3 hinge lines would have plunged down to the E within the E to SE dipping plane of the slide zone. It has been argued in Chapter 5 that the D_3 strain ellipsoid (after D_4 deformation) has the X axis, or the extension direction, vertical. When the effect of F_4 folding is removed, the X direction of the D_3 strain ellipsoid trends approximately NW-SE within the plane of the slide zone and is approximately co-linear to its dip direction. Thus it appears that D_3 extension was approximately northwesterly directed up the (pre-D_4) dip of the slide zone. D_3 extension and transport of the Glenfinnan Division rocks towards the NW (as opposed to down the dip of the slide zone towards the SE) accounts for the juxtaposition of the high grade migmatitic Lochailort pelite (= Glenfinnan Division) on top of the underlying, non-migmatitic rocks of the Morar Division at Lochailort.

In post-D_3/pre-D_4 times the structures in the Glenfinnan Division were essentially flat lying major F_3 folds related to syn-metamorphic sliding in a NW direction. In the Loch Eil Division, where D_4 "steep belt" effects are minimal. the Loch Eil Division psammites contain a major F_3 synform, the Druim Beag synform, which has a near horizontal hinge line and shallow SE dipping axial plane, faces up to the NW. However, Strachan (1985) working in a larger area, noted that the F_3 Druim Beag synform and other F_3 folds are generally upright and fairly open (Strachan,1985. Table 2).

The E-W variation of F_3 fold geometry is best explained in terms of simple shear producing initially upright folds which tighten, amplify and rotate to become recumbent with increasing shear strain (Fig.80)(Sanderson, 1979: Ramsay et al.1983). In the west of the area, as a result of intense NW directed D_3 shear strain the D_3 Sgurr Beag slide developed below the F_3 recumbent sheath folds in the Glenfinnan Division.

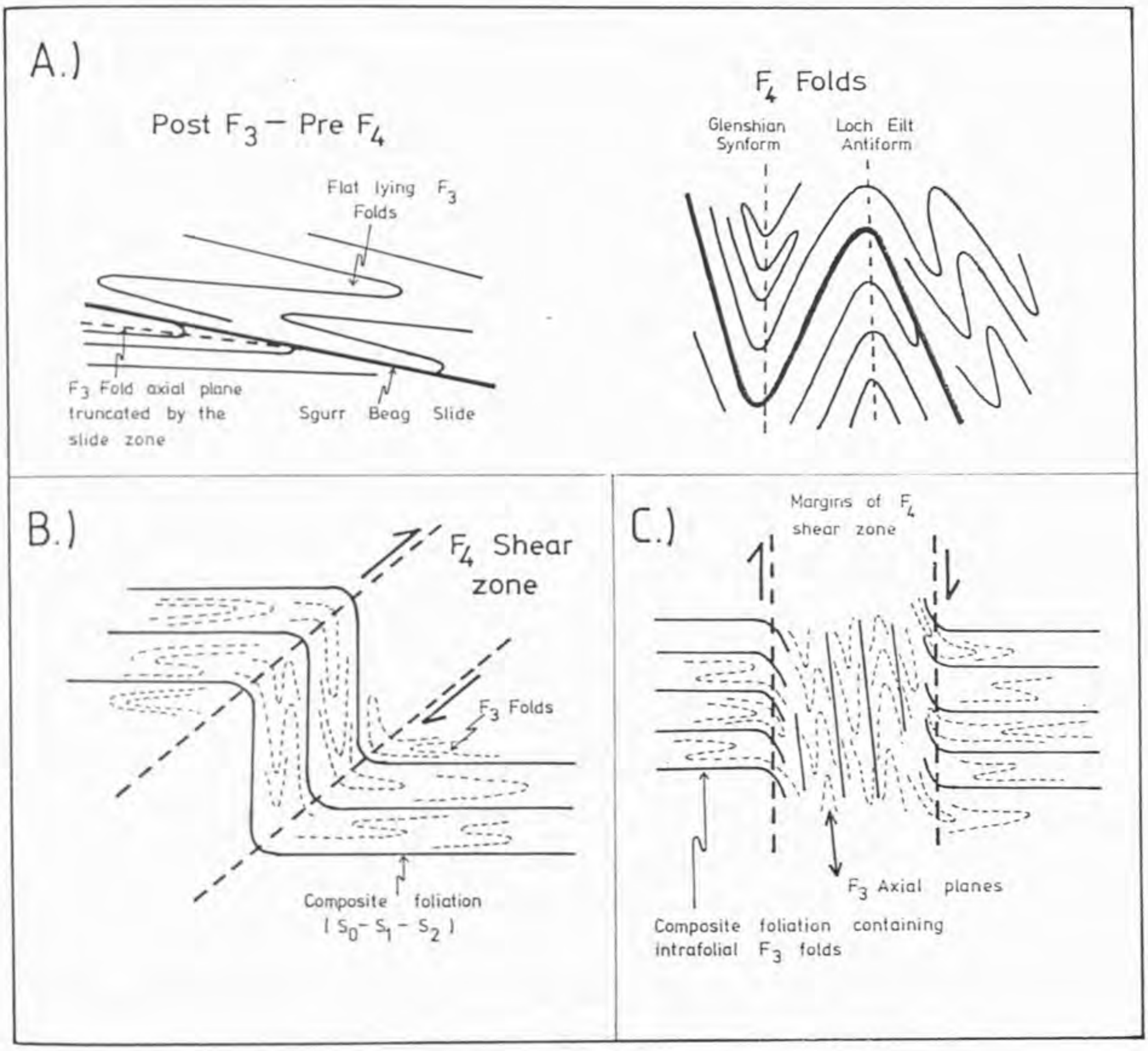
The "steep belt" is the result of intense D_4 deformation which has produced the vertical disposition of the earlier flat-lying F_3 folds. D_4 deformation is also responsible for the deformation of the post-D_3 suite of microdiorite sheet intrusions (Chapter 8.2). The "steep belt" is approximately coincident with the area between the Sgurr Beag slide in the west and the Loch Quoich line in the east and it is worth considering why the "steep belt" is restricted in width with such a strong geometrical contrast between the interior and exterior of the belt. It is possible that the "steep belt" could be some form of major simple shear zone such as sketched diagrammatically in Figure 81b. In such a major F_4 simple shear zone there need not be major F_4 folds within the margins of the shear; and indeed, folds which are unequivocally F_4 folds between the limits of the steep belt have not been observed. However, as sketched in Figure 81b the margins of the simple shear zone (which would be the axial planes of major F_4 folds) are strongly oblique to the axial planes of F_3 folds. To fit the geometry observed across

Figure 80.



Diagrammatic sketch to show the development of D_3 structures as a result of simple shear deformation.

Figure 81.

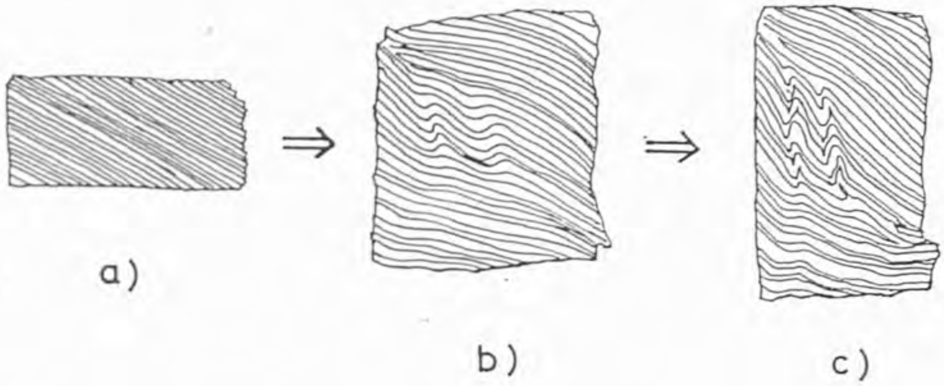


Sketches to show possible modes of formation of the D_4 "steep belt".(see text for discussion).

the steep belt, the margins of the simple shear zone would have to be virtually co-planar with the axial planes of the F_3 folds within the shear zone (Fig.81c), a situation which would require very high levels of shear strain (eg. $\gamma > 5$). The "D_4 simple shear zone" (ie. the "steep belt") margins are vertical and the shear zone width is some 10-15 Km. so that vertical movements across the simple shear would have to be in the order of 50++ Km. Such movement with inherent changes of metamorphic grade on either side of the shear zone, are inconceivable. Thus it seems that the "steep belt" cannot be made to fit any simple model of a large scale simple shear zone.

Alternatively, the production of the "steep belt" can be considered to be the product of horizontally directed pure shear— a large scale example of crenulation folding, small scale examples of which have been modelled by Cobbold et al.(1971. text-fig.17) and Cosgrove (1976,Plate 2a,b,c.) (see Fig.82). Both of these models require that the initial layering is slightly oblique (20°) to the direction of maximum compression. The end results of the experiments (with horizontal compression) are packages of tight to nearly isoclinal folds with near vertical axial planes within generally shallowly dipping rocks. When scaled up to be analogous with the "steep belt", the buckles correspond to major F_4 folds and the banding or layering, to a composite foliation with almost intrafolial F_3 folds This analogy has a horizontal NW-SE directed compression during D_4 which compresses a composite foliation with almost intrafolial F_3 folds which dip gently (20°) to the SE. There is no evidence that major F_3 folds are involved in major re-folds within the Glenfinnan Division eastwards from the outcrop of the Sgurr Beag slide at the eastern end of Loch Eilt, therefore if the analogue (see Fig.82) is reasonable then this area must be on the steep eastern limb of a major upright fold (the F_4 Loch Eilt antiform) the steep nature of which begins to decrease in the Glenfinnan village-Loch Quoich line area. In this analogue it is assumed that the folds which form could vary in amplitude along their hinge lines and could form enechelon, so that the "steep belt" may be a highly variable structure along its length. Chapter 9 considers the D_3 sliding and nappe formation and D_4 pure shear compression in its regional, crustal context.

Figure 82.



Traced from Cosgrove (1976. Plate 2 a,b,c)

Sketches to show the development of upright crenulation folds as a result of horizontal compression of a slightly inclined multi-layer. (see text for discussion).

CHAPTER 8.

Basic and metabasic rocks.

This chapter is not intended to be a comprehensive study of the basic and metabasic rocks of the area; rather it is an attempt to describe the distinguishing criteria (petrographic, geochemical and structural) of the different suites of basic and metabasic rocks and to supply geochemical data which may be of use to others. (All basic and metabasic geochemical analyses together with sample grid references are included in Appendices 2 & 4).

8.1) Microdiorite petrology and geochemistry.

Microdiorite sheet intrusions are common along the road section (A830.) from Loch Eilt eastwards through most of the Glenfinnan Division. Fewer microdiorites have been found further east and in the hillsides to the north.

The microdiorite suite contains a range of petrographic types all of which have been subjected to various amounts of metamorphism and deformation. Figure 83 lists major and some trace elements of 30 analysed samples. Samp-

Figure 83.

Goechemical analyses of microdiorite samples.

Microdiorites

	33/229	57/234	71/237	74/241	83/255	85/259	91/318	93/335	99/418
P	0.215	0.250	0.005	0.195	0.315	0.485	0.285	0.375	0.355
Si	56.825	58.615	72.985	63.355	54.850	50.895	56.220	56.425	51.350
Al	15.975	17.690	15.985	16.740	15.450	16.240	15.770	15.450	11.230
Mg	5.950	4.095	0.295	3.175	6.890	7.890	6.060	6.710	8.455
Mn	0.135	0.125	0.070	0.090	0.145	0.155	0.180	0.120	0.135
Fe	7.365	7.025	1.350	5.275	7.805	9.040	8.440	7.610	9.425
Ti	1.005	1.040	0.200	0.840	1.200	1.770	1.330	1.270	1.620
Ca	6.655	4.785	1.630	4.585	8.260	7.960	6.825	6.565	8.310
K	1.610	2.755	3.630	1.705	1.370	1.795	1.340	1.850	1.215
Na	4.390	4.270	5.055	4.670	3.440	3.380	3.350	3.925	3.220
Total	100.060	100.655	101.205	100.635	99.735	99.610	99.850	100.310	99.305
Zr	168	190	141	177	180	230	228	263	155
Y	23	26	25	23	23	32	123	110	28
Rb	42	107	100	53	27	74	59	44	28
Nb	14	12	14	12	5	11	14	15	9
Sr	592	733	559	826	874	781	661	793	689
Th	6	12	6	11	6	2	5	5	0

	112/503	113/503	114/503	115/506	117/512	118/514	125/546	142/654	151/721
P	0.260	0.275	0.195	0.205	0.220	0.220	0.160	0.250	0.375
Si	58.620	62.770	59.310	52.730	55.230	50.250	60.535	57.590	50.910
Al	16.900	17.495	15.575	13.875	15.675	12.075	15.705	16.125	16.245
Mg	3.450	2.170	5.915	8.480	6.275	12.230	5.830	5.030	6.125
Mn	0.140	0.100	0.120	0.155	0.145	0.250	0.110	0.115	0.135
Fe	7.480	5.720	5.905	9.200	8.180	10.560	5.380	6.345	8.835
Ti	1.250	0.880	0.985	1.520	1.315	1.225	0.890	1.055	1.230
Ca	5.690	4.075	5.370	8.320	3.595	7.885	4.880	5.575	8.455
K	2.330	1.960	1.830	1.365	1.950	1.725	1.130	1.785	1.375
Na	4.130	5.395	4.065	2.905	3.960	1.895	4.035	4.595	4.380
Total	100.270	100.850	99.260	98.660	96.545	98.310	98.665	98.570	98.055
Zr	200	255	154	162	205	135	135	211	176
Y	34	24	21	32	30	28	19	25	27
Rb	65	51	53	37	95	80	50	49	31
Nb	14	17	13	5	15	11	12	11	12
Sr	672	835	560	465	344	282	511	748	934
Th	5	7	2	2	6	4	5	4	4

	154/721	155/723	156/724	158/728	180/985	182/986	202/1063	218/1089	219/1092
P	0.400	0.265	0.215	0.24	0.360	0.260	0.295	0.550	0.190
Si	53.680	59.650	53.520	57.915	52.430	51.630	51.685	48.480	49.285
Al	18.440	18.515	14.570	17.085	15.475	15.080	13.270	11.725	11.775
Mg	4.610	2.395	8.755	4.265	7.790	8.575	11.420	11.960	11.745
Mn	0.140	0.070	0.145	0.105	0.150	0.160	0.120	0.200	0.190
Fe	8.505	5.215	8.415	6.465	8.455	9.145	8.010	10.060	11.130
Ti	1.580	0.750	1.185	0.840	1.295	1.220	1.150	1.165	1.195
Ca	4.585	4.765	7.380	4.595	8.140	9.080	6.410	11.275	9.360
K	1.255	1.465	1.140	3.200	1.420	1.360	2.300	1.700	1.590
Na	4.625	5.515	3.100	3.825	3.325	2.320	3.370	0.985	2.865
Total	97.815	98.595	98.625	98.535	98.835	98.815	98.030	98.100	99.390
Zr	224	195	149	177	180	136	146	137	153
Y	41	17	21	25	27	24	22	28	18
Rb	55	53	39	150	36	42	50	73	34
Nb	18	22	9	10	18	14	7	10	6
Sr	829	1017	527	736	879	490	545	838	308
Th	2	4	2	2	6	5	3	7	-1

Major elements: See note on the following page.

Figure 83.
(continued).

Geochemical analyses of microdiorite samples.

Microdiorites

	226/1092	227/1092	228(1)/1092	228(2)/1092	229(1)/1092	229(2)/1092	229(3)/1092	229(4)/1092
P	0.425	0.350	0.485	0.435	0.390	0.365	0.425	0.485
Si	51.955	49.350	51.535	50.565	48.950	49.570	49.225	50.665
Al	15.855	14.745	14.930	13.870	11.980	12.090	13.810	14.110
Mg	6.970	8.705	8.290	9.880	12.565	12.215	9.190	9.445
Mn	0.135	0.180	0.165	0.160	0.170	0.170	0.150	0.150
Fe	7.790	10.035	8.835	9.140	9.330	9.290	8.700	8.485
Ti	1.430	1.270	1.250	1.235	1.095	1.155	1.335	1.180
Ca	8.530	8.850	9.555	0.265	10.405	10.235	9.940	9.380
K	1.680	1.890	1.400	1.200	1.375	1.495	1.310	1.360
Na	4.085	3.510	3.060	2.255	1.605	1.765	3.420	2.845
Total	98.845	98.575	99.515	99.000	97.865	98.360	97.510	98.110
Zr	156	144	169	153	131	118	152	168
Y	34	24	28	23	24	24	26	22
Rb	36	60	43	32	36	48	40	43
Nb	19	8	14	5	10	7	11	14
Sr	975	915	1089	1070	702	719	1128	1133
Th	8	5	12	10	7	13	10	11

Major element analyses (ie. P to Na) are listed as Wt.% oxides (P_2O_5, SiO_2, Al_2O_3, MgO, MnO, Fe_2O_3, TiO_2, CaO, K_2O, Na_2O).

les 228/1092 and 229/1092 demonstrate the geochemical variation across the complete width of a 21 cm. wide sheet with intensely schistose margins and a much less schistose interior.

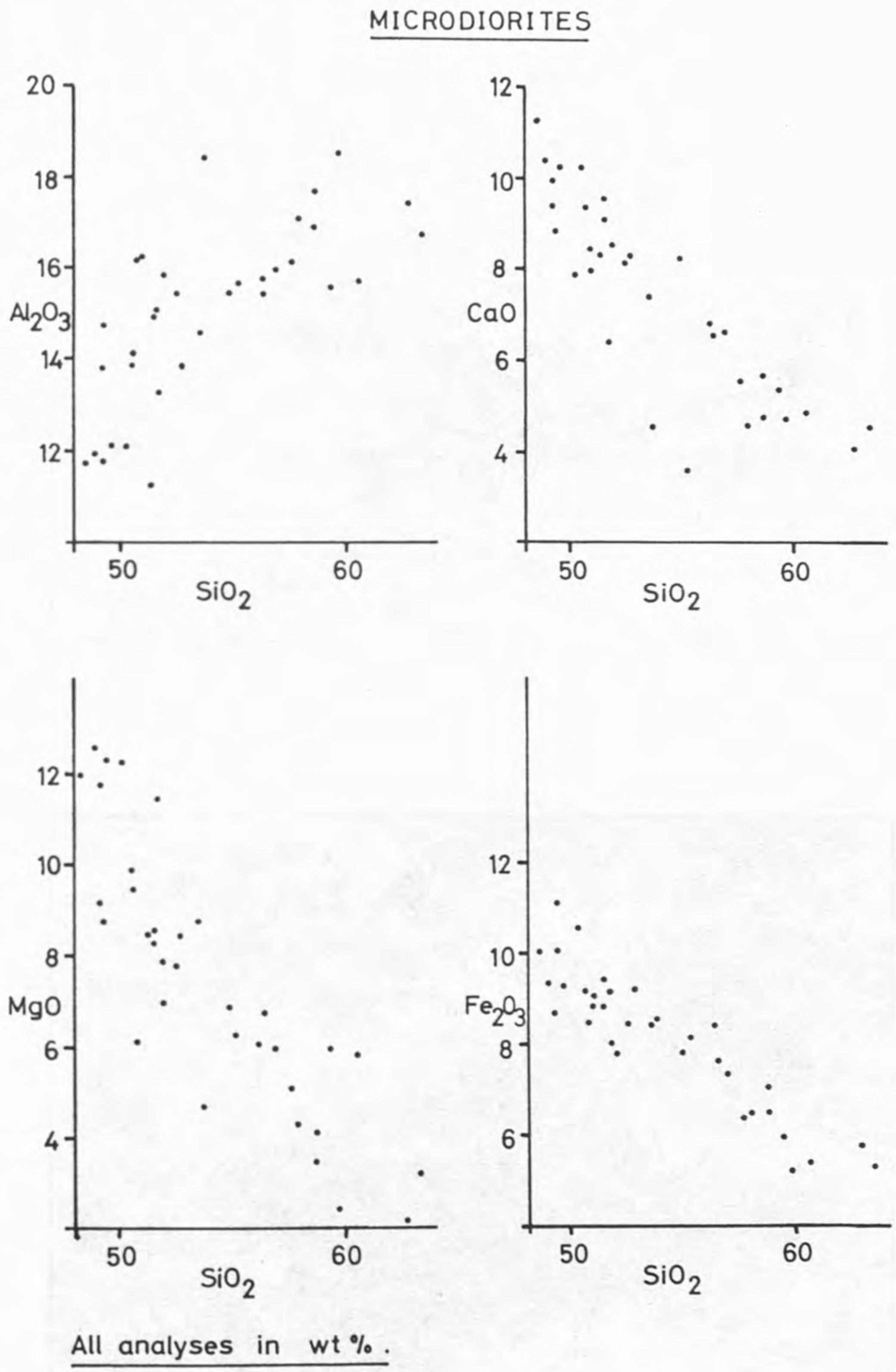
The microdiorite suite consists of a transition from fine grained basic (basaltic) rocks to andesitic and more acidic intrusives. Figure 84 shows the variation of Al_2O_3, CaO, MgO and Fe_2O_3 plotted against SiO_2. The rocks appear to be part of a high-alumina, calc-alkaline suite. The usual mineral assemblage is Feldspar (mainly plagioclase) + Amphibole + Biotite ± Quartz.

Texturally the rocks are usually fine grained and granoblastic, preserving few if any igneous features. Some large plagioclase crystals (1-2 mm.) may be relict igneous phenocrysts which are partly engulfed and corroded by the fine felsic groundmass (Fig.85a). In most sections the plagioclase is andesine with a more albitic rim. Amphibole and biotite laths usually define a schistosity which is often strongly developed; the laths can occur singly or as elongate, spindle shaped clusters up to 1-2 mm. long, biotite is often seen replacing amphibole. The amphibole is hornblende, occurring as laths with a very pale light green, brown-green pleochroic scheme. Frequently the amphiboles contain an inner core which is dusty, brownish with rods of dark brown material (? iron ore. see Fig.85b). The core has a more stubby shape and the rods of (?) iron ore are often normal to the length of the enclosing elongate hornblende. The core and mantle are optically continuous. The texture appears to be the product of conversion of (stubby) pyroxenes to hornblende, with the "rods" representing exsolution products or relics of a diallage type of texture. It the precursor was pyroxene then it has been converted to hornblende by preferential growth on the prism faces (010) of the pyroxenes.

Smith (1979) notes that towards the west coast igneous textured microdiorites occur. He defines an area of amphibolite facies microdiorites, the western margin of which crudely coincides with the western edge of the "steep belt". Noteworthy is the lack of displacement of this margin by the D_3 Sgurr Beag slide. The eastern limit of amphibolite facies microdiorites extends to the Great Glen fault.

Metamorphism and deformation of microdiorites may produce geochemical changes and a preliminary examination of this problem was attempted by analysing samples 228/1092 and 229/1092 (see Fig.83). These traverse the complete width of a 21 cm. wide microdiorite which has highly schistose margins 2-3 cm. thick in contrast to the moderately schistose interior. Slight changes in chemistry can be detected which vary symmetrically from the middle to the edges of the sheet. The analyses of 228(1) to 229(4) are listed in Figure 83 so as to record the geochemistry across the sheet

Figure 84.

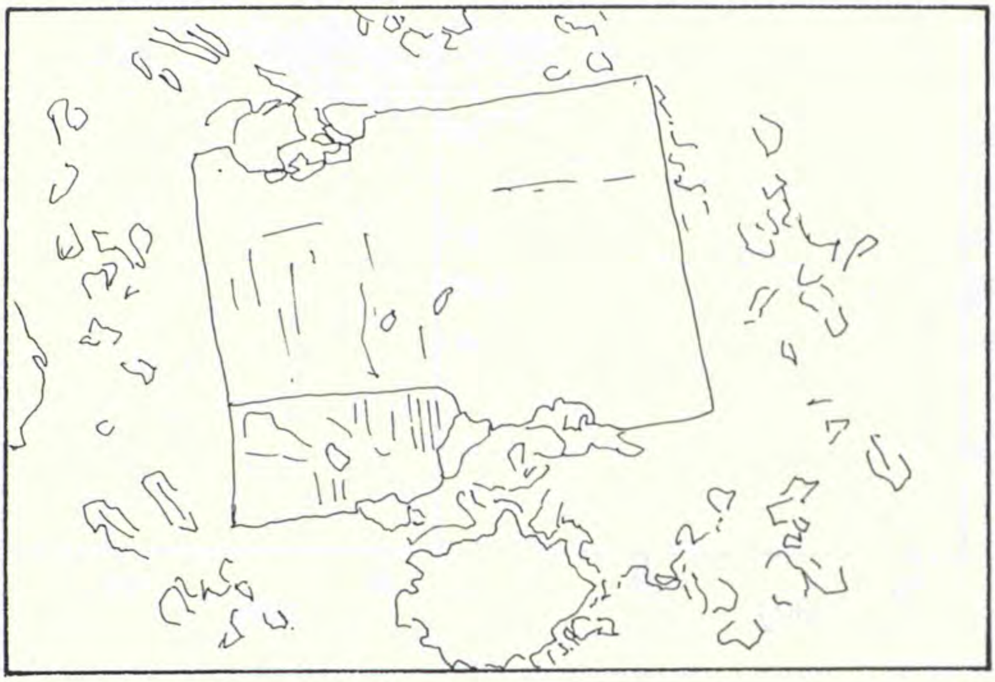


Graphs of Al_2O_3, CaO, MgO and Fe_2O_3 versus SiO_2 in microdiorites.

Figure 85. Photomicrographs of microdiorite textures.

Relict igneous plagioclase phenocrysts.

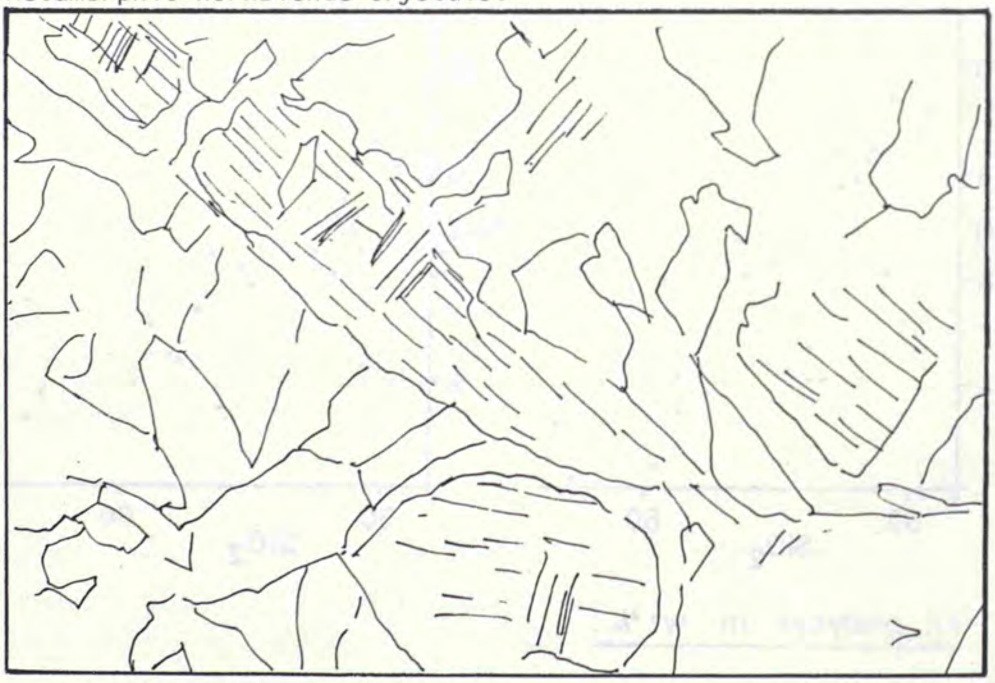
(a)



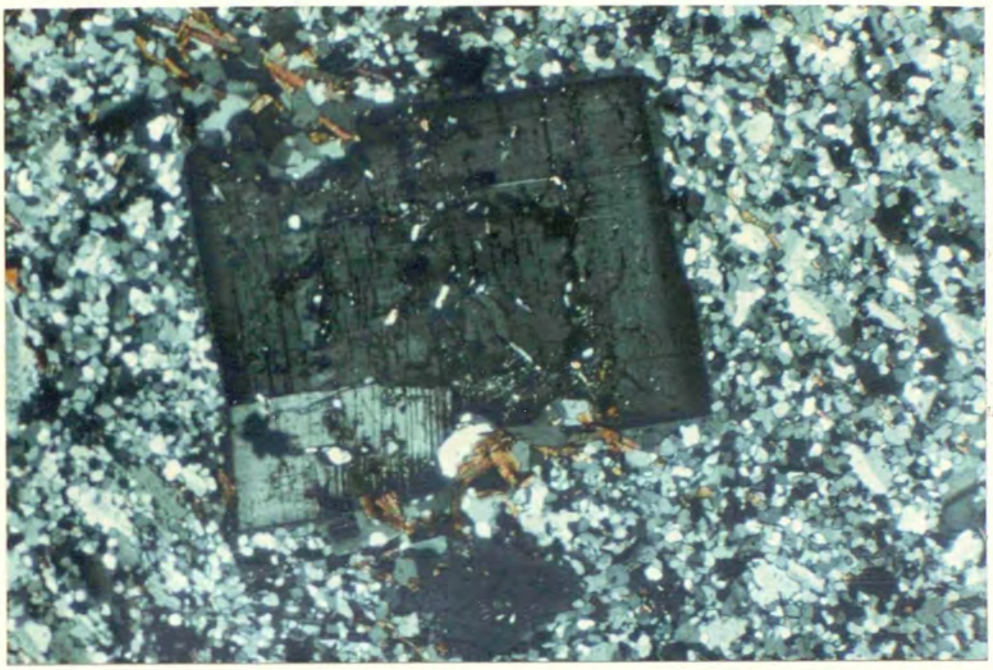
(Exp.71/237). crossed polars. 0 mm. 1

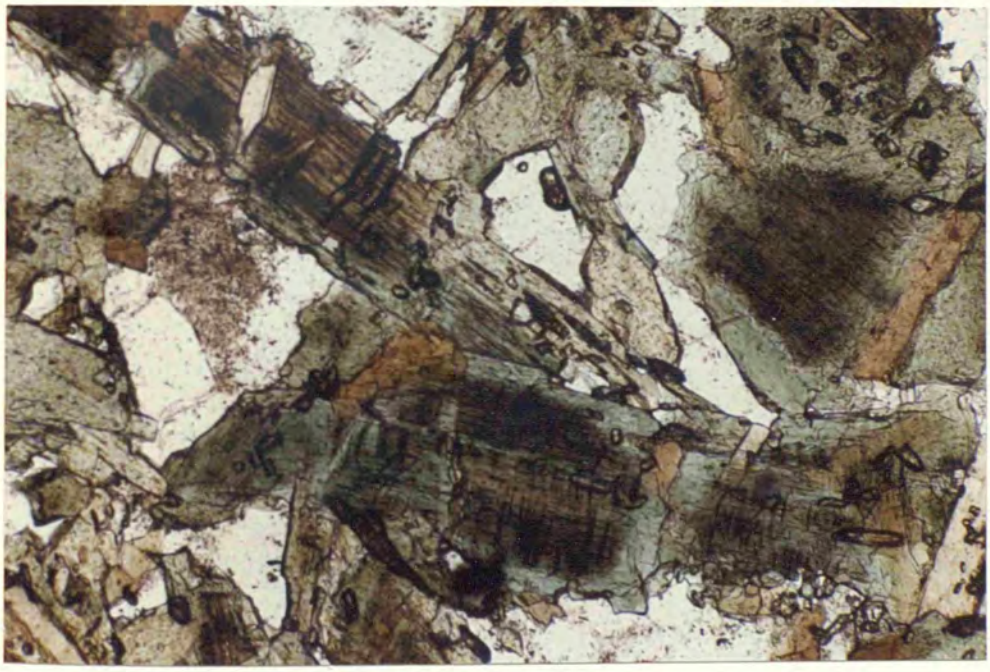
Metamorphic hornblende crystals.

(b)



(Exp.226/1092). plane polars. 0 mm. 1





from one margin to the other. Each analysis corresponds to a strip of rock approximately 3 cm. wide. Moving towards the edge of the sheet there is a slight increase in SiO_2, P_2O_5, Al_2O_3 and Na_2O and a slight decrease in MgO, Fe_2O_3 and CaO. The variations in geochemistry coincide with the variations of intensity of schistosity. The margins of the microdiorite are much more intensely schistose and rich in biotite. The variation may be due to chemical re-adjustments during metamorphism and heterogeneous deformation but such variations could be a product of whatever process of igneous differentiation or partial melting that produced the geochemical vatiation within the suite as a whole; the margins of this intrusion are slightly more acidic than its core.

8.2) Microdiorite structure.

8.2a) Introduction.

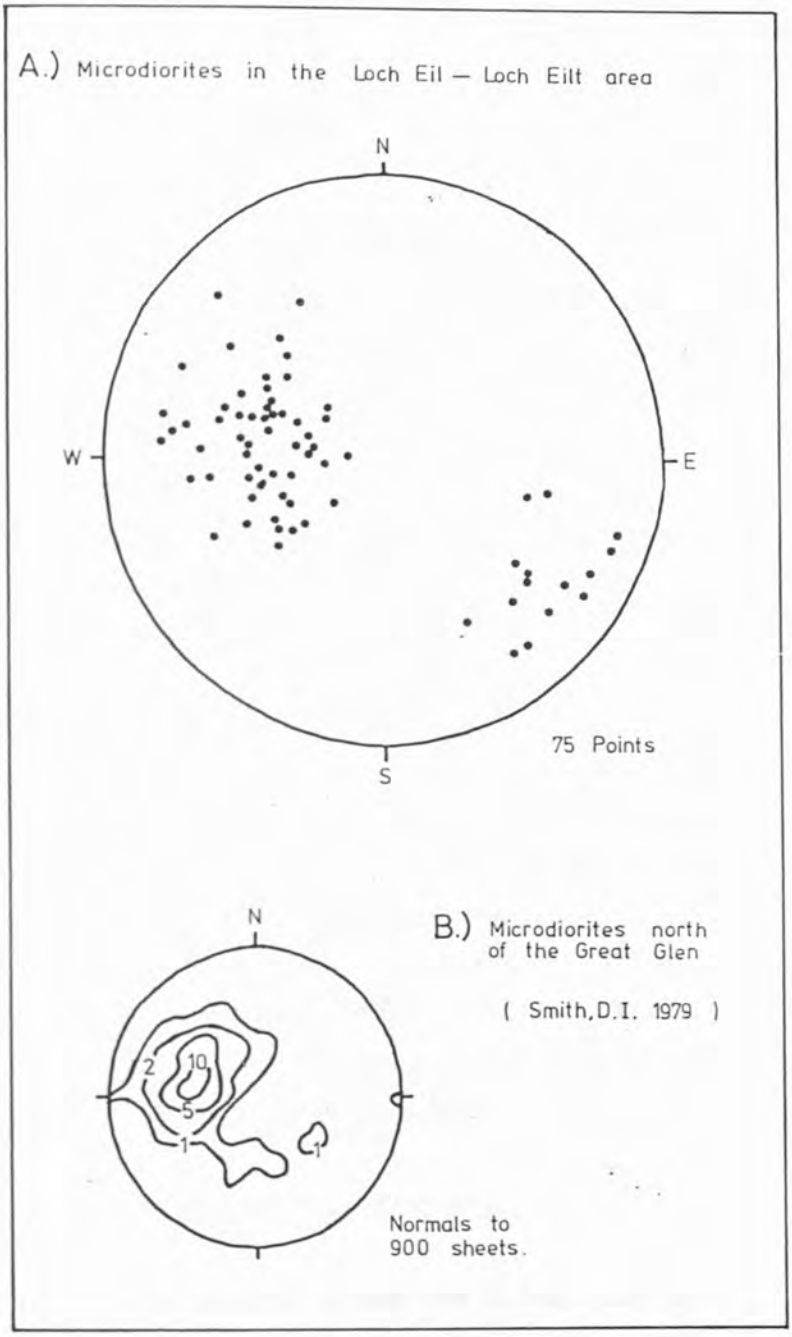
It is obvious in the field that most, if not all, of the microdiorite sheet intrusions have been deformed and metamorphosed, however Johnson & Dalziel (1966) claimed that this deformation and metamorphism was not accompanied by penetrative deformation in the country rocks and thus post-dated the regional folding. They further argued that the metamorphism of the microdiorites could be due to such factors as strain energy due to deformation, heat of intrusion, thermal state of the host rocks, load pressure and water vapour pressure; however these explanations usually require some form of special pleading and do not seem to fit many of the observations cited below.

Talbot (1983) using data collected from the road section from Lochailort to Loch Eil and from Loch Sunart outlined the deformation history of both the microdiorites and the adjacent Moine country rocks. Many of his field observations agree with those given below but some, together with his regional correlations and extrapolations, are at odds with the field evidence outlined herein.

8.2b) Microdiorite sheet geometry.

Microdiorites occur as parallel sided sheet intrusions from 5 cm. to 2 metres thick which dip variably to the ESE or WNW (Fig.86). Co-planar multiple intrusions are quite common but there are very few examples of sheets cutting each other and no evidence that the steeper sheets consistently cut the shallower ones or vice versa (cf. Talbot,1983.p142). There

Figure 86.



Stereonets to show the orientation of microdiorite sheets.

are no obvious geochemical differences between steeply or shallowly dipping sheets nor any geochemical variation which can be related to geographical position. Geochemical analysis strongly suggests a single consanguinous origin for all of the microdiorites which have been analysed (Fig.83).

The A 830 road section provides a well exposed E-W traverse across all of the Glenfinnan Division and part of the adjacent Morar and Loch Eil Divisions. Along the road section regional changes of microdiorite sheet orientation have been observed (Map 7) and plotted stereographically (Fig. 86a). From the western end of the traverse at Loch Eilt to Glenfinnan village, the sheets have a range of dips up to 75° either eastwards or westwards. Relatively flat lying sheets are quite common, many of these are gently folded, having vertical NNE-SSW trending axial planes and sub-horizontal hinge lines. Eastwards from Glenfinnan village the dip of the sheets tend to be lower and predominantly eastwards (dip direction is approx. 110°). Many of these sheets seem to be co-planar with a well developed set of easterly dipping joint planes. There is , however, no conjugate westerly dipping set of joints into which some of the other sheets could have been intruded. Figure 86b,taken from Smith (1979) shows the orientation of microdiorite sheets north of the Great Glen fault.

Map 7 shows that the maximum amount of dip decreases progressively eastwards across the section. It must be emphasised that there is no relationship (antithetic or otherwise) between the sheet orientation and the position of F_3 major folds (compare Maps 3,4 & 7) and no geometrical evidence to support the idea of "counter folds" (Talbot,1983.p.143). There are both easterly and westerly dipping sheets on both limbs of all the major F_3 folds west of Glenfinnan village (see Map 7).

8.2c) Microdiorite internal geometry.

Most of the sheets exposed along the A 830 road are schistose, the schistosity being oblique and usually steeper than the sheet margins. The schistosity displays variable intensity across many of the sheets, usually it is more intensely developed near the margins and weaker or not developed in the centre. However there are some sheets containing schistose layers gradational into non-schistose layers all of which are co-planar to the sheet margins.

The steeply dipping sheets contain a schistosity which is always steeper than the sheet margins. In thin sheets the schistosity is generally planar and dips approximately 5-10° more steeply than the sheet margins. Near the margins of the thicker sheets the schistosity is frequently sigmoidal, being nearly co-planar with, but steeper than the margins. In the centre of the sheets the schistosity is more highly oblique (see Fig.87a & c), indeed, where the sheets have very steep dips the internal schistosity can be so steep as to pass through vertical to dip in the opposite direction (see Map 7, western side).

Although this is easily the most common type of sigmoidal geometry in the steeply dipping sheets, another geometry also occurs. There are sheets where the schistosity, which is steeper than the sheet margins, is strongly oblique at both margins with the obliquity decreasing as the intensity of the schistosity decreases towards the centre of the sheet (Fig.87b; exp.1092, samples 228(1) to 229(4)). Unlike the much more common geometry where the oblique schistosity is asymptotic to the sheet margins. This geometry has not yet been recorded by Talbot (1983). No examples have been observed where the extremely steep or vertical sheets have an internal schistosity which is co-planar with the sheet margins.

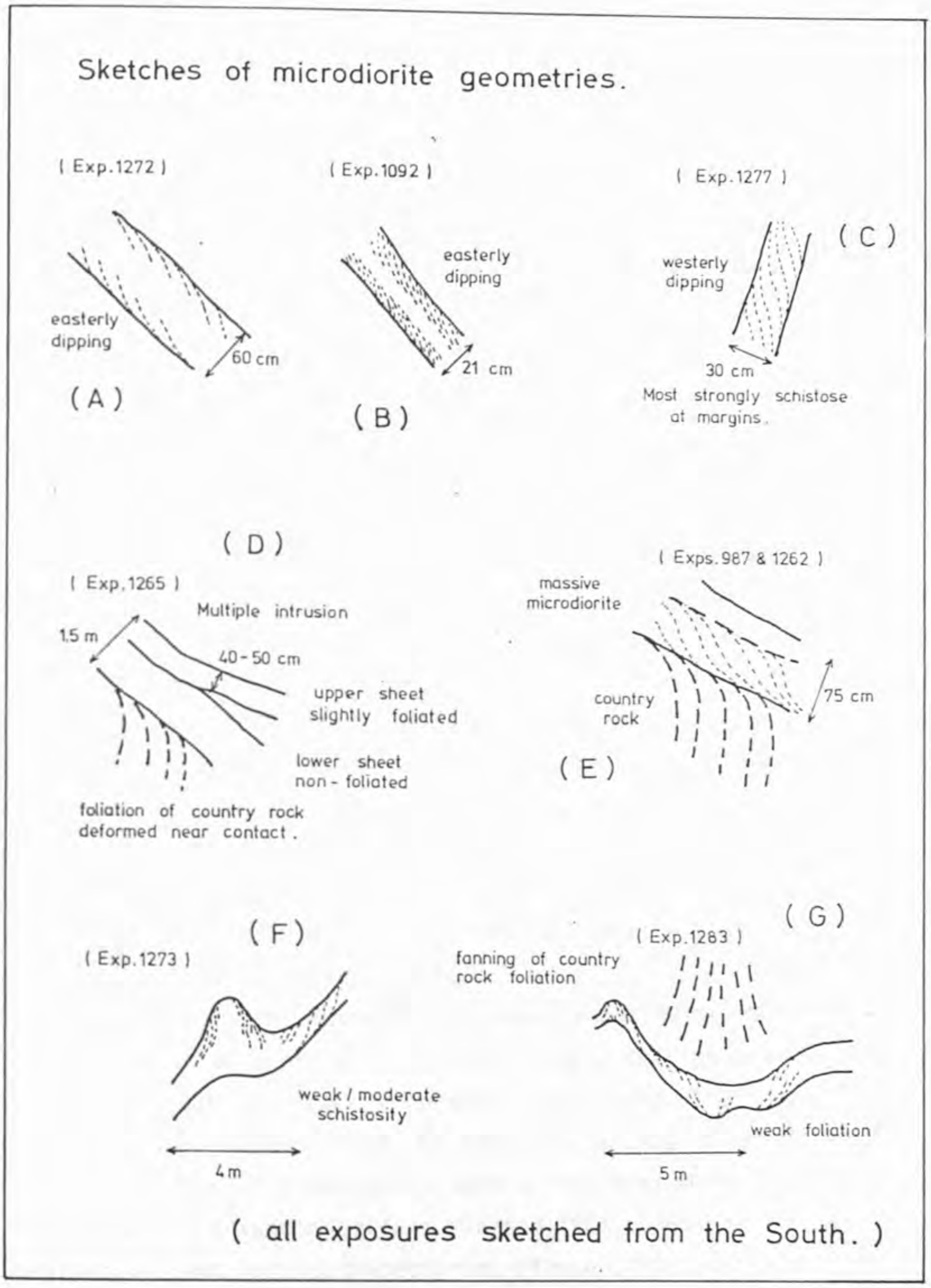
In the less steeply dipping and sub-horizontal sheets the internal schistosity is usually about 10-20° steeper than the sheet margins (see Map 7). Only very rarely do the less steeply dipping sheets contain a weak sub-vertical internal schistosity. Shallowly dipping sheets are quite often non-schistose.

Some of the less steeply dipping microdiorites are folded, having vertical NNE-SSW trending axial planes and sub-horizontal hinge lines. In the hinge zones of these folds the internal schistosity is axial planar and sub-vertical. On the limbs of the folds the schistosity is steeper than the sheet margins and usually sigmoidally asymptotic to them (Fig.87f & g). The schistosity is symmetrical about the fold axial plane.

There are no examples where the folds of the microdiorites fold the oblique, often sigmoidal schistosity. Two examples have been noted where the steep internal schistosity is very sporadically kinked by small irregular crenulation folds with sub-horizontal axial planes (Exp.1285, NM 853818 and Exp.1274, NM 819824).

Recent road widening at the head of Loch Eil has revealed an unusual, exceedingly thin (?) microdiorite (Exp.1039, NM 971792), cited by Talbot (1981,p.140,text-fig.3c) which may be recumbently folded, however there are no signs of similar recumbent folds in the enclosing Loch Eil Division psammites. The intrusion, which is rich in muscovite, may not be a microdiorite (pers.comm. Dr.D.Powell).

Figure 87.



Sketches of microdiorite geometries.
(Exp.1272)
easterly dipping
60 cm
(A)
(Exp.1092)
easterly dipping
21 cm
(B)
(Exp.1277)
(C)
westerly dipping
30 cm
Most strongly schistose at margins.
(D)
(Exp.1265)
Multiple intrusion
1.5 m
40 - 50 cm
upper sheet slightly foliated
lower sheet non - foliated
foliation of country rock deformed near contact.
(Exps.987 & 1262)
massive microdiorite
country rock
75 cm
(E)
(F)
(Exp.1273)
weak / moderate schistosity
4 m
(G)
(Exp.1283)
fanning of country rock foliation
weak foliation
5 m
(all exposures sketched from the South.)

8.2d) Mechanism of deformation.

Johnson & Dalziel (1966) claimed that deformation and metamorphism of the microdiorites was not accompanied by deformation in the country rocks. However close examination of the microdiorites exposed along the A 830 road section reveals that at some exposures the country rock can be seen to have undergone deformation related to the intrusion and deformation of the adjacent microdiorites. Figures 87d & e show deformation of the country rock which could be the result of sinistral simple shear across the sheet. Direct measurement of the amount of any simple shear displacement which may have occurred was not possible because of the general absence of reliable markers in the country rocks. Figure 87g shows that the country rock foliation has been deformed during the folding of the microdiorite sheet. The microdiorite folds are probably not the product of a phase of buckle folding. If the lithological banding had been vertical prior to buckle folding of the microdiorites, then, distant from the intrusions one would not expect the lithological banding to have re-orientated since it would have been normal to the direction of maximum compression. However near to the actively buckling intrusions, as a result of the heterogeneous stress field, one would expect to observe rotation of lithological banding. The vast majority of the intrusions do not show local re-orientations of the country rock lithological banding.

If the lithological banding had not been vertical prior to the active buckle folding, then after "active" buckling of the intrusions one would not expect to see mirror image lithological banding/sheet margin geometries at easterly and westerly dipping sheets.

These arguments tend to suggest that the microdiorite deformation was not simply the result of a phase of "active" buckle folding.

It is tempting,but probably erroneous, to equate the convergent fan of foliation in the country rock in the core of the microdiorite folds and the divergent fan of schistosity within the microdiorite with either a Class 1b/Class 3 dip-isogon pattern (Ramsay,1967.p.365) or with a pattern of divergent/convergent strain trajectories (Ramsay,1967.p.405).

However, observation of the geometry of the schistosity within the folded microdiorites (Fig.87f & g) leads to one very important conclusion. For whilst it is possible to argue that the schistosity pattern on both limbs of the folds illustrated and in all of the dipping planar sheets could be the product of simple shear, the observation that in the hinge of the microdiorite folds there exists an axial planar internal schistosity in an area where simple shear is not solely responsible for the generation of the schistosity within microdiorites (cf. Smith,1979.p.690) and that at least some of the fabric has been generated as a result of approximately horizontal E-W bulk shortening (ie. pure shear).

For the vast majority of sheets the geometry of the internal schistosity in relation to the sheet walls is explainable in terms of heterogeneous simple shear. Shear strain, concentrated at the sheet margins, produces a more intense schistosity which lies more nearly co-planar to the margins (eg. Fig.87a & c). However such a mechanism will not produce the geometry seen at Exp.1092 (Fig.87b). Here the more highly schistose margins of the microdiorite are more strongly oblique to the sheet walls than the less schistose interior of the sheet.

The symmetry of the internal schistosity seen in Figures 87f & 87g indicates that the geometry of the non-planar (folded) sheets is at least partly as a result of folding. It has already been argued that the deformation was not totally by a mechanism of active buckle folding. Therefore, the folds must have a component of horizontal, approximately E-W bulk shortening (ie. pure shear) which has passively rotated the limbs of the folds into their present orientation. To initiate and accentuate such folds the sheets cannot have been perfectly planar prior to the pure shear bulk shortening. The non-planar nature may be the result of initial intrusive perturbations or the result of slight buckling at the beginning of the bulk shortening, however because of the general lack of marginal re-orientation of the lithological banding it is assumed that any initial buckling which may have occurred must have been of very low amplitude.

The symmetrical disposition of the internal schistosity in the folded microdiorites (Fig.87f & g) suggests that the bulk strain that deformed the microdiorite suite was irrotational with horizontal, approximately E-W bulk shortening (Z axis of the strain ellipsoid).

The calculated line of intersection of the internal fanning cleavage planes around the fold hinges is sub-horizontal, trending NNE-SSW and is coincident with the hinge line of the microdiorite folds. This coincidence, it is argued below, suggests that the initial undulatory nature of the microdiorite sheet is a product of buckle folding rather than an original igneous phenomena. Buckle folding of any planar element will produce a fold whose hinge lies within the XY plane of the strain ellipsoid. Only if the plane originally lay in the YZ plane of the strain ellipsoid will the hinge line of the fold lie parallel to Y. It is interesting to consider the orientation of a fanning cleavage in relation to the folded surface in which it is contained. If the hinge line of the fold lies within the XY plane of the strain ellipsoid but not co-linear with the Y axis, one assumes that the cleavage fan is more likely to be geometrically related to the fold and its local stress field rather than the regionally applied stress field (see Fig.88). However folds with a cleavage which is not perfectly symmetrical to the fold are known (see Fig.44 and Evans,1963).

This line of argument leads to the conclusion that the cleavage fan around microdiorite fold hinges cannot be used to define the unique orientation of the regional strain ellipsoid but can be used to define the Z axis of the strain ellipsoid (and σ_1 in the case of irrotational strain).

Deformation of an originally undulatory sheet intrusion by pure shear bulk shortening may lead to the passive amplification of a perturbation of the sheet but the hinge line of the "fold" need not have any geometrical relationship to the XY plane of the strain ellipsoid. Where the "fold" hinge line lies relatively close to the orientation of the XY plane of the regional strain ellipsoid, the layer competency contrasts may refract stress, but in the majority of cases, where the hinge line of the "fold" lies well away from the XY plane of the strain ellipsoid one would not expect cleavage orientation to be related to the orientation of the "fold". Furthermore, if the "folds" are the amplification of original igneous perturbations then one might expect a series of such folds to have random orientations.

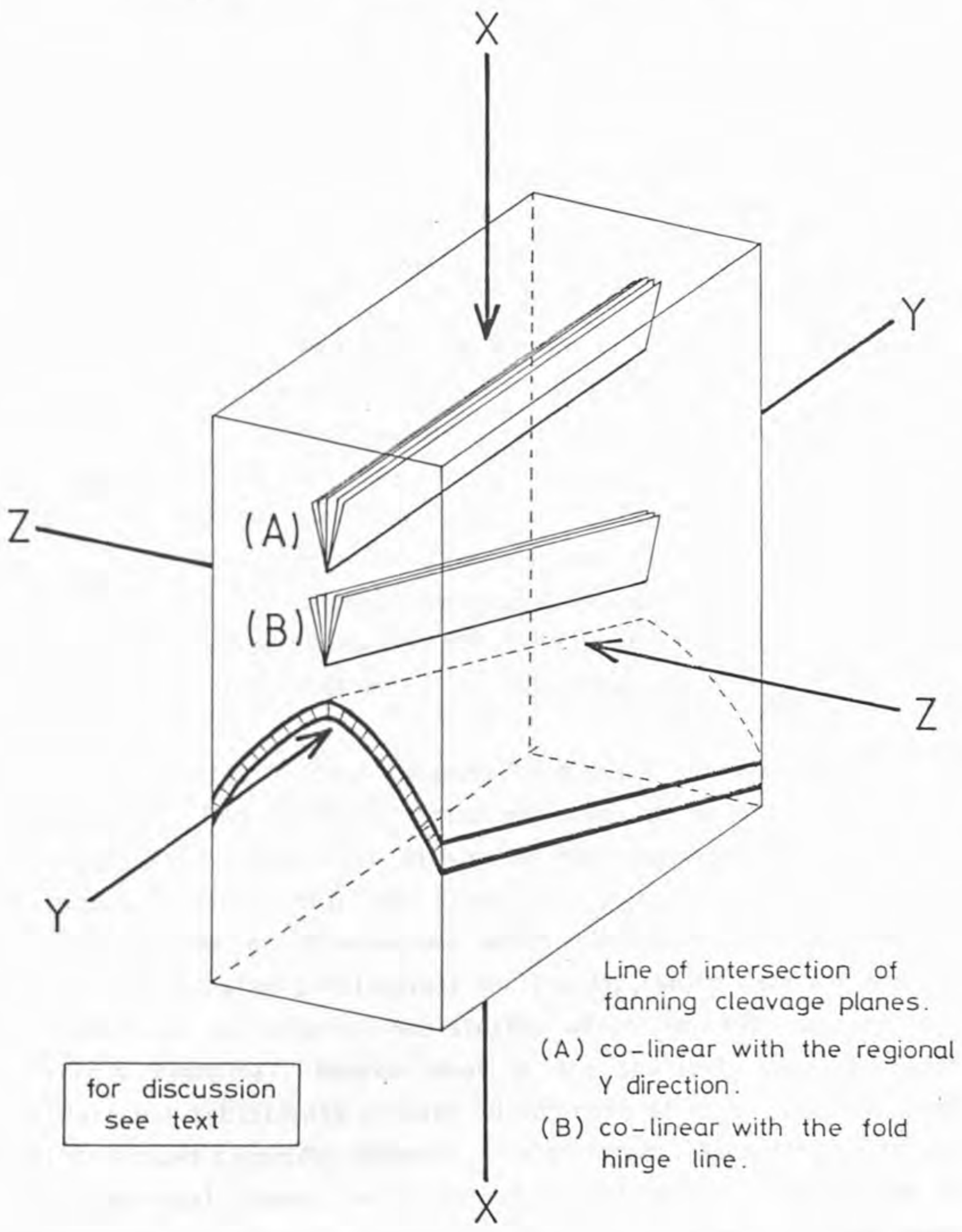
Thus general considerations of the mechanism of deformation of the microdiorites and the enclosing country rocks suggest deformation as a result of horizontally directed pure shear compression which may, possibly, have been preceeded by very low amplitude buckle folding of the more flat lying microdiorite sheets. Likewise it seems that, on a regional scale, deformation was not the result of simple shear although simple shear across individual microdiorite sheets may have occured during rotation of the sheets as a result of horizontal E-W pure shear compression.

It has proved impossible to define more precisely an exact mechanism of deformation of the microdiorite sheets since a number of parameters remain unknown. This problem is considered below.

It has been observed that the most steeply dipping sheets occur in the west of the area and maximum sheet dip progressively decreases eastwards. This may be an indication that more horizontally directed pure shear compression with associated passive rotation of planes towards vertical has occured in the west of the area, alternatively it may be a reflection of the initial orientation of the sheets. It is argued elsewhere (Chapter 7.5) that the regional "steep belt" was produced by intense D_4 deformation in the form of horizontally E-W directed compression which was more intense in the west of the area than in the east. Since the predominant deformation

Figure 88.

Sketch to show the possible relationships between fanning cleavage planes and fold hinge lines.



Diagrammatic sketch to show possible relationships between folded microdiorite sheets and their associated internal cleavage fans of the microdiorites also occured during D_4 (see Chapter 8.2e) it appears that the regional variation in microdiorite sheet dip reflects the intensity of their deformation and passive rotation, rather than their initial orientations.

It is not known how strongly the present geometry of schistosity in both folded and planar, dipping sheets reflects its original form and how much of its present geometry is as a result of subsequent deformation. likewise it is not known whether the schistosity formed rapidly during a relatively small increment of deformation and subsequently behaved as a passive marker or whether the schistosity formed incrementally along with increments of deformation and rotation of the sheets, much in the same way as schistose shear zones are believed to form (Ramsay & Graham, 1970).

Talbot (1983) presents a mechanism of deformation of the microdiorite sheets in which he believes that the internal schistosity initially forms in an orientation normal to the Z axis of the irrotational strain ellipsoid (Fig.89a). It is only with subsequent rotation of the sheets as a result of continued horizontal compression that the schistosity is deformed into its sigmoidal shape (Fig.89b). He argues that, for example, in a sheet which is rotating anticlockwise, opposing points on the walls are displaced antithetically (clockwise) and as a result the internal schistosity is distorted into S-shaped sigmoids. He argues that the poles to the sheet walls will rotate and move along a structural movement path towards Z and poles to different planar elements of the sigmoidal schistosity within the sheets will lie along what he calls the "apparent structural movement path (ASMOP)".

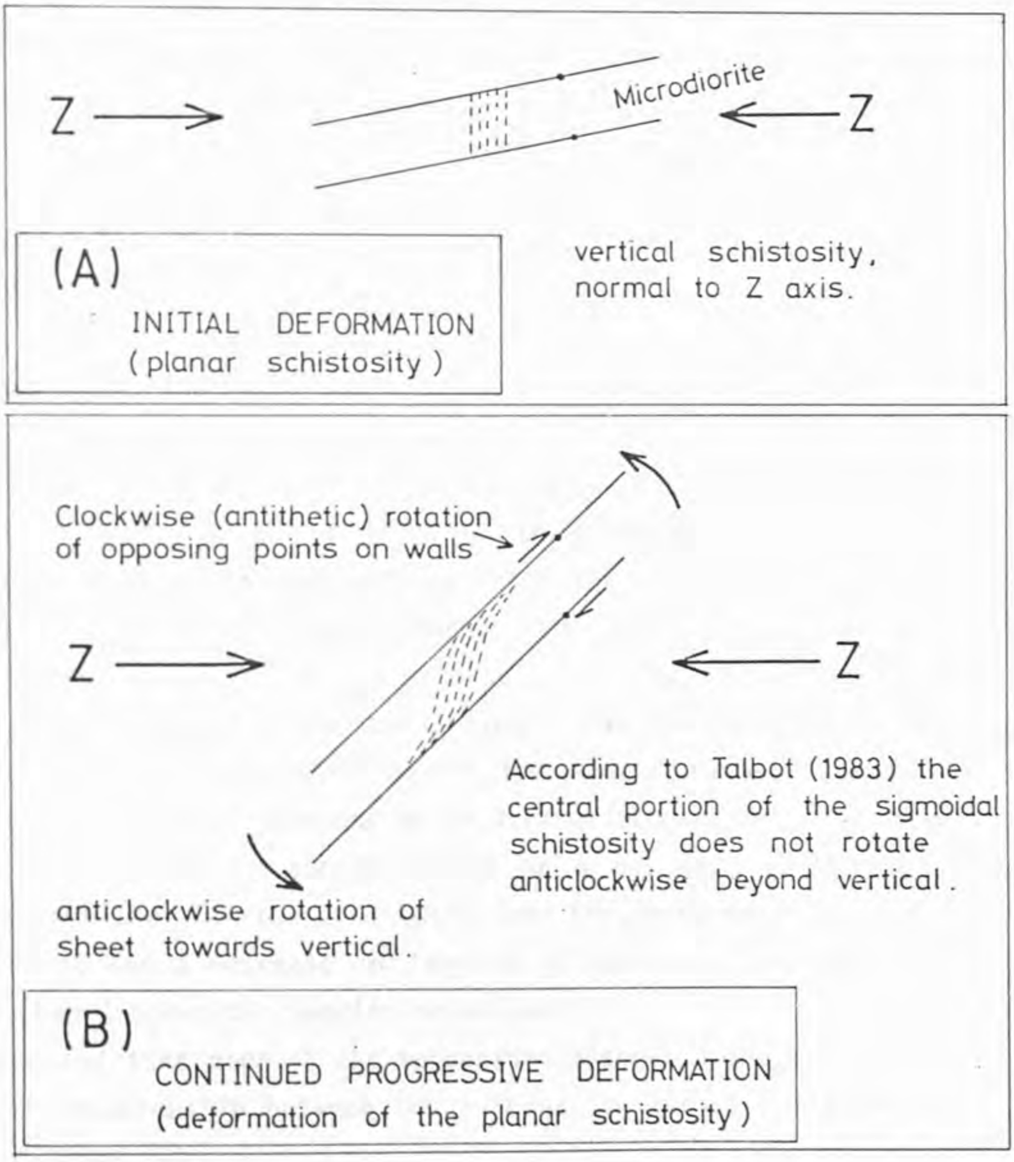
The internal schistosity pole furthest from the sheet margins will lie closest to Z or ZY. Thus on this argument all poles to the internal schistosity and sheet walls should lie within the same quadrant of the strain ellipsoid and never cross a principal plane.

There are a series of observations which contradict the theories of Talbot (1983). Sheets with a relatively shallow dip, which have not rotated much should possess an internal schistosity which is still sub-vertical and is strongly sigmoidal. However most of the shallowly dipping sheets contain an internal schistosity dipping 10-20° more steeply than the sheet walls. Well developed sigmoidal geometry is also common. According to Talbot (1983), near vertical sheets, originally lying within or near to the XY plane should have an internal schistosity which is co-planar to the sheet margins. This geometry has not been observed.

According to the theory of Talbot (1983) and taking the specific example of a vertical XY plane, the internal schistosity in steeply dipping sheets

Figure 89.

Sketches to show interpretations proposed by Talbot (1983). (See text for discussion).



should not pass through vertical to dip in the opposite direction to the sheet margins (Fig.89b). Such a geometry would have to be the result of rotation of the principal axes of the strain ellipsoid. However, adjacent very steeply dipping sheets (both easterly and westerly) often contain an oblique sigmoidal internal schistosity which dips steeply in the opposite direction (see Map 7 and Fig.87c), implying on the one hand a clockwise rotation of the principal axes of the strain ellipsoid and on the other hand an anticlockwise rotation: clearly indicating that the model proposed by Talbot (1983) does not hold true.

In the hinge areas of the microdiorite folds where limb rotation and simple shear due to interlayer slip can be discounted, the pattern of internal schistosity (Figs.87f & g) indicates that the cleavage did not form perfectly co-planar to the fold axial planes (Class 2 fold, Ramsay,1967). The fan of internal schistosity indicates a heterogeneous strain pattern and it is possible that the sigmoidal pattern of internal schistosity typical of microdiorite fold limbs and virtually all of the planar dipping sheets was produced during the formation of the schistosity (ie. cleavage refraction) and is not the result of modification of an originally planar schistosity.

If after formation, the limbs had then passively rotated towards vertical as a result of continued horizontal compression then one would expect the angular relationships between the sigmoidal internal schistosity and the sheet walls to be retained. As the dip of the sheet increased one would expect the internal sigmoidal schistosity which formed initially in a sub-vertical orientation, to be rotated through vertical to dip steeply in the opposite direction. This is commonly seen (see Map 7, western side). In the near vertical sheets one would expect the horizontally directed compression to produce layer thinning of the sheets, especially as the sheets have acted as the incompetent units during deformation. The thinning or flattening of the sheets should result in a descease in the angular discordance between the internal schistosity and the sheet margins. Evidence of such flattening and a decrease in "angular discordance" has not however been observed (Map 7 shows the angular relationships).

It must be noted that none of the mechanisms discussed above can explain the geometrical relationship between the internal schistosity and the sheet walls shown in Figure 87b.

From the discussion above it is obvious that the mechanism of deformation proposed by Talbot (1982 & 1983) does not fit a large number of field observations, and whilst it is difficult to propose a detailed mechanism of deformatiom which fits all the observations it would appear that a mechanism involving horizontal E-W directed pure shear compression of variably dipping incompetent microdiorite sheets in which local heterogeneous strain in and adjacent to the microdiorite sheets produced a sigmoidal internal schistosity subsequently modified during (?) passive rotation associated with the E-W directed pure shear compression.

8.2e) Age of intrusion and deformation of the microdiorites.

No absolute isotopic age determinations have been reported for any microdiorite sheets. Smith (1979,p.688) notes that "their age in relation to major granitic intrusions is unequivocal, they freely cut the Cluanie granite [417 Ma. U-Pb zircon, Pidgeon & Aftalion,1978] but are cut by veins extending from the Strontian complex [435±10 Ma. U-Pb zircon, Pigeon & Aftalion,1978 and 421±10 Ma. and 407±18 Ma. K-Ar biotite, Miller & Brown, 1965] as well as occurring as inclusions within the granite. No members of the microdiorite suite intersect the Ross of Mull granite [414±3 Ma. Rb-Sr whole rock, Halliday et al.1979] but several examples of microdiorites cutting the adjacent Moine schists and hornfelsed along with them are known". Microdiorites frequently cut "late" pegmatites (450±10 Ma. Rb-Sr muscovite and U-Pb zircon, van Breemen et al.1974).

There is no structural evidence to suggest that there are several generations of microdiorite intrusions which have different structural histories. All the lines of evidence point towards intrusion after D_3 deformation and followed by deformation during D_4. If the microdiorites had been intruded prior to D_3, then one would expect to see a change of sheet and internal schistosity orientation across the axes of the major F_3 folds and changes of metamorphic grade and intensity of deformation in microdiorites across the D_3 Sgurr Beag slide, none of which was seen. Smith (1979 text-fig.3) shows that the distribution pattern of microdiorites is not displaced by movement on the D_3 Sgurr Beag slide.

Brewer et al.(1979) reported Rb-Sr ages of 467±20 Ma. to 413±17 Ma. from pelites in the Lochailort to Glenfinnan area and related these ages to cooling of the Moine rocks, after Caledonian metamorphism, by a process of sequential uplift along discrete zones (slides?) beginning in the west. The D_3 Sgurr Beag slide is one, if not the major, of these discrete slide zones. The Glen Dessary syenite, intruded at 456±5 Ma. (van Breemen et al. 1979) was intruded prior to D_3 deformation (Roberts et al.1984 and Baird,1985). Thus indirect isotopic age determinations and structural correlations place the post-D_3 age of intrusion and subsequent D_4 deformation of the microdiorites as late Caledonian. There is no isotopic or structural support for a 750 Ma. ("Morarian") age for any of the microdiorites (cf. Talbot,1983).

The orientation of the D_3 strain ellipsoid changes progressively across the area. In the west, the Z axis of the regional strain ellipsoid is coincident with the Z axis of the microdiorite strain ellipsoid, but while the Z axis of the microdiorite strain ellipsoid remains constant across the area, the orientation of the D_3 strain ellipsoid changes progressively (Fig.90 and Schematic Section 1), again indicating that the deformation of the microdiorites occurred after D_3 regional deformation.

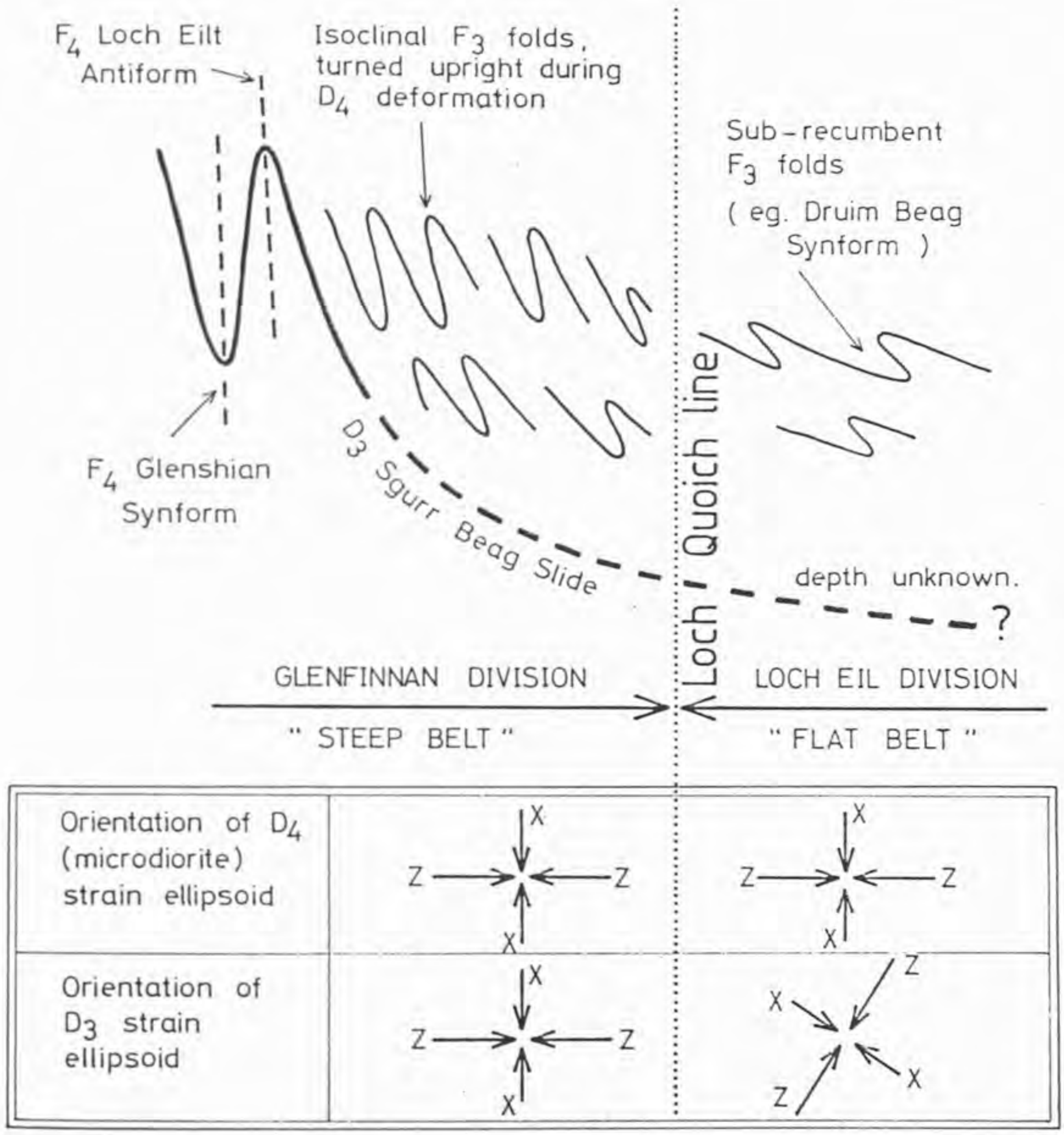
The Z axis of the D_4 strain ellipsoid and the microdiorite deformation strain ellipsoid are coincident throughout the area (Fig.90). It has been argued (Chapter 7.5) that D_4 deformation has produced the Glenfinnan Division "steep belt" and that this deformation was much less intense further eastwards in the Loch Eil Division "flat belt". The increase in D_4 strain westwards coincides with the occurrence of more steeply dipping microdiorites.

If most of the microdiorites had been intruded into a post-D_3/pre-D_4 joint set which dipped shallowly to the ESE in pre-D_4 times, then ESE-WNW directed D_4 compression which was more intense within the Glenfinnan Division "steep belt" would have been responsible for the rotation and increase of dip of the sheets with more rotation in the "steep belt". Sheets not intruded into the joints and dipping westwards would have similarly steepened in dip during D_4 deformation, especially within the "steep belt". Flat sheets would have responded to horizontally directed compression by folding and developing near vertical fold axial planes. In this regard D_4 deformation has produced the major F_4 Loch Eilt antiform and the F_4 Glenshian synform (see Chapter 7). Around these major folds, depending on the mechanism of folding, one would expect to see variations of microdiorite sheet and joint set orientations. In the area under consideration, ie. on the eastern limb of the F_4 Loch Eilt antiform, horizontally directed compression has produced a fairly wide range of sheet orientations. It is presumed that on the western limb of the tight to isoclinal F_4 Loch Eilt antiform similar amounts of horizontally directed compression has produced an equally large range of sheet orientations. West of the axial plane of the F_4 Loch Eilt antiform microdiorite sheets dip to the east at 50-60° (Grid ref. NM 787832) and have typically sigmoidal internal schistosity (Dr.D.Powell pers.comm.). Whilst the range of sheet geometries in this fold limb is not known, it is noted that the geometry described is consistent with pre-D_4 intrusion and subsequent D_4 deformation involving horizontal, WNW-ESE directed pure shear compression.

On a regional scale, the areas to the west of the axial plane of the F_4 Glenshian synform and to the east of the Loch Eilt antiform share the same long limb geometry. It is only in the area between these two major F_4 fold axial planes that the F_4 short limb geometry occurs and it is only with information collected in this area that a more detailed investigation of the mechanism of F_4 folding can be made. It is assumed that if the joint set observed in the area is of post-D_3/pre-D_4 age that this will show a

Figure 90.

See text for discussion.



Sketches to show the formation of the D_4 "steep belt" and the orientations of the D_3 and D_4 strain ellipsoids across the area.

change of orientation on either side of the major F_4 fold axial planes.

The occurrence of a post-D_3/pre-D_4 joint set intruded by a set of pre-D_4 igneous sheets, some of which still preserve good igneous textures implies a postD_3/pre-D_4 period during which the country rocks cooled, were uplifted, and behaved in a brittle manner before the onset of D_4 deformation and metamorphism of the microdiorite sheets under more ductile and metamorphic conditions.

Thus the structural development of the D_3 Sgurr Beag slide and its deformation into its present upright (D_4) attitude cannot be considered to be part of a continuous process of progressive deformation.

It can also be noted that the easterly dipping post-D_3/pre-D_4 joint set has a similar orientation to the plane of the Moine thrust, a fact which may or may not have chronological significance.

8.3) Amphibolites.

There are two groups of amphibolites in the area (see Chapter 5). Both garnetiferous amphibolites and hornblende schists are found in the Glenfinnan Division but only hornblende schists have been found in the Loch Eil Division. Both groups contain a strong L/S fabric which is co-planar with the dominant early foliation (S_1 or S_2 ?). Both groups occur within the Ardgour granitic gneiss but it has not been possible to ascertain their relative ages or even whether they were intruded pre-D_1 and pre-migmatisation, or post-D_1/pre-D_2 and subsequently deformed and metamorphosed during the re-working of the migmatitic gneiss. Neither group was found in the Morar Division — a feature held to be distinctive of the Morar Division in the SW Northern Highlands (Powell,1974).

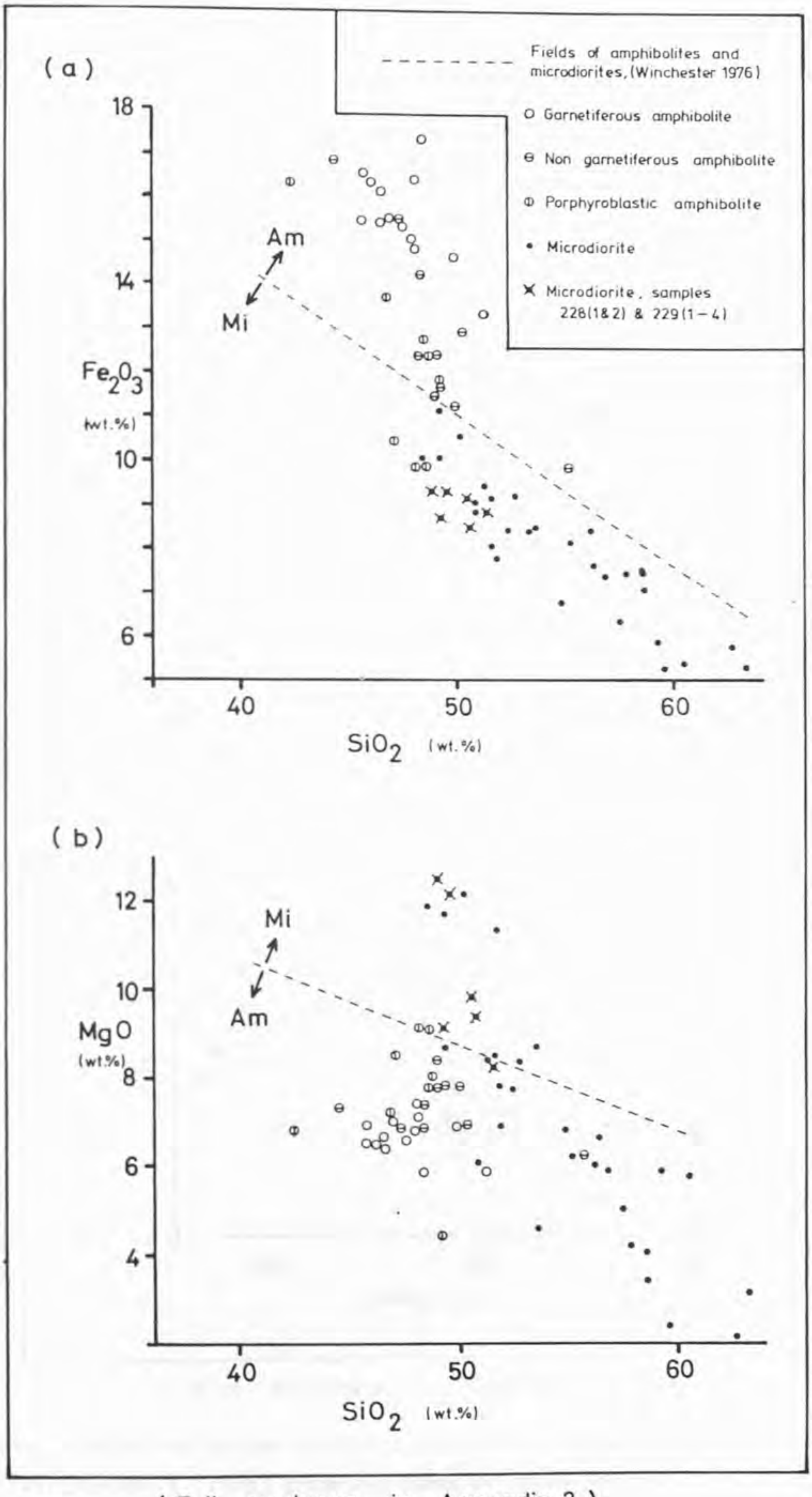
The garnetiferous amphibolites occur as small isolated boudins which cannot be proved to be intrusive. A sedimentary origin has been suggested (Johnstone et al.1969) however Winchester (1976), on geochemical grounds, favoured an igneous origin as a phase of tholeiitic magmatism. Further north in the Fannich area of Ross-shire, Winchester has recorded thin sheets of amphibolite in rocks correlated with both the Morar and Glenfinnan Divisions of the SW Northern Highlands.

Winchester uses plots of Fe_2O_3 (total); MgO; TiO_2; and p.p.m. Y against SiO_2 to distinguish between microdiorites (which he termed meta-appinites) and amphibolites. Figure 91 is plotted using data from rocks collected in the thesis mapping area (Figs.83 & 92) and incorporates Winchester's boundary lines between amphibolites and microdiorites.

Examination of Figure 91 shows that the local data is in broad agreement with that of Winchester (1976) however Y (p.p.m.) plotted against % SiO_2. Winchester has used a plot of the ratio of Nb/Y against the ratio of Zr/P_2O_5 to distinguish between amphibolites from metasediments correlated with the Morar and Glenfinnan Divisions in western Ross-shire. All of the thesis area amphibolites plot within Winchesters' "Glenfinnan tholeiitic area" (Fig.93).

In Figures 91 & 93 the Glenfinnan area amphibolites have been sub-divided into garnetiferous amphibolites. hornblende schists and porphyroblastic hornblende schists. It is clear from these graphs and geochemical tables that the different types of "amphibolite" cannot be distinguished geochemically although the garnetiferous amphibolite textures and slightly different metamorphic mineralogy may be a reflection of original igneous differences such as grain size and phyric/aphyric nature or they may be a result

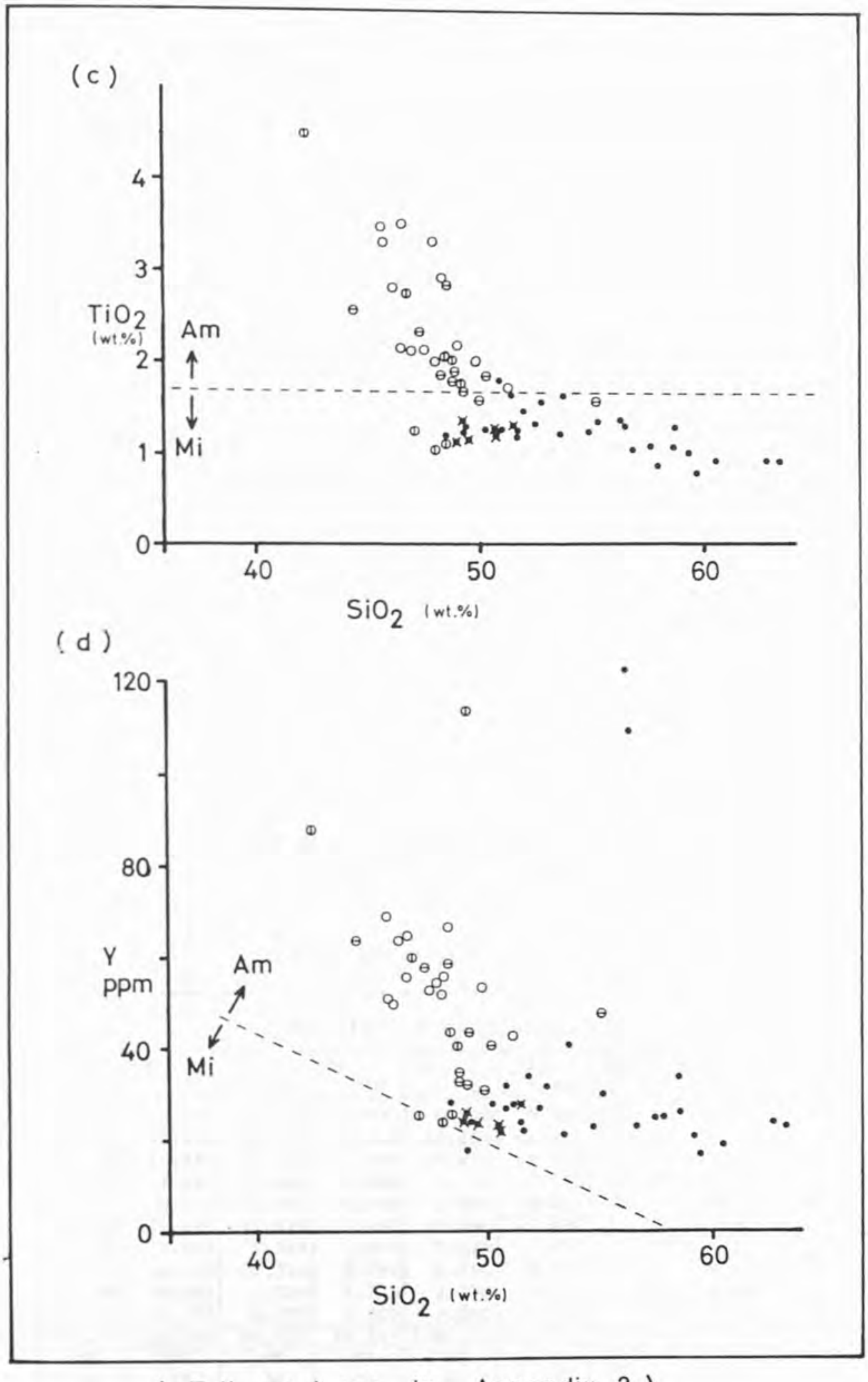
Figure 91.



(Full analyses in Appendix 2)

Graphs plotted using the author's data.
Compositional fields from Winchester, 1976.

Figure 91.
(continued).



(Full analyses in Appendix 2)

Graphs plotted using the author's data. Included on the graphs are Winchester's (1976) proposed compositional fields.

Figure 92.

Geochemical analyses of Amphibolite samples.

Garnetiferous Amphibolites

	4/6	41/230	67/235	143/654	147/669	152/721	159/740	165/817
P	0.445	0.320	0.370	0.235	0.220	0.215	0.405	0.330
Si	46.480	48.335	47.920	48.035	46.460	49.865	51.240	46.130
Al	13.600	13.660	14.080	13.880	13.840	13.770	14.255	13 640
Mg	6.460	5.900	6.840	7.135	6.705	6.960	5.920	6.545
Mn	0.255	0.295	0.230	0.240	0.260	0.230	0.215	0.260
Fe	15.470	17.360	15.010	16.420	16.185	14.635	13.360	16.340
Ti	3.495	2.880	3.280	2.165	2.105	1.965	1.685	2.785
Ca	8.550	8.485	9.970	9.705	10.670	9.245	10.060	10.840
K	3.090	2.380	2.190	0.600	0.860	1.445	0.995	0.695
Na	0.885	0.690	1.110	1.720	1.390	0.790	1.240	1.425
Total	98.745	100.400	100.990	100.140	98.700	99.150	99.425	98.980
Zr	231	177	178	123	134	106	99	179
Y	65	67	55	56	56	54	43	64
Rb	157	104	91	19	10	49	47	8
Nb	9	5	5	3	5	1	4	3
Sr	34	32	39	116	135	87	221	90
Th	0	1	0	3	1	1	4	5

	Gt. Amph.					Hb. Sch.		
	166/843	169/900	178/971	184/1006	209/1082	75/243	172/952	221/1089
P	0.195	0.200	0.435	0.270	0.250	0.220	0.260	0.275
Si	45.740	46.905	45.615	43.015	47.530	55.170	47.270	44.370
Al	13.370	14.120	14.890	14.305	13.775	15.160	13.660	13.895
Mg	6.975	7.070	6.590	7.480	6.620	6.310	6.905	7.350
Mn	0.275	0.265	0.250	0.240	0.335	0.170	0.245	0.270
Fe	16.580	15.500	15.495	14.835	15.345	9.880	15.515	16.855
Ti	3.285	2.080	3.445	1.965	2.105	1.540	2.310	2.515
Ca	10.950	9.740	11.035	9.840	10.370	8.650	10.055	9.420
K	0.895	1.340	0.645	1.215	1.230	1.170	1.220	2.135
Na	1.435	1.480	0.820	0.815	1.675	2.890	1.135	1.280
Total	99.700	98.695	99.215	98.990	99.230	101.160	98.570	96.355
Zr	125	120	266	125	126	180	136	141
Y	51	50	69	52	53	48	58	64
Rb	17	43	12	33	27	38	27	121
Nb	3	8	1	3	3	4	0	3
Sr	72	141	114	137	152	165	88	196
Th	0	1	3	0	1	5	3	3

Hornblende Schists

	20/118	22/138	23/143	54/232	56/232	73/240	108/470	127/549
P	0.175	0.335	0.180	0.235	0.200	0.180	0.140	0.115
Si	49.955	4 .290	48.860	48.310	50.260	48.915	49.220	49.810
Al	14.540	14.665	16.125	14.890	15.250	14.825	14.760	15.360
Mg	7.850	7.445	8.460	6.900	6.980	7.870	7.895	9.465
Mn	0.195	0.235	0.190	0.195	0.200	0.215	0.230	0.200
Fe	11.280	14.240	11.475	12.420	12.990	12.420	11.760	10.305
Ti	1.540	2.835	1.745	1.830	1.820	1.850	1.640	1.155
Ca	9.910	10.440	10.310	8.250	6.490	9.710	11.920	9.610
K	1.350	0.985	1.110	2.725	3.730	2.555	0.915	0.990
Na	2.440	1.175	0.905	2.405	2.200	2.425	1.995	2.660
Total	99.250	100.650	99.375	98.165	100.100	100.970	100.465	99.670
Zr	98	205	96	96	101	75	98	57
Y	31	59	35	44	41	33	22	25
Rb	48	15	36	169	260	115	15	47
Nb	1	5	3	9	11	3	3	1
Sr	116	122	65	80	75	318	321	315
Th	1	1	0	4	1	3	0	4

Major elements: See note on the following page.

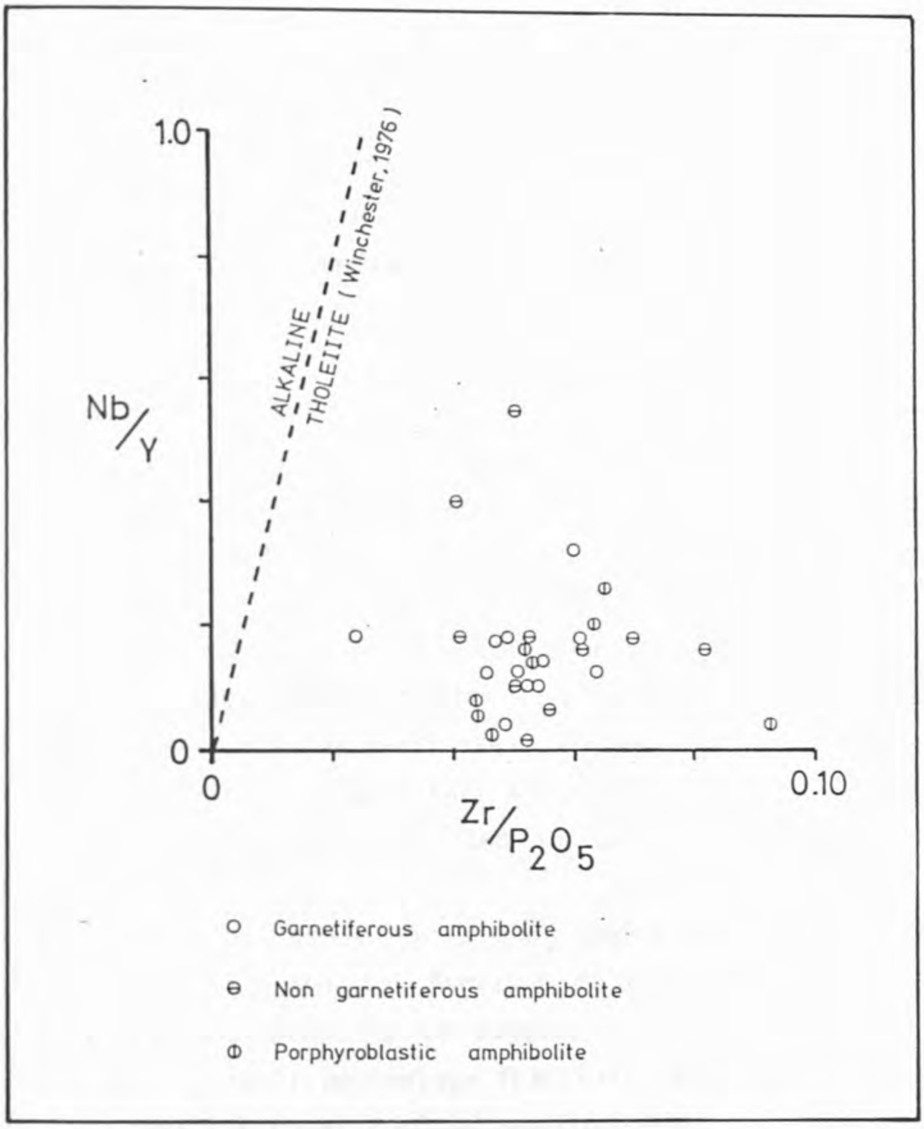
Figure 92.
(continued).

Geochemical analyses of Amphibolite samples.

	← Porphyroblastic Hornblende Schists →							
	28/174	88/281	89/293	92/333	94/346	124/543	128/549	132/581
P	0.580	0.340	0.130	0.200	0.110	0.130	0.130	0.220
Si	42.280	46.745	48.575	49.200	48.465	47.050	48.080	48.720
Al	14.375	14.155	16.790	14.800	14.385	16.700	16.785	14.910
Mg	6.850	7.230	9.165	4.485	7.815	8.585	9.200	8.080
Mn	0.325	0.240	0.175	0.205	0.240	0.210	0.170	0.220
Fe	16.340	13.720	9.890	11.815	12.745	10.455	9.895	12.390
Ti	4.415	2.710	1.085	1.685	2.010	1.225	1.065	1.995
Ca	10.040	10.510	11.480	12.025	10.695	12.145	10.895	10.485
K	2.655	1.855	0.790	0.645	1.150	0.980	0.700	1.190
Na	1.230	1.565	2.155	1.760	1.745	2.250	2.155	1.795
Total	99.100	99.070	100.225	100.825	99.365	99.735	99.070	100.710
Zr	378	216	57	94	104	69	57	119
Y	88	60	26	114	44	26	24	41
Rb	90	85	30	17	33	35	31	85
Nb	11	6	1	1	1	2	1	3
Sr	108	217	155	129	181	141	255	303
Th	0	0	1	3	1	1	0	3

Major element analyses (ie. P to Na) are listed as Wt. % oxides.
(P_2O_5, SiO_2, Al_2O_3, MgO, MnO, Fe_2O_3, TiO_2, CaO, K_2O, Na_2O).

Figure 93.



Graph of Nb/Y versus Zr/P_2O_5 for amphibolites collected from the Glenfinnan area.

of the different suites having undergone different metamorphic histories although this cannot be ascertained in the Glenfinnan area.

8.4) Camptonites and Dolerites.

Throughout the area there are a number of vertical basic dykes which trend mostly E-W with some trending from NW-SE through to NE-SW. All of these dykes have chilled margins and are undeformed. A small boss 250 metres in diameter (Exp.258/1139 NM 82958375) appears to be the feeder to an E-W trending dyke some 500 metres further east (Exp 257/1136 NM 83528390).

In thin section it is obvious that these dykes form two very distinct suites. There is a suite of dolerites and coarse basalts with large plagioclase, titan augite and olivine (sometimes serpentinised) phenocrysts. These rocks are part of a suite of Hebridean Tertiary Alkali Basalts (Morrison et al.1980). The other suite is finer grained, usually altered, and typified by the late stage growth of fine needle shaped, cinnamon coloured alkali amphiboles (Barkevikite/Kaersutite). These camptonite dykes are part of a Carboniferous-Permian suite (Gallagher,1963: Speight & Mitchell, 1979 and Baxter & Mitchell,1984). The two suites can be readily distinguished geochemically (Fig.94).

The camptonites have extremely low % SiO_2 and very high values of the trace elements Zr, Rb, Nb and Sr. Morrison et al.(1980) have used this abundance of incompatible elements to suggest that the camptonites have been produced by a very small percentage fractional melt of the underlying mantle. They claim that the removal of incompatible elements from the mantle source in Carboniferous-Permian times depleted the mantle so that subsequent partial melting in Tertiary times produced a melt which was depleted in incompatible elements. It is questionable if the mantle remains stable over such an interval and it is also questionable if the removal of such a small percentage of (camptonitic) material would noticably change the bulk chemistry of the mantle so as to affect subsequent much later partial melting. However it must be noted that whatever mechanisms of formation are discussed, the trace and major element geochemistry readily distinguishes between the two suites of dykes.

Figure 94.

Geochemical analyses of Camptonites and Dolerites.

← Permo-Carboniferous Camptonites →

	100/429	101/429	106/467	109/472	126/549	274/1182	275/1228	276/1228	145/656
P	0.605	0.785	0.745	0.710	1.260	0.920	0.585	0.600	0.950
Si	39.920	41.995	42.065	38.290	38.535	43.255	43.870	43.270	31.435
Al	11.655	13.815	12.350	12.020	11.855	12.875	14.080	14.220	13.635
Mg	7.670	6.420	9.590	7.840	11.240	9.730	7.820	8.110	6.160
Mn	0.170	0.150	0.170	0.170	0.200	0.185	0.170	0.220	0.160
Fe	13.255	13.730	11.200	12.160	11.495	12.060	10.780	10.880	18.655
Ti	2.940	3.135	2.645	2.880	2.655	2.945	2.550	2.495	2.105
Ca	11.590	10.530	11.475	13.660	14.730	12.475	10.620	11.030	11.385
K	1.985	2.680	3.285	2.340	1.870	0.745	2.015	1.890	2.405
Na	1.635	1.985	0.965	1.505	0.470	3.520	2.960	3.155	0.035
Total	91.430	95.220	94.490	91.570	94.300	98.710	95.455	95.885	86.345
Zr	250	322	338	291	306	288	275	293	141
Y	22	28	30	27	30	33	31	33	32
Rb	52	67	93	61	45	20	66	64	65
Nb	74	100	125	60	118	84	107	117	34
Sr	1254	998	1093	784	1984	641	1120	1105	120
Th	6	5	11	3	8	6	6	9	0

← Tertiary Dolerites →

	78/245	79/246	98/400	107/469	242/1094	257/1136	258/1139		96/357
P	0.240	0.180	0.130	0.330	0.420	0.245	0.155		0.760
Si	54.695	54.410	57.615	46.430	48.520	46.850	55.650		59.410
Al	12.350	15.835	15.030	13.760	15.300	14.045	14.965		14.890
Mg	9.325	5.965	6.930	7.120	7.535	9.330	5.120		5.745
Mn	0.165	0.140	0.150	0.240	0.145	0.210	0.140		0.090
Fe	12.135	10.115	9.795	10.270	10.315	12.150	10.160		5.595
Ti	1.840	1.255	1.115	1.570	2.095	1.760	1.295		0.995
Ca	8.065	10.135	7.350	13.620	10.595	10.705	8.520		3.975
K	0.740	0.310	1.080	0.390	1.035	0.440	0.625		3.385
Na	2.735	2.695	3.025	2.290	3.085	2.385	3.190		4.100
Total	102.685	101.020	100.210	96.030	99.045	98.120	99.830		98.940
Zr	102	62	72	94	133	87	79		269
Y	24	20	22	21	27	20	20		19
Rb	21	1	32	4	18	4	11		59
Nb	10	7	8	19	26	11	4		21
Sr	351	307	622	463	534	391	318		1392
Th	0	0	1	0	3	1	0		12

145/656 Highly altered (?) Camptonite

96/357 "Granitic" sheet

Major element analyses (ie. P to Na) are listed as Wt.% oxides.
(P_2O_5, SiO_2, Al_2O_3, MgO, MnO, Fe_2O_3, TiO_2, CaO, K_2O, Na_2O).

C H A P T E R 9.

Local structure and its regional and crustal context.

9.1) Synopsis of local structure.

This chapter presents a synthesis of the structural development of the local sub-areas described in detail in Chapter 5, the structural development of the Sgurr Beag slide (Chapter 7) and the structure of the microdiorite suite (Chapter 8). Subsequently, models of the crustal structure of the NW Scottish Highlands are discussed in context with the local structure.

Before discussing the observed structure it is worth noting that sedimentary structures are generally not found in the area, and where they were observed deformation is still relatively intense. The deformation has produced and then deformed highly planar fabrics and it is quite possible that early phases of deformation have not been observed or have been mis-correlated.

The earliest observed structures are the S_1 fabric and the possible F_1 minor folds in and adjacent to the Ardgour granitic gneiss. This fabric is a planar mica fabric, which in the Ardgour granitic gneiss contains co-planar pegmatites with biotite selvages. The planar migmatitic fabric dominant in the Druim na Saille pelite is also of this age. All indications are that this fabric has been produced during a phase of high grade metamorphism and deformation. However, no major F_1 folds have been mapped within the area. Minor F_1 folds have been described elsewhere (Powell,1974.table.2), but only in the extreme west of the Moine nappe have major F_1 folds been described. These are held to be responsible for the interleaving of the

Moine schists with the Lewisian basement (Ramsay,1958).

The planar penetrative mica fabric ubiquitous in the metasediments of the Glenfinnan Division could be an S_1 fabric. It cannot be related to any major fold in the area west of the Beinn an Tuim fault. It could, however, equally well be an S_2 fabric.

The D_2 phase of deformation has produced a pair of major folds in the Ardgour granitic gneiss and in the metasediments on Beinn an Tuim (Map 3). These major folds are isoclinal over most of their outcrop and show very complex minor fold patterns which cannot be explained by variations in the intensity of subsequent deformation. The F_2 Beinn an Tuim synform passes from a relatively open upward facing fold in the Loch Eil Division psammites in the east, tightening and becoming more steeply plunging along its axial plane towards the west. The hinge line of the fold, still opening towards the east, locally passes through vertical so that at its western end it has become an antiformal structure.

The F_2 Meall nan Damh fold on a regional scale, further south, is an antiform, the hinge of which plunges to the south. Traced northwards along its axial plane trace into the region under consideration, the hinge of the fold has increased its plunge to become reclined and further north, near the Beinn an Tuim fault locally the hinge line has gone beyond vertical so that the fold is locally a synform.

The D_2 deformation in the Ardgour granitic gneiss and associated metasediments is typified by the extremely strong development of a penetrative S_2 fabric which has transposed most of the bedding and early (S_1) fabric, often making the identification of F_2 folds difficult.

Contrasting with the major tight to isoclinal F_2 folds with their strongly developed axial planar fabrics in the Ardgour granitic gneiss and adjacent metasediments, no major F_2 folds with strongly developed axial planar penetrative fabrics have been mapped in the Loch Eil Division. The Beinn an Tuim synform in the Loch Eil Division is fairly open and does not contain a strong axial planar penetrative fabric: the planar mica fabric is, at least locally, of S_1 age and folded by F_2 folds. West of the Beinn an Tuim fault the Glenfinnan Division metasediments contain a strongly developed pre-D_3 axial planar penetrative mica fabric which in more pelitic lithologies is often migmatitic and which could be of S_2 age. If it is an S_1 fabric then west of the Beinn an Tuim fault there is no evidence for any re-working of the S_1 fabric by the D_2 deformation, a phase of deformation which is very intense east of the Beinn an Tuim fault. If it is an S_2 fabric, it presumably has re-worked the S_1 fabric so that it is no longer separable or distinguishable.

The D_3 phase of deformation has produced most of the major folds which occur in the area. These have axial planes which are vertical and trend generally NE-SW. The fold hinges plunge at variable angles towards the NE. Related minor folds are typified by associated tight crenulations of the pre-existing planar mica fabrics in pelitic and semi-pelitic layers.

The amount of D_3 deformation decreases eastwards through the Glenfinnan Division and into the Loch Eil Division. The Sgurr Beag slide is a D_3 ductile slide zone which was produced during very intense D_3 deformation. Eastwards from the slide the major F_3 folds change from isoclinal to tight folds, minor folds become more common, the amount of boudinage decreases and the plunge of minor F_3 fold hinges also decreases. In the extreme east, within the Loch Eil Division the Druim Beag synform is a tight to isoclinal, sub-recumbent F_3 major fold (Schematic Section 1).

The trace of the Sgurr Beag slide re-appears at the western end of Loch Eilt and again further west at Lochailort as a result of the slide being folded by a set of major tight to isoclinal F_4 folds (Powell et al,1981 text-fig.1; Baird,1982.text-fig.2 and Schematic Section 1). Some of the minor crenulation folds in the west of the area could relate to this D_4 phase of deformation, although it seems more likely that the same minor folds are F_3 in age (Chapter 5.6e).

It is interesting to consider the progressive development of folds during D_3. In the east of the area the hinge lines are sub-horizontal. Moving westwards their plunge increases until it is sub-vertical in the west of the Glenfinnan Division. If this is a result of progressive deformation with rotation of hinge lines towards the X axis of the D_3 strain ellipsoid as deformatiom proceeds then <u>at present</u> the extension direction (X) of the D_3 strain ellipsoid is vertical. Rotation and vertical extension on such a large scale (for example the Sgurr a Mhuidhe synform can be traced for at least 12 Km. to Glen Dessary in the NE (Roberts et al.1984. text-fig.1)) poses major problems in terms of tectonics and crustal thickness.

D_4 deformation has produced the F_4 Loch Eilt antiform which folds the D_3 Sgurr Beag slide and associated F_3 folds (see Chapter 7). The F_4 Loch Eilt antiform has a vertical NE-SW trending axial plane with a sub-horizontal hinge line. Some indication of the pre-D_4 geometry of the area can be obtained by extending the section horizontally in a NW-SE direction (ie. normal to the XY plane of the D_4 strain ellipsoid) assuming that there has been little, if any, rotation of the F_4 hinge lines during D_4 deformation. It could be argued that some or all of the rotation of F_3 hinge lines occured during D_4 deformation. If this was the case then rotation of F_3 hinge lines during D_4 deformation would have been towards vertical.

However, the hinge lines of folds formed during D_4 deformation have remained sub-horizontal and not rotated.

With F_4 folds removed, the Sgurr Beag slide and the axial planes of associated F_3 folds dip gently to the east. Further east, beyond the eastern limits of F_4 folding the F_3 Bruim Beag synform in the Loch Eil Division still retains its sub-recumbent easterly dipping form, however Strachan (1985) notes that, in general, within the Loch Eil Division, F_3 folds are fairly upright and open.

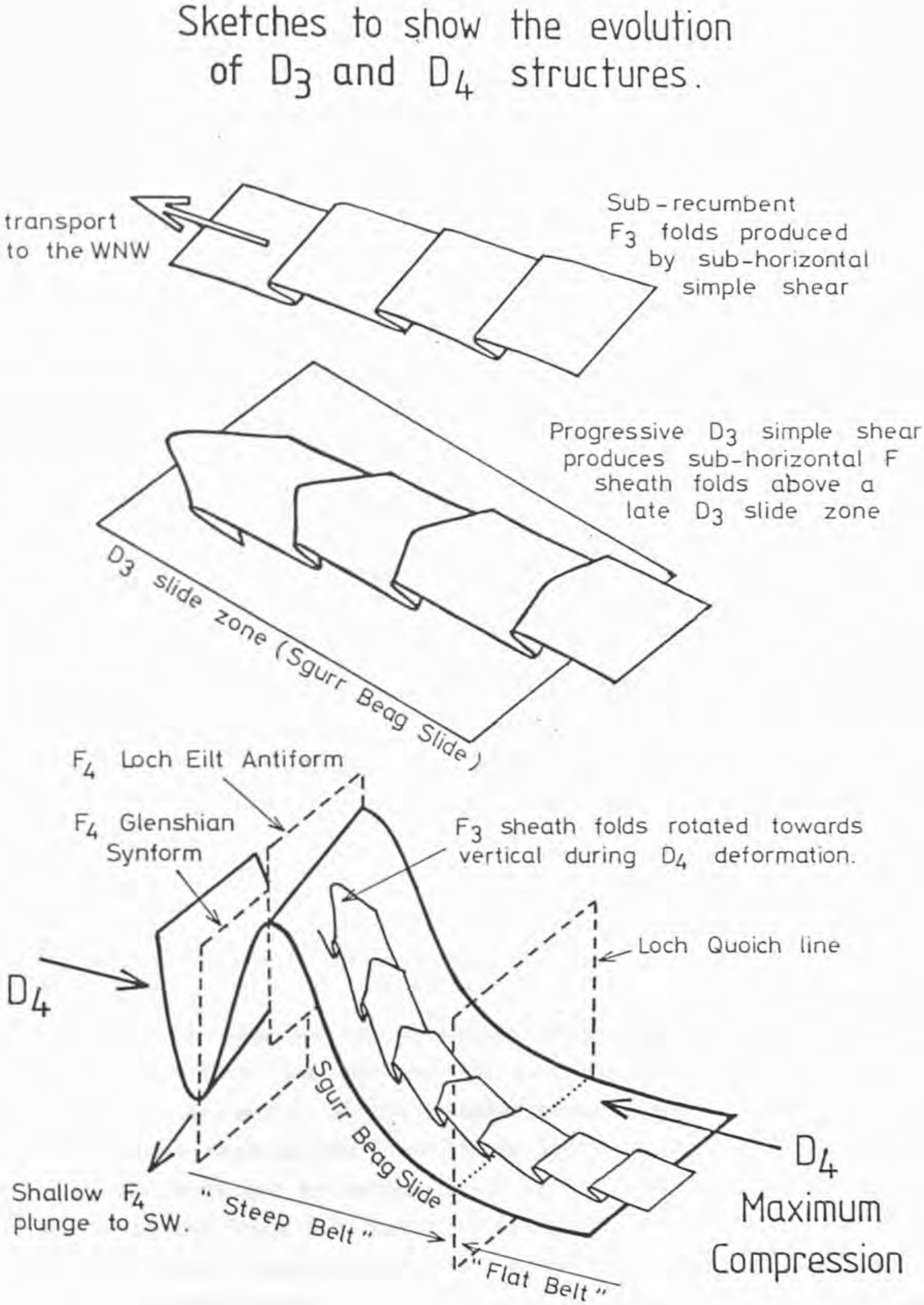
The D_3 extension direction, defined as the direction to which F_3 fold hinges rotate with progressive deformation, within the XY plane of the D_3 strain ellipsoid, prior to D_4 deformation lay within the shallowly dipping F_3 fold axial planes and trended shallowly upwards to the W or NW (Fig.95).

Such a stack of sub-recumbent folds, lying on top of a ductile shear zone, with approximately up-dip extension and transport may be the product of sub-horizontal NW-SE directed simple shear, material being transported up the Sgurr Beag slide towards the NW over the underlying Morar Division rocks. Relatively low levels of D_3 simple shear well above and to the east of the slide zone (ie. within the Loch Eil Division) would have produced more upright F_3 Folds (see Fig.80). If a D_3 pure shear mechanism is invoked then σ_1 must have been relatively steep — plunging to the NW.

D_4 deformation rotated the sub-recumbent F_3 major folds into their present upright and reclined orientations, thus producing the "steep belt" which is roughly coincident with the Glenfinnan Division. D_4 deformation also produced major F_4 upright folds of the D_3 Sgurr Beag slide. It could be argued that the steepening of the axial planes of the F_3 major folds was the result of passive rotation of the material on the limbs of the major F_4 folds, however, within the post-D_3/pre-D_4 microdiorite sheet intrusions deformational patterns produced during D_4 (Chapter 8) reveal that D_4 deformation was the result of horizontally WNW-ESE directed pure shear compression.

In the west of the region there is a well developed suite of open F_5 minor crenulation folds, the hinges of which plunge uniformly to the SE which seem to be related to a major open fold which folds the trace of the D_3 Sgurr Beag slide and the trace of the F_4 Loch Eilt antiform which folds the slide (Baird,1982.text fig.4. see Appendix 6). No folds of F_5 age were observed in the vicinity of the Ardgour granitic gneiss but in the Loch Eil Division a set of large open warps and buckles have been observed, the hinges of which plunge to the east and which could also be of F_5 age.

Figure 95.


Sketches to show the evolution
of D_3 and D_4 structures.
transport
to the WNW
Sub - recumbent
F_3 folds produced
by sub-horizontal
simple shear
Progressive D_3 simple shear
produces sub-horizontal F
sheath folds above a
late D_3 slide zone
D_3 slide zone (Sgurr Beag Slide)
F_4 Loch Eilt Antiform
F_4 Glenshian
Synform
F_3 sheath folds rotated towards
vertical during D_4 deformation.
Loch Quoich line
D_4
Sgurr Beag Slide
D_4
Maximum
Compression
Shallow F_4
plunge to SW.
" Steep Belt "
" Flat Belt "

The Loch Quoich line marks the eastern limit of F_4 folding (Fig.95). East of the line F_3 folds retain their pre-D_4 "flat belt" geometry. Since the structural history of the rocks on either side of the line is identical, the Loch Quoich line cannot be a slide zone separating rocks with a basement/cover relationship (cf. Piasecki & van Breemen,1979.p.142). Likewise it cannot be an unconformity with Loch Eil Division sediments resting on previously deformed and metamorphosed basement Glenfinnan Division rocks (cf. Lambert et al.1979. and Talbot,1983).

Several major points can be made from the discussion of the local structure outlined above.

1) The Morar, Glenfinnan and Loch Eil Division rocks have undergone the same sequence of deformation events.
2) The intensity of deformation during any one phase is variable across the area.
3) Major F_3 folds, prior to D_4 deformation, were probably sub-recumbent with shallowly eastward dipping axial planes.
4) The Loch Quoich line is the eastern limit of F_4 major folding.
5) The "steep belt" is the result of D_4 deformation which steepened up the axial planes of the earlier sub-recumbent F_3 major folds.
6) The Ardgour granitic gneiss and adjacent metasediments have been deformed by the only major F_2 folds in the area, and whilst they have been subjected to the most intense D_2 deformation they still retain an S_1 fabric and evidence for D_1 deformation and migmatisation.

9.2) Regional and crustal considerations.

A local structural sequence has been established which relates the geometry, relative structural age and absolute geochronological age, obtained both directly and indirectly, of the tectonic features exposed at the surface. However these observations shed little light on the deeper crustal structure of the area and by extrapolation the crustal structure of the Caledonides NW of the Great Glen fault. To establish an idea of the crustal structure in this area requires answers to a number of fundamental questions, some of which are noted below:-

1) The Moine thrust carries rocks of the Moine nappe WNW over the Hebridean foreland craton. At what depth in the crust is the Moine thrust in this area?
2) How far eastwards underneath the Moine thrust does the undeformed Lewisian foreland basement extend?
3) Does the Caledonian deformation seen within the Moine nappe extend downwards into the rocks below the Moine thrust?
4) How has the lower crust within the Caledonian orogen responded to deformation?
5) What was the original orientation of the Sgurr Beag slide and how much transport has occurred along it?
6) What was the source of the Lewisian rocks which are found along the Sgurr Beag slide and within the rest of the Moine succession?

The answers to many of these questions cannot be obtained using solely the techniques of structural and lithostratigraphical mapping which have been used extensively since the late 1950's. Even structural correlations between the areas have proved equivocal since, according to published descriptions, many of the areas have suffered differing numbers of deformational events. Isotopic studies which would have facilitated correlations were generally not available.

More recently our understanding of the deep continental structure of the NW Scottish Highlands has been advanced by the use of the geophysical techniques of aeromagnetic and gravity anomaly surveying, electrical conductivity, seismic refraction and most importantly seismic reflection. Unfortunately these techniques have not been used extensively in the southern Moine area (Mallaig-Glenfinnan-Fort William). Consequently any attempt to establish the deep structure of this area relies on two principal assumptions. Firstly, that one can correlate geophysical information to geological structure or lithostratigraphy and secondly, that the geological crustal structure postulated for the NW Scottish Caledonides can be extrapolated to the southern Moine area.

The Lithospheric Seismic Profile in Britain (L.I.S.P.B; Bamford et al. 1977) provides a P-wave velocity profile from < 6.2 Km/sec to 6.4 Km/sec occurs at a depth of approximately 10 Km. A sizeable increase in velocity from 6.4-6.7 Km/sec up to 8.0 Km/sec occurs across a surface dipping very shallowly to the SE at a depth of 26-29 Km. which has been interpreted as the Moho.

The discontinuity at 10 Km. is unlikely to represent the junction between the Moine cover and Lewisian basement since this junction is so highly sliced and infolded as to bring up Lewisian inliers to the surface at many places to the east of the Moine thrust. In a thin skinned interpretation of the Moine thrust zone, Coward (1980) places the Moine thrust at this level, whereas in a thick skinned interpretation of the Moine thrust zone, Soper & Barber (1982) incorporating evidence from a survey of electrical conductivity (Hutton et al.1980) postulate that the Moine thrust is sigmoidal, dipping steeply (40-50°) eastwards, flattening out along the Moho. Similarly they interpret the Naver and Meadie slides above the Moine thrust as a steep, sigmoidal thrusts branching from the Moho floor thrust. They interpret the 10 Km. P-wave velocity discontinuity as a Caledonian amphibolite-granulite metamorphic transition.

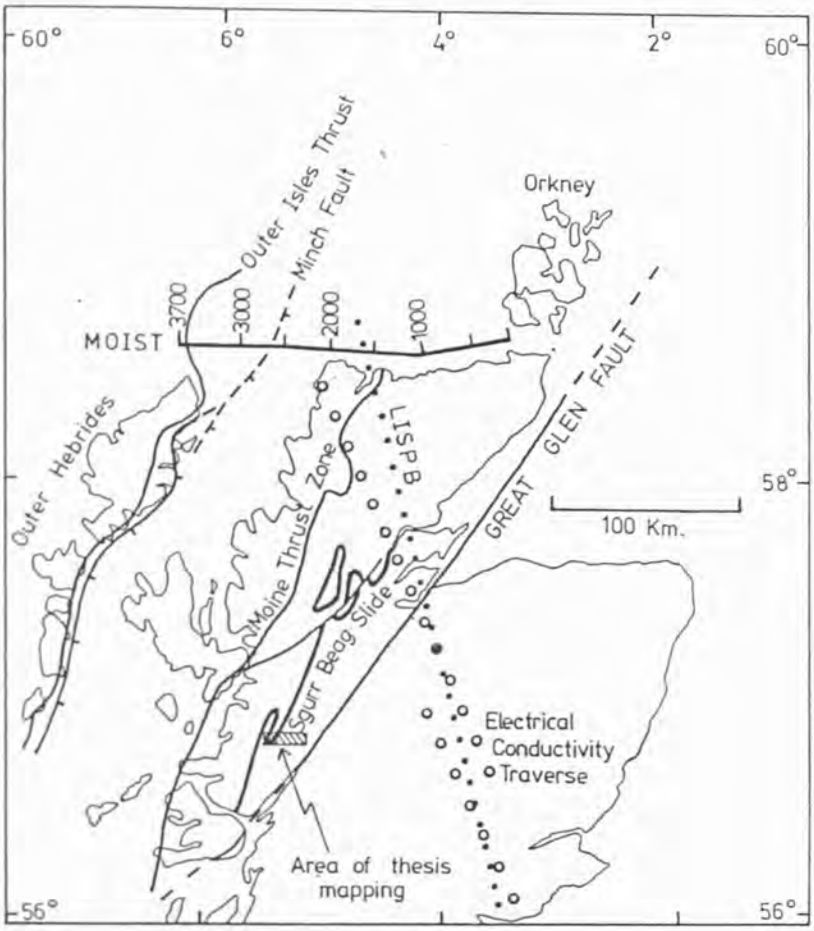
A marine seismic line, the Moine and Outer Isles Seismic Traverse (MOIST) shot off the north coast of Scotland (Fig.96a) shows a number of important seismic reflectors. An interpretation of the data (Fig.96b. taken from Smythe et al.1982 and Brewer & Smythe,1984) shows that the most prominent reflector is sub-horizontal, occurring at a depth of about 25 Km, which is interpreted as the Moho. In the east of the section a set of prominent easterly dipping reflectors terminate or flatten at a depth of 17-20 Km whereas further west similarly orientated reflectors descend to, and in places cut, the Moho (Fig.96b). Whilst this profile provides basic, fundamental geophysical information, its geological interpretation is equivocal. Brewer & Smythe (1984), noting the difficulty in extrapolating onland geological structures northwards to the MOIST profile, suggest that the Moine thrust is either the westernmost of the set of easterly dipping reflectors in the east of the section which flatten or terminate at a depth of 17-20 Km, or alternatively it is a reflector lying further east which structurally overlies the easterly dipping reflectors. Brewer & Smythe (1984) favour this second alternative and suggest that the easterly dipping reflectors below the Moine thrust may correspond to off-shore sedimentary rocks imbricated against the Lewisian basement edge.

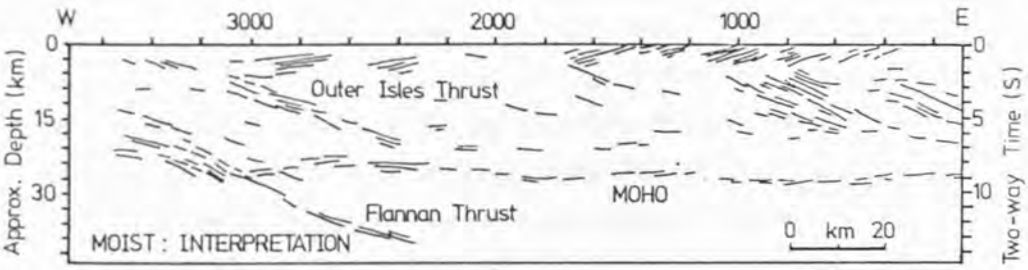
Coward (1980,1983) and Butler & Coward (1984) in proposing a thin skinned model for the Moine thrust, suggest that the thick skinned model (eg. Soper & Barber,1982) cannot explain palinspastic re-constructions which imply that the foreland Cambro-Ordovician sedimentary cover sequence must have extended at least 54 Km. east of its present position and would have been underlain by Lewisian basement at its time of deposition. Butler & Coward (1984) concluded that this Lewisian basement, stripped of its sedimentary cover must remain under the flat lying Moine thrust, and that the easterly dipping seismic reflectors in the lower crust are shear zones within the Lewisian basement produced as a sequence of foreland propagating ductile

Figure 96.

Map of Northern Scotland to show the lines of major geophysical traverses. (modified after Brewer & Smythe,1984).

(a)





(b) Interpretation of the principal seismic reflectors on MOIST. (after Brewer & Smythe,1984).

thrusts during (late) Caledonian deformation. The gravity and magnetic anomalies typical of the undeformed Lewisian foreland extend 20-30 Km. eastwards underneath the Moine thrust (Watson & Dunning,1979) and may define the eastern edge of the foreland basement. Eastwards, of this, according to Butler & Coward (1984), the Lewisian basement has been involved in Caledonian deformation. The existence of the Outer Isles and Flannan thrusts within the foreland (Smythe et al.1982) indicates that the boundary of the Caledonian orogen is not easily defined.

Having established a number of models for the crustal structure of the NW Scottish Caledonides one has to ask whether these models can be extrapolated to the area of the SW Moines. Cross sections through the SW Moines show complex fold geometries (Powell,1974. and Schematic Section 1) in which the Sgurr Beag slide is intensely refolded (Powell et al.1981: Baird, 1982) and contrast markedly with sections through the northern Moines, where Soper & Barber (1982) considered the Naver and Meadie slides as relatively simple sigmoidal ductile thrust zones branching from a floor thrust.

The Sgurr Beag slide in the SW Moines is a relatively flat lying Caledonian ductile shear zone (Powell et al.1981: Baird,1982: Coward,1983) along which displacement of 25 Km. has been postulated (Lambert et al.1979). More recently in the central Moines an estimate of 50 Km. displacement has been made (Kelley & Powell,1985). In a slide zone with overall sigmoidal geometry different levels of erosion will expose differently orientated portions of the slide so that the Sgurr Beag slide could be a relatively high level, flat lying portion of a sigmoidal slide zone. However this geometry seems unlikely as the deformation associated with the relatively flat lying Sgurr Beag slide occurs at relatively high metamorphic grade (Powell et al.1981). Therefore one must assume a real difference in geometry between the Sgurr Beag slide and the slides within the northern Moines. This, of course, assumes that the Naver and Meadie slides are represented by two of the easterly dipping reflectors dipping steeply and sigmoidally eastwards. However in a thin skinned interpretation, these problems do not arise since the steep easterly dipping reflectors lie below the Moine thrust. The slides within the Moine nappe, if they are major structures, must be relatively flat lying. Thus it would seem that correlations between the structure of the northern and SW Moines are more easily made if one assumes a thin skinned model.

The Sgurr Beag slide in the SW Moines attained its present orientation as a result of D_4 horizontally directed heterogeneous pure shear. Consideration of strain incompatibility leads one to suggest that the D_4 shortening has probably taken place over a flat or gently dipping decollement plane.

Since the Sgurr Beag slide is deformed and folded by D_4 Caledonian Deformation, the decollement plane is most probably the underlying Moine thrust. Attempts to balance sections in the Loch Quoich-Loch Garry area (A.M.Roberts & D.Barr pers.comm.) have not been able to quantify the depth to decollement with sufficient accuracy to distinguish between thick and thin skinned tectonic models.

Recently it has been shown that the fabrics of the Sgurr Beag slide are earlier than those of the Moine thrust mylonites (Kelley & Powell,1985).

It could be argued that the D_4 deformation and folding of the Sgurr Beag slide occurred as the Moine nappe was carried, piggy-back fashion, along the Moine thrust over an underlying ramp. In thrust terminology (see Butler,1982) the F_4 folds would be hanging wall anticlines. However if the F_4 folds in the SW Moines were generated in this manner then one would expect that the rocks of the Moine nappe, including the Naver and Meadie slides in the northern Moines would be similarly deformed which is not the case.

Differences between the structures of the northern and SW Moines are further emphasised by an examination of the MOIST profile (Fig.96b) which shows a series of westerly dipping reflectors in the upper crust. Smythe et. al.(1982) and Brewer & Smythe (1984) interpret these reflectors as half grabens filled with late Palaeozoic and younger sediments. Each half graben being bounded on the west by an easterly dipping thrust which has been re-activated as a normal fault. In the SW Moines there is no evidence of either sediment filled half grabens or large scale re-activation of slides such as the Sgurr Beag slide as major normal extensional faults.

From the discussion above it is clear that in attempts to construct a detailed model for the crustal structure of the SW Moines one needs geophysical evidence, especially seismic reflection profiles and seismic refraction surveys of the area and an improved technique for correlating between the geophysical and geological evidence.

CHAPTER 10.

Summary of conclusions.

The conclusions are best summed up in the form of a history of deposition, deformation and metamorphism.

The rocks of the Morar and Glenfinnan Divisions were deposited as a series of mixed feldspathic sandstones and mudstones. The Glenfinnan Division passes upwards, probably without a stratigraphical break, and certainly without a tectonic break, into the monotonous feldspathic sandstones of the Loch Eil Division. A granitic body, which subsequently has been transformed into the Ardgour granitic gneiss, was intruded along and near to the junction of the Glenfinnan and Loch Eil Divisions.

The first phase of deformation (D_1) and metamorphism (M_1) produced a partial anatectic melt of the granite, converting it into a granitic gneiss. Pegmatites, which may be pre-D_1 or syn-D_1, are co-planar to the S_1 gneissose foliation and contain well developed biotite selvages. Major F_1 folds have not been mapped.

The Ardgour granitic gneiss is intensely deformed by major F_2 folds. The gneiss has been extensively re-worked during M_2 and it is possible that some new migmatitic material was generated co-planar to the S_2 gneissose foliation. The adjacent Loch Eil Division metasediments contain an early schistosity (S_1) which is folded by F_2 folds. Where D_2 deformation is intense and the rocks more pelitic, F_2 folds contain an axial planar migmatitic fabric (S_2). Elsewhere in the Glenfinnan Division the metasediments contain an "early" penetrative schistosity which is often migmatitic in texture, but probably the product of solid state mass transfer rather than partial anatexis. This "early" schistosity may be S_1, S_2 or a composite S_1/S_2 foliation.

Major F_1 and F_2 folds have not been found away from the Ardgour granitic gneiss, over most of the area bedding is virtually co-planar with the "early" schistosity. To the west of the mapped area M_1 is lower grade than M_2, but M_1 rises eastwards to produce partial anatexis at Glenfinnan. Possibly only east of Sgurr a Mhuidhe is M_1 grade higher than M_2. Isograd surfaces from Sgurr a Mhuidhe westwards were established using calc-silicats rocks and are thought to be M_2 surfaces, apparently co-planar to the composite bedding/schistosity.Both M_2 isograd surfaces and the composite bedding/

schistosity were probably flat lying prior to subsequent deformation.

D_3 deformation has produced a series of major tight to isoclinal F_3 folds which were flat lying and had axial planes with shallow dips to the ESE. Locally D_3 deformation intensified and developed into a tectonic slide, the Sgurr Beag slide, which dipped shallowly to the ESE. As D_3 deformation intensified westwards towards the Sgurr Beag slide major F_3 folds tightened and their hinge lines rotated from nearly horizontal NE-SW trending, towards an E-W or WNW-ESE trend. The fold hinges rotated towards the extension and transport direction of the Sgurr Beag slide, towards the W or WNW. Subsequently D_4 deformation has re-orientated both the hinge lines and axial planes of the F_3 folds.

The Sgurr Beag slide separates rocks of the Morar and Glenfinnan Divisions which have the same pre-sliding deformation and metamorphic history, thus it does not obscure or delimit an orogenic front separating rock groups with basement and cover relationships. F_3 folds have folded the earlier (M_1 or M_2) isograd pattern and have only re-worked the metamorphic assemblage within the high strain areas of the slide zone. Large intrusive pegmatites are common, many are folded by F_3 folds but some may be post-F_3 in age.

Uplift, cooling and jointing occurred after D_3. The joint suite dipped shallowly eastwards and was intruded by a series of microdiorite sheets. Fewer numbers of these sheets had an initial dip westwards.

D_4 deformation is intense in the east of the Morar Division and throughout the Glenfinnan Division. It is much less intense to the east of this zone, called the "highly inclined" or "steep belt". D_4 deformation has produced a few major, nearly isoclinal F_4 folds with vertical NE-SW trending axial planes and nearly horizontal fold hinge lines. In the "steep belt", this phase of upright folding has re-orientated all the earlier planar and linear elements so that the major F_3 folds now have vertical axial planes trending NE-SW. Where major F_3 fold hinges were nearly horizontal NE-SW trending, they have remained so. However in areas of more intense D_3 deformation where F_3 fold hinges plunged down to the E or ESE they have been rotated during F_4 folding so that they now plunge steeply towards the NE, within F_3 vertical axial planes.

Beyond the eastern limit of the "steep belt", which roughly coincides with the lithological change from the Glenfinnan to Loch Eil Divisions and has been termed the Loch Quoich line, D_4 deformation is weak and major F_3 folds have retained their nearly flat lying axial plane disposition.

D_4 deformation has been responsible for the deformation of the micro-diorite suite. Within the "steep belt" intense D_4 horizontally NW-SE directed compression has rotated microdiorites away from the maximum compression direction towards a vertical NE-SW trend. East of the "steep belt" the microdiorites retain relatively shallow dips. The microdiorites were incompetent relative to the country rocks during deformation, so that even in areas of low levels of D_4 deformation the microdiorites are quite intensely deformed.

D_5 deformation has produced a series of open major F_5 folds with NW-SE trending axial planes. Minor open crenulations in the extreme west of the area are associated with these major F_5 folds. In the Loch Eil Division a series of open large scale warps have hinge lines which plunge shallowly eastwards and have been correlated as F_5 folds.

The deformation phases D_3, D_4, and D_5 have occurred under conditions of metamorphism sufficiently high to recrystallise mica crystals after their deformation.

The Moine succession of the area has been intruded by two suites of undeformed camptonites and dolerites of Carboniferous/Permian and Tertiary ages respectively.

Appendix 1. References.

Aftalion,M. & van Breemen,O (1980).

U-Pb, Zircon, Monazite and Rb-Sr Whole rock Systematics of Granitic Gneiss and Psammitic to Semi-Pelitic host Gneiss from Glenfinnan, North-western Scotland.
Contrib.Mineral.Petrol. 72, 87-98.

Althaus,E.,Nitsch,K.H.,Karothe,E. & Winkler,H.G.F. (1970).

An experimental re-examination of the upper stability limit of muscovite plus quartz.
Neues Jahrb.Mineral.Monatshefte. 7, 325-336.

Anderson,D.E. & Olympio,J.C. (1977).

Progressive homogenisation of metamorphic garnets, South Morar, Scotland: evidence of volume diffusion.
Can.Mineral. 15, 205-216.

Ashworth,J.R. (1972).

Myrmekites of exsolution and replacement origins.
Geol.Mag. 109, 45-62.

Bailey,E.B. (1950).

The Structural history of Scotland.
Int.geol.Congr. 18th Rep. 230-255.
_____(1955).
Moine tectonics and metamorphism in Skye.
Trans.goel.Soc.Edinburgh. 16, 93-166.

Baird,A.W (1982).

The Sgurr Beag Slide within Moine rocks at Loch Eilt, Inverness-shire.
J.geol.Soc.London. 139, 647-654.
____(1985).
Discussion of the structural setting and tectonic significance of the Glen Dessary Syenite, Inverness-shire.
J.geol.Soc.London. 142, 713-718.

Bamford,D.,Nunn,K.,Prodehl,C.& Jacob,B. (1977).

LISPB - III. Upper crustal structure of Northern Britain.
J.geol.Soc.London. 133, 481-8.

Barr,D. (1983).

Geochemical constraints on migmatization in the Moine rocks of Scotland.
In: Migmatites, melting and metamorphism. eds. Atherton,M.P. & Gribble, C.D. (Shiva geology series 1983) p,297-8.

Barrow,G. (1912).

On the Geology of Lower Dee-side and the Southern Highland Border,
Proc.Geol.Assoc.London. 23, 274-290.

____(1925).

In discussion of Tilley,C.E. A Preliminary Survey of Metamorphic Zones in the Southern Highlands of Scotland.
Q.J.geol.Soc.London. 81, 100-112.

Baxter,A.N. & Mitchell,J.G. (1984).

Camptonite-Monchiquite dyke swarms of Northern Scotland; Age relationships and their implications.
Scott.J.Geol. 20, 297-308.

Beckinsale,R.D. & Obradovich,J.D. (1973).

Potassium-Argon ages for minerals from the Ross of Mull, Argyllshire, Scotland.
Scott,J.Geol. 9, 147-156.

Brewer,J.A. & Smythe,D.K. (1984).

MOIST and the continuity of crustal reflector geometry along the Caledonian-Appalachian orogen.
J.geol.Soc.London. 141, 105-120.

Brewer,M.S.,Brook,M. & Powell,D. (1979).

Dating of the tectono-metamorphic history of the southwestern Moine, Scotland.
In: Harris,A.L et al. (1979).p,129-137.

Brook,M.,Powell.D. & Brewer,M.S. (1976).

Grenville age for rocks in the Moine of northwestern Scotland.
Nature.London. 260, 515-517.
____,____&____ (1977).
Grenville events in Moine rocks of the Northern Highlands,Scotland.
J.geol.Soc.London. 133, 489-496.

Brown,R.L. (1964).

The structure and metamorphism of east Moidart and west Sunart, Argyll and Inverness-shire.
Univ.Edinburgh.PhD. thesis(unpubl.).
___,Dalziel,I.W.D. & Johnson,M.W.R. (1970).
A review of the structure and stratigraphy of the Moinian of Ardgour, Moidart and Sunart: Argyll and Inverness-shire.
Scott.J.Geol. 6, 309-335.

Butler,B.C.M. (1965).

A chemical study of some rocks of the Moine Series of Scotland.
Q.J.geol.Soc.London. 121, 163-208.

Butler,R.W.H. (1982).

The terminology of structures in thrust belts.
J.Struct.Geol. 4, 239-245.
___& Coward,M.P. (1984).
Geological constraints, structural evolution, and deep geology of the Northwest Scottish Caledonides.
Tectonics. 3, 347-365.

Charnley.N.C. (1976).

Geochemical studies of Moine rocks in Western Inverness-shire.
Unpubl.PhD.thesis. Bedford College,Univ. of London.

Chayes,F. (1952).

Staining of potash feldspars with Sodium cobaltinitrite in thin section.
Am.Mineral. 37, 337-340.

Cheeney,R.F. & Matthews,D.W. (1965).

The structural evolution of the Tarskavaig and Moine nappes in Skye.

Scott.J.Geol. 1, 256-281.

Clark,G.C. (1961).

Structural and metamorphic history of the Lochailort (Moidart) area.
Univ.Edinburgh. PhD.thesis (unpubl.).

Clifford,T.N. (1957).

The stratigraphy and structure of part of the Kintail district of southern Ross-shire: its relation to the Northern Highlands.
Q.J.geol.Soc.London. 117, 57-92.

Cobbold,P.R.,Cosgrove,J.W. & Summers,J.M. (1971).

The development of internal structures in deformed anisotropic rocks.
Tectonophysics. 12, 23-53.

Cosgrove,J.W. (1976).

The formation of crenulation cleavage.
J.geol.Soc.London. 132, 155-178.

Coward,M.P. (1980).

The Caledonian thrusts and shear zones of N.W. Scotland.
J.Struct.Geol. 2, 11-17.
____(1983).
The thrust and shear zones of the Moine thrust zone and the NW Scottish Caledonides.
J.geol.Soc.London. 140, 795-811.

Dalziel,I.W.D. (1963a).

A structural study of the granite gneiss and associated Moinian rocks between Loch Shiel and Loch Eil, Argyll and Inverness-shire.
Univ.Edinburgh. PhD.thesis (unpubl.).
____(1963b).
Zircons from the granitic gneiss of western Ardgour, Argyll : their bearing on its origin.
Trans.geol Soc.Edinburgh. 19, 349-362.
___(1966).
A structural study of the granitic gneiss of western Ardgour, Argyll

and Inverness-shire.
Scott.J.Geol. 2, 125-152.

___ & Johnson,M.W.R. (1963).
Evidence for the geological dating of the granitic gneiss of western Ardgour.
Geol.Mag. 100, 244-254.

___ & Brown,R.L. (1965).
The structural dating of sillimanite-grade metamorphism of the Moines in Ardgour (Argyll) and Moidart (Inverness-shire).
Scott.J.Geol. 1, 304-311.

Deer,W.A.,Howie,R.A. & Zussman,J. (1966).
An introduction to the Rock-forming minerals. Longman, London. 528pp.

Drever,H.E. (1940).
The geology of Ardgour, Argyllshire.
Trans.R.Soc.Edinburgh. 60, 141-170.

Esang,C.B. & Piper,J.D.A. (1984).
Palaeomagnetism of Caledonian intrusive suites in the Northern Highlands of Scotland; constraints to tectonic movements within the Caledonian orogenic belt.
Tectonophysics. 104, 1-34.

Evans,A.M. (1963).
Conical folding and Oblique structures in Charnwood Forest, Leicestershire.
Proc.Yorkshire.geol.Soc. 34, 67-79.

Fleuty,M.J. (1964).
The description of folds.
Proc.Geol.Assoc.London. 75, 461-489.

Flinn,D. (1962).
On folding during three dimensional Progressive deformation.
Q.J.geol.Soc.London. 118, 385-433.

Frödin,G. (1922).
On the Analogies between the Scottish and Scandinavian Portions of the

Caledonian Mountain Range.
Bull.geol.Instn.Univ.Upsala. 18, 199-238.

Gallacher,M.J. (1963).
Lamprophyre dykes from Argyll.
Mineralog.Mag.London. 33, 415-430.

Giletti,B.J.,Moorgath,S. & Lambert,R.St.J. (1961).
A geochronological study of the metamorphic complexes of the Scottish Highlands.
Q.J.geol.Soc.London. 117, 233-272.

Glazner,A.F. (1980).
Frequency distribution of plagioclase extinction angles: precision of the Michel-Levy technique.
Am.Mineral. 65, 1050-1052.

Gould,D. (1966).
Geochemical and mineralogical studies of the granitic gneiss and associated rocks of western Ardgour, Argyll.
Univ.Edinburgh. PhD.thesis (unpubl.).

Halliday,A.N.,Aftalion,M.,van Breemen,O. & Jocelyn,J. (1979).
Petrogenetic significance of Rb-Sr and U-Pb isotopic systems in the 400 Ma. old British Isles granitoids and their hosts.
In : Harris,A.L. et al. (1979).p,653-661.

Harris,A.L. & Pitcher,W.S. (1975).
The Dalradian Supergroup.
In : A correlation of Precambrian rocks in the British Isles;
Eds. Harris,A.L.,Shackleton,R.M.,Watson,J.V.,Downie,C.,Harland,W.B. & Moorbath,S. Spec.Rep.geol.Soc.London 6, 52-75.

Harris,A.L.,Holland,C.H. & Leake,B.E.(eds.). (1979).
The Caledonides of the British Isles - Reviewed.
Spec.Publ.geol.Soc.London. 8. 768pp.

Harry,W.T. (1953).

The composite granite gneiss of western Ardgour, Argyll.
Q.J.geol Soc.London. 109, 285-308.

Holmes,A. (1960).

A revised geological time-scale.
Trans.geol.Soc.Edinburgh. 17, 183-216.

Horne,J. (1930).

In : Chapters on the Geology of Scotland. Peach,B.N. & Horne,J.
Oxford University Press. (p,200-1).

Hoschek,G. (1969).

The stability of Staurolite and Chloritoid and their Significance in Metamorphism of Pelitic Rocks.
Contrib.Mineral.Petrol. 22, 208-232.

Howkins,J.B. (1961).

The structural and metamorphic history of Moidart, southwest Inverness-shire.
Univ.Edinburgh. PhD.thesis (unpubl.).

Hutton,D.H.W. (1979).

Tectonic slides: A Review and Reappraisal.
Earth.Sci.Rev. 15, 151-172.

Hutton,V.R.S.,Ingham,M.R. & Mbipom,E.W. (1980).

An electric model of the crust and upper mantle in Scotland.
Nature.London. 287, 30-33.

Johnson,M.W.R. & Dalziel,I.W.D. (1966).

Metamorphosed Lamprophyres and the late thermal history of the Moines.
Geol.Mag. 103, 240-249.

Johnstone,G.S.,Smith,D.I. & Harris,A.L. (1969).

The Moinian Assemblage of Scotland.
Amer.Assoc.Petrol.Geol. Memoir 12, 159-180.

Kelley,S.P. & Powell,D. (1985).

Relationships between marginal thrusting and movement on major, internal shear zones in the Northern Highland Caledonides, Scotland.
J.Struct.Geol. 7, 161-174.

Kennedy,W.Q. (1946).

The Great Glen fault.
Q.J.geol.Soc.London. 102, 41-76.
____(1949).
Zones of progressive regional metamorphism in the Moine schists of the Western Highlands of Scotland.
Geol.Mag. 86, 43-56.
____(1955).
The tectonics of the Morar anticline and the problem of the North-west Caledonian front.
Q.J.geol.Soc.London. 110, 357-390.

Kulp,J.L. (1959).

Geological time scale.
Bull.geol.Soc.Am. 70, 1634.

Lambert,R.St.J. (1958).

A metamorphic boundary in the Moine schists of the Morar and Knoydart districts of Inverness-shire.
Geol.Mag. 95, 177-194.
____(1969).
Isotopic studies relating to the Precambrian history of the Moinian of Scotland.
Proc.geol.Soc.London. 1652, 243-245.
____ , Winchester,J.A. & Holland,J.G. (1979).
Time, space and intensity relationships of the Precambrian and lower Palaeozoic metamorphisms of the Scottish Highlands.
In : Harris,A.L. et al. (1979).p,363-367.

Lawrie,T.R.M. et al. (1961).

Scottish Highlands and Islands.
Summ.Prog.geol.Surv.G.B. for 1960.

Leedal,G.P. (1952).

The Cluanie igneous intrusion. Inverness-shire and Ross-shire.
Q.J.geol.Soc.London. 108, 35-62.

Long,L.E. & Lambert,R.St.J. (1963).

Rb-Sr isotope ages from the Moine Series.
In : Johnson,M.W.R. & Stewart,F.H. (eds.). The Britist Caledonides.
Oliver & Boyd, Edinburgh, p,217-247.

MacGregor,A.G. (1948).

Resemblances between Moine and "Sub Moine" metamorphic sediments in the Western Highlands of Scotland.
Geol.Mag. 85, 265-275.

MacQueen,J.A. & Powell,D. (1977).

Relationships between deformation and garnet growth in Moine(Precambrian) rocks of western Scotland.
Bull.geol.Soc.Am. 88, 235-240.

Miller,J.A. & Brown,P.E. (1965).

Potassium-argon studies in Scotland.
Geol.Mag. 102, 106-134.

Moorbath,S, (1969).

Evidence for the age of deposition of the Torridonian sediments of north-west Scotland.
Scott.J.Geol. 5, 154-170.

Morrison,M.A.,Thompson,R.N.,Gibson,I.L. & Marriner,G.F. (1980).

Lateral chemical heterogeneity in the Palaeocene upper mantle beneath the Scottish Hebrides.
Philos.Trans.R.Soc.London. Series A. 297, 229-244.

Olimpio,J.C. & Anderson,D.E. (1978).

The relationship between chemical and textural (optical) zoning in metamorphic garnets, South Morar, Scotland.
Am.Mineral. 63, 677-689.

Peach,B.N. (1892).

Annual Rep.Geol.Surv.G.B. (p,262).

____(1913).

The Geology of Central Ross-shire (Explan. Sheet 82) Mem.Geol.Survey.

____, & Wilson,J.S.G. (1904).

West Highland district. Older igneous rocks in metamorphic series, Ardgour area.

Summ.Prog.geol.Surv.G.B. (for 1903),68.

Phillips,F.C. (1951).

Apparent Coincidences in the Life History of the Moine Schists.

Geol.Mag. 88,225-235.

Phillips,W.E.A.,Stillman,C.J. & Murphy,T. (1976).

A Caledonian plate tectonic model.

J.geol.Soc.London. 132, 579-609.

Piasecki,M.A.J. (1980).

New light on the Moine rocks of the Central Highlands of Scotland.

J.geol.Soc.London. 137, 41-59.

____, & van Breemen,O. (1979).

The "Central Highland Granulites" cover-basement tectonics in the Moine.

In : Harris,A.L. et al. (1979) .p,139-144.

Pigeon,R.T. & Aftalion,M. (1978).

Cogenetic and inherited zircon U-Pb systems in granites: Palaeozoic granites of Scotland and England.

In : Bowes,D.R. & Leake,B.E. (eds.). Crustal evolution in northwestern Britain and adjacent regions. Geol.J.Spec.Issue No.10, 183-248.

Powell,D. (1963).

The Stratigraphy, Structure and Metamorphism of the area around Lochailort, Inverness-shire.

Univ.Coll.Wales,Aberystwyth. PhD.thesis (unpubl.).

____(1964).

The stratigraphical succession of the Moine schists around Lochailort (Inverness-shire) and its regional significance.

Proc.Geol.Assoc.London. 75, 223-246.

Powell,D. (1966).
The structure of the south-eastern part of the Morar antiform, Inverness-shire.
Proc.Geol.Assoc.London. 77, 79-100.
____(1974).
Stratigraphy and structure of the western Moine and the problem of Moine orogenesis.
J.geol.Soc.London. 130, 575-593.
____, & MacQueen,J.A. (1976).
Relationships between garnet-shape, rotational inclusion fabrics and strain in some Moine metamorphic rocks of Skye, Scotland.
Tectonophysics. 35, 391-402.
____,Baird,A.W.,Charnley,N.C. & Jordan,P.J. (1981).
The metamorphic environment of the Sgurr Beag Slide; a major crustal displacement zone in Proterozoic, Moine rocks of Scotland.
J.geol.Soc.London. 138. 661-673.
____,Brook,M. & Baird,A.W. (1983).
Structural dating of a Precambrian pegmatite in Moine rocks of northern Scotland and its bearing on the "Morarian Orogeny".
J.geol.Soc.London. 140, 813-824.

Raith,M. (1976).
The Al-Fe(III) Epidote Miscibility Gap in a Metamorphic Profile Through the Penninic Series of the Tauern Window, Austria.
Contrib.Mineral.petrol. 57, 99-117.

Ramsay,J.G. (1958).
Moine-Lewisian reletions at Glenelg, Inverness-shire.
Q.J.geol.Soc.London. 113, 487-523.
____(1963).
Structure and Metamorphism of the Moine and Lewisian rocks of the Northwest Caledonides. In ; Johnson,M.W.R & Stewart,F.H. (eds.). The British Caledonides. Oliver & Boyd, Edinburgh. p,143-170.
____(1967).
Folding and fracturing of Rocks.
M^{c}Graw Hill, New York. 568 pp.
____, & Graham,R.H. (1970).
Strain variations in shear belts.
Can.J.Earth Sci. 7, 786-813.

Ramsay,J.G.,Casey,M. & Kligfield,R. (1983).

Shear zones and nappe tectonics in the Helvetic fold/thrust belt.
Geology. 11, 439-442.

Rathbone,P.A. & Harris,A.L. (1979).

Basement-cover relationships at Lewisian inliers in the Moine rocks.
In : Harris,A.L. et al. (1979). p,101-107.

____,Coward,M.P. & Harris,A.L.(1983).
Cover and basement: A contrast in style and fabrics.
Geol.Soc.AM.Mem. 158, 213-223.

Read,H.H. (1934).

Age Problems of the Moine Series.
Geol.Mag. 71, 302-316.

Richey,J.E. et al. (1932 and 1934-6).

Scotland, West Highland District.
Summ.Progr.geol.Surv.G.B. (for 1931 and 1933-5).

____, & Kennedy,W.Q. (1939).
The Moine and Sub Moine Series of Morar, Inverness-shire.
Bull.geol.Surv.G.B. 2, 26-45.

Roberts,A.M. & Harris,A.L. (1983).

The Loch Quoich Line - a limit of early Palaeozoic crustal reworking of the Northern Highlands of Scotland.
J.geol.Soc.London. 140, 883-892.

____,Smith,D.I. & Harris,A.L. (1984).
The structural setting and tectonic significance of the Glen Dessary Syenite, Inverness-shire.
J.geol.Soc.London. 141, 1033-1042.

Sanderson,D.J. (1979).

The transition from upright to recumbent folding in the Variscan fold belt of southwest England: a model based on the kinematics of simple shear.
J.Struct.Geol. 1, 171-180.

Smith,D.I. (1979).

Caledonian minor intrusions of the N.Highlands of Scotland.
In : Harris,A.L. et al. (1979). p,683-697.

Smythe,D.K.,Dobinson,A.,M^{c}Quillan,R.,Brewer,J.A.,Matthews,D.H.Blundell,D.J. & Kelk,B. (1982).

Deep structure of the Scottish Caledonides revealed by the MOIST reflection profile.
Nature.London. 299, 338-340.

Soper,N.J. & Barber,A.J. (1982).

A model for the deep structure of the Moine thrust zone.
J.geol.soc.London. 139, 127-138.

Spear,E.S. (1977).

Evidence for a miscibility gap in plagioclase feldspar in the composition range An.39-An.88.
Yearb.Carnegie Inst.Washington. 76, 619-621.

Speight,J.M. & Mitchell,J.G. (1979).

The Permo-Carboniferous dyke-swarm of northern Argyll and its bearing on dextral displacement on the Great Glen Fault.
J.geol.Soc.London. 136, 3-11.

Steiger,R.H. & Jager,E. (1977).

Subcommission on geochronology: convention on the use of decay constants in geo- and cosmochronology.
Earth planet.Sci.Lett. 36, 359-362.

Strachan,R.A. (1982).

Tectonic sliding within the Moinian Loch Eil Division near Kinlocheil, W. Inverness-shire.
Scott.J.Geol. 18, 187-203.

___(1985).

The stratigraphy and structure of the Moine rocks of the Loch Eil area, West Inverness-shire.

Talbot,C.J. (1982).

Obliquely foliated dikes as deformed incompetent single layers.
Bull.geol.soc.Am. 93, 450-460.

____(1983).
Microdiorite sheet intrusions as incompetent time- and strain-markers in the Moine assemblage NW of the Great Glen fault, Scotland.
Trans.R.Soc.Edinburgh. Earth Sciences, 74, 137-152.

Tanner,P.W.G. (1971).

The Sgurr Beag Slide - A major tectonic break within the Moinian of the Western Highlands of Scotland.
Q.J.geol.Soc.London. 126, 435-463.

____(1976).
Progressive Regional Metamorphism of Thin Calcareous Bands from the Moinian Rocks of N.W.Scotland.
J.Petrol. 17, 100-134.

____,Johnstone,G.S.,Smith,D.I. & Harris,A.L. (1970).
Moinian Stratigraphy and the problem of the central Ross-shire Inliers.
Bull.geol.Soc.Am. 81, 299-306.

____, & Miller,R.G. (1980).
Geochemical evidence for loss of Na and K from Moinian calc-silicate pods during prograde metamorphism.
Geol.Mag. 117, 267-275.

Tyler,I.M. & Ashworth,J.R. (1982).

Sillimanite - potash feldspar assemblages in Graphitic pelites, Strontian area, Scotland.
Contrib.Mineral.Petrol. 81, 18-29.

van Breemen,O.,Pigeon,R.T. & Johnson,M.W.R. (1974).

Precambrian and Palaeozoic pegmatites in the Moines of northern Scotland.
J.geol.Soc.London. 130, 493-507.

____,Halliday,A.N.,Johnson,M.W.R. & Bowes,D.R. (1978).
Crustal additions in late Precambrian times.
In : Bowes,D.R. & Leake,B.E. (eds.). Crustal evolution in northwestern Britain and adjacent regions. Geol.J.Spec.Issue. 10, 82-106.

van Breemen,O.,Aftalion,M.,Pankhurst,R.J. & Richardson,S.W. (1979).

Age of the Glen Dessary Syenite, Inverness-shire: diachronous Palaeozoic metamorphism across the Great Glen.
Scott.J.Geol. 15, 49-62.

Van der Voo,R. & Scotese,C. (1981).

Palaeomagnetic evidence for a large ($\sim$2000 Km) sinistral offset along the Great Glen fault during Carboniferous time.
Geology. 9, 583-589.

von Platen,H. & Holler,H. (1967).

Anatexis des Stainzer Plattengneises von Koralpe, Steiermark, bei 2,4,7 und 10 kbH_2O-D_{ruck}.
Neues Jahrb.Mineral.Abhandlungen. 106, 106-130.

Watson.J.V. & Dunning,F.W. (1979).

Basement-cover relations in the British Caledonides.
In : Harris,A.L. et al (1979). p,67-91.

Wenk,C. & Keller,F. (1969).

Isograde in Amphibolitserien der Zentralalpen.
Schweiz.mineral.petrogr.Mitt. 49, 157-198.

Winchester,J.A. (1972).

The Petrology of Moinian Calc Silicate Gneisses from Fannich Forest, and their Significance as Indicators of Metamorphic Grade.
J.Petrol. 13, 405-424.

____(1974a).

The zonal pattern of regional metamorphism in the Scottish Caledonides.
J.geol.Soc.London. 130, 509-524.

____(1974b).

The control of the whole rock content of CaO and Al_2O_3 on the occurence of the Al_2SiO_5 polymorphs in amphibolite facies pelites.
Geol.Mag. 111, 205-211.

____(1975).

Epidotic calc-silicates of Arnipol Type - a widely distributed minor sedimentary facies of the Moinian Assemblage, Scotland.
Geol.Mag. 112, 175-181.

Winchester,J.A. (1976).

Different Moinian amphibolite suites in northern Ross-shire.
Scott.J.Geol. 12, 187-204.

____, & Whittles,K.H. (1979).
The pattern of three - dimensional metamorphisn in the Killin area, Inverness-shire: a direct method of determining the thickness of metamorphism in the Caledonides.
In : Harris,A.L. et al. (1979). p,345-349.

Winkler,H.G.F. (1966).

Der Prozess der Anatexis: Seine Bedeuting fur die Genese der Migmatite
Tschermaks Mineral.Petrogr.Mitt. 11, 266-287.

____(1979).
Petrogenesis of Metamorphic Rocks, (5th Edition)Springer-Verlag:New York.

Appendix 2.

All the listed major element geochemical analyses are recorded as weight percent oxides. (ie. P to Na).
(P_2O_5, SiO_2, Al_2O_3, MgO, MnO, Fe_2O_3, TiO_2, CaO, K_2O, Na_2O).

For a summary of analytical technique see:

Norrish,K & Hutton,J.T. (1969).

An accurate X-ray spectrographic method for the analysis of a wide range of geochemical samples.
Geochim.Cosmochim.Acta., 33, 431-453.

Appendix 2.

Calc - silicate rocks.

	34/229	36/229	37/229	40/229	48/231	81/250	90/301	103/440	104/451
P	0.055	0.110	0.100	0.095	0.060	0.050	0.030	0.040	0.030
Si	81.630	86.290	78.560	84.580	88.330	81.590	95.425	89.395	64.095
Al	9.075	7.320	8.485	5.625	5.830	9.680	2.635	5.560	15.515
Mg	0.655	0.280	0.810	0.500	0.280	0.700	0.170	0.225	1.665
Mn	0.140	0.020	0.195	0.155	0.130	0.090	0.020	0.045	0.230
Fe	1.815	1.065	4.555	2.885	1.320	1.415	0.460	0.775	4.345
Ti	0.470	0.285	0.415	0.205	0.250	0.225	0.110	0.130	1.270
Ca	3.240	2.730	6.240	4.870	3.410	4.295	0.740	2.355	11.210
K	0.960	0.470	0.400	0.165	0.300	1.210	0.480	0.465	0.590
Na	1.185	1.375	0.610	0.285	0.480	1.060	0.495	0.775	1.220
Total	99.230	99.955	100.340	99.355	100.370	100.325	100.075	99.755	100.445
An%	Alt.	69	57	Alt.	72?	Alt.	Alt.	42+	70
Ca/Al	0.357	0.373	0.735	0.866	0.585	0.444	0.281	0.424	0.723

	114/505	119/517	122/537	131/573	134/582	167/883	170/948	174/956	175/961
P	0.090	0.140	0.050	0.050	0.120	0.070	0.180	0.120	0.090
Si	78.640	74.240	82.080	79.650	76.680	75.025	69.920	72.030	74.210
Al	10.965	12.490	9.620	11.770	10.970	11.755	15.770	13.760	13.210
Mg	0.535	0.860	0.380	0.330	0.345	1.140	0.980	1.410	1.265
Mn	0.080	0.190	0.080	0.130	0.125	0.295	0.290	0.370	0.315
Fe	2.235	3.290	1.410	1.290	2.205	3.340	3.920	2.720	3.590
Ti	0.345	0.600	0.300	0.220	0.675	0.430	0.680	0.550	0.575
Ca	3.235	6.520	2.970	3.350	7.110	6.650	9.930	4.900	5.660
K	0.850	0.810	0.610	0.380	0.240	0.180	0.180	1.580	0.920
Na	3.325	1.280	2.340	3.270	1.525	1.020	0.390	1.600	1.290
Total	100.290	100.420	99.840	100.450	99.985	99.895	101.940	99.020	101.125
An%	Alt.	Alt.	47	47	Alt.	83	76	72	89
Ca/Al	0.295	0.522	0.309	0.285	0.648	0.566	0.630	0.356	0.428

Major elements are listed as Wt.% oxides.

Appendix 2.

Calc - silicate rocks.

	187/1031	200/1050	204/1065	207/1068	208/1071	212/1087	213/1087	215/1088
P	0.130	0.095	0.035	0.070	0.115	0.070	0.140	0.060
Si	81.610	78.940	78.255	74.575	73.340	78.670	73.605	78.500
Al	8.960	11.475	11.325	13.725	12.075	9.940	13.935	12.000
Mg	0.560	0.420	0.580	0.665	0.890	0.410	0.540	0.370
Mn	0.310	0.360	0.230	0.305	0.610	0.170	0.240	0.240
Fe	2.470	1.955	1.930	2.490	4.460	1.910	2.730	1.940
Ti	0.340	0.375	0.510	0.470	0.445	0.570	0.855	0.560
Ca	4.260	5.890	5.970	7.065	6.310	3.160	5.865	3.380
K	0.750	0.480	0.075	0.435	0.530	0.880	1.295	0.790
Na	0.440	0.310	0.250	0.780	0.610	1.950	1.135	2.520
Total	99.820	100.295	99.155	100.580	99.390	97.730	100.340	100.370
An%	77	74	89	87	76	48+	88	48
Ca_{Al}	0.475	0.513	0.527	0.514	0.523	0.318	0.421	0.282

	225/1091	252/1110	253/1115	254/1119	255/1123	256/1126	259/1145		Average
P	0.150	0.060	0.080	0.160	0.060	0.050	0.160		0.100
Si	78.905	75.730	84.810	71.120	81.210	80.445	73.290		76.520
Al	10.380	12.675	9.185	13.930	10.015	10.265	14 385		12.040
Mg	0.425	0.475	0.315	0.990	0.465	0.695	0.700		0.700
Mn	0.100	0.310	0.155	0.430	0.460	0.695	0.200		0.320
Fe	2.060	1.735	1.295	3.420	1.835	2.740	2.795		2.600
Ti	0.415	0.480	0.280	0.610	0.410	0.430	0.640		0.510
Ca	2.110	4.470	3.680	7.160	4.650	5.020	7.235		5.440
K	0.660	0.880	0.670	0.790	0.275	0.195	0.645		0.640
Na	3.105	1.600	0.420	0.710	0.550	0.410	0.825		1.000
Total	98.305	98.395	100.890	99.330	99.925	100.935	100.860		99.910
An%	35	88	88	85	75	84	92		
Ca_{Al}	0.203	0.353	0.401	0.514	0.464	0.489	0.503		

Major elements are listed as Wt.% oxides.

Appendix 2.

← Garnetiferous Amphibolites →

	4/6	41/230	67/235	143/654	147/669	152/721	159/740	165/817
P	0.445	0.320	0.370	0.235	0.220	0.215	0.405	0.330
Si	46.480	48.335	47.920	48.035	46.460	49.865	51.240	46.130
Al	13.600	13.660	14.080	13.880	13.840	13.770	14.255	13 640
Mg	6.460	5.900	6.840	7.135	6.705	6.960	5.920	6.545
Mn	0.255	0.295	0.230	0.240	0.260	0.230	0.215	0.260
Fe	15.470	17.360	15.010	16.420	16.185	14.635	13.360	16.340
Ti	3.495	2.880	3.280	2.165	2.105	1.965	1.685	2.785
Ca	8.550	8.485	9.970	9.705	10.670	9.245	10.060	10.840
K	3.090	2.380	2.190	0.600	0.860	1.445	0.995	0.695
Na	0.885	0.690	1.110	1.720	1.390	0.790	1.240	1.425
Total	98.745	100.400	100.990	100.140	98.700	99.150	99.425	98.980
Zr	231	177	178	123	134	106	99	179
Y	65	67	55	56	56	54	43	64
Rb	157	104	91	19	10	49	47	8
Nb	9	5	5	3	5	1	4	3
Sr	34	32	39	116	135	87	221	90
Th	0	1	0	3	1	1	4	5

← Gt. Amph. → ← Hb. Sch. →

	166/843	169/900	178/971	184/1006	209/1082	75/243	172/952	221/1089
P	0.195	0.200	0.435	0.270	0.250	0.220	0.260	0.275
Si	45.740	46.905	45.615	43.015	47.530	55.170	47.270	44.370
Al	13.370	14.120	14.890	14.305	13.775	15.160	13.660	13.895
Mg	6.975	7.070	6.590	7.480	6.620	6.310	6.905	7.350
Mn	0.275	0.265	0.250	0.240	0.335	0.170	0.245	0.270
Fe	16.580	15.500	15.495	14.835	15.345	9.880	15.515	16.855
Ti	3.285	2.080	3.445	1.965	2.105	1.540	2.310	2.515
Ca	10.950	9.740	11.035	9.840	10.370	8.650	10.055	9.420
K	0.895	1.340	0.645	1.215	1.230	1.170	1.220	2.135
Na	1.435	1.480	0.820	0.815	1.675	2.890	1.135	1.280
Total	99.700	98.695	99.215	98.990	99.230	101.160	98.570	96.355
Zr	125	120	266	125	126	180	136	141
Y	51	50	69	52	53	48	58	64
Rb	17	43	12	33	27	38	27	121
Nb	3	8	1	3	3	4	0	3
Sr	72	141	114	137	152	165	88	196
Th	0	1	3	0	1	5	3	3

← Hornblende Schists →

	20/118	22/138	23/143	54/232	56/232	73/240	108/470	127/549
P	0.175	0.335	0.180	0.235	0.200	0.180	0.140	0.115
Si	49.955	4 .290	48.860	48.310	50.260	48.915	49.220	49.810
Al	14.540	14.665	16.125	14.890	15.250	14.825	14.760	15.360
Mg	7.850	7.445	8.460	6.900	6.980	7.870	7.895	9.465
Mn	0.195	0.235	0.190	0.195	0.200	0.215	0.230	0.200
Fe	11.280	14.240	11.475	12.420	12.990	12.420	11.760	10.305
Ti	1.540	2.835	1.745	1.830	1.820	1.850	1.640	1.155
Ca	9.910	10.440	10.310	8.250	6.490	9.710	11.920	9.610
K	1.350	0.985	1.110	2.725	3.730	2.555	0.915	0.990
Na	2.440	1.175	0.905	2.405	2.200	2.425	1.995	2.660
Total	99.250	100.650	99.375	98.165	100.100	100.970	100.465	99.670
Zr	98	205	96	96	101	75	98	57
Y	31	59	35	44	41	33	22	25
Rb	48	15	36	169	260	115	15	47
Nb	1	5	3	9	11	3	3	1
Sr	116	122	65	80	75	318	321	315
Th	1	1	0	4	1	3	0	4

Major elements are listed as Wt.% oxides.

Appendix 2.

← Porphyroblastic Hornblende Schists. →

	28/174	88/281	89/293	92/333	94/346	124/543	128/549	132/581
P	0.580	0.340	0.130	0.200	0.110	0.130	0.130	0.220
Si	42.280	46.745	48.575	49.200	48.465	47.050	48.080	48.720
Al	14.375	14.155	16.790	14.800	14.385	16.700	16.785	14.910
Mg	6.850	7.230	9.165	4.485	7.815	8.585	9.200	8.080
Mn	0.325	0.240	0.175	0.205	0.240	0.210	0.170	0.220
Fe	16.340	13.720	9.890	11.815	12.745	10.455	9.895	12.390
Ti	4.415	2.710	1.085	1.685	2.010	1.225	1.065	1.995
Ca	10.040	10.510	11.480	12.025	10.695	12.145	10.895	10.485
K	2.655	1.855	0.790	0.645	1.150	0.980	0.700	1.190
Na	1.230	1.565	2.155	1.760	1.745	2.250	2.155	1.795
Total	99.100	99.070	100.225	100.825	99.365	99.735	99.070	100.710
Zr	378	216	57	94	104	69	57	119
Y	88	60	26	114	44	26	24	41
Rb	90	85	30	17	33	35	31	85
Nb	11	6	1	1	1	2	1	3
Sr	108	217	155	129	181	141	255	303
Th	0	0	1	3	1	1	0	3

Major elements are listed as Wt.% oxides.

Appendix 2.

Microdiorites

	33/229	57/234	71/237	74/241	83/255	85/259	91/318	93/335	99/418
P	0.215	0.250	0.005	0.195	0.315	0.485	0.285	0.375	0.355
Si	56.825	58.615	72.985	63.355	54.850	50.895	56.220	56.425	51.350
Al	15.975	17.690	15.985	16.740	15.450	16.240	15.770	15.450	11.230
Mg	5.950	4.095	0.295	3.175	6.890	7.890	6.060	6.710	8.455
Mn	0.135	0.125	0.070	0.090	0.145	0.155	0.180	0.120	0.135
Fe	7.365	7.025	1.350	5.275	7.805	9.040	8.440	7.610	9.425
Ti	1.005	1.040	0.200	0.840	1.200	1.770	1.330	1.270	1.620
Ca	6.655	4.785	1.630	4.585	8.260	7.960	6.825	6.565	8.310
K	1.610	2.755	3.630	1.705	1.370	1.795	1.340	1.850	1.215
Na	4.390	4.270	5.055	4.670	3.440	3.380	3.350	3.925	3.220
Total	100.060	100.655	101.205	100.635	99.735	99.610	99.850	100.310	99.305
Zr	168	190	141	177	180	230	228	263	155
Y	23	26	25	23	23	32	123	110	28
Rb	42	107	100	53	27	74	59	44	28
Nb	14	12	14	12	5	11	14	15	9
Sr	592	733	559	826	874	781	661	793	689
Th	6	12	6	11	6	2	5	5	0

	112/503	113/503	114/503	115/506	117/512	118/514	125/546	142/654	151/721
P	0.260	0.275	0.195	0.205	0.220	0.220	0.160	0.250	0.375
Si	58.620	62.770	59.310	52.730	55.230	50.250	60.535	57.590	50.910
Al	16.900	17.495	15.575	13.875	15.675	12.075	15.705	16.125	16.245
Mg	3.450	2.170	5.915	8.480	6.275	12.230	5.830	5.030	6.125
Mn	0.140	0.100	0.120	0.155	0.145	0.250	0.110	0.115	0.135
Fe	7.480	5.720	5.905	9.200	8.180	10.560	5.380	6.345	8.835
Ti	1.250	0.880	0.985	1.520	1.315	1.225	0.890	1.055	1.230
Ca	5.690	4.075	5.370	8.320	3.595	7.885	4.880	5.575	8.455
K	2.330	1.960	1.830	1.365	1.950	1.725	1.130	1.785	1.375
Na	4.130	5.395	4.065	2.905	3.960	1.895	4.035	4.595	4.380
Total	100.270	100.850	99.260	98.660	96.545	98.310	98.665	98.570	98.055
Zr	200	255	154	162	205	135	135	211	176
Y	34	24	21	32	30	28	19	25	27
Rb	65	51	53	37	95	80	50	49	31
Nb	14	17	13	5	15	11	12	11	12
Sr	672	835	560	465	344	282	511	748	934
Th	5	7	2	2	6	4	5	4	4

	154/721	155/723	156/724	158/728	180/985	182/986	202/1063	218/1089	219/1092
P	0.400	0.265	0.215	0.24	0.360	0.260	0.295	0.550	0.190
Si	53.680	59.650	53.520	57.915	52.430	51.630	51.685	48.480	49.285
Al	18.440	18.515	14.570	17.085	15.475	15.080	13.270	11.725	11.775
Mg	4.610	2.395	8.755	4.265	7.790	8.575	11.420	11.960	11.745
Mn	0.140	0.070	0.145	0.105	0.150	0.160	0.120	0.200	0.190
Fe	8.505	5.215	8.415	6.465	8.455	9.145	8.010	10.060	11.130
Ti	1.580	0.750	1.185	0.840	1.295	1.220	1.150	1.165	1.195
Ca	4.585	4.765	7.380	4.595	8.140	9.080	6.410	11.275	9.360
K	1.255	1.465	1.140	3.200	1.420	1.360	2.300	1.700	1.590
Na	4.625	5.515	3.100	3.825	3.325	2.320	3.370	0.985	2.865
Total	97.815	98.595	98.625	98.535	98.835	98.815	98.030	98.100	99.390
Zr	224	195	149	177	180	136	146	137	153
Y	41	17	21	25	27	24	22	28	18
Rb	55	53	39	150	36	42	50	73	34
Nb	18	22	9	10	18	14	7	10	6
Sr	829	1017	527	736	879	490	545	838	308
Th	2	4	2	2	6	5	3	7	1

Major elements are listed as Wt.% oxides.

Appendix 2.

Microdiorites

	226/1092	227/1092	228(1)/1092	228(2)/1092	229(1)/1092	229(2)/1092	229(3)/1092	229(4)/1092
P	0.425	0.350	0.485	0.435	0.390	0.365	0.425	0.485
Si	51.955	49.350	51.535	50.565	48.950	49.570	49.225	50.665
Al	15.855	14.745	14.930	13.870	11.980	12.090	13.810	14.110
Mg	6.970	8.705	8.290	9.880	12.565	12.215	9.190	9.445
Mn	0.135	0.180	0.165	0.160	0.170	0.170	0.150	0.150
Fe	7.790	10.035	8.835	9.140	9.330	9.290	8.700	8.485
Ti	1.430	1.270	1.250	1.235	1.095	1.155	1.335	1.180
Ca	8.530	8.850	9.555	0.265	10.405	10.235	9.940	9.380
K	1.680	1.890	1.400	1.200	1.375	1.495	1.310	1.360
Na	4.085	3.510	3.060	2.255	1.605	1.765	3.420	2.845
Total	98.845	98.575	99.515	99.000	97.865	98.360	97.510	98.110
Zr	156	144	169	153	131	118	152	168
Y	34	24	28	23	24	24	26	22
Rb	36	60	43	32	36	48	40	43
Nb	19	8	14	5	10	7	11	14
Sr	975	915	1089	1070	702	719	1128	1133
Th	8	5	12	10	7	13	10	11

Major elements are listed as Wt.% oxides.

Appendix 2.

← Permo-Carboniferous Camptonites →

	100/429	101/429	106/467	109/472	126/549	274/1182	275/1228	276/1228	145/656
P	0.605	0.785	0.745	0.710	1.260	0.920	0.585	0.600	0.950
Si	39.920	41.995	42.065	38.290	38.535	43.255	43.870	43.270	31.435
Al	11.655	13.815	12.350	12.020	11.855	12.875	14.080	14.220	13.635
Mg	7.670	6.420	9.590	7.840	11.240	9.730	7.820	8.110	6.160
Mn	0.170	0.150	0.170	0.170	0.200	0.185	0.170	0.220	0.160
Fe	13.255	13.730	11.200	12.160	11.495	12.060	10.780	10.880	18.655
Ti	2.940	3.135	2.645	2.880	2.655	2.945	2.550	2.495	2.105
Ca	11.590	10.530	11.475	13.660	14.730	12.475	10.620	11.030	11.385
K	1.985	2.680	3.285	2.340	1.870	0.745	2.015	1.890	2.405
Na	1.635	1.985	0.965	1.505	0.470	3.520	2.960	3.155	0.035
Total	91.430	95.220	94.490	91.570	94.300	98.710	95.455	95.885	86.345
Zr	250	322	338	291	306	288	275	293	141
Y	22	28	30	27	30	33	31	33	32
Rb	52	67	93	61	45	20	66	64	65
Nb	74	100	125	60	118	84	107	117	34
Sr	1254	998	1093	784	1984	641	1120	1105	120
Th	6	5	11	3	8	6	6	9	0

← Tertiary Dolerites →

	78/245	79/246	98/400	107/469	242/1094	257/1136	258/1139		96/357
P	0.240	0.180	0.130	0.330	0.420	0.245	0.155		0.760
Si	54.695	54.410	57.615	46.430	48.520	46.850	55.650		59.410
Al	12.350	15.835	15.030	13.760	15.300	14.045	14.965		14.890
Mg	9.325	5.965	6.930	7.120	7.535	9.330	5.120		5.745
Mn	0.165	0.140	0.150	0.240	0.145	0.210	0.140		0.090
Fe	12.135	10.115	9.795	10.270	10.315	12.150	10.160		5.595
Ti	1.840	1.255	1.115	1.570	2.095	1.760	1.295		0.995
Ca	8.065	10.135	7.350	13.620	10.595	10.705	8.520		3.975
K	0.740	0.310	1.080	0.390	1.035	0.440	0.625		3.385
Na	2.735	2.695	3.025	2.290	3.085	2.385	3.190		4.100
Total	102.685	101.020	100.210	96.030	99.045	98.120	99.830		98.940
Zr	102	62	72	94	133	87	79		269
Y	24	20	22	21	27	20	20		19
Rb	21	1	32	4	18	4	11		59
Nb	10	7	8	19	26	11	4		21
Sr	351	307	622	463	534	391	318		1392
Th	0	0	1	0	3	1	0		12

145/656 Highly altered (?) Camptonite

96/357 "Granitic" sheet

Major elements are listed as Wt.% oxides.

Appendix 3.

Modal analyses of Loch Eil Division Psammites.

	18/109	29/180	35/229	38/229	39/229	86/276	87/279	90/301	95/335
QUARTZ	45.3	61.8	79.5	56.8	80.2	64.0	77.1	87.2	79.2
FELDSPAR	45.4	30.2	9.2	10.5	10.8	3.1	4.4	6.4	8.8
MUSCOVITE	0.4	0.0	3.3	17.4	5.9	25.7	18.4	2.1	0.1
BIOTITE	8.7	3.6	7.5	14.9	2.5	3.5	0.1	0.3	1.6
GARNET	0.0	0.0	0.0	0.0	0.0	2.0	0.0	0.0	0.0
EPIDOTE	0.1	3.3	0.0	0.0	0.1	0.0	0.0	3.1	9.2
CHLORITE	0.0	0.6	0.6	0.0	0.2	1.2	0.0	0.4	0.0
CALCITE	0.0	0.0	0.1	0.0	0.3	0.0	0.0	0.0	0.0
ACCESSORIES	0.1	0.5	0.0	0.4	0.0	0.5	0.0	0.5	1.1

	97/376	102/436	111/482	114/503	120/520	123/539	130/572	190/1032	191/1032
QUARTZ	37.1	25.2	59.2	50.4	41.4	71.5	53.1	65.8	81.7
FELDSPAR	34.5	30.9	32.1	31.0	46.1	19.0	20.3	25.1	2.3
MUSCOVITE	4.7	21.0	0.7	0.3	1.0	3.5	7.4	5.2	14.6
BIOTITE	23.5	19.0	5.9	2.6	9.4	5.8	13.1	3.2	1.2
GARNET	0.0	0.0	0.0	0.8	0.0	0.0	0.0	0.0	0.0
EPIDOTE	0.0	0.2	0.8	12.2	0.7	0.0	0.2	0.1	0.0
CHLORITE	0.0	0.0	1.2	0.8	0.2	0.0	5.5	0.1	0.0
CALCITE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
ACCESSORIES	0.2	3.7	0.1	1.9	1.2	0.2	0.4	0.5	0.2

	192/1032	193/1035	194/1036	195/1037	196/1038	197/1039		Average
QUARTZ	50.0	72.7	20.6	65.8	66.6	79.0		61.30
FELDSPAR	18.4	5.6	77.2	14.1	18.8	16.2		22.01
MUSCOVITE	22.8	14.6	0.8	19.2	3.4	4.1		8.20
BIOTITE	7.2	6.6	0.3	0.5	9.2	0.3		6.27
GARNET	0.0	0.0	0.0	0.0	0.0	0.0		0.17
EPIDOTE	0.0	0.0	0.0	0.0	0.0	0.0		1.25
CHLORITE	0.3	0.2	0.7	0.3	0.3	0.0		0.53
CALCITE	0.0	0.0	0.0	0.0	0.0	0.0		0.02
ACCESSORIES	1.3	0.3	0.4	0.1	1.7	0.4		0.61

500 Points per section

Appendix 3.

Modal analyses of Glenfinnan Division Psammites.

	9/60	26/164	31/214	44/230	148/672	161/783	177/963	188/1031	205/1065	214/1087
QUARTZ	44.6	76.8	48.6	54.0	40.0	73.0	65.8	54.6	71.0	63.8
FELDSPAR	38.6	19.2	36.6	23.2	28.2	24.2	29.0	29.8	18.0	30.8
MUSCOVITE	0.0	0.0	0.0	18.2	21.6	1.0	1.0	9.0	5.8	1.0
BIOTITE	16.8	4.0	14.4	4.2	9.6	1.8	4.0	6.4	2.6	4.4
GARNET	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
EPIDOTE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
CHLORITE	0.0	0.0	0.0	0.0	0.0	0.0	0.2	0.0	0.4	0.0
CALCITE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
ACCESSORIES	0.0	0.0	0.4	0.4	0.6	0.0	0.0	0.2	0.2	0.0

	236/1093	238/1093	260/1163	261/1164	262/1165	263/1166	264/1167	265/1168	266/1169
QUARTZ	64.8	22.2	58.1	68.5	68.5	75.6	50.7	70.9	67.5
FELDSPAR	23.4	57.4	32.7	26.4	26.8	14.5	41.5	22.9	24.7
MUSCOVITE	2.8	0.0	0.0	1.2	1.5	6.6	0.4	2.2	2.4
BIOTITE	8.8	20.2	8.4	3.3	2.8	3.1	7.0	4.0	3.0
GARNET	0.0	0.0	0.0	0.0	0.0	0.0	0.3	0.0	0.8
EPIDOTE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
CHLORITE	0.0	0.0	0.1	0.1	0.1	0.2	0.0	0.0	1.2
CALCITE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
ACCESSORIES	0.2	0.2	0.8	0.5	0.3	0.0	0.1	0.0	0.4

500 Points per section

Appendix 3.

Modal analyses of Morar Division Psammites.

	240/1094	244/1097	267/1170	268/1171	269/1172	270/1173	271/1174	272/1175	273/1176
QUARTZ	54.2	61.8	76.8	61.2	61.9	68.3	60.3	58.4	68.2
FELDSPAR	32.8	29.4	20.8	27.6	33.8	27.7	35.6	34.9	22.3
MUSCOVITE	2.4	1.6	0.9	0.4	2.4	0.1	1.0	0.0	3.5
BIOTITE	10.6	6.6	0.7	10.1	1.9	3.6	3.1	5.2	5.5
GARNET	0.0	0.0	0.1	0.3	0.0	0.0	0.0	0.0	0.1
EPIDOTE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
CHLORITE	0.0	0.2	0:4	0.1	0.0	0.3	0.0	1.5	0.3
CALCITE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
ACCESSORIES	0.0	0.4	0.3	0.3	0.0	0.0	0.0	0.0	0.1

500 Points per section.

Appendix 3.

Modal analyses of Calc-silicate rocks.

	34/229	36/229	37/229	40/229	103/440	104/451	105/451	119/519
QUARTZ	79.1	75.3	55.6	77.5	79.0	37.3	42.7	44.3
PLAGIOCLASE	10.4	18.4	22.8	5.5	15.2	18.7	18.8	32.3
AMPHIBOLE	0.0	4.0	8.4	1.2	0.0	10.9	13.1	5.0
BIOTITE	0.0	0.0	0.0	0.0	0.4	0.0	0.0	0.0
MUSCOVITE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
GARNET	0.1	0.0	7.2	0.3	0.0	4.1	1.9	3.4
PYROXENE	0.0	0.0	1.9	1.3	0.0	0.0	0.0	0.0
CALCITE	0.0	0.0	0.4	0.7	0.0	0.0	0.0	0.0
EPIDOTE	0.0	0.0	0.0	0.0	5.2	0.0	0.0	0.0
CLINOZOISITE	9.9	1.5	2.5	13.2	0.0	26.8	21.6	11.7
CHLORITE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	1.7
ACCESSORIES	0.5	0.0	1.2	0.3	0.0	2.2	1.9	1.6

	131/573	134/582	48/231	81/250	167/883	170a/948	170b/948	174/956
QUARTZ	51.8	56.4	83.8	68.2	48.9	52.8	47.8	70.9
PLAGIOCLASE	36.9	12.2	8.3	24.2	26.5	33.0	29.7	15.7
AMPHIBOLE	2.5	0.0	1.7	0.4	10.2	7.3	5.6	5.1
BIOTITE	0.0	0.0	0.1	0.9	0.0	0.2	0.1	0.0
MUSCOVITE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
GARNET	3.8	0.0	0.3	0.0	3.7	1.8	12.5	6.5
PYROXENE	0.0	0.0	0.1	0.0	0.0	0.0	0.0	0.0
CALCITE	0.3	0.2	0.1	0.0	0.0	0.0	0.0	0.0
EPIDOTE	0.0	30.6	0.0	0.0	0.0	0.0	0.0	1.2
CLINOZOISITE	3.4	0.0	5.4	5.4	9.3	1.9	3.4	0.0
CHLORITE	0.0	0.2	0.0	0.0	0.4	0.0	0.0	0.0
ACCESSORIES	0.6	0.6	0.2	0.9	1.0	3.0	0.9	0.6

	175/961	187/1031	200/1050	204/1065	207/1068	208/1071	212/1087	213/1087
QUARTZ	43.5	66.0	63.3	60.5	50.4	53.4	58.2	59.3
PLAGIOCLASE	37.9	23.1	26.5	28.0	36.0	20.0	32.9	29.1
AMPHIBOLE	9.5	3.6	2.7	5.6	1.9	3.7	0.1	0.1
BIOTITE	1.1	0.1	0.0	2.5	0.0	0.0	0.0	1.7
MUSCOVITE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.7
GARNET	4.5	3.7	4.7	1.2	5.7	16.4	2.1	3.1
PYROXENE	0.0	0.0	0.4	0.0	0.0	1.6	0.0	0.0
CALCITE	0.1	0.8	0.0	0.0	2.0	0.0	0.0	0.4
EPIDOTE	0.5	0.0	0.0	0.0	0.0	0.0	0.0	0.9
CLINOZOISITE	0.0	1.7	0.5	0.4	1.3	0.5	1.3	0.0
CHLORITE	1.3	0.3	0.3	0.5	1.5	1.7	3.5	3.1
ACCESSORIES	1.6	0.7	1.6	1.3	1.2	2.7	1.9	1.6

	215/1088	225/1091	252/1110	253/1115	254/1119	255/1123	256/1126	259/1145
QUARTZ	54.3	52.3	51.3	70.2	47.9	74.2	69.5	56.2
PLAGIOCLASE	36.7	37.9	41.1	24.8	36.7	20.3	21.2	33.1
AMPHIBOLE	0.7	0.1	2.8	0.3	6.7	0.3	0.4	3.5
BIOTITE	1.3	4.4	0.7	0.3	0.0	0.0	0.0	0.0
MUSCOVITE	1.4	1.7	0.5	0.3	0.3	0.3	0.2	0.6
GARNET	2.8	0.8	0.8	0.7	6.6	2.0	4.0	2.6
PYROXENE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
CALCITE	0.0	0.4	0.0	0.1	0.0	0.0	0.0	0.0
EPIDOTE	0.0	0.0	0.0	2.5	0.0	0.0	2.1	0.0
CLINOZOISITE	0.1	0.0	0.5	0.0	0.4	1.3	0.0	2.0
CHLORITE	1.9	1.5	0.7	0.3	0.9	1.1	2.0	1.1
ACCESSORIES	0.8	0.9	1.6	0.5	0.5	0.5	0.6	0.6

1000 Points per section.

Appendix 3.

Modal analyses of striped and semi-pelitic rocks.

	← Glenfinnan Division →						
	8/60 ●	13/90 ●	19/112 ●	21/138 ●	32/217 ●	136/588 ●	138/643 X
QUARTZ	31.2	22.2	29.8	32.6	24.5	16.2	49.2
FELDSPAR	28.4	45.8	33.2	37.2	43.2	47.0	14.8
BIOTITE	20.8	23.4	25.0	10.2	1.4	25.6	9.6
MUSCOVITE	14.0	8.2	11.2	10.6	9.4	10.2	26.0
GARNET	3.8	0.0	0.0	0.0	0.0	0.0	0.0
IRON ORES	1.2	0.4	0.4	0.4	0.2	0.2	0.4
CHLORITE	0.0	0.0	0.2	8.8	20.8	0.0	0.0
ACCESSORIES	0.6	0.0	0.2	0.2	0.0	0.8	0.0

	← Glenfinnan →			← Loch Eil →		
	146/664 X	163/799 X	201/1086 ●	110/481 ●	121/525 ●	129/550 ●
QUARTZ	55.8	41.8	29.6	29.4	23.0	43.6
FELDSPAR	27.8	18.8	46.0	36.2	12.4	14.2
BIOTITE	6.6	18.4	17.8	32.0	35.4	21.2
MUSCOVITE	9.6	17.6	5.6	1.2	28.4	13.8
GARNET	0.2	0.2	0.0	0.0	0.2	2.4
IRON ORES	0.0	0.0	0.0	0.4	0.4	4.8
CHLORITE	0.0	0.0	0.8	0.0	0.0	0.0
ACCESSORIES	0.2	0.6	0.2	0.8	0.2	0.0
SILLIMANITE		2.6				

500 Points per section

● Semi-pelitic rock.

X Striped lithology.

Appendix 3.

Modal analyses of Glenfinnan Division Pelites.

	10/64	12/70	16/106	17/107	24/143	25/155	42/230	43/230	47/231
QUARTZ	15.6	23.8	20.8	23.4	20.8	18.6	40.8	20.6	21.6
FELDSPAR	26.4	31.4	52.4	39.8	46.2	47.6	35.6	34.8	18.2
BIOTITE	30.0	19.8	23.6	27.6	24.8	17.8	14.4	23.2	23.6
MUSCOVITE	23.2	16.6	1.4	7.6	5.8	10.0	8.8	18.0	34.8
GARNET	3.6	5.4	0.0	0.6	1.8	6.8	0.0	3.0	0.4
CHLORITE	0.6	1.2	0.2	0.0	0.0	0.2	0.2	0.2	0.0
ACCESSORIES	0.6	0.6	1.4	0.6	0.2	0.0	0.2	0.0	0.2
IRON ORES	0.0	1.2	0.2	0.4	0.2	1.0	0.0	0.2	1.2
SILLIMANITE	0.0	0.0	0.0	0.0	0.2	0.0	0.0	0.0	0.0

	135/583	82a/251	82b/251	Total 82/251	137/609	153/721	157/727	160/776	162/794
QUARTZ	29.0	5.8	17.4	11.6	44.8	24.8	31.2	7.8	30.2
FELDSPAR	0.2	28.0	48.0	38.0	26.0	37.4	11.8	23.6	18.6
BIOTITE	22.8	51.6	17.4	34.5	0.4	32.4	18.0	8.2	12.6
MUSCOVITE	17.0	0.2	0.6	0.	0.4	1.6	37.6	31.2	32.4
GARNET	0.6	2.6	1.6	2.2	15.8	3.0	0.2	8.4	3.8
CHLORITE	0.0	0.0	0.0	0.0	0.4	0.0	0.0	15.8	0.0
ACCESSORIES	0.2	0.2	0.6	0.4	0.4	0.4	0.4	0.0	0.4
IRON ORES	0.2	0.8	0.4	0.6	0.8	0.4	0.8	1.6	0.8
AMPHIBOLE	0.0	10.4	12.6	11.5	11.0	0.0	0.0	0.0	0.0
CLINOZOISITE	0.0	0.4	1.4	0.9	0.0	0.0	0.0	0.0	0.0
STAUROLITE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.4	0.0
SILLIMANITE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	1.0	1.2

	164/802	168/883	171/948	173/954	176/961	179/971	181/986	186/1012	189/1031
QUARTZ	31.2	30.0	28.4	45.4	36.8	11.8	20.6	19.0	31.0
FELDSPAR	27.2	16.2	14.6	20.4	27.4	63.0	17.8	27.2	28.6
BIOTITE	15.2	22.6	15.0	11.2	10.6	21.2	33.8	30.2	18.8
MUSCOVITE	18.4	30.8	35.6	19.0	24.2	1.6	25.4	21.6	15.6
GARNET	5.4	0.0	5.4	2.4	0.0	0.2	1.8	1.2	1.4
CHLORITE	0.0	0.0	0.0	0.2	0.2	2.0	0.0	0.0	0.6
ACCESSORIES	1.2	0.2	0.2	0.0	0.2	0.2	0.4	0.6	0.2
IRON ORES	1.4	0.2	0.8	0.4	0.6	0.0	0.0	0.2	1.8

500 Points per section

Appendix 3.

Modal analyses of Glenfinnan Division Pelites.

	199/1045	201/1050	203/1064	206/1065	211/1086	217/1089	223/1090	224/1090	232/1092	234/1093
QUARTZ	31.4	37.6	38.8	26.0	21.0	26.6	32.2	23.4	18.6	28.2
FELDSPAR	18.2	12.0	18.4	33.0	36.2	17.0	39.2	41.0	26.8	19.0
BIOTITE	15.6	19.6	15 2	28.2	20.0	13.2	22.8	17.6	31.4	14.6
MUSCOVITE	31.6	29.6	25.4	11.4	14.6	39.4	4.2	15.0	21.6	31.4
GARNET	0.6	0.0	0.0	0.2	5.0	0.0	0.4	1.6	0.8	2.6
CHLORITE	0.2	0.0	0.0	0.6	0.8	0.0	0.8	0.6	0.4	0.6
ACCESSORIES	0.4	0.2	0.0	0.2	0.2	0.2	0.2	0.2	0.2	0.0
IRON ORES	1.2	1.0	0.8	0.2	1.6	3.6	0.2	0.6	0.0	2.2
SILLIMANITE	0.8	0.0	0.4	0.2	0.6	0.0	0.0	0.0	0.2	1.4

	235/1093	237/1093	243/1096	245/1104	246/1104	247/1104	248/1104	249/1104	250/1104
QUARTZ	9.0	20.4	42.0	29.6	25.2	23.4	15.6	22.8	22.6
FELDSPAR	45.8	40.2	37.2	27.4	44.4	39.6	36.4	47.4	40.0
BIOTITE	39.0	25.4	6.6	22.6	24.6	31.4	33.8	21.2	24.0
MUSCOVITE	5.8	10.2	11.8	11.6	5.2	5.4	11.2	8.0	12.0
GARNET	0.0	2.0	0.2	5.4	0.2	0.2	1.4	0.0	0.8
CHLORITE	0.0	1.0	1.8	0.2	0.2	0.0	0.0	0.4	0.4
IRON ORES	0.2	0.6	0.0	1.8	0.0	0.0	1.0	0.0	0.2
ACCESSORIES	0.2	0.0	0.2	0.0	0.0	0.0	0.6	0.2	0.0
STAUROLITE	0.0	0.0	0.0	0.2	0.0	0.0	0.0	0.0	0.0
SILLIMANITE	0.0	0.2	0.2	1.2	0.2	0.0	0.0	0.0	0.0

500 Points per section

Appendix 3.

Modal analyses of Ardgour granitic gneiss.

	6/60	7/60	14/92	15/93	27/164	53/232	55/232	58/235	59/235
QUARTZ	28.2	30.6	36.4	40.0	38.6	36.6	14.6	40.0	15.6
PLAGIOCLASE	25.0	29.8	46.6	36.4	15.8	39.4	41.8	42.6	31.2
K-FELDSPAR	30.2	21.6	12.4	16.2	19.2	10.6	5.0	1.4	53.2
MYRMEKITE	6.4	2.0	0.0	2.0	4.4	0.4	0.0	0.0	0.0
GARNET	0.0	0.0	0.0	0.0	0.0	0.6	0.0	1.8	0.0
BIOTITE	9.4	15.4	0.4	2.6	13.6	11.0	37.0	13.6	0.0
MUSCOVITE	0.0	0.4	2.4	1.4	8.0	0.0	0.0	0.0	0.0
CHLORITE	0.0	0.2	1.4	1.4	0.0	0.2	0.0	0.0	0.0
'EPIDOTE'	0.0	0.0	0.2	0.0	0.0	0.4	1.2	0.4	0.0
ACCESSORIES	0.8	0.0	0.2	0.0	0.4	0.8	0.4	0.2	0.0

	68/235	70/235	77/244	277/1235	278/1235	280/1235	281/1235	Peg
QUARTZ	35.4	28.0	30.6	24.6	29.6	26.6	32.8	23.8
PLAGIOCLASE	46.8	53.8	48.6	27.8	41.0	28.0	23.6	38.0
K-FELDSPAR	1.6	1.0	0.8	28.0	2.6	6.6	16.0	30.2
MYRMEKITE	0.0	0.0	0.0	3.8	2.6	0.6	1.4	5.2
GARNET	1.4	2.0	2.0	0.0	1.6	0.0	0.2	0.0
BIOTITE	14.0	15.2	17.4	14.8	24.6	29.2	26.0	2.0
MUSCOVITE	0.0	0.0	0.0	0.8	0.0	8.8	0.0	0.8
CHLORITE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
'EPIDOTE'	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
ACCESSORIES	0.6	0.0	0.6	0.2	0.0	0.2	0.0	0.0

500 Points per section

Appendix 4.

ABBREVIATIONS :

GF.	Glenfinnan Division.
LE.	Loch Eil Division.
Morar	Morar Division.
Ps.	Psammitic rock.
Pel.	Pelitic rock.
Striped.	Striped and semi-pelitic rock.
Calc-sil.	Calc-silicate rock.
A.g.g	Ardgour granitic gneiss.
Gt.Amph.	Garnetiferous amphibolite.
Hb.Sch.	Hornblende schist.
Porph.Hb.Sch.	Porphyroblastic hornb-ende schist.
Micro.	Microdiorite. (Caledonian)
Camptonite.	Camptonite. (Permo-Carb.)
Dolerite.	Dolerite. (Tertiary)

NOTE ON SAMPLE AND LOCALITY NUMBERING.

(Exp.234/1169). 1^{st} number is the sample number.
2^{nd} number is the locality number used on field slips and note books.

(Exp.159 NM 93568323). 1^{st} number is the locality number
2^{nd} number is the Grid Reference.

Appendix 4.

Sample Number	Exp. Number	Grid Reference	Lithology
1		Not known	
3		Not Known	
5		Not known	
7	60	NM 92988167	A.g.g.
9	60	NM 92988167	A.g.g.
11		Not known	
13	90	NM 93268233	GF.Striped
15	93	NM 94028222	A.g.g.
17	107	NM 94288236	GF.Pel.
19	112	NM 94478272	GF.Striped.
21	138	NM 92538163	GF.Striped.
23	143	NM 93078106	Hb.Sch.
25	155	NM 92188185	GF.Pel.
27	164	NM 92278193	A.g.g.
29	180	NM 94808232	LE.Ps.
31	214	NM 94538049	GF.Ps.
33	229	NM 90898047	Micro.
35	229	NM 90898047	LE.Ps.
37	229	NM 90898047	Calc-sil.
39	229	NM 90898047	LE.Ps.
41	230	NM 91068047	Gt.Amph.
43	230	NM 91068047	GF.Pel.
45	230	NM 91068047	GF.Ps.
47	231	NM 91278033	GF.Pel.
49	231	NM 91278033	Calc-sil.
51	231	NM 91278033	GF.Ps.
53	232	NM 91448030	A.g.g.
55	232	NM 91448030	A.g.g.
57	234	NM 91448028	Micro.
59	235	NM 91778020	A.g.g.
61	235	NM 91778020	Gt.Amph.
63	235	NM 91778020	Gt.Amph.
65	235	NM 91778020	Gt.Amph.
67	235	NM 91778020	Gt.Amph.
69	235	NM 91778020	Gt.Amph.
71	237	NM 91947998	Micro.
73	240	NM 92207994	Hb.Sch.
75	243	NM 92917977	Hb.Sch.
77	244	NM 93047988	A.g.g.
79	246	NM 93837978	Dolerite.
81	250	NM 95037921	Calc-sil.
83	255	NM 96278040	Micro.
85	259	NM 96288021	Micro.
87	279	NM 96187993	LE.Ps.
89	293	NM 96437961	Porph.Hb.Sch.
91	318	NM 96638152	Micro.
93	335	NM 97218231	Micro.
95	355	NM 98008148	LE.Ps.
97	375	NM 97678087	LE.Ps.
99	418	NM 97827976	Micro.

Sample Number	Exp. Number	Grid Reference	Lithology
2		Not known	
4	6	NM 91778020	Gt.Amph.
6	60	NM 92988167	A.g.g.
8	60	NM 92988167	GF.Striped
10	64	NM 93068148	GF.Pel.
12	70	NM 92738124	GF.Pel.
14	92	NM 94098228	A.g.g.
16	106	NM 94268229	GF.Pel.
18	109	NM 94398205	LE.Ps.
20	118	NM 93878124	Hb.Sch.
22	138	NM 92538163	Hb.Sch.
24	143	NM 93078106	GF.Pel.
26	164	NM 92278193	GF.Ps.
28	174	NM 94548203	Porph.Hb.Sch.
30	209	NM 95128234	GF.Pel.
32	217	NM 94628025	GF.Striped.
34	229	NM 90898047	Calc-sil.
36	299	NM 90898047	Calc-sil.
38	229	NM 90898047	LE.Ps.
40	229	NM 90898047	Calc-sil.
42	230	NM 91068047	GF.Pel.
44	230	NM 91068047	GF.Ps.
46	231	NM 91278033	GF.Ps.
48	231	NM 91278033	Calc-sil.
50	231	NM 91278033	Hb.Sch.
52	231	NM 91278033	GF.Ps.
54	232	NM 91448030	Hb.Sch.
56	232	NM 91448030	Hb.Sch.
58	235	NM 91778020	A.g.g.
60	235	NM 91778020	Gt.Amph.
62	235	NM 91778020	A.g.g.
64	235	NM 91778020	Gt.Amph.
66	236	NM 91778020	Gt.Amph.
68	235	NM 91778020	A.g.g.
70	235	NM 91778020	Gt.Amph.
72	240	NM 92207974	A.g.g.
74	241	NM 92407944	Micro.
76	244	NM 93047988	A.g.g.
78	245	NM 93637989	Dolerite.
80	248	NM 94327954	GF.Ps.
82	251	NM 95287917	GF.Pel.
84	259	NM 96288021	Granitic sheet.
86	276	NM 96077967	LE.Ps.
88	281	NM 96457988	Porph.Hb.Sch.
90	301	NM 96508053	LE.Ps.
92	333	NM 96918212	Porph.Hb.Sch.
94	346	NM 97778216	Porph.Hb.Sch.
96	357	NM 98188198	'Granitic'sheet.
98	400	NM 98258038	Dolerite.
100	429	NM 98107936	Camptonite.

Appendix 4.

Sample Number	Exp. Number	Grid Reference	Lithology
101	429	NM 98107936	Camptonite.
103	440	NM 98678008	Calc-sil.
105	451	NM 98927913	Calc-sil.
107	469	NM 99328017	Dolerite.
109	472	NM 99088030	Camptonite.
111	482	NM 98648152	LE.Ps.
113	503	NM 99948098	Micro.
114	503	NM 99948098	Micro.
116	510	NN 00017995	Fault breccia.
118	514	NN 00337940	Micro.
120	520	NN 00528110	LE.Ps.
122	537	NN 00528176	LE.Ps.
124	543	NN 00828291	Porph.Hb.Sch.
126	549	NN 01838161	Camptonite.
128	549	NN 01838161	Porph.Hb.Sch.
130	572	NN 01958098	LE.Ps.
132	581	NM 96118091	Porph.Hb.Sch.
134	582	NM 96068083	Calc-sil.
136	588	NM 94938288	GF.Striped.
138	643	NM 93178333	GF.Striped.
140	648	NM 92838258	Fault breccia.
142	654	NM 92728288	Micro.
144	656	NM 92798278	Fault breccia.
146	664	NM 92828311	LE.Striped.
148	672	NM 92828337	GF.Ps.
150	673	NM 92818345	Pegmatite.
152	721	NM 91538328	Gt.Amph.
154	722	NM 91618309	Micro.
156	724	NM 90218085	Micro.
158	728	NM 90408088	Micro.
160	776	NM 90788368	GF.Pel.
162	794	NM 89538437	GF.Pel.
164	802	NM 88728424	GF.Pel.
166	843	NM 89618348	Gt.Amph.
168	883	NM 88478358	GF.Pel.
170	948	NM 86378363	Calc-sil.
172	952	NM 86098412	Hb.Sch.
174	956	NM 87148173	Calc-sil.
176	961	NM 86818197	GF.Pel.
178	971	NM 85908237	Gt.Amph.
180	985	NM 86318155	Micro.
182	986	NM 86658152	Micro.
184	1006	NM 85338289	Gt.Amph.
186	1012	NM 86158263	GF.Pel.
188	1031	NM 84588166	GF.Ps.
190	1032	NN 02237992	LE.Ps.
192	1034	NN 00037993	LE.Ps.
194	1036	NM 99427941	LE.Ps.
196	1038	NM 98037978	LE.Ps.
198	1039	NM 97127927	Pegmatite.

Sample Number	Exp. Number	Grid Reference	Lithology
102	436	NM 98247998	LE.Ps.
104	451	NM 98927913	Calc-sil.
106	467	NM 98737952	Camptonite.
108	470	NM 99728014	Hb.Sch.
110	481	NM 98638132	LE.Striped.
112	503	NM 99948098	Micro.
114	505	NM 99938087	LE.Ps.(Arnipol).
115	506	NM 99928077	Micro.
117	512	NN 00187972	Micro.
119	517	NN 01088016	Calc-sil.
121	525	NN 00108207	LE.Striped.
123	539	NN 00548217	LE.Ps
125	546	NN 01688200	Micro.
127	549	NN 01838161	Camptonite.
129	550	NN 01848158	LE.Striped
131	573	NN 02138064	Calc-sil.
133	582	NM 96068083	Thrust cataclasite.
135	583	NM 95978053	GF.Pel.
137	609	NM 94728295	GF.Pel.
139	645	NM 92938356	GF.Ps.
141	649	NM 92838266	Micro.
143	654	NM 92728288	Gt.Amph.
145	656	NM 92798278	Alt.?Camptonite.
147	669	NM 94208313	Gt.Amph.
149	672	NM 92828337	GF.Ps.
151	721	NM 91538328	Micro.
153	721	NM 91538328	GF.Pel.
155	723	NM 91648286	Micro.
157	727	NM 90238146	GF.Pel.
159	740	NM 90138227	Gt.Amph.
161	783	NM 90328428	GF.Ps.
163	799	NM 89018453	LE.Striped.
165	817	NM 89888237	Gt.Amph.
167	883	NM 88478358	Calc-sil.
169	900	NM 86538357	Gt.Amph.
171	948	NM 86378363	GF.Pel.
173	954	NM 86078444	GF.Pel.
175	961	NM 86818197	Calc-sil.
177	963	NM 86788247	GF.Ps.
179	971	NM 85908237	GF.Pel.
181	986	NM 86638152	GF.Pel.
183	989	NM 85888147	GF.Ps.
185	1011	NM 86308254	GF.Ps.
187	1031	NM 84588166	Calc-sil.
189	1031	NM 84588166	GF.Pel.
191	1033	NN 02127894	LE.Ps.
193	1035	NN 00317902	LE.Ps.
195	1037	NM 98887947	LE.Ps.
197	1039	NM 97127927	LE.Ps.
199	1045	NM 84628213	GF.Pel.

Appendix 4.

Sample Number	Exp Number	Grid Reference	Lithology
200	1050	NM 84078233	Calc-sil.
202	1063	NM 83888182	Micro.
204	1065	NM 84138288	Calc-sil.
206	1065	NM 84138288	GF.Pel.
208	1071	NM 84428337	Calc-sil.
210	1086	NM 83588198	LE.Striped.
212	1087	NM 83448197	Calc-sil.
214	1087	NM 83448197	GF.Ps.
216	1088	NM 83358199	Pegmatite.
218	1089	NM 83238203	Micro.
220	1089	NM 83238203	HB.Sch.
222	1089	NM 83238203	GF.Ps.
224	1090	NM 81378206	GF.Pel.
226	1092	NM 83068207	Micro.
228	1092	NM 83028208	Micro.
230	1092	NM 83028208	Micro.
232	1092	NM 83028208	GF.Pel.
234	1093	NM 92988209	GF.Pel.
236	1093	NM 82978209	GF.Ps.
238	1093	NM 82948210	GF.Ps.
240	1094	NM 82678206	Morar Ps.
242	1094	NM 82678206	Dolerite.
244	1097	NM 81978235	Morar Ps.
246	1104	NM 84388379	GF.Pel.
248	1104	NM 83488379	GF.Pel.
250	1104	NM 83488379	GF.Pel.
252	1110	NM 82838248	Calc-sil.
254	1119	NM 83818324	Calc-sil.
256	1123	NM 84068361	Calc-sil.
258	1139	NM 82958375	Dolerite.
260	1163	NM 84638309	GF.Ps.
262	1165	NM 84358317	GF.Ps.
264	1167	NM 84068327	GF.Ps.
266	1169	NM 84128347	GF.Ps.
268	1171	NM 83878349	Morar Ps.
270	1173	NM 83738357	Morar Ps.
272	1175	NM 83528364	Morar Ps.
274	1182	NM 90858120	Camptonite.
276	1228	NM 85648135	Camptonite.
278	1235	NM 91778020	A.g.g.
280	1235	NM 91778020	A.g.g.
282	1254	NM 83448197	Calc-sil.
284	1255	NM 82838248	Calc-sil.
286	1261	NM 82138204	Micro.

Sample	Exp	Grid Reference	Lithology
201	1050	NM 84078233	GF.Pel.
203	1064	NM 83718227	GF.Pel.
205	1065	NM 84138288	GF.Ps.
207	1068	NM 84288345	Calc-sil.
209	1082	NM 84608375	Gt.Amph.
211	1086	NM 83588198	GF.Pel.
213	1087	NM 83448197	Calc-sil.
215	1088	NM 83358199	Calc-sil.
217	1089	NM 83238203	GF.Pel.
219	1089	NM 83238203	Micro.
221	1089	NM 83238203	Hb.Sch.
223	1090	NM 81378206	GF.Pel.
225	1091	NM 83108208	Calc-sil.
227	1092	NM 83068207	Micro.
229	1092	NM 83028208	Micro.
231	1092	NM 83028208	Pegmatite.
233	1093	NM 82988209	GF.Ps.
235	1093	NM 82988209	GF.Pel.
237	1093	NM 82948210	GF.Pel.
239	1094	NM 82678206	Morar Pel.
241	1094	NM 82678206	Morar Ps.
243	1096	NM 82138204	Morar Pel.
245	1104	NM 84388379	GF.Pel.
247	1104	NM 84388379	GF.Pel.
249	1104	NM 83488379	GF.Pel.
251	1104	NM 83488379	GF.Ps.
253	1115	NM 83708270	Calc-sil.
255	1123	NM 84068361	Calc-sil.
257	1136	NM 83528390	Dolerite.
259	1145	NM 82938398	Calc-sil.
261	1164	NM 84478304	GF.Ps.
263	1166	NM 84158320	GF.Ps.
265	1168	NM 84038345	GF.Ps.
267	1170	NM 83938348	Morar Ps.
269	1172	NM 83808352	Morar Ps.
271	1174	NM 83608360	Morar Ps.
273	1176	NM 83418367	Morar Ps.
275	1228	NM 85648135	Camptonite.
277	1235	NM 91778020	A.g.g.
279	1235	NM 91778020	Pegmatite.
281	1235	NM 91778020	A.g.g.
283	1254	NM 83448197	Calc-sil.
285	1256	NM 83198265	Calc-sil.
287	1263	NM 90828050	Micro.

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The metamorphic environment of the Sgurr Beag Slide; a major crustal displacement zone in Proterozoic, Moine rocks of Scotland

D. Powell, A. W. Baird, N. R. Charnley & P. J. Jordan

SUMMARY: New evidence from the SW Northern Highlands of Scotland establishes the southerly extension of the Sgurr Beag Slide. Interpretation of changes in the mineralogy of calc-silicate rocks, particularly the use of the anorthite content of plagioclase feldspar as an index of metamorphic grade, together with consideration of the texture and mineralogy of pelitic rocks, suggests that the slide juxtaposes crustal segments of different metamorphic grade and history. Movement on the slide post-dates a major phase of regional metamorphism and two phases of regional deformation of Precambrian age, was synchronous with a regional fold-forming event and regional metamorphism at or before c. 467 Ma (Caledonian), and was followed by regional folding.

Models of the thermal effects of displacement across syn-metamorphic shear zones are investigated with particular reference to the roles of thermal relaxation and levels of shear strain. A model based on the development of a ductile, syn-metamorphic, asymmetric shear zone best explains the metamorphic patterns associated with the Sgurr Beag Slide. An original low to moderate easterly dip is implied, with north-westerly upthrusting of the eastern crustal block.

The Sgurr Beag Slide was first recognized in the Kinloch Hourn area of western Scotland (Fig. 1, inset). It is a major tectonic break since it has been traced for some 100 km through Moine rocks of the north-western part of the orthotectonic zone of the Caledonide orogenic belt (Tanner 1970; Tanner *et al.* 1970), and is associated with the emplacement of slivers of Archaean to early Proterozoic Lewisian basement at high stratigraphic levels within the younger Proterozoic Moine cover (Tanner *et al.* 1970; Rathbone & Harris 1979).

Separation of some of the Moine rocks into the Morar and Glenfinnan divisions by the Sgurr Beag Slide has been advocated because of the inferred large displacement upon the slide (Johnstone 1975). Despite the lack of quantitative data concerning both the age and displacement of the slide, speculations have been made on the ages of the Morar and Glenfinnan divisions on the basis of correlations with the less well known Grampian Highland 'Moine' and a few radiometric age determinations. The Glenfinnan division is claimed to be the oldest metamorphic, and by inference sedimentary, complex of the Moine (Piasecki & van Breemen 1979; van Breemen *et al.* 1978) which because of a Rb/Sr whole-rock isochron age of 1028 ± 43 Ma for a granitic gneiss body within it (Brook *et al.* 1976) is claimed to be Grenvillian. In contrast, the Morar division is thought to represent a post-Grenvillian complex (Piasecki & van Breemen 1979) despite isotopic evidence to the contrary (Brewer *et al.* 1979). Such speculations require the Sgurr Beag Slide to constitute or hide an orogenic front.

The amount of displacement across the Sgurr Beag Slide is difficult to establish because of its essentially conformable and syn-metamorphic nature and the lack of well defined markers. Consideration of supposed Precambrian metamorphic zones and their apparent offsetting are claimed to indicate some 25 km of thrust displacement (Lambert *et al.* 1979). However, these authors gave no detailed, definitive analysis of the effects of the slide on metamorphic assemblages which provides the basis for their conclusions, nor did they analyse the associated structures.

The relative and absolute age of movement on the slide is a subject of debate. Relative to local sequences of deformation it is held to be pre-F_2 or syn-F_2 at Kinloch Hourn (Tanner 1970), syn-F_3 in relation to the Morar/Glenfinnan deformation sequence (Powell 1974), or possibly coeval with the formation of the primary mylonites of the Moine Thrust and thus Caledonian in age (Powell 1974; van Breemen *et al.* 1974; Brewer *et al.* 1979; Mendum 1979). Such suggestions, however, have depended largely on long-range correlations of the deformation sequences between local areas of complex histories lacking readily identifiable time markers.

Recognition of the Sgurr Beag Slide

Major practical problems in recognizing the presence of the Sgurr Beag Slide are: the lack of well defined stratigraphical markers in most of the Moine which might permit correlations across the slide; the general lack of discontinuities along its known course (Tanner *et al.* 1970); and the absence of easily identifiable characteristic fabrics (a consequence of its syn-

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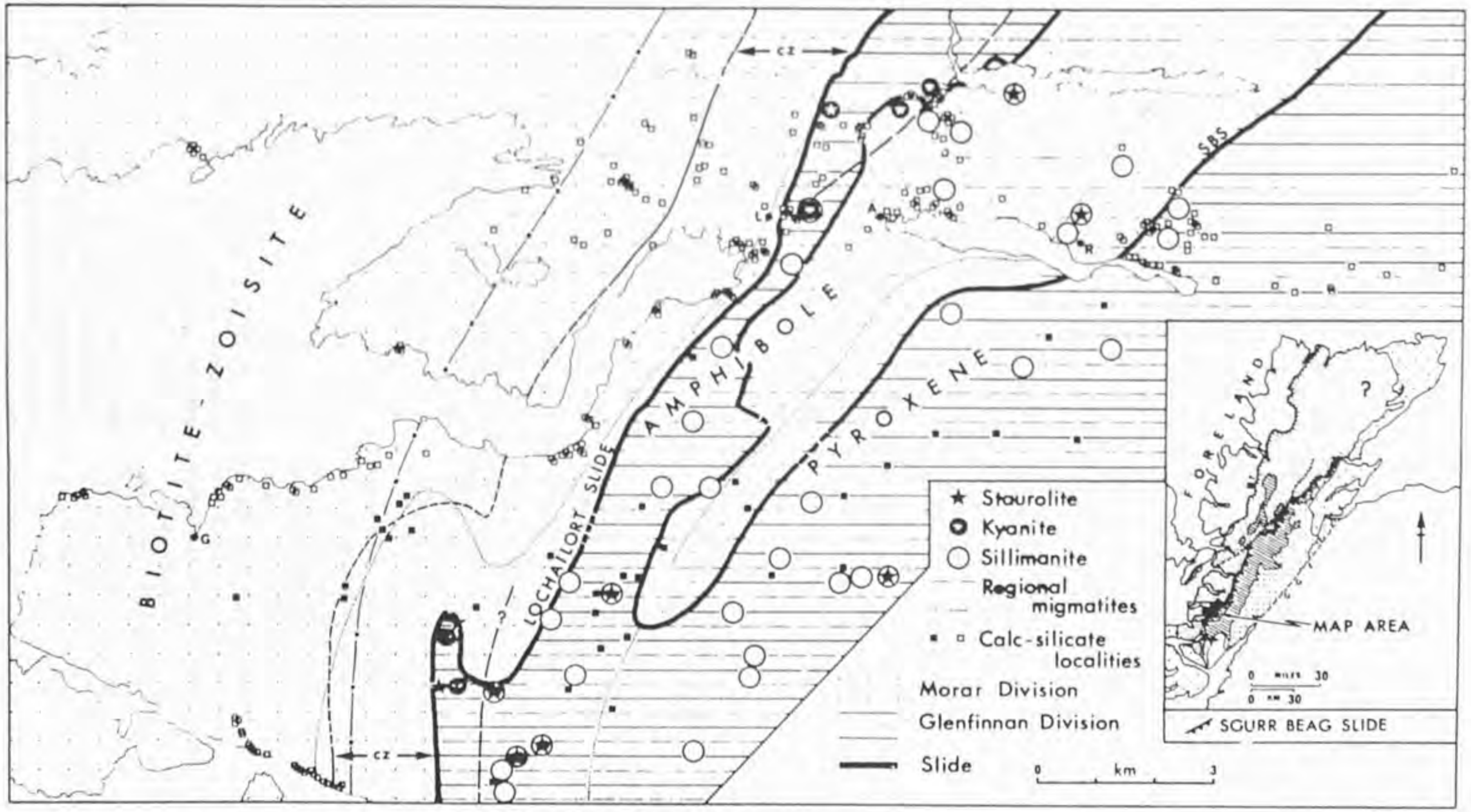

FIG. 1. Location and metamorphic features of the Morar/Glenfinnan area. Open squares: calc-silicate localities (Bedford College collection); solid squares: calc-silicate localities from Howkins (1961) and Clark (1961). Dotted lines: limits of calc-silicate mineral assemblage fields modified from Kennedy (1949). Dot/dash line: western limit of amphibolite facies assemblages in microdiorites (Smith 1980). Dashed line: western limit of silimanite in pelitic rock cz = clinozoisite zone (see text). A = Arieniskill, G = Glenuig, L = Lochailort, R = Ranochan.

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metamorphic and ductile nature). Furthermore, there are difficulties involved in establishing metamorphic grade within and adjacent to the slide in rocks which, because of the chemical nature of their pelitic members, rarely develop alumino-silicate minerals (Winchester 1974).

Slivers of Lewisian basement provide convincing evidence for the presence of the slide at Kinloch Hourn and elsewhere. Where such Lewisian rocks are absent, extrapolation of the slide has relied heavily upon identification of Moine rocks as belonging to either the Morar or Glenfinnan divisions, on the assumption that it should everywhere separate these units (Fig. 1, inset; Johnstone 1975) and on the interpretation of 'platy zones' as slide zone indices (Rathbone & Harris 1979; Mendum 1979).

As a direct consequence of these problems, the southerly extension of the Sgurr Beag Slide has not previously been established firmly, though various courses have been advocated (Tanner *et al.* 1970; Powell 1974; Rathbone & Harris 1979). The new observations presented here provide evidence for the course of the slide and for metamorphic grade contrasts across it. Interpretation of the nature of these changes suggest syn-metamorphic shear displacements of an earlier metamorphic complex across outcrops of the slide repeated by major folds.

Metamorphic considerations

In the Morar/Glenfinnan area there is an increase in metamorphic grade from W–E over a horizontal distance of *c.* 18 km (Kennedy 1949; Powell 1974; Charnley 1976). This is indicated by: the incoming of staurolite, kyanite and sillimanite (mostly fibrolite) in pelitic rocks (Fig. 1); the incoming of migmatitic rocks (Fig. 1); changes in the mineralogy of calc-silicate rocks (Kennedy 1949; Charnley 1976); and prograde metamorphic changes in the mineralogy of intrusive microdiorites (Smith 1979).

Calc-silicate rocks usually occur in Moine rocks as thin, conformable ribs or lenses and represent either original sediments or early concretions. The use of the mineralogical changes in these calc-silicate lenses as indicators of metamorphic grade has been attempted by several authors (Kennedy 1949; Soper & Brown 1971; Winchester 1972, 1974; Charnley 1976; Tanner 1976; Tanner & Miller 1980). This work demonstrates that the incoming and outgoing progressively of biotite, amphibole and pyroxene is related not only to increasing metamorphic grade but also to whole-rock chemistry (using as a simple index the ratio CaO/Al_2O_3). An investigation of the metamorphic, mineralogical and chemical changes in calc-silicate rocks from the Morar/Glenfinnan area is in progress. It is not the purpose of this paper to present our results in detail but the early deductions it allows are given in outline because they have considerable bearing on the course, nature and relative age of the Sgurr Beag Slide.

A total of 210 calc-silicate samples have been examined and of these 107 have been analysed for their major element composition. The anorthite content of the plagioclase in 163 of the samples has been determined optically.

The following main mineral assemblages are recognized:

1a. Biotite + zoisite ± clinozoisite ± calcite + garnet + plagioclase + quartz.
b. Biotite + clinozoisite ± calcite + garnet + plagioclase + quartz.
2a. Amphibole ± biotite ± zoisite + garnet + plagioclase + quartz.
b. Amphibole ± biotite + clinozoisite + garnet + plagioclase + quartz.
3a. Amphibole ± pyroxene ± zoisite + garnet + plagioclase + quartz.
b. Amphibole ± pyroxene + clinozoisite + garnet + plagioclase + quartz.
4a. Biotite + epidote ± calcite ± garnet + plagioclase + quartz.
b. Amphibole + epidote ± calcite ± garnet plagioclase + quartz.

Chlorite in some cases replaces amphibole, and/or biotite and/or garnet.

Assemblages 1a–2a–3a are, on textural evidence and their spatial distribution, prograde, but also dependent on the whole rock CaO/Al_2O_3 ratio, a relationship which supports the conclusions of earlier workers. In assemblages 1b, 2b and 3b, clinozoisite occurs in place of zoisite. Assemblages 4a and 4b form a distinct sub-group which coexists with assemblages 1, 2 and 3 but appears not to have responded so readily to changes in metamorphic conditions. It is not therefore considered further. Those calc-silicate rocks (not containing epidote) analysed have CaO/Al_2O_3 ratios ranging from 0.3 to 1.003.

Calc-silicate rocks with CaO/Al_2O_3 ratios of 0.353 and above contain amphibole and/or pyroxene in environments where pelitic rocks contain staurolite, kyanite and sillimanite (Fig. 1). Where alumino-silicates are absent on a regional scale, biotite plus zoisite characterize calc-silicate rocks with CaO/Al_2O_3 ratios of less than 1.003. Thus within the constraints of these ratios, biotite plus zoisite assemblages indicate garnet grade conditions, whereas amphibole and/or pyroxene reflect kyanite grade and above.

In samples from the lower grade, western half of the area, sub- to euhedral generally acicular zoisite and clinozoisite are characteristic and clinozoisite does not occur without zoisite. The zoisite has predominantly anomalous grey or grey-brown birefringence colours, while those of clinozoisite are grey-brown–deep blue–bright blue. To the E of the western edge of the clinozoisite zone (Figs 1 & 2) clinozoisite becomes

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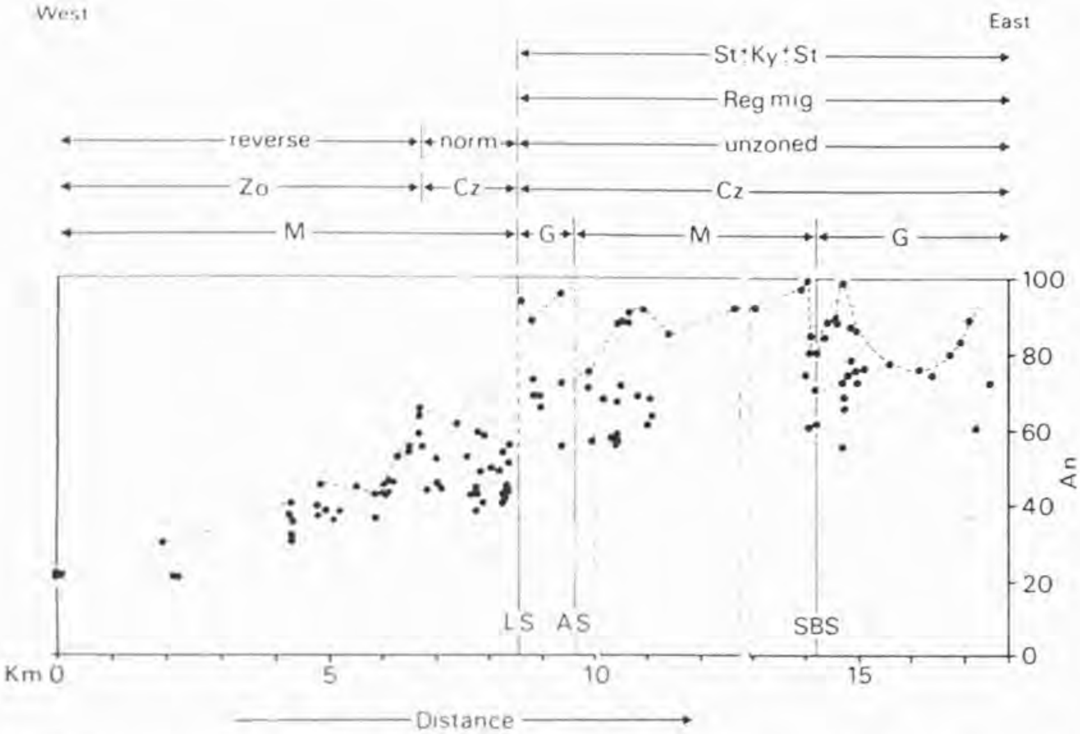

FIG. 2. Composite profile of the variation in An content of plagioclase in calc-silicate rocks and other metamorphic features, across the area. St, Ky, Si = occurrence of staurolite, kyanite and sillimanite in pelitic rocks. Reg mig = extent of regional migmatites. Unzoned; norm; reverse = predominant types of zoning in plagioclase in calc-silicate rocks. cz = clinozoisite predominant; zo = zoisite predominant in calc-silicate rocks. G = rocks of Glenfinnan division; M = rocks of Morar division. SBS = Sgurr Beag Slide; AS = Arieniskill Slide; LS = Lochailort Slide.

predominant with zoisite occurring in only a few samples. Here the interference colours of clinozoisite are exclusively grey–bright blue–yellow, the crystals are irregular in shape and, on textural evidence, may result from breakdown of plagioclase and/or zoisite.

Plagioclase feldspar is a dominant mineral phase in the calc-silicates examined, usually constituting *c.* 30% of the rock. Optical estimates of the anorthite content of plagioclase, using the Michel-Levy method, indicate a variation in different rock samples ranging from An_{20} to An_{97}. Microprobe analyses have been made on 12 of the samples in order to check the plagioclase compositions obtained by optical means. These indicate that the Michel-Levy results give a minimum estimate of An content tending to be only slightly lower (within An_6) than chemical analyses. The general patterns of plagioclase behaviour outlined later are not affected by these differences.

The variation in plagioclase composition across the area is not random; there is clearly a general, and in places gradual, increase in An content of plagioclase from W to E, i.e. with the overall increase in metamorphic grade (Fig. 2). Furthermore, Fig. 3 provides strong evidence that An content is independent of the whole-rock CaO/Al_2O_3 ratio; samples with similar CaO/Al_2O_2 ratios show An contents as low as 20 and as high as 94 O.C.; highest values of An content occur in calc-silicates with CaO/Al_2O_3 ratios of 0.35–0.76, which were collected from areas where adjacent or nearby pelitic rocks contain staurolite and/or kyanite, and/or fibrolite; lowest values for An occur where CaO/Al_2O_3 ratios range from 0.39 to 0.83 and come from western localities of low grade pelitic rocks.

The conclusion that plagioclase composition is essentially independent of CaO/Al_2O_3 ratio is further borne out on the local scale. To the NW of Ranochan (Fig. 1), 9 calc-silicate samples collected within an area of approximately 1 km^2 show a range in the An content of plagioclase between samples of 68–100 and a range in CaO/Al_2O_3 ratios of 0.40–0.72; An values above 80 occur in rocks with CaO/Al_2O_3 ratios between 0.40 and 0.60 while, for An values below 80, the same ratios range from 0.51 to 0.72.

The same conclusion can be drawn from the results of a parallel study in the Knoydart area *c.* 12 km to the N of the Morar/Glenfinnan area (P. Jordan). Here

Appendix 5.

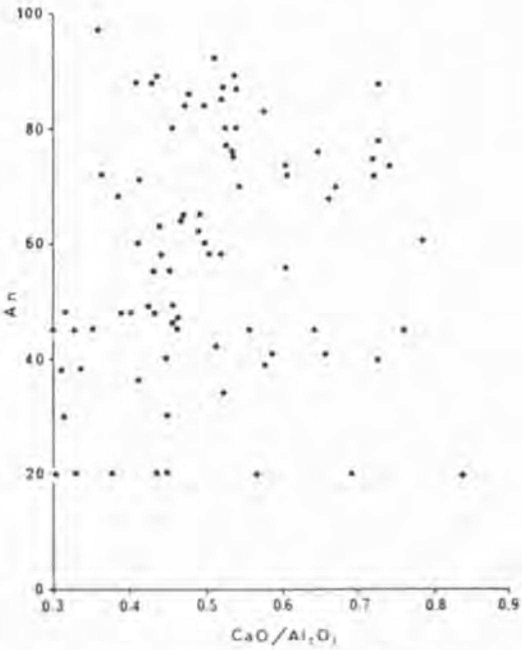

FIG. 3. Plot of anorthite content of plagioclase in calc-silicate rocks against whole-rock CaO/Al_2O_3 ratio (wt%).

64 microprobe analyses of plagioclase compositions in 11 calc-silicate rocks collected from a 5 km traverse show a progressive upgrade increase in the An content from 11 to 89 and a variation in CaO/Al_2O_3 ratios of 0.27–0.58. An values of 70–90 occur in rocks with CaO/Al_2O_3 between ratios 0.31 and 0.50, and An values of 10–30 where this ratio is 0.27–0.53. Within these groupings there is likewise no direct relationship.

The results of our analyses of calc-silicate rocks indicate no direct or simple correlation between An content and the amounts of other major oxides present, with the notable exception of Na_2O. There are clear indications that increasing An content coincides with decreasing Na_2O in the whole-rock, but both parameters are a function of metamorphic grade (Fig. 4); Na_2O is evidently lost from the whole-rock (see also Tanner & Miller 1980). Despite the lack of precision inherent in the determination of An content of plagioclase by the Michel-Levy method (see, however, Glazner 1980) it appears from Figs 1 & 2 and the above discussion that the anorthite content of plagioclase in calc-silicate rocks is monitoring changes in metamorphic grade both on the regional and more local scale. The local progressive changes in An content with distance (Fig. 2), particularly where sampling is intense, suggest that plagioclase composition (i.e. *maximum* An content) provides a sensitive index of metamorphic grade. In this respect it is noteworthy that while the range of An content exhibited at given distances in Fig. 2 may be due to measuring errors and/or some, as yet unidentified, chemical control, the changes in maximum An content with distance are followed, in general terms sympathetically, by changes in the minimum recorded An content; both values show a general W to E increase.

The variation in the maximum An content of plagioclase in calc-silicates and their mineralogy across the area (Figs 1 & 2), indicate a progressive increase in grade from W to E which is clearly interrupted in two places (near Lochailort and Arieniskill) where the An content of plagioclase suddenly changes. Around Lochailort a jump from An_{56} to An_{94} takes place across the boundary between rocks which are referred

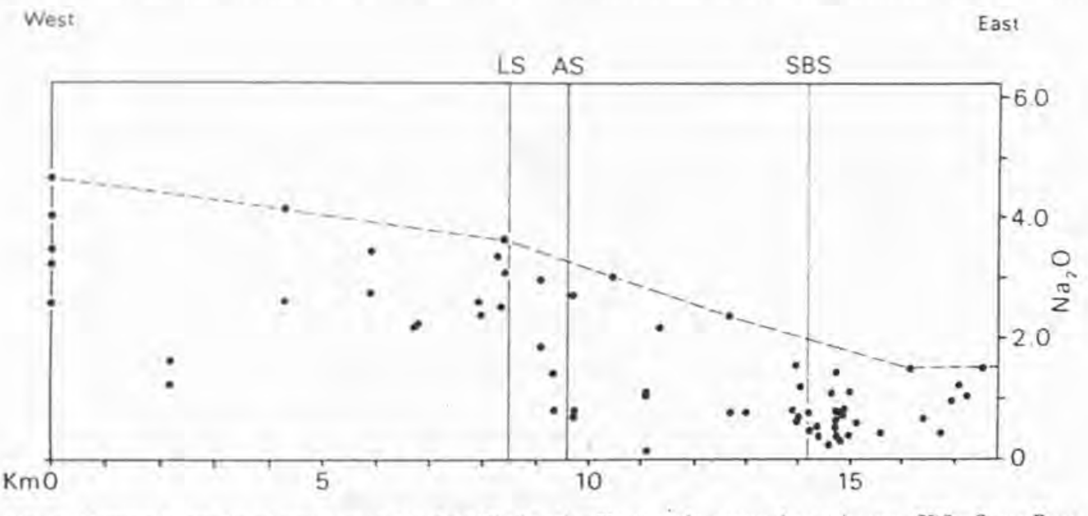

FIG. 4. Composite profile of variation in Na_2O (wt%) in calc-silicate rocks across the study area. SBS = Sgurr Beag Slide; AS = Arieniskill Slide; LS = Lochailort Slide.

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to the Glenfinnan division (the Lochailort Pelite of Powell 1964) and those considered as belonging to the Morar division (the Ardnish Psammite of Powell 1964). The jump here also coincides with the incoming eastwards of migmatitic rocks, and more approximately, with the incoming of amphibole as a common constituent of calc-silicate rocks, the incoming of alumino-silicate minerals in some pelitic rocks, and a change in the predominant nature of zoning in plagioclase in calc-silicate rocks (Figs 1 & 2).

It might be argued that these jumps in composition reflect the gap in An values for plagioclase between 55 and 70 recorded in calc-silicate rocks by Kennedy (1949) and Tanner (1976). However, some 25% of our optically determined results lie within this range of composition, as do 18% of the microprobe analyses, suggesting that no such gap exists in rocks from the area. The intensity of sampling would appear to be more representative and spatially more intensive than that of the earlier work. In view of all these changes we conclude that a rapid increase in metamorphic grade takes place across the psammite/pelite junction.

Immediately to the W of this junction a 2 km wide zone, characterized by a progressively *westward* increase in An content and the occurrence of anhedral clinozoisite (blue-yellow) rather than zoisite, interrupts the gradual eastward rise in grade in prograde assemblages lying further to the W (Fig. 2). In pelitic rocks along the junction and within this 'clinozoisite zone' (Fig. 1), garnet, biotite and muscovite are, however, stable.

Further to the E, near Arieniskill (Figs 1 & 2), a similar, though less marked, change in An content and thus metamorphic grade, is centred on the junction between predominantly psammitic and pelitic formations (the Arieniskill Psammitic and Lochailort Pelitic groups of Powell 1964). Likewise, near Ranochan, a change in An content coincides with a psammite-pelite junction (Figs 1 & 2).

Each of these three junctions have been, or are, from stratigraphical and/or structural considerations, thought to mark, or to be near, tectonic slide zones (Tanner *et al.* 1970; Powell 1974; and Baird, pers. comm.). Following, in part, and modifying the interpretation of the regional structure given by Powell (1974), and on the evidence given in this earlier work, together with our conclusion regarding the metamorphic pattern, it is here suggested that each of the junctions traces the southerly course of the Sgurr Beag Slide repeated across the area by major folds (Figs 1 & 5).

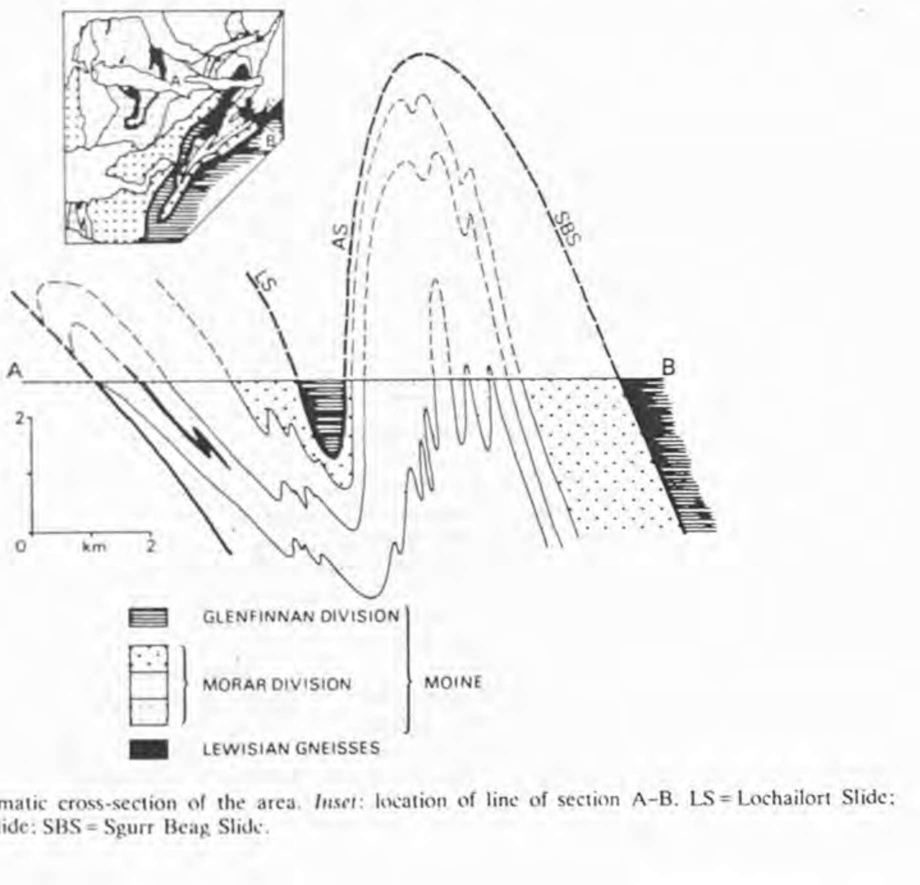

FIG. 5. Semi-schematic cross-section of the area. *Inset*: location of line of section A–B. LS = Lochailort Slide; AS = Arieniskill Slide; SBS = Sgurr Beag Slide.

Appendix 5.

Metamorphic history of area in relation to formation of Sgurr Beag Slide

In common with relationships described from elsewhere (Tanner 1970; Rathbone & Harris 1979), no cataclastic or classical mylonitic features are found associated with the Sgurr Beag Slide in the Morar/Glenfinnan area. However, metamorphic changes associated with the slide are: a drop in the An content of plagioclase; the incoming of clinozoisite in place of zoisite in many calc-silicate rocks; the juxtaposition of migmatitic against non-migmatitic rocks; and the incoming of staurolite, kyanite and sillimanite. All of these changes do not, however, occur at each of the three levels of exposure of the slide. In particular, a large jump in An content only occurs at the Lochailort and Arieniskill slides; clinozoisite (blue-yellow) occurs in many, but not all, calc-silicate rocks to the E of the Lochailort Slide; alumino-silicates and staurolite occur on both sides of the Arieniskill Slide; and migmatitic features characterize all rocks to the E of the Lochailort Slide. Characteristic structural features of the slide are rapid increases in strain at Lochailort (Rathbone & Harris 1979), extreme attenuation of stratigraphical formations to the N of Arieniskill (Powell 1974), and changes in the profile geometry of folds, together with truncation of lithological units at Ranochan (Baird, pers. comm.).

From the evidence presented it is clear that the Sgurr Beag Slide is *not* a post-metamorphic feature. Because at Lochailort it emplaces high grade migmatitic rocks against non-migmatitic rocks of lower grade, without severe retrogression of pelitic and calc-silicate rocks, it either translated already metamorphosed rocks during a subsequent syn-metamorphic 'sliding' event, or it translated rocks at high temperatures, bringing them against cooler rocks lying to the W. In view of the progressive drop in An content, and thus metamorphic grade, through the 'clinozoisite zone' immediately W of Lochailort, together with the incoming of clinozoisite and loss of zoisite, which are both interpreted as retrograde features, and the sudden 'jump' in metamorphic grade at the psammite/pelite junction which is witnessed by not only An content of plagioclase, but also pelitic mineralogy, it appears that movement on the slide displaced already metamorphosed rocks. The abruptness of the metamorphic changes across the slide seemingly precludes the thermal overprinting to be expected if hot rocks were emplaced against cooler, during a single prograde metamorphic event.

A model for the metamorphic environment of the Sgurr Beag Slide

The patterns of metamorphic changes across slide zones of the Morar/Glenfinnan area are similar to those reported for a ductile shear zone in the Precambrian rocks of Greenland (Grocott 1979). An important difference, however, is the asymmetric nature of the Morar/Glenfinnan examples. Strain studies of rocks adjacent to and within the Lochailort Slide zone (here the 'clinozoisite zone'), using changes of grain size, rotation of quartz veins, and changes in the geometrical relationships of cross-bedding, led Rathbone & Harris (1979) to suggest an asymmetric distribution of shear strain across the zone. Shear strains, assumed to have developed during 'sliding', progressively increase from W to E within a zone *c.* 400–500 m wide. If these observations are correct it would appear that the Lochailort Slide zone (and by correlation the Arieniskill and Sgurr Beag slide zones) are asymmetric ductile shear zones.

A model of the possible patterns of metamorphic grade developed during, and directly following, syn-metamorphic shearing within an asymmetric shear zone is given in Fig. 6. The model assumes a strain rate sufficient to allow metamorphic reactions to proceed; adjustment of the thermal instabilities resulting from displacement of isotherms (thermal relaxation of Oxburgh & Turcotte 1974 and Grocott 1979); and essentially parallel isotherms both before displacement and after 'thermal relaxation'. Such a model predicts fields of possible retrograde and prograde metamorphism, the boundaries of which are asymmetric to the shear zone boundaries (cf. Grocott 1979) as a consequence of the parallel nature of the relaxed isotherms and the similar geometry of the displaced isotherms (Fig. 6*b*). The patterns of relative metamorphic grade at different levels of erosion through such a model are shown in Figs 6 & 7. In ductile shear zones it can be argued that metamorphic reactions will not proceed in the absence of deformation (Beach 1976; Grocott 1979); thus retrogressive or progressive reactions may be limited to the shear zone alone (Fig. 8*a*). In the case of asymmetric shear zones, however, it is unlikely that such reactions will take place below a critical value of shear strain and thus it follows that the final metamorphic pattern would be similar to those shown in Fig. 8*b*.

The analysis of changes in thermal conditions resulting from overthrust faulting given by Oxburgh & Turcotte (1974) is compatible with the modelling discussed here in as much as they predict substantial, comparatively rapid readjustment (thermal relaxation) of geotherms displaced at thrust planes. Relaxation following rapid displacement gives temperatures at the thrust plane of 0.75–0.5 of the initial temperature of the base of the thrust slab after 4 Ma depending upon the depth of the thrust plane (Oxburgh & Turcotte, fig. 3). Further, they predicted a return to pre-thrusting temperatures in the rocks overlying the thrust plane following an initial drop in temperature, but this recovery takes over 60 Ma. In rocks lying below the thrust a continual increase in temperature

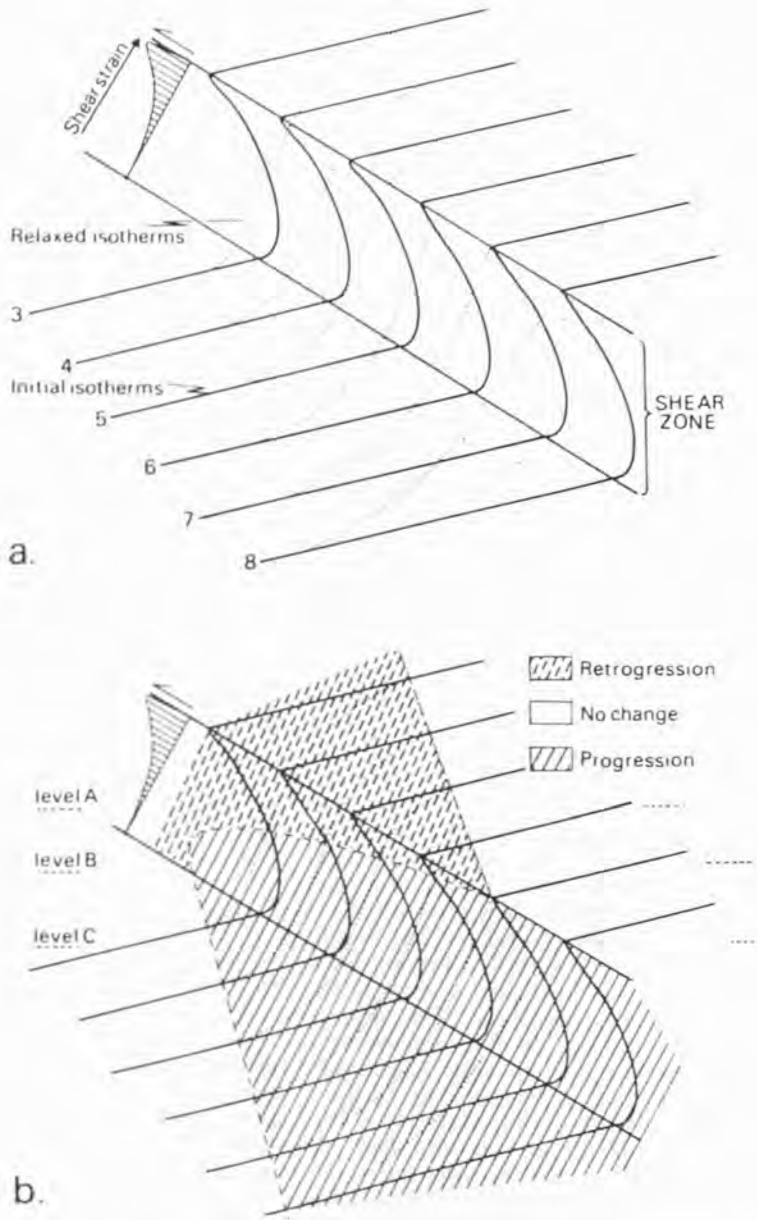

FIG. 6. Pattern of isotherms after displacement and thermal relaxation across a model asymmetric shear zone. *a*, Positions of displaced and 'relaxed' isotherms. Numerical values for isotherms indicate relative temperature values. *b*, Fields of potential prograde and retrograde metamorphism. A, B, C = levels of erosion referred to in Fig. 7.

follows immediately upon displacement. Oxburgh & Turcotte's analysis, however, assumes pre-thrusting temperatures of 0°C at the top of the rock pile underlying the thrust; displacement across a single thrust plane; and does not consider, in detail, the possible effects of erosion and strain.

The geological constraints in the case of the Sgurr Beag Slide at Lochailort (Fig. 1) would appear to require displacement across a *c.* 2 km wide ductile shear zone and pre-thrusting temperatures immediately above the shear zone at which kyanite and staurolite are stable, i.e. *c.* 600°C. Below the shear

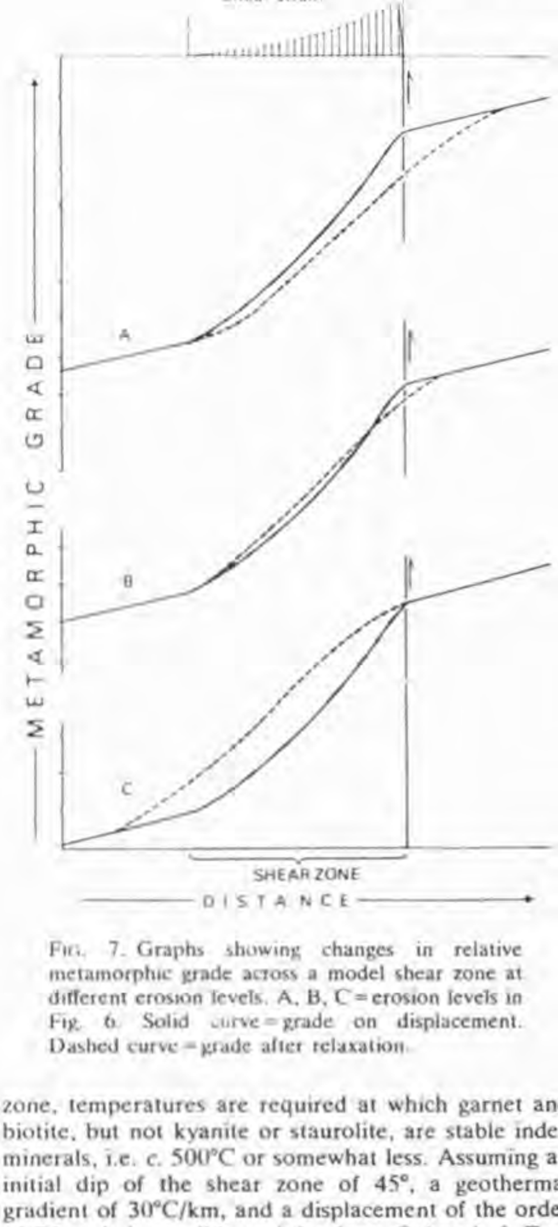

FIG. 7. Graphs showing changes in relative metamorphic grade across a model shear zone at different erosion levels. A, B, C = erosion levels in Fig. 6. Solid curve = grade on displacement. Dashed curve = grade after relaxation.

zone, temperatures are required at which garnet and biotite, but not kyanite or staurolite, are stable index minerals, i.e. *c.* 500°C or somewhat less. Assuming an initial dip of the shear zone of 45°, a geothermal gradient of 30°C/km, and a displacement of the order of 7 km, it is possible to infer from Oxburgh & Turcotte's analysis a rapid drop in temperature to *c.* 510°C at the top of the shear zone following displacement and then a slow rise in temperature returning to 600°C. At the bottom of the shear zone temperatures would rapidly rise to *c.* 590°C and slowly continue to rise thereafter to an equilibrium geothermal temperature of *c.* 690°C. While these actual temperatures may not be accurate, given the problems involved in establishing temperatures from metamorphic mineral assemblages, it is important to note that for thrusts developed at depths greater than 10 km the values of the temperature *differences* upon relaxation are not greatly affected by depth, geothermal gradient and thus starting temperatures. However, a change in the amount of displacement proportionally increases or decreases these temperature differences as doeş, for a given displacement, a change in dip of the shear zone.

It thus appears possible, within a short period of time following displacement (about 4 Ma), to generate temperature fluctuations which might drive both pro- and retrograde reactions in different parts of a ductile shear zone. Following Oxburgh & Turcotte we note that the value of the temperature differences upon relaxation would be increased if the erosion rate consequent upon displacement and therefore uplift were sufficiently high to affect thermal relaxation. In addition, the increase in temperature at the top of the shear zone following the initial rapid cooling might not occur under such conditions and, given rapid erosion, it follows that any phase changes resulting from the temperature drop might be preserved.

Characteristic of the metamorphic profiles derived from the model (Fig. 6) is a jump in 'grade' at either the shear zone boundaries or one boundary and the plane of critical shear strain (Fig. 8). Such rapid changes in 'grade' are clearly observed across the slide zones in the Morar/Glenfinnan area (Fig. 2), but the geometry of these profiles is not, in more detail, similar to those of the model. Moreover, changes in the model with regard to the relative attitudes of the isotherms and the shear zone boundaries, the geometry of the isotherms (i.e. similar rather than parallel), and the value chosen for the critical shear strain, do not produce major changes in the geometry of the 'grade' profiles and thus do not provide more precise analogues for the Morar/Glenfinnan examples.

In the Morar/Glenfinnan area there is geological and isotopic evidence for two major phases of regional metamorphism (Powell 1974; van Breemen *et al.* 1974; Brewer *et al.* 1979). The earliest of these (here denoted M1), on isotopic evidence, is likely to have been Precambrian, whilst the younger (here M2), was Caledonian. Textural and mineralogical evidence related to structural observations (Powell & MacQueen 1976; MacQueen & Powell 1977; Anderson & Olimpio 1977; Brook *et al.* 1977) indicates garnet grade M1 overprinted by chlorite-garnet grade M2 in the area to the W of Lochailort, whereas to the E, M1 and M2 both appear to have been high grade.

In order to accommodate the effects of a pre-shearing metamorphism, the model outlined above may be modified (Fig. 9). The relative attitudes of the M1 and M2 isotherms chosen are those which most closely fit the observed M1–M2 grade relationships in the area. Combining the effects of an earlier metamorphism with the factors outlined in the previous model, the resultant 'grade' profiles at different erosion surfaces are those given in Fig. 10. These differ significantly

Appendix 5.

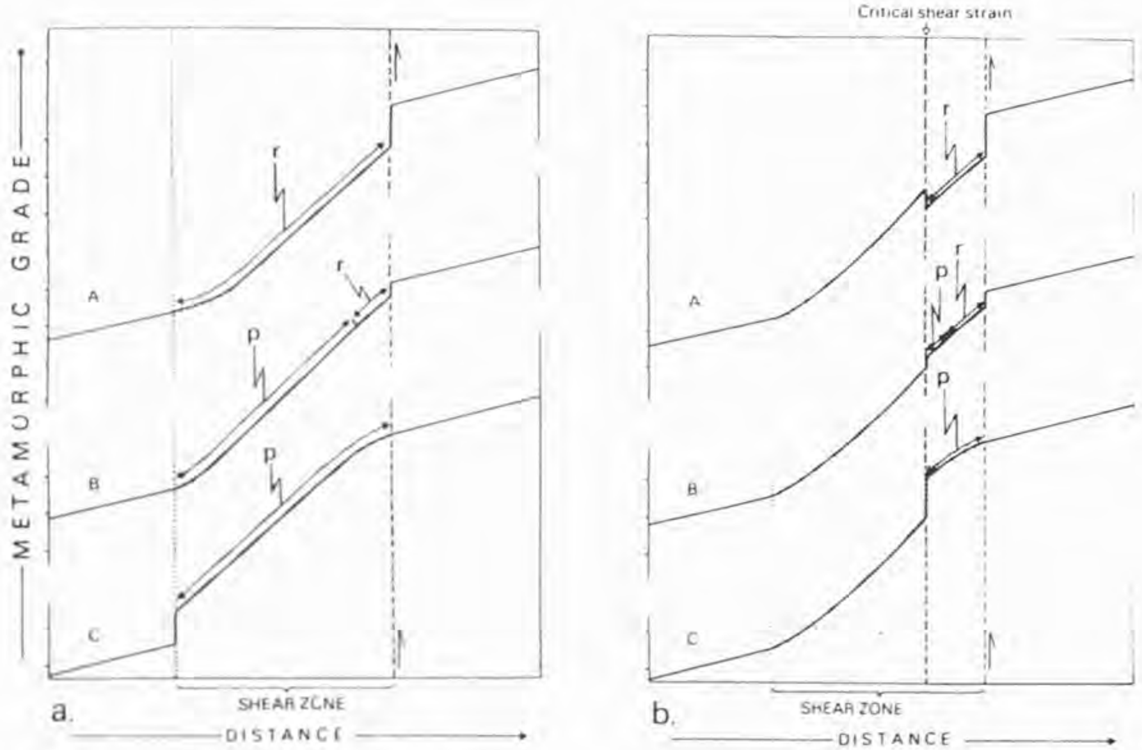

FIG. 8. Graphs showing possible resultant metamorphic grade profiles across a model shear zone. *a.* Changes in grade at different erosion levels assuming metamorphic reactions occur within entire shear zone. *b.* Changes in grade assuming metamorphic reactions proceed only above a critical shear strain. *r* = retrograde metamorphism; *p* = prograde metamorphism. A, B, C = erosion levels of Fig. 6.

from those of the 'single metamorphism' model because, outside the zone where the shear strain is equal to, or greater than critical, the resultant 'grade' would be that attained during M1, not M2.

The metamorphic profiles at erosion surfaces A and E (Fig. 10) are remarkably similar to those observed at the Lochailort, Arieniskill, and Sgurr Beag slides (Fig. 2), especially if the compositional change in plagioclase in calc-silicate rocks is proportional to shear strain as well as temperature.

Regional implications

A fundamental problem in understanding the orogenic history of the Moine rocks of the Northern Highlands of Scotland has been the lack of substantive geological evidence for major crustal reworking of an older metamorphic complex by Caledonian events. Such evidence might provide a geological framework for the groupings of isotopic dates, at *c.* 1004, 780–730, and 470–400 Ma. Polyphasal deformation sequences are recorded for many areas, but these vary from 4 phases around Carn Chuinneag (Wilson & Shepherd 1979) and Morar/Moidart (Brown *et al.* 1970; Powell 1974), to 6 phases in Glenelg (Barber & May 1976) and Strathconnon/Glen Affric (Tobisch *et al.* 1970). Correlation of these sequences has proved difficult because of the problems involved in recognizing common, regionally developed, distinctive geological events which might provide time markers. Polyphase metamorphism has also been recognized (Tobisch *et al.* 1970; Powell 1974; Winchester 1974; Fettes 1979), and while an earlier, Precambrian, metamorphism followed by Caledonian re-heating are supported by isotopic evidence, a detailed understanding of their interplay and relationships to deformation sequences has not emerged.

Taken with earlier work, the present study provides evidence for intensive crustal reworking associated with the development of the Sgurr Beag Slide. It strongly suggests that, on a regional scale, an early metamorphic complex which had suffered at least two phases of pervasive deformation, was subsequently reworked by 3 phases of ductile deformation, the earliest of which was syn-metamorphic and involved

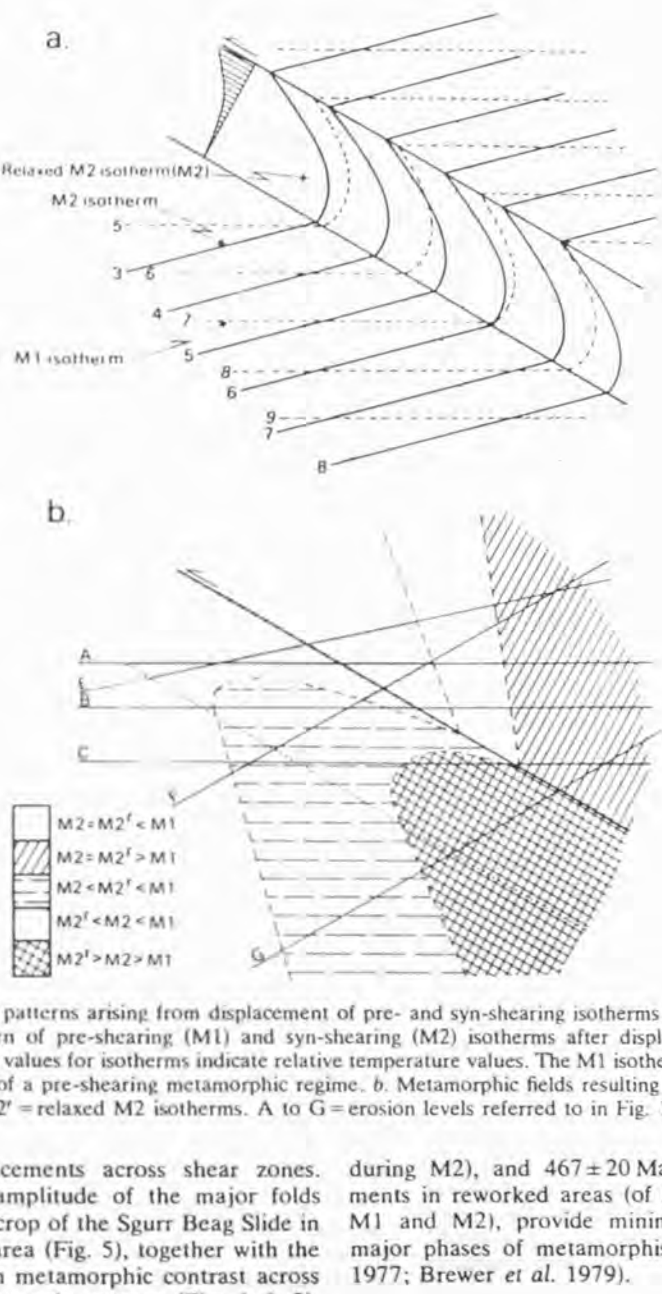

FIG. 9. Metamorphic patterns arising from displacement of pre- and syn-shearing isotherms across an asymmetric shear zone. *a*. Pattern of pre-shearing (M1) and syn-shearing (M2) isotherms after displacement and thermal relaxation. Numerical values for isotherms indicate relative temperature values. The M1 isotherms are those relating to the establishment of a pre-shearing metamorphic regime. *b*. Metamorphic fields resulting from model in *a*. M1 and M2 as above, M2' = relaxed M2 isotherms. A to G = erosion levels referred to in Fig. 10.

regional crustal displacements across shear zones. Consideration of the amplitude of the major folds which duplicate the outcrop of the Sgurr Beag Slide in the Morar/Glenfinnan area (Fig. 5), together with the progressive reduction in metamorphic contrast across the slide in the more easterly outcrop (Figs 1 & 2), suggests that the slide was generated with a moderate to shallow, easterly dip.

The grouping of the orogenic events outlined above clearly provides a geological framework for the isotopic evidence if it is accepted that Rb-Sr whole-rock isochron ages of 1004±28 Ma for pelitic metasediments from the W of the study area (of medium metamorphic grade during M1, and low grade during M2), and 467±20 Ma for similar metasediments in reworked areas (of high grade during both M1 and M2), provide minimum ages for the two major phases of metamorphism (Brook *et al.* 1976, 1977; Brewer *et al.* 1979).

Conclusions

The pattern of changes in the anorthite content of plagioclase feldspar in calc-silicate rocks across an 18 km section of the Northern Highlands, together with parallel changes in the mineralogy and texture of pelitic rocks, strongly suggest that the anorthite con-

Appendix 5.

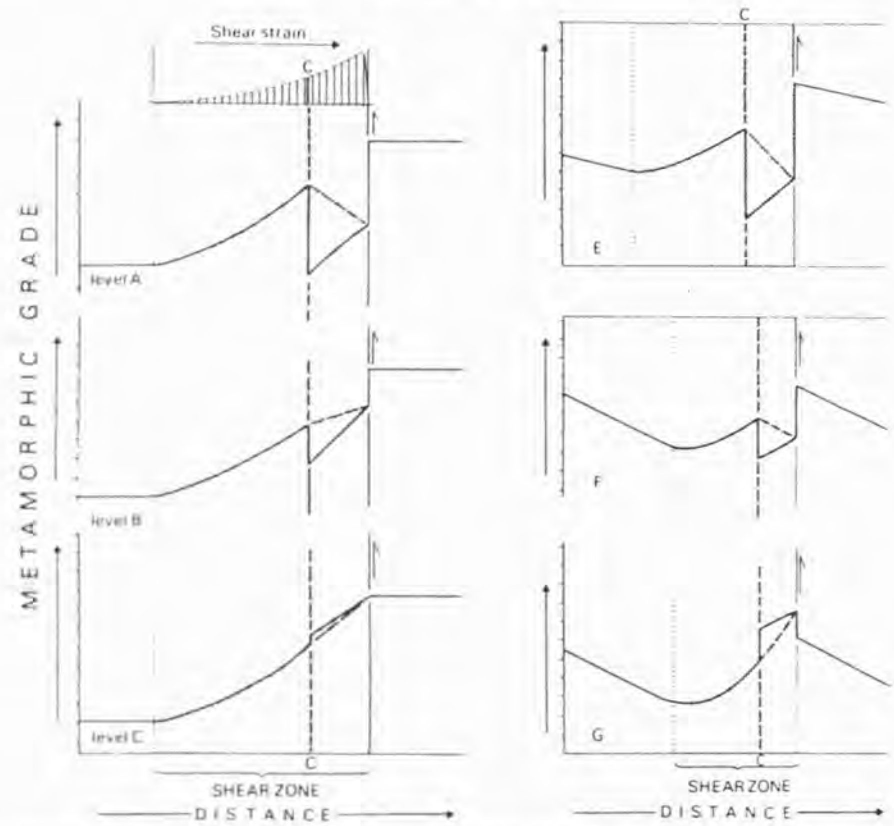

FIG. 10. Graphs showing possible resultant metamorphic grade profiles across model shear zone. A to G = erosion levels of Fig. 9. c = critical shear strain. Solid curves = metamorphic grade assuming complete adjustment of grade in zone where shear strains exceed critical value. Dashed curves = metamorphic grade assuming adjustment is proportional to level of shear strain.

tent of plagioclase in such rocks provides a sensitive index of relative metamorphic grade.

Abrupt changes in plagioclase compositions, and thus metamorphic grade, coincide with the presence of slide zones. In the light of structural data and the interpreted nature of the changes in metamorphic grade, such slide zones are most readily explained as asymmetric syn-metamorphic shear zones.

A model of the metamorphic patterns resulting from temperature fluctuations produced as a result of crustal displacement along these syn-metamorphic shear zones is compatible with the analysis given by Oxburgh & Turcotte (1974). However, the model also suggests that the level of shear strains may influence metamorphic reactions.

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References

ANDERSON, D. E. & OLYMPIO, J. C. 1977. Progressive homogenisation of metamorphic garnets, S. Morar, Scotland. *Can. Mineral.* **15**, 205–16.

BARBER, A. J. & MAY, F. 1976. The history of the Western Lewisian in the Glenelg Inlier, Lochalsh, Northern Highlands. *Scott. J. Geol.* **12**, 35–50.

BEACH, A. 1976. The interrelations of fluid transport, deformation, geochemistry and heat flow in early Proterozoic shear zones in the Lewisian complex. *Philos. Trans. R. Soc. London*, **A280**, 569–604.

BREWER, M. S., BROOK, M. & POWELL, D. 1979. Dating of the tectono-metamorphic history of the southwestern Moine, Scotland. *In*: HARRIS, A. L. *et al.* (eds.). *The Caledonides of the British Isles—Reviewed.* Spec. Publ. geol. Soc. London, **8**, 129–37.

BROOK, M., POWELL, D. & BREWER, M. S. 1976. Grenville age for rocks in the Moine of north-western Scotland. *Nature, London,* **260,** 515–17.

——, —— & —— 1977. Grenville events in Moine rocks of the Northern Highlands, Scotland, *J. geol. Soc. London,* **133,** 489–96.

Brown, R. L., Dalziel, I. W. D. & Johnson, M. R. W. 1970. A review of the structure and stratigraphy of the Moinian of Ardgour, Moidart and Sunart—Argyll and Inverness-shire. *Scott. J. Geol.* **6,** 309–35.

Charnley, N. R. 1976. *Geochemical studies of the Moine rocks in western Inverness-shire.* Thesis, PhD, Univ. London (unpubl.).

Clark, G. C. 1961. *Structural and metamorphic history of the Lochailort (Moidart) area.* Thesis, PhD, Univ. Edinburgh (unpubl.).

Fettes, D. J. 1979. A metamorphic map of the British and Irish Caledonides. *In*: Harris, A. L., Holland, C. H. & Leake, B. E. (eds.) *The Caledonides of the British Isles—Reviewed.* Spec. Publ. geol. Soc. London, **8,** 307–21.

Glazner, A. F. 1980. Frequency distribution of plagioclase extinction angles: precision of the Michel-Levy technique. *Am. Mineral.* **65,** 1050–52.

Grocott, J. 1979. Controls of metamorphic grade in shear belts. *Rapp. Grønlands geol. Unders.* **89,** 47–61.

Howkins, J. B. 1961. *The structural and metamorphic history of Moidart, southwest Inverness-shire.* Thesis, PhD, Univ. Edinburgh (unpubl.).

Johnstone, G. S. 1975. The Moine Succession. *In*: Harris, A. L. *et al.* (eds.). A correlation of the Pre-Cambrian rocks in the British Isles. *Spec. Rep. geol. Soc. London,* **6,** 30–42.

Kennedy, W. Q. 1949. Zones of progressive regional metamorphism in the Moine Schists of the Western Highlands of Scotland. *Geol. Mag.* **86,** 43–56.

Lambert, R. St. J., Winchester, J. A. & Holland, J. G. 1979. Time, space and metamorphisms of the Scottish Highlands. *In*: Harris, A. L. *et al.* (eds.). *The Caledonides of the British Isles—Reviewed.* Spec. Publ. geol. Soc. London, **8,** 363–67.

MacQueen, J. A. & Powell, D. 1977. Relationships between deformation and garnet growth in Moine (Precambrian) rocks of western Scotland. *Bull. geol. Soc. Am.* **88,** 235–40.

Mendum, J. R. 1979. Caledonian thrusting in NW Scotland. *In*: Harris, A. L. *et al.* (eds.). *The Caledonides of the British Isles—Reviewed.* Spec. Publ. geol. Soc. London, **8,** 291–97.

Oxburgh, E. R. & Turcotte, D. L. 1974. Thermal gradients and regional metamorphism in overthrust terrains with special reference to the eastern Alps. *Schweiz. mineral. petrogr. Mitt.* **54,** 641–62.

Piasecki, M. A. J. & van Breemen, O. 1979. A Morarian age for the 'younger Moines' of central and western Scotland. *Nature, London,* **278,** 734–36.

—— & —— 1979. The 'Central Highland Granulities': cover-basement tectonics in the Moine. *In*: Harris, A. L., Holland, C. H. & Leake, B. E. (eds.). *The Caledonides of the British Isles—Reviewed.* Spec. Publ. geol. Soc. London, **8,** 139–44.

Powell, D. 1964. The stratigraphical succession of the Moine Schists around Lochailort (Inverness-shire), and its regional significance. *Proc. Geol. Assoc. London,* **75,** 223–50.

—— 1974. Stratigraphy and structure of the western Moine and the problem of Moine orogenesis. *J. geol. Soc. London,* **130,** 575–93.

—— & MacQueen, J. A. 1976. Relationships between garnet shape, rotational inclusion fabrics and strain in some Moine metamorphic rocks of Skye, Scotland. *Tectonophysics,* **35,** 391–402.

Rathbone, P. A. & Harris, A. L. 1979. Basement-cover relationships at Lewisian inliers in the Moine rocks. *In*: Harris, A. L. *et al.* (eds.). *The Caledonides of the British Isles—Reviewed.* Spec. Publ. geol. Soc. London **8,** 101–7.

Smith, D. I. 1979. Caledonian minor intrusions of the N. Highlands of Scotland. *In*: Harris, A. L., Holland, C. H. & Leake, B. E. (eds.). *The Caledonides of the British Isles—Reviewed.* Spec. Publ. geol. Soc. London, **8,** 683–97.

Soper & Brown 1971. Relationship between metamorphism and migmatism in the northern part of the Moine Nappe. *Scott. J. Geol.* **7,** 305–25.

Tanner, P. W. G. 1970. The Sgurr Beag Slide—a major tectonic break within the Moinian of the Western Highlands of Scotland. *Q. J. geol. Soc. London,* **126,** 435–63.

—— 1976. Progressive regional metamorphism of thin calcareous bands from the Moinian rocks of N.W. Scotland. *J. Petrol.* **17,** 100–34.

——, Johnstone, G. S., Smith, D. I. & Harris, A. L. 1970. Moinian stratigraphy and the problems of the central Ross-shire inliers. *Bull. geol. Soc. Am.* **81,** 299–306.

—— & Miller, R. G. 1980. Geochemical evidence for loss of Na and K from Moinian calc-silicate pods during prograde metamorphism. *Geol. Mag.* **117,** 267–75.

Tobisch, P. T., Fleuty, M. J., Merh, S. S., Mukhopadhyay, D. & Ramsay, J. G. 1970. Deformational and metamorphic history of Moinian and Lewisian rocks between Strathconan and Glen Affric. *Scott. J. Geol.* **6,** 243–65.

van Breemen, O., Pidgeon, R. T. & Johnson, M. R. W. 1974. Precambrian and Palaeozoic pegmatites in the Moines of northern Scotland. *J. geol. Soc. London,* **130,** 493–508.

——, Halliday, A. N., Johnson, M. R. W. & Bowes, D. R. 1978. Crustal additions in late Precambrian times. *In*: Bowes, D. R., Holland, C. H. & Leake, B. E. (eds.). *Crustal evolution in northwestern Britain and adjacent regions.* Geol. J. Spec. Iss. **10,** 82–106.

Wilson, D. & Shepherd, J. 1979. The Carn Chuinneag granite and its aureole. *In*: Harris, A. L., Holland, C. H. & Leake, B. E. (eds.). *The Caledonides of the British Isles—Reviewed.* Spec. Publ. geol. Soc. London, **8,** 669–75.

Winchester, J. A. 1972. The petrology of Moinian calc-silicate gneisses from Fannich Forest, and their significance as indicators of metamorphic grade. *J. Petrol.* **13,** 405–24.

—— 1974. The zonal pattern of regional metamorphism in the Scottish Caledonides. *J. geol. Soc. London,* **130,** 509–24.

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Appendix 6.

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The Sgurr Beag Slide within Moine rocks at Loch Eilt, Inverness-shire

A. W. Baird

SUMMARY: The Sgurr Beag Slide is shown to be a syn- to late-tectonic movement zone of local F_3 age within a regionally developed five phase deformation sequence. It truncates both lithological units and a major F_2 synform, and is associated with the intensification of an F_3 crenulation fabric, the widespread development of boudinage and the tightening of interlimb angles of major and minor F_3 folds. The Moine rocks of the Glenfinnan and Morar Divisions on either side of the slide share the same pre-sliding history of deformation and metamorphism; they both contain minor F_2 folds with strongly developed axial planar mica foliations enhanced in pelitic lithologies by migmatitic segregations, together with less abundant minor F_1 folds. The slide is refolded by isoclinal major F_4 folds and by major open F_5 folds.

The existence of a major tectonic break within the Moine Succession at Kinloch Hourn (Fig. 1) 30 km NE of Loch Eilt was first noted by Tanner (1970). The break, termed the Sgurr Beag Slide at Kinloch Hourn, separates rocks grouped as the Morar and Glenfinnan Divisions (Johnstone *et al.* 1969; Tanner *et al.* 1970) which were thought to be in continuous stratigraphical succession (Brown *et al.* 1970; Powell 1974). Locally, and only to the N of Kinloch Hourn, small tectonic slices of Lewisian basement rocks occur along and near the junction of the Morar and Glenfinnan divisions. The presence of these slices, at high stratigraphical levels within the Morar succession, not only provides evidence of the tectonic origin of the slide but also shows it to be a major feature (Tanner *et al.* 1970; Rathbone & Harris 1980).

The slide is characterized by being virtually bed parallel over most of its known outcrop. It shows no mylonitic or cataclastic features (Tanner 1970), although grain size reduction has been recorded at some localities (Rathbone & Harris 1980). It has been inferred to be a syn-metamorphic, ductile shear zone (Rathbone & Harris 1980; Powell *et al.* 1981). As a consequence of these features it is often extremely difficult to establish its position and structural setting especially where basement Lewisian material is not found within the slide zone.

In the Lochailort–Loch Eilt area (Fig. 2) the Lochailort Pelite (Powell 1964) has been correlated with the Glenfinnan Division (Brown *et al.* 1970; Powell 1974) and the rocks E of the Lochailort Pelite with the Morar Division (Powell 1964, 1974). Powell (1964) noted no disconformity between the Lochailort Pelite and the adjacent rocks and considered the junction to be entirely sedimentary; however, Rathbone & Harris (1980) indicated that there are large strain variations across the junction at Lochailort and considered it to be a southerly extension of the Sgurr Beag Slide.

The Morar/Glenfinnan junction is found on both sides of a major fold, the Glenshian synform (Powell 1966; Powell *et al.* 1981) and therefore it is inferred that this fold post-dates the formation of the slide. Extrapolation of this slide to the Sgurr Beag Slide at Kinloch Hourn requires that it has also been folded by the Loch Eilt antiform (Powell 1974), its trace being repeated in the eastern Loch Eilt area (Figs 1 & 2) and continuing northeastwards to Kinloch Hourn (Powell *et al.* 1981; and IGS 1" and 1:50000 Geological Maps 61 and 62 W, Scotland).

It is the aim of this paper to describe the nature of the Sgurr Beag Slide where it outcrops to the N of eastern Loch Eilt; to establish it as not only a zone of high strain but also a tectonic break across which translation and truncation have occurred; to discuss its local structural setting; and to discuss its implications in the interpretation of local and regional fold 'phase correlations together with the deformation history of the Morar and Glenfinnan divisions.

Lithological considerations

E of the slide

The rocks of the Glenfinnan Division E of the slide are a series of banded and striped psammites and semi-pelites with few pelitic horizons (Fig. 3). All the units strike NE–SW and are nearly vertical. Lithological layering, probably tectonically modified bedding, is planar; no other sedimentary structures have been observed. This may be a reflection of the original conditions of deposition or, more likely, sedimentary structures may have been obliterated by high strain throughout the area. The pelitic rocks contain a well-developed, planar, migmatitic fabric consisting of small, oblate, quartzo-feldspathic segregations, locally co-planar with the axial planes of F_2 folds. As the F_2 folds are isoclinal, the migmatitic fabric is mostly concordant with bedding. Psammitic and striped lithological units also carry an axial planar fabric, comprising orientated micas and oblate quartz grains,

Appendix 6.

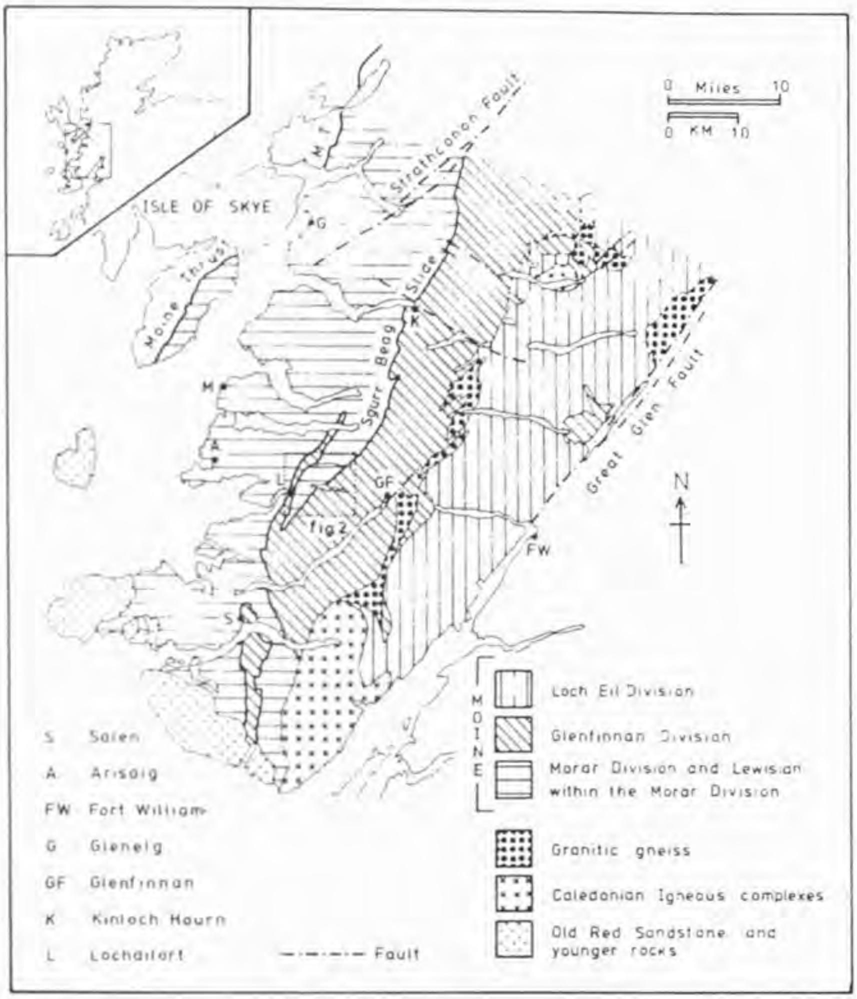

FIG. 1. Location map of the study area (modified after Johnstone *et al.* 1969).

which is axial planar to local minor F_2 folds. The aspect ratio of quartz grains is never more than two to one.

W of the slide

W of the slide (Fig. 3) a series of near-vertical, psammitic and striped psammitic, semi-pelitic and pelitic lithologies strike from NE-SW to N–S and pass westwards into a pelitic lithology (the Ranochan Pelite of Powell 1964). These rocks have planar modified bedding but no other unequivocal sedimentary structures have been observed. Oblate, planar migmatitic fabrics, which are axial planar to local minor F_2 folds and folded by F_3 folds, are present in the pelitic portions of the striped lithological units.

The rocks to the E and W of the slide are mineralogically similar. Psammites are light grey-cream coloured and usually contain more than 50% quartz with smaller quantities of feldspar, both plagioclase and K-feldspar, together with mica, usually both biotite and muscovite. Biotite is often evenly disseminated throughout the rock, whereas muscovite is more commonly found with biotite in thin micaceous partings between the quartzo-feldspathic layers. Accessory minerals include garnet, chlorite, calcite, iron ore, apatite, sphene and zircon. Striped and semi-peltic rocks in both divisions contain the same mineral assemblages as the psammitic rocks with a higher percentage of micas and feldspars and lesser amounts of quartz. The pelitic lithologies are medium to dark grey containing approximately equal proportions of quartz,

Appendix 6.

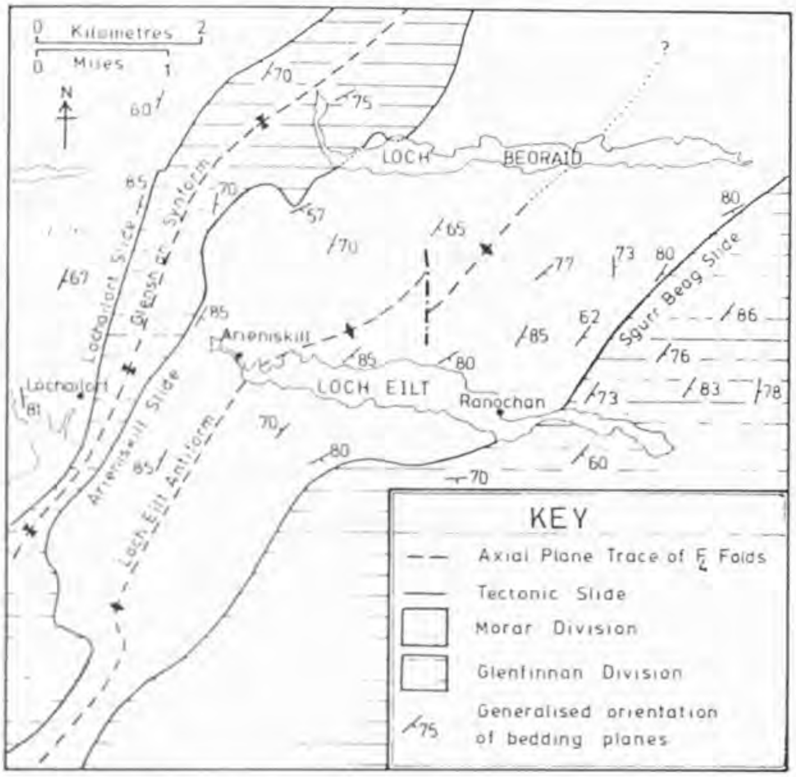

FIG. 2. Map of the Loch Eilt area showing essential structure. See Fig. 1 for location.

feldspar (plagioclase and K-feldspar) and mica (biotite and muscovite) with much less garnet and accessory amounts of chlorite, sillimanite, staurolite, iron ore, apatite, sphene and zircon. In no lithologies are there changes of grain size related to distance from the slide.

Sillimanite is formed sporadically in pelites on both sides of the slide, indicating high grade metamorphism; however, alumino-silicates are too uncommon to enable mapping of isograds. Calc-silicate rocks occur on both sides of the slide and the variable mineral parageneses in these have been used to establish metamorphic grade and patterns in the area (Powell *et al.* 1981). No pure quartzites occur, such as those held to be one of the features which distinguishes Morar and Glenfinnan Division rocks (Johnstone *et al.* 1969), but amphibolites, which in the south-western Moine are believed to characterize the Glenfinnan Division (Powell 1974), are restricted to the rocks E of the slide.

There are thus no marked mineralogical or lithological contrasts between the rocks on either side of the slide; however, the psammitic and striped lithologies W of the slide are folded by a major fold which is cut out by the slide (Fig. 3), which, therefore, on geometrical grounds alone must be a discontinuity. The detailed structural changes across the slide are discussed below.

Structural considerations

Neither cataclastic and mylonitic deformation textures nor widespread hydrothermal, retrogressive features, such as might relate to a high level thrust, are apparent in either the field or in thin section (see also Tanner 1970; Rathbone & Harris 1980). Thus, it follows that the slide, apparent because of its mapped geometrical relationships, was a result of ductile deformation which was either followed by post-sliding recrystallization, which overprinted and destroyed cataclastic and/or mylonitic textures, or which proceeded under conditions of dynamic recrystallization which prevented severe grain size reduction. Quantitative estimates of strain associated with the slide zone are impossible to obtain because no suitable strain markers are present.

E of the slide

In the area from Loch Eilt to Glenfinnan village (Fig. 4) the predominant folds are F_1 in age. Large

Appendix 6.

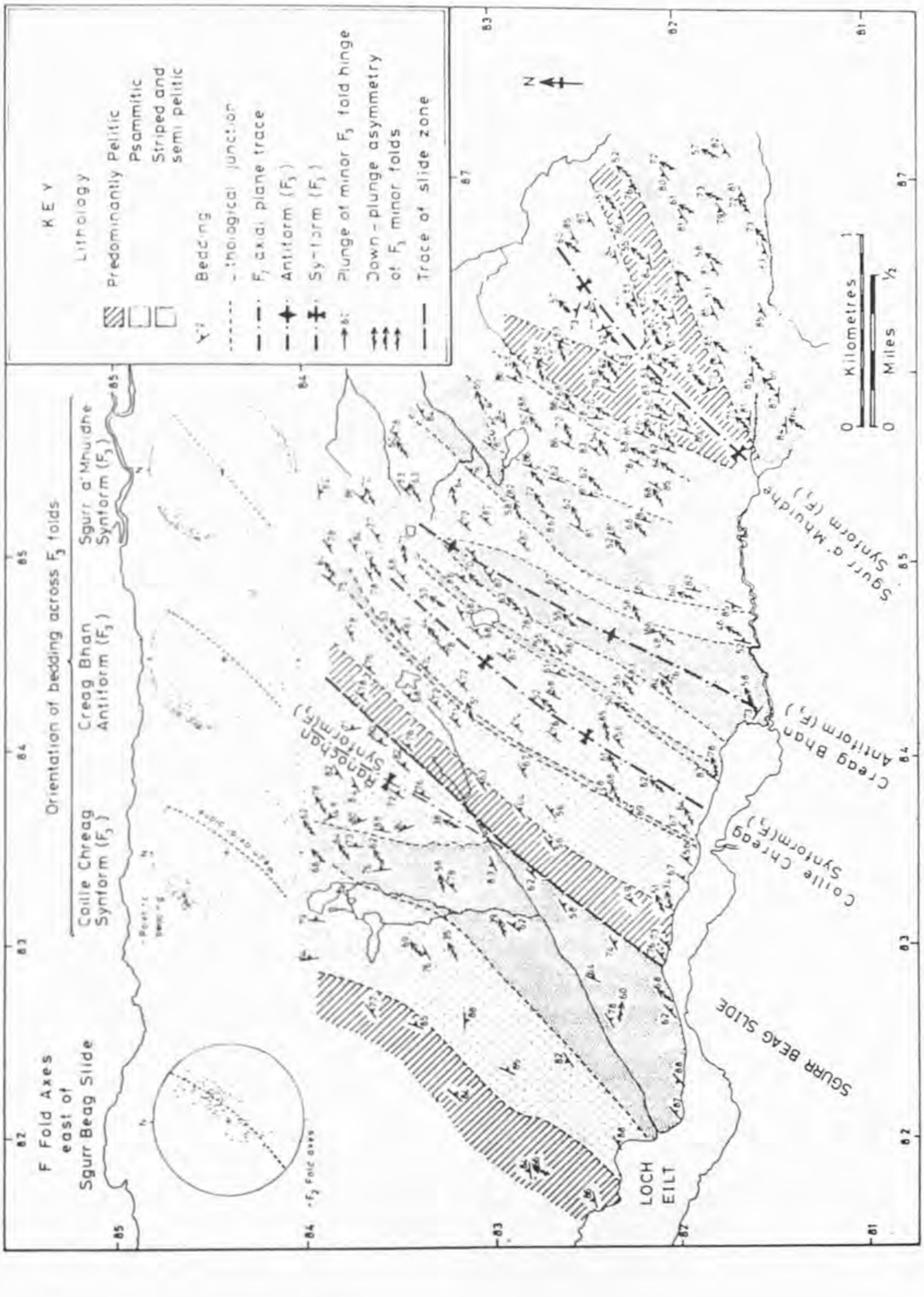

FIG. 3. Detailed map of the Moine rocks at the eastern end of Loch Eilt

Appendix 6.

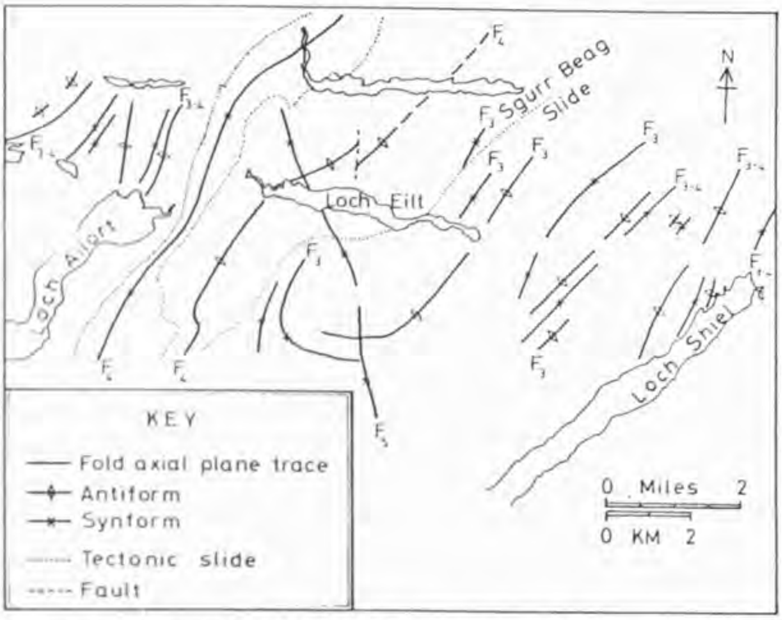

FIG. 4. Fold axial plane traces between Loch Shiel (Glenfinnan) and Loch Ailort (partly after Brown *et al.* 1970 and Powell 1974).

scale, tight to nearly isoclinal folds with NE–SW trending, nearly vertical axial planes and very steep NE plunges are present, together with abundant associated minor folds which have axial planar, tight mica crenulation fabric. These structures deform tight to isoclinal minor F_2 folds which have variable, steep to moderate, plunges towards the NE. The F_2 folds have either an associated axial planar penetrative mica fabric or an axial planar migmatitic fabric. Rarely, minor F_1 isoclines with an axial planar fabric are seen refolded by folds of the F_2 phase.

A progressive change in the structural geometry, especially in the structures of F_3 age, can be traced westwards into the slide zone. The Sgurr à Mhuidhe synform (Fig. 3) is a relatively open F_3 synform which plunges steeply to the NE. The complementary F_3 Creag Bhan antiform to the W is tighter, and farther W the F_3 Coille Chreag synform is isoclinal. The westward tightening of F_3 interlimb angles is indicated on the stereographic projections included in Fig. 3. F_3 minor folds also tighten westwards towards the slide and the number of minor F_3 folds decreases rapidly in this direction so that the rocks become markedly planar.

Moving W from the Sgurr à Mhuidhe synform (Fig. 3), boudinage of psammitic beds becomes progressively more common. Boudin pods containing isoclinal F_2 folds can be found near the slide and occasionally F_3 minor folds are boudinaged and thinned. These features are taken to indicate that boudinage occurred during or after the F_3 deformation.

The tightening of F_3 fold interlimb angles, the increase in the amount of boudinage, and the reduction in the number of minor F_3 folds approaching the slide can all be attributed to an increase of strain, with rotation of planar elements into the extension field of the strain ellipsoid (Flinn 1962). Minor F_2 fold hinges (stereographic projection, Fig. 3) are co-axial with the intersection of bedding and the migmatitic fabric and lie near the mean F_3 fold axial plane. There are, however, no progressive changes in orientation of F_2 hinge directions within this mean plane as the slide is approached, nor are there in the orientation of minor F_3 fold hinges and axial planes.

W of the slide

There is a major F_3 synform, the Ranochan synform, immediately to the W of the slide zone in the N of the area (Fig. 3). The western limb of this contains numerous open to tight minor F_3 folds which plunge moderately to the NE (Fig. 3) and fold the migmatitic fabric related to occasional tight F_2 minor folds. Moving eastwards across the axial plane trace of the Ranochan synform the F_3 minor folds have the opposite sense of vergence and have tighter interlimb angles. Slightly farther E the F_3 folds become even more tight and much less frequent (over a distance of 10–15 m). There is no progressive re-orientation of minor F_3 fold hinge directions or axial plane trends as the slide is approached. Within the pelitic unit adjacent to the slide, migmatitic segregations are tightly folded and are associated with very tight mica crenulations and strong boudinage of F_2 isoclines. These features indicate that F_3 strain increased progressively eastwards to the slide. The vergence of minor F_3 folds is spatially related to the Ranochan synform; it is therefore concluded that both the Ranochan synform and the slide formed during this phase of deformation. The evidence presented above seems incompatible with the Ranochan synform being cut out by a later, unrelated tectonic break, since sliding unrelated to and after F_3 deformation would be expected progressively to re-orientate minor F_3 folds. Nevertheless, the axial plane trace of the Ranochan synform is clearly oblique to the trend of the lithological divisions mapped (Fig. 3) and can be traced southwards to the pelite unit adjacent to the slide but not through it and into the rocks of the Glenfinnan Division; thus the F_3 Ranochan synform is cut out and replaced southwards by the slide (see Fig. 3).

Southwestwards along the N side of Loch Eilt (Fig. 3) there is no evidence for the structural cut out of the Ranochan synform. The rocks of the Morar Division contain minor F_3 folds with asymmetry indicating the presence of a major F_3 synform to the E. Likewise, immediately E of the slide the asymmetry of minor F_3 folds indicates a major F_3 synform lying to the E (the Coille Chreag synform). Thus, an E–W traverse along the northern side of Loch Eilt across the slide zone does not, from minor F_3 fold vergence, give indication of the structural cut out of the Ranochan synform by the slide, a situation which may well occur elsewhere along the slide. Since it can be assumed (where plunge direction is constant, see Fig. 3) that synforms necessarily have complementary antiforms, then it also follows that at least one antiform between the Ranochan synform and the Coille Chreag synform has been cut out or replaced by the slide.

Revision of the regional deformation sequence

As a result of this study and that of Powell *et al.* (1981) the regional sequence of fold phases previously recognized (Brown *et al.* 1970; Powell 1974) requires revision. The Glenshian synform, mapped by Powell (1974) as a regional F_3 fold with associated tight to isoclinal crenulation folds, clearly folds the Sgurr Beag Slide (Figs 2 and 4; see also Powell *et al.* 1981). The Sgurr Beag Slide, however, cross cuts the F_3 Ranochan synform (this paper) and thus the Glenshian synform must be regarded as a later regional F_4 fold. It is noteworthy that the F_3 and F_4 folds are regionally co-axial and co-planar and both have associated minor tight crenulation folds. Thus, they are impossible to distinguish in the field except where their relationship to the sliding event can be established, or refolded folds can be seen. For this reason the major folds immediately W of Glenfinnan (Fig. 4) could be of either F_3 or F_4 age. They are therefore categorized as F_{3-4} folds.

Later than all these structures is a phase of small scale open folds, F_5, with axial plane traces trending NNW–SSE and hinges plunging moderately to the SSE. These folds appear to be minor folds associated with the large open folds due S of Loch Eilt (Fig. 4) mapped by Brown *et al.* (1970) as F_3 and by Powell (1974) as F_4.

Regional implications

The slide at the eastern end of Loch Eilt is, on the basis of lithological mapping (IGS 1" and 1:50 000 Geological maps 61 and 62 W, Scotland) a southerly extension of the Sgurr Beag Slide at Kinloch Hourn. As at Kinloch Hourn it separates rock units which can be regarded on lithological grounds as belonging to the Morar and Glenfinnan Divisions (Tanner *et al.* 1970) but at Loch Eilt, unlike Kinloch Hourn, it is demonstrably a discordant structure. Further W, at Arieniskill and Lochailort (Fig. 2), it has been suggested that the slide is repeated by major F_4 folds (Powell *et al.* 1981). Thus, the Sgurr Beag Slide can be traced for at least 75 km along its outcrop within the southwestern Moine area.

Several important points arise from this and other work (Powell *et al.* 1981). (a) The deformation (F_3) associated with the formation of the slide deforms already migmatitic rocks. (b) There is no production of new migmatite associated with the F_3 deformation phase, but adjacent to the slide extreme crenulation of the pre-existing migmatitic fabric and rotation of this fabric towards the F_3 axial planes can be seen. (c) The slide displaces a pre-existing metamorphic pattern (Powell *et al.* 1981). (d) The regional sequence of fold phases previously recognized (Brown *et al.* 1970; Powell 1974) has to be revised. (e) The pre-sliding deformation and metamorphic history is the same on both sides of the slide and therefore the slide cannot be readily interpreted as a boundary between basement and cover sequences. Thus, the correlations of the Grampian Division of the Central Highland Granulites lying to the SE of the Great Glen Fault with the Morar Division, and of the Central Highland Division (considered to be older basement rocks lying beneath the Grampian Division) with the Glenfinnan Division (Piasecki 1980) are ill-founded unless the slide has juxtaposed once spatially distant rocks which had suffered tectono-metamorphic events differing only in their age.

Conclusions

The Sgurr Beag Slide in the SW Moine area, as elsewhere, is a major structural feature. It is a syn- to late-tectonic zone of movement associated with a zone of high strain and locally truncating lithostratigraphic units and regional F_3 folds. The absence of associated

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strain markers and fabrics, together with the recrystallization associated with later phases of deformation makes it very difficult to establish, in detail, a mechanism for formation of the slide. The slide separates rocks which contain the same pre-sliding structural history, thus it does not delimit or obscure an orogenic front separating rock groups with basement and cover relationships. Movement post-dated the local migmatization event but was synchronous with and followed by metamorphic activity.

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References

BROWN, R. L., DALZIEL, I. W. D. & JOHNSON, M. R. W. 1970. A review of the structure and stratigraphy of the Moinian of Ardgour, Moidart and Sunart-Argyll and Inverness-shire. *Scott. J. Geol.* **6**, 309–35.

FLINN, D. 1962. On folding during three-dimensional progressive deformation. *Q. J. geol. Soc. London*, **118**, 385–433.

JOHNSTONE, G. S., SMITH, D. I. & HARRIS, A. L. 1969. The Moinian Assemblage of Scotland. *In*: KAY, M. (ed.), *North Atlantic—geology and continental drift*. Mem. Am. Assoc. Petrol. Geol. **12**, 159–80.

PIASECKI, M. A. J. 1980. New light on the Moine rocks of the Central Highlands of Scotland. *J. geol. Soc. London*, **137**, 41–59.

POWELL, D. 1964. The stratigraphical succession of the Moine Schists around Lochailort (Inverness-shire) and its regional significance. *Proc. Geol. Assoc. London*, **75**, 223–50.

—— 1966. The structure of the south-eastern part of the Morar Antiform, Inverness-shire. *Proc. Geol. Assoc. London*, **77**, 79–100.

—— 1974. Stratigraphy and structure of the Western Moine and the problem of Moine orogenesis. *J. geol. Soc. London*, **130**, 575–93.

——, BAIRD, A. W., CHARNLEY, N. R. & JORDAN, P. J. 1981. The metamorphic environment of the Sgurr Beag Slide; a major displacement zone in Proterozoic, Moine rocks of Scotland. *J. geol. Soc. London*, **138**, 661–73.

RATHBONE, P. A. & HARRIS, A. L. 1980. Basement-cover relationships at Lewisian inliers in the Moine rocks. *In*: HARRIS, A. L., HOLLAND, C. H. & LEAKE, B. E. (eds). *The Caledonides of the British Isles—reviewed*. Spec. Publ. geol. Soc. London, **8**, 101–7.

TANNER, P. W. G. 1971. The Sgurr Beag Slide, a major tectonic break within the Moinian of the Western Highlands of Scotland. *Q. J. geol. Soc. London*. **126**, 435–63.

——, JOHNSTONE, G. S., SMITH, D. I. & HARRIS, A. L. 1970. Moinian stratigraphy and the problem of the central Ross-shire inliers. *Bull. geol. Soc. Am.* **81**, [illegible].

VAN BREEMEN, O., PIDGEON, R. T. & JOHNSON, M. R. W. 1974. Precambrian and Palaeozoic pegmatites in the Moines of northern Scotland. *J. geol. Soc. London*, **130**, 493–507.

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Structural dating of a Precambrian pegmatite in Moine rocks of northern Scotland and its bearing on the status of the 'Morarian Orogeny'

D. Powell, M. Brook & A. W. Baird

SUMMARY: A cross-cutting pegmatite emplaced into low grade, non-migmatitic, Proterozoic, Moine metasediments on the western sea-board of Inverness-shire, Scotland, gives Rb-Sr muscovite ages of 746–776 Ma which are interpreted as minimum estimates for the age of formation of the pegmatite. Intrusion followed at least one phase of folding and schistosity production in the country rocks and an early phase of syn-kinematic garnet growth, correlated with the regional D_2 tectono-metamorphic event. The pegmatite was folded and suffered considerable grain size reduction under metamorphic conditions during the regional D_3 deformation episode.

The dilational character and low grade context of the pegmatite show that it was not a product of melting in place during metamorphism of the adjacent Moine Schists. Its lack of initial, syn-kinematic crystallization textures and fabrics, together with the absence of evidence that might relate its emplacement to any regional phase of compressive ductile deformation, would be compatible with a considerable time lapse between its intrusion and the earlier deformation and metamorphism of the host rocks. We have, however, no direct evidence for the extent of this interval.

The relationship of the pegmatite to the earlier deformation and metamorphism of the host rocks provides clear evidence for substantial Precambrian orogenic activity but the age of this activity, whether 'Morarian' and/or 'Grenvillian', remains a matter of interpretation.

The Moine rocks of Scotland comprise a major assemblage of Proterozoic metasediments which lie within the Orthotectonic zone of the British Caledonides. Their orogenic history has recently been interpreted as involving 'Grenvillian'—c. 1000 Ma, 'Morarian'—c. 780 Ma and Caledonian—c. 470 to 420 Ma, episodes. Within the Moine outcrop of the Northern Highlands (Fig. 1) evidence for Precambrian orogenic activity is derived solely from isotopic analyses, with that for 'Grenvillian' events being based on the interpretation of Rb-Sr whole rock isochrons for the Ardgour granitic gneiss and Moine metasediments, as dating regional metamorphism at c. 1004 Ma (Brewer *et al.* 1979). Evidence for a 'Knoydartian' (Bowes 1968) or 'Morarian' (Lambert 1969) orogeny is derived from interpretation of Rb-Sr and U-Pb analyses of muscovite and monazite separates from pegmatites lying within high grade, migmatitic Moine rocks, and Rb-Sr whole-rock regression lines for Moine semi-pelites. Ages of 615 to 780 Ma are held to date regional metamorphism during which the pegmatites were secreted (van Breemen *et al.* 1978; Piasecki & van Breemen 1979; Lambert *et al.* 1979; Piasecki *et al.* 1981).

In the Northern Highlands geological evidence, other than isotopic, for such a complex history is equivocal and largely circumstantial. Whilst the evident, polyphasal nature of both the deformation and regional metamorphism of the Moine rocks (Ramsay 1963; Tobisch *et al.* 1970; Brown *et al.* 1970; Tanner 1970; Powell 1974; Lambert *et al.* 1979; Powell *et al.* 1981) could relate to two or more cycles of orogenic activity, a polyphase history does not, in itself, constitute *a priori* evidence for this without recognition of internal unconformities, or of widespread igneous activity which clearly separates groups of events related to orogeny. In this respect it is noteworthy that no unconformities have yet been recognized within the Moine rocks of the Northern Highlands; indeed, it would appear that all the major lithological divisions of at least the southwestern Moine share the same deformational and metamorphic history (Baird 1982; Roberts & Harris in press; I. L. Millar, pers. comm.). Further, the only igneous rocks so far recognized that might suggest separation of major orogenic cycles are the Glen Dessary Syenite, the Carn Chuinneag Granite and the Strath Halladale Granite (Fig. 1). Acceptance of an age for intrusion of the Glen Dessary Syenite at 456 ± 5 Ma (van Breemen *et al.* 1979) implies that the deformation and amphibolite facies metamorphism that followed the intrusive event are of Caledonian age, but it provides only a minimum age for the pre-intrusion deformation and regional metamorphism. The Carn Chuinneag Granite intruded already deformed low-grade Moine metasediments and was subsequently deformed under amphibolite facies metamorphic conditions (Long & Lambert 1963; Shepherd 1973). The U-Pb zircon and Rb-Sr whole-rock ages of 555 ± 10 Ma and 550 ± 10 Ma held to date intrusion, therefore, again indicate

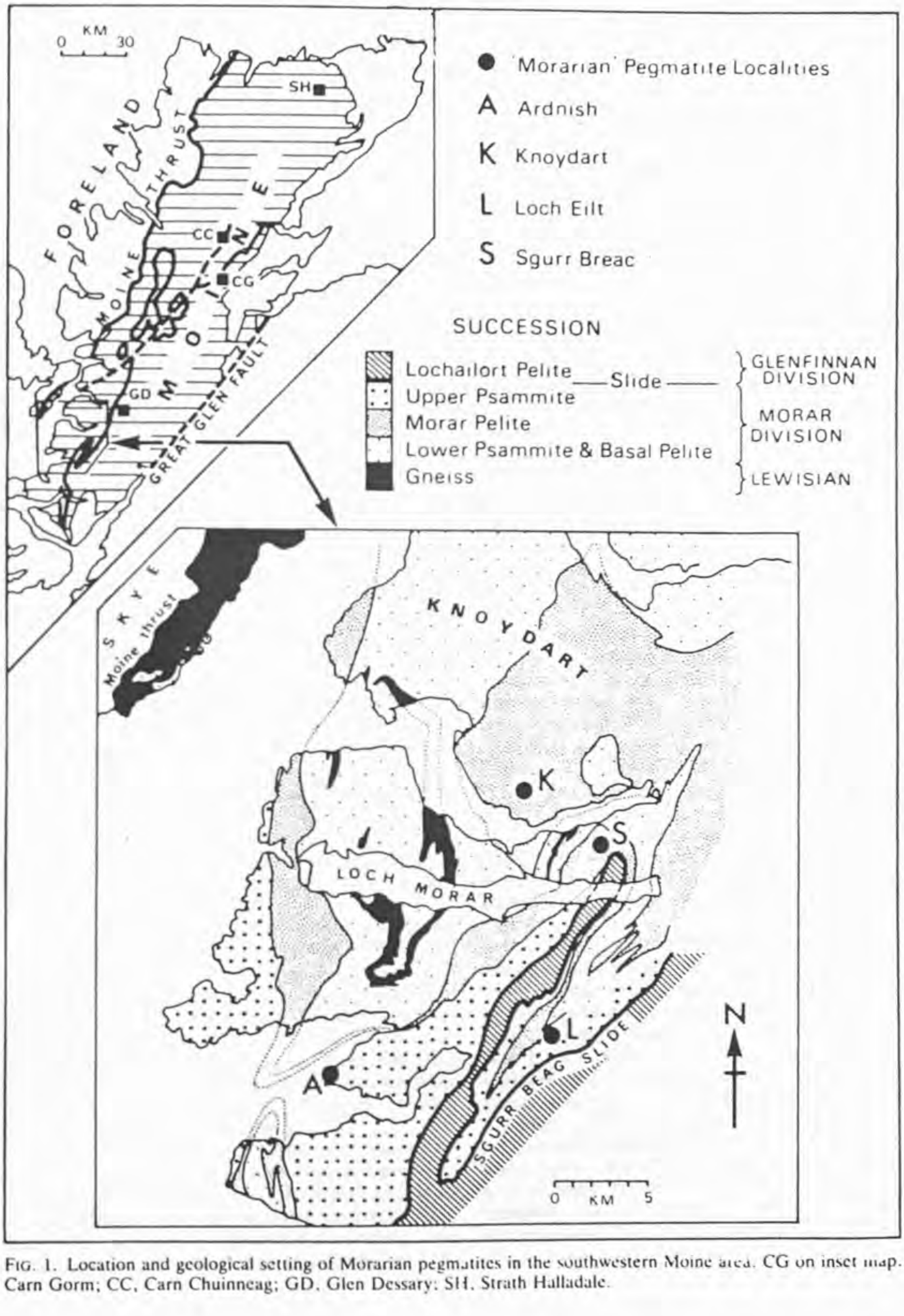

FIG. 1. Location and geological setting of Morarian pegmatites in the southwestern Moine area. CG on inset map, Carn Gorm; CC, Carn Chuinneag; GD, Glen Dessary; SH, Strath Halladale.

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post-granite Caledonian orogenic activity and provide a minimum age for early deformation and metamorphism (Long & Lambert 1963, Pidgeon & Johnson 1974). In both of these situations the earlier tectono-metamorphic activity could be early 'Caledonian' or Precambrian.

Acceptance of an Rb-Sr whole-rock isochron age of 649 ± 30 Ma for intrusion of the late phase of the Strath Halladale granite complex, which intrudes metasediments correlated with the Moine, appears, however, to demonstrate substantial Precambrian polyphase deformation and regional metamorphism (B. C. Lintern & M. Brook, pers. comm.).

Taken overall, the isotopic evidence for Precambrian thermal activity is substantial, but its precise age, complexity and nature remain a matter of interpretation and thus require further investigation.

'Morarian' pegmatites

In the Moine rocks of the Northern Highlands, pegmatites giving Rb-Sr muscovite ages of *c.* 750 Ma appear to be locally rare but are recorded from widely separated localities, *viz.* Knoydart, Sgurr Breac, Loch Eilt, Carn Gorm and, with the present example, Ardnish (Fig. 1). With the exception of the Ardnish body, the 'Morarian' pegmatites occur in high-grade, migmatitic country rocks and are described as concordant to lithological banding and migmatitic foliation (Gilletti *et al.* 1961, Long & Lambert 1963). The Carn Gorm and Loch Eilt examples are folded and the latter boudinaged (Long & Lambert 1963; van Breemen *et al.* 1974). Because of their mineralogy (quartz + muscovite + plagioclase + garnet ± K-feldspar ± biotite ± apatite ± beryl ± tourmaline), their concordance with migmatitic foliation, their deformed nature and the absence of associated larger bodies of possible parent igneous material, they have been interpreted as products of regional metamorphism (Giletti *et al.* 1961; Long & Lambert 1963; van Breemen *et al.* 1974).

In view of the assumed high blocking temperatures for radiogenic Sr migration in muscovite (500°C), the Rb-Sr ages, supported by similar U-Pb ages for monazite and zircon in the Sgurr Breac and Loch Eilt examples, are held to date crystallization, and thus metamorphism, at 780 ± 50 Ma (van Breemen *et al.* 1978). On this basis a 'Morarian' orogeny, affecting the Moine rocks of the Northern Highlands, has been accepted by many workers (most recently by Lambert *et al.* 1979 and Piasecki *et al.* 1981).

It is against this background that we present new evidence for the geological relationships of a previously unrecorded, well exposed, 'Morarian' pegmatite from the Moine rocks of western Inverness-shire. Unlike the previously described examples, the Ardnish pegmatite clearly cross-cuts bedding, an early schistosity and minor isoclinal folds in the country rocks; is clearly dilational in character; and is emplaced into low grade, non-migmatitic metasediments.

The nature and field relationships of the Ardnish pegmatite

The Ardnish pegmatite is a discontinuous sheet, 0.25 cm–1.25 m thick, exposed in a shore section on the south-eastern limb of the Morar antiform (Nat. Grid Ref. NM 6947 8139). It is tightly folded, as are the surrounding interbanded psammitic, semi-pelitic and pelitic metasediments and, in detail, it clearly transgresses the lithological banding (modified bedding) of the country rocks (Fig. 2). The sense of transgression is most clearly seen in low strain areas which coincide with fold hinge zones (Fig. 2) where, in outcrop, the pegmatite sheet lies clockwise to the lithological banding. On the fold limbs this relationship is generally impossible to detect, because the pegmatite sheet and lithological layering have been brought into approximate parallelism and both are severely modified by the later deformation. Locally, however, the same clockwise sense is observed (Fig. 2).

These relationships, together with the positioning and sense of vergence of folds in the country rocks adjacent to the pegmatite throughout most of its outcrop, accord with a simple model of folding of a sheet emplaced obliquely, but at a small angle, to the layering (Fig. 3).

Internally, the pegmatite is strongly deformed, exhibiting an L = S fabric with the long axes of porphyroclasts defining a lineation lying essentially coaxial with the hinges of tight-to-isoclinal folds of the pegmatite and its host rocks (Figs 2 & 4). The planar element of the fabric is oblique to the pegmatite margins and axial planar to the tight-to-isoclinal folds (Fig. 2).

The pegmatite comprises the assemblage quartz + muscovite + microcline + plagioclase + garnet, all of which occur both as large porphyroclasts (occasionally up to 1 to 5 cm in length), and as very much smaller crystals which form a fine-grained schistose matrix with grain size varying from about 1.0 to 0.02 mm (Fig. 5). Garnet occurs in two distinct size populations—as large porphyroclasts up to 3 mm in diameter, containing large inclusions of quartz and occasional plagioclase around which the matrix schistosity wraps (Fig. 5B), and as small 0.8 to 0.05 mm inclusion-free crystals, which were derived by fragmentation of porphyroclasts and have since partly recrystallized (Fig. 5C). Microcline and plagioclase similarly occur as large porphyroclasts with incomplete

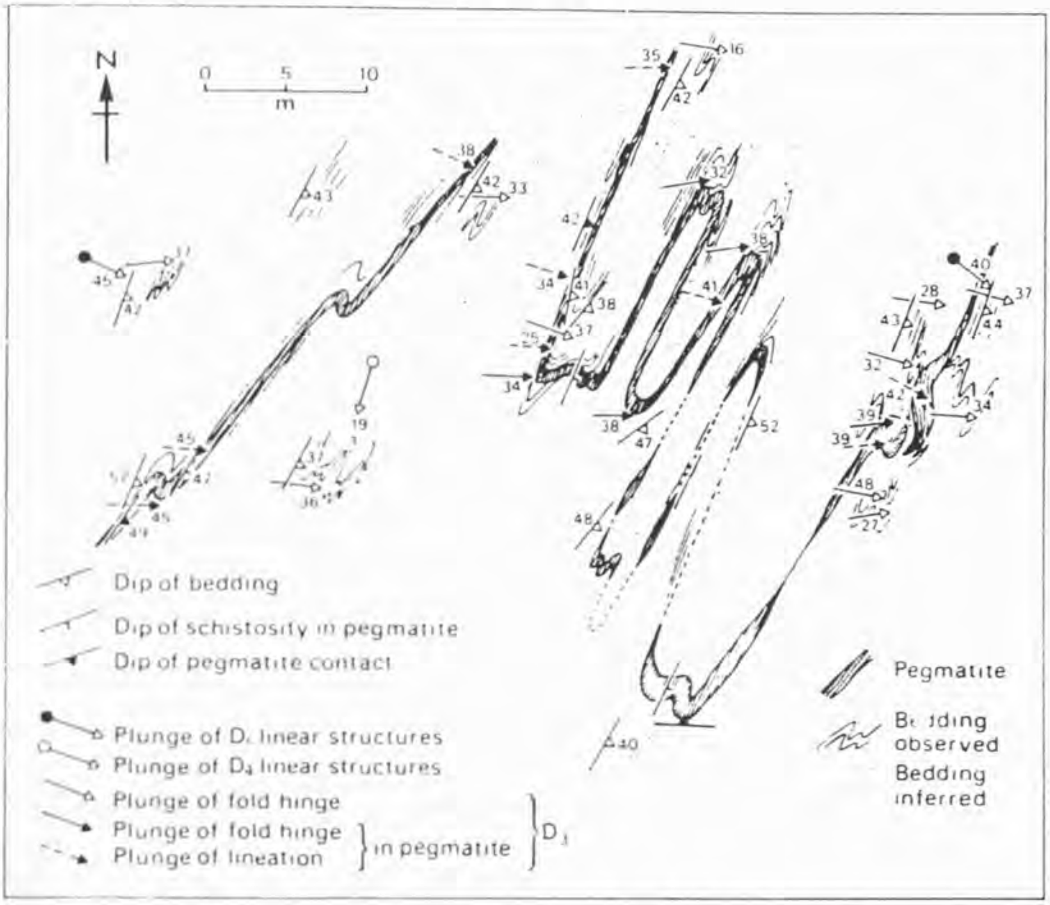

FIG. 2. Geological map of the Ardnish pegmatite locality. Ornament within the pegmatite indicates the general attitude of the grain-size reduction fabric (see text). D_3 to D_5 refer to deformation phases referred to in the text.

'haloes' or 'tails' of small, relatively strain-free sub-grains (Fig. 5A, C & D) and as components of the general fine-grained matrix. Porphyroclasts of quartz exhibit varying degrees of strain, as witnessed by undulose extinction and polygonization. They show a range of grain sizes from large porphyroclasts up to 4 mm in length, down to small, largely strain-free, crystals of the matrix (Fig. 5A). Quartz also forms stringers and ribbons of crystal aggregates parallel to the schistosity. Muscovite is present as porphyroclasts (up to 5 cm long), and as smaller crystals of the matrix (Fig. 5A, B & D).

All of the textural features demonstrate that an originally coarse-grained mineral assemblage has suffered severe grain-size reduction in response to the deformation that produced the folds of the pegmatite and host rocks. However, during deformation the temperatures, pressures and strain rates were such as to allow dynamic recrystallization of all constituent mineral phases. The textures are very similar to those described from mylonites but here there is no evidence for a strain regime dominated by simple shear.

The field and microscope studies have revealed no evidence for folding of, or production of, deformation fabrics within the pegmatite earlier than those so far described. The deformation phase that produced the predominant folds at Ardnish was the first to affect the pegmatite.

The isotopic age of the pegmatite

Four samples of muscovite books measuring up to 4 cm in diameter were collected from the pegmatite over a distance of approximately 50 m.

Appendix 7.

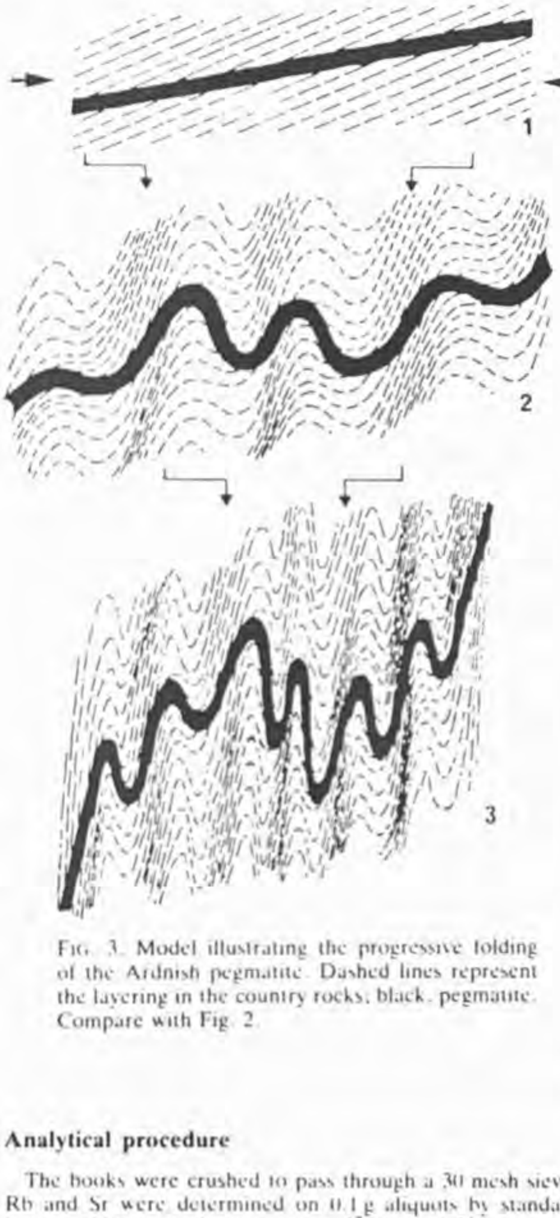

FIG. 3. Model illustrating the progressive folding of the Ardnish pegmatite. Dashed lines represent the layering in the country rocks; black, pegmatite. Compare with Fig. 2.

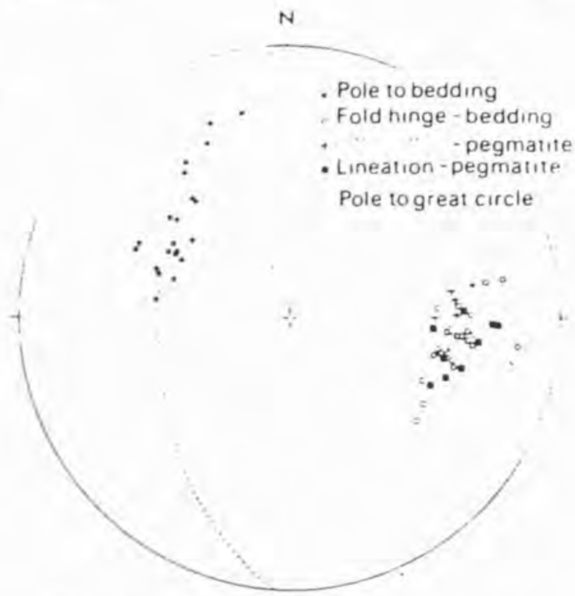

FIG. 4. Orientation of structural elements at the Ardnish pegmatite locality. Note the slight misfit between hinges of folds affecting the pegmatite and those of bedding. This reflects the cross-cutting nature of the pegmatite. Lower hemisphere equal angle projection.

Analytical procedure

The books were crushed to pass through a 30 mesh sieve. Rb and Sr were determined on 0.1 g aliquots by standard isotope dilution techniques using ^{87}Rb- and ^{84}Sr-enriched spikes. Each sample was digested in HF and HNO_3. Rb and Sr were separated from other elements in ion-exchange columns and loaded as phosphate on tantalum filaments. Isotopic measurements were made using a 30 cm radius 90° magnetic field sector mass spectrometer operated at 8 Kv acceleration voltage.

Analytical errors are estimated to be ±2% of ^{87}Rb/^{86}Sr and ±0.04% of ^{87}Sr/^{86}Sr at the 95% confidence level. The initial ^{87}Sr/^{86}Sr was taken to be 0.720 when calculating the age of the muscovites, but in view of the high (>50) ^{87}Sr/^{86}Sr values the choice of this initial ratio is not critical to the age obtained.

Potassium analyses were made in triplicate on separate aliquots of the muscovites. After dissolution in HF and $HClO_4$ acids, the samples were taken up in water. Aliquots of these solutions were mixed with a lithium internal standard and the potassium was determined on an Instrumentation Laboratory 243 digital flame photometer.

Argon measurements were made in duplicate by standard isotope dilution techniques involving fusion of the sample by induction heating in a bakeable vacuum system and mixing the liberated radiogenic ^{40}Ar with a known amount of ^{38}Ar spike or tracer. Isotopic ratio measurements were made in an AEI MS10 mass spectrometer run in the static mode and fitted with a digital output system. Errors are at the 95% confidence level and take into account uncertainties in the potassium analyses, the argon spike calibration, mass spectrometric ratio determinations, and the error enhancement involved in correcting for atmospheric argon.

The Rb/Sr analyses on the four muscovite samples give ages ranging from 776 Ma to 746 Ma (Table 1). Within error, samples 1, 2 and 4 are indistinguishable in age from one another, but applying the critical value test (Dalrymple & Lanphere 1969), sample 3 is significantly younger. This is also true of the K-Ar ages where sample 3 again yields a significantly younger age (Table 1).

The younger ages from sample 3 indicate that there has been partial loss of radiogenic strontium and partial, if not complete, radiogenic argon loss from the

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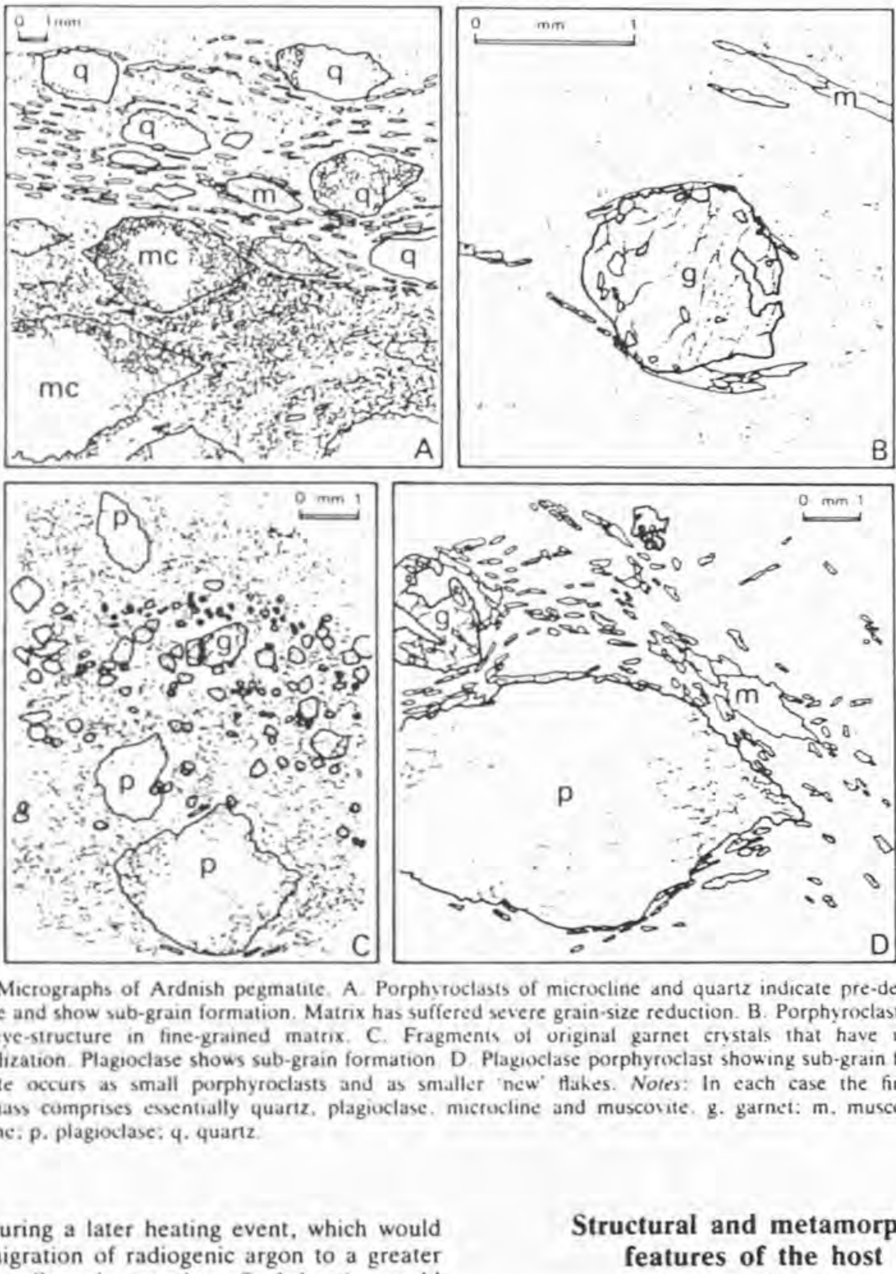

FIG. 5. Micrographs of Ardnish pegmatite. A. Porphyroclasts of microcline and quartz indicate pre-deformation grain size and show sub-grain formation. Matrix has suffered severe grain-size reduction. B. Porphyroclast of garnet within eye-structure in fine-grained matrix. C. Fragments of original garnet crystals that have undergone recrystallization. Plagioclase shows sub-grain formation. D. Plagioclase porphyroclast showing sub-grain formation. Muscovite occurs as small porphyroclasts and as smaller 'new' flakes. *Notes*: In each case the fine-grained groundmass comprises essentially quartz, plagioclase, microcline and muscovite. g, garnet; m, muscovite; mc, microcline; p, plagioclase; q, quartz.

muscovite during a later heating event, which would affect the migration of radiogenic argon to a greater degree than radiogenic strontium. Such heating could result from the Caledonian metamorphism recorded in the general area of the pegmatite locality (Brewer *et al.* 1979).

Since all the muscovite crystals were collected within a distance of approximately 50 m, it is possible that the other samples also suffered this disturbance but to a lesser degree. Argon has certainly been significantly outgassed but it is more difficult to assess to what degree radiogenic strontium has migrated.

Structural and metamorphic features of the host metasediments

The pegmatite lies within the biotite–zoisite zone of metamorphism as defined by the mineral assemblages of calc-silicate rocks (Kennedy 1949; Winchester 1974; Tanner 1976; Tanner & Miller 1980; Powell *et al.* 1981), which is equivalent to the Barrovian garnet zone as witnessed by the pelite paragenesis. Thus, the rocks are within the lower amphibolite facies of regional metamorphism.

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TABLE 1. *Rb-Sr and K-Ar analyses of muscovite from the Ardnish pegmatite*

No.	*Rb ppm*	*Sr ppm*	*$^{87}Rb/^{86}Sr$*	*$^{87}Sr/^{86}Sr$*	*Age (Ma)*
1	1900	7.206	4453	50.103	776 ± 15
2	1920	7.167	4541	50.333	765 ± 15
3	1890	6.872	4665	50.411	746 ± 15
4	1888	7.022	4648	51.566	766 ± 15
	% K	*% Atm^{40}Ar*	*nl/g Rg^{40}Ar*		*Age (Ma)*
1	8.49	25.4	186.12		491 ± 13
2	8.67	13.2	193.28		498 ± 12
3	8.79	7.4	157.19		410 ± 10
4	8.51	25.0	184.94		488 ± 10

Constants used in the calculations: $\lambda^{87}Rb = 1.42 \times 10^{-11} a^{-1}$; $\lambda^{40}K\beta = 4.962 \times 10^{-10} a^{-1}$; $\lambda^{40}Ke = 0.581 \times 10^{-10} a^{-1}$; samples 1 to 4 are of separate muscovite books.

The dominant structures of the host rocks are a set of asymmetric, tight-to-isoclinal, reclined folds with axial-planar quartz veins and coaxial lineations defined by crenulation and schistosity-bedding intersection. An associated axial planar fabric varies from a penetrative schistosity, defined by aligned micas in psammitic lithologies, to a crenulation schistosity in pelitic; a co-planar grain-size reduction fabric occurs in high strain zones. Garnet porphyroblasts in pelite (Fig. 6) are strongly wrapped by the axial planar fabric and contain inclusions of quartz and opaques. The inclusions form sigmoidal, symmetric, inclusion fabrics of rotational type (Powell & Treagus 1970). There is no continuity of the inclusion fabrics with the external schistosity and the grain size of included quartz is markedly greater than that of quartz in the matrix (Fig. 6). These observations, together with the occurrence of rotational garnet porphyroblasts with opposite senses of rotation lying only 1 cm apart along the same schistosity plane (S_3 in Fig. 6), indicate that the schists underwent garnet grade, syn-kinematic metamorphism *before* production of the predominant fold structures and their attendant planar fabrics. It is reasonable to assume that the schistosity, which is crenulated by the predominant set of folds, was coeval with, or formed earlier than, this early phase of garnet growth. Also, because this early schistosity, as well as minor isoclinal folds, are cross-cut by the pegmatite it is reasonable to deduce at least one phase of pre-pegmatite deformation and metamorphism.

The predominant folds of the country rocks thus deform not only the bedding, an early schistosity, and a suite of cross-cutting quartz veins which cut the pegmatite, but also minor, tight-to-isoclinal folds of bedding which have axial-planar quartz veins. They are themselves deformed by open minor folds with coaxial crenulation lineations (Figs 2 & 4).

The evidence given above suggests the following history of events:

a. deformation producing the earliest recognizable minor folds with attendant axial planar schistosity and quartz veins;
b. pegmatite intrusion;
c. formation of quartz veins;
d. deformation producing the predominant folds and axial planar fabrics; and
e. deformation producing sporadic open folds and crenulation lineations.

In this history the relationship between the early regional metamorphism and the earliest tectonic event is equivocal, but it is clear that the predominant set of folds (d. above) was produced under ductile conditions when microcline, plagioclase, muscovite and garnet were stable. The textural and structural features within *both* the pegmatite and the country rocks show that garnet was present before this deformation and, with lack of evidence to the contrary, it appears that its growth in the country rocks was coeval with the pre-pegmatite deformation phase. In view of the lack of a tectonic fabric in the pegmatite which pre-dates that produced during the development of the predominant folds, garnet growth in the pegmatite cannot be directly related to that in the country rocks. Indeed, the latter must be earlier. We have, however, no direct evidence to indicate the magnitude of the time lapse, either between these two periods of growth, or between the early phase of deformation and metamorphsim, and pegmatite emplacement.

Regional considerations

The deformation and metamorphic history of the Moine rocks of the area surrounding the pegmatite locality has recently been revised to include five main

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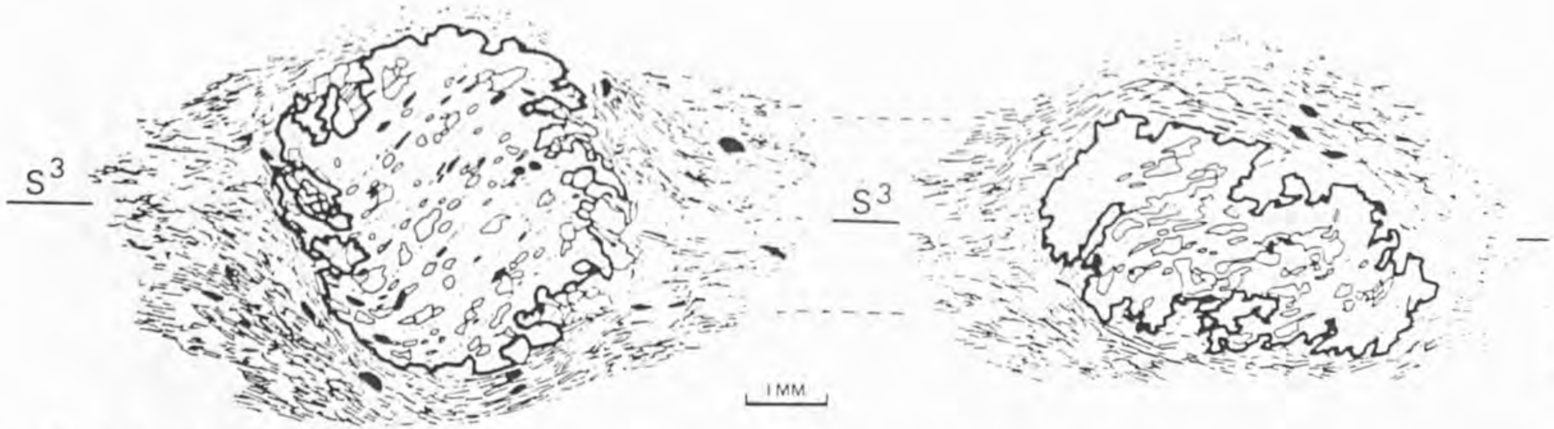

FIG. 6. Garnet porphyroblasts from host rocks. Inclusions of quartz and opaques define rotational inclusion fabrics. The fine-grained matrix comprises biotite, muscovite, quartz, plagioclase and opaques. Lines denote cleavage trends in individual mica flakes. The two garnet crystals illustrated occur within the same thin section and lie along the same plane of schistosity (S3) but show opposite senses of rotation.

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phases of deformation and two of metamorphism (Powell *et al.* 1981; Baird 1982; *cf* Powell 1974). Two main groups of events are separated by a phase of major crustal displacement across ductile shear zones—that which produced the Sgurr Beag Slide (Tanner 1970; Rathbone & Harris 1979). The earliest deformation (D1) produced major isoclinal folds and rare minor folds (Ramsay 1960; Powell 1974) but these have their main expression along the southwestern seaboard in Glenelg, Knoydart and Morar. The second deformation (D2) was synchronous with regional metamorphism and gave rise to major and minor folds, the latter being particularly common (Brown *et al.* 1970; Powell & MacQueen 1976; MacQueen & Powell 1977). Structures associated with the third deformation (D3) include steeply-plunging, reclined, major and minor folds which were probably coeval with formation of major 'slide' zones, including the Sgurr Beag Slide, and regional metamorphism (Powell *et al.* 1981; Baird 1982). D4 comprises regionally developed major and minor folds with generally shallowly inclined hinges which fold the Sgurr Beag Slide (Powell *et al.* 1981). D5 produced open-to-tight minor folds which are commonly conjugate in form and which have a regional southeasterly plunge. Locally this full sequence of events is not always readily apparent—fold geometry and the type of associated minor structures are *not* everywhere diagnostic of a particular deformation phase. The full sequence only becomes apparent with regional knowledge and by reference to the formation of the Sgurr Beag Slide and similar major structures, and to metamorphic fabrics.

The pegmatite locality is situated in ground which hitherto has not been investigated since the original survey by the Institute of Geological Sciences in the late 1930s. It is currently the subject of detailed mapping and whilst a thorough understanding of the correlation of the local structures with the regional scheme requires substantiation, tentative conclusions can be made.

In the Skye-Morar area garnet growth in the Moine Schists was syn-D2 in age (Powell & MacQueen 1976; MacQueen & Powell 1977). Rotational inclusion fabrics are typical and widespread, and thus the growth of garnet in the country rocks of the pegmatite locality at Ardnish is correlated with the D2 events. In recently mapped areas 3 km from the locality, D3 folds are reclined, SE- to ESE-plunging and are refolded by sub-horizontally plunging NE- to SW-oriented D4 folds. They deform a pre-existent strong mica schistosity. Thus, by analogy, the predominant folds at the pegmatite locality appear to correlate with the regional D3 deformation (Fig. 4). In this respect it is noteworthy that D3 is considered to be coeval with formation of the Sgurr Beag Slide (Baird 1982) which, in places, is associated with dynamic recrystallization fabrics similar to those produced in the Ardnish pegmatite.

Textural evidence given above shows that garnet was an original component of the pegmatite, so it remains possible that pegmatite formation could have occurred immediately after D2. If, however, we are correct in our assumption that formation of the early minor folds, early schistosity, and garnet growth in the host rocks, were all related in time, then it follows that the cross-cutting relationships and dilational character of the pegmatite, its lack of mica-rich marginal selvedges, its low grade metamorphic context and the rarity of similar pegmatites both locally and regionally could imply a not inconsiderable time gap and derivation from a source beyond the present level of exposure.

Derivation of the pegmatite in Ardnish from a deeper crustal region that was undergoing metamorphism whilst the Ardnish country rocks were already cooling from the same metamorphic episode cannot be ruled out. The rocks of the pegmatite locality could represent a higher level in the metamorphic pile than the source region. However, deeper levels in the same metamorphic complex are probably represented in Knoydart 18 km to the NE (Fig. 1), where a 'Morarian' pegmatite (Giletti *et al.* 1961) is emplaced in high-grade, migmatitic Moine metasediments. The oldest Rb-Sr ages reported for muscovite books within this pegmatite (766 ± 16 Ma recalculated from Giletti *op. cit.*) are indistinguishable from those for the Ardnish pegmatite and thus it could be argued that rocks in the two localities cooled through the blocking temperatures for Sr migration (*c.* 500°C), at the same time. This would not be expected if the rocks in Knoydart represent a deeper crustal level where the pegmatite was of local derivation during metamorphism—its muscovite cooling age might be expected to record later rather than coeval uplift. Unfortunately, the lack of precision in Rb-Sr mineral dating does not allow separation of such possibly closely spaced events. The similarities in mineralogy, original grain size and textures of the pegmatites, together with their lack of any indications of strain-controlled original mineral growth fabrics (Giletti *op. cit.*; James 1976), tends, however, to support the view that both pegmatites post-date, or represent a late static phase of, an early episode of regional metamorphism. It is pertinent to add that garnet is found in pegmatites of purely igneous origin (Deer *et al.* 1966; Leake 1968).

Conclusions

The Ardnish pegmatite is mineralogically similar and yields a compatible Rb-Sr muscovite age, to both the Knoydart pegmatite and the Loch Eilt pegmatite lying 12 km to the E (van Breemen *et al.* 1974). All three pegmatites Fig. 1) lie structurally beneath the Sgurr Beag Slide within metasediments that have been correlated with the Morar Division (Johnstone *et al.* 1969). The pegmatites form part of a suite previously

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regarded as metamorphic in origin and related to the development of high grade migmatitic rocks within the Moine during late Proterozoic orogenesis. The Loch Eilt example is described as pre-F2 in structural age (van Breemen *et al.* 1974) but it appears from our interpretation of the Ardnish locality that the folds of the pegmatite at Loch Eilt do not equate with the regional D2; they are most likely to be D4 in age. If this interpretation is shown to be correct, the problem of having different tectono-metamorphic histories affecting Morar Division rocks which lie at the same structural level would be resolved.

Consideration of the spatial distribution of isotopic ages reported for 'Morarian' pegmatites throughout the Northern Highlands (Giletti *et al.* 1961; Long & Lambert 1963; van Breemen *et al.* 1974, 1978) reveals a progressive decrease in maximum apparent ages from W to E, that is with increasing levels of Caledonian deformation and metamorphism (Brewer *et al.* 1979). The Ardnish pegmatite is the most westerly yet discovered and is emplaced in rocks of low metamorphic grade; it gives the oldest Rb-Sr age so far reported for muscovite. Providing that the timing of emplacement was not diachronous, the oldest age of 776 ± 15 Ma for the Ardnish pegmatite should be regarded as a minimum age for emplacement of the pegmatite suite as a whole. Our interpretation of the youngest Rb-Sr and K-Ar muscovite ages from Ardnish suggests that the apparently systematic regional changes in age may be due to partial, but increasing, disturbance during Caledonian metamorphic activity. U-Pb studies on 'Morarian' pegmatites (van Breemen *et al.* 1978) indicate an upper intersect on concordia at about 815 Ma, which may be a better indication of the age of emplacement of the Morarian pegmatite suite.

Despite the problems involved in interpretation of the isotopic data and the origin of the 'Morarian' pegmatites, the present study gives direct geological evidence for the development of a tectono-metamorphic complex earlier than *c.* 780 Ma whether or not our correlations of deformation phases are correct. The time interval between formation of the early structures (D2) with accompanying lower amphibolite facies mineral assemblages and intrusion of the Ardnish pegmatite is difficult to assess directly. However, isotopic evidence for a 1004 ± 28 Ma regional, syn-D2 metamorphism of nearby Moine metasediments (Brook *et al.* 1976, 1977; Brewer *et al.* 1979) could suggest a time-lapse of some 250 million years. A metamorphic origin for the 'Morarian' pegmatites has yet to be proved, as has their association with a major phase of compressive ductile deformation that might relate to orogenic activity.

There is thus no unequivocal evidence for a 'Morarian Orogeny' in Moine rocks of the southwestern Moine area and by implication elsewhere in the Northern Highlands.

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References

BAIRD, A. W. 1982. The Sgurr Beag Slide within Moine rocks at Loch Eilt, Inverness-shire. *J. geol. Soc. London*, **139**, 647–53.

BOWES, D. R. 1968. The absolute time scale and the subdivision of Precambrian rocks in north-western Scotland. *Forh. geol. Foren. Stockholm*, **90**, 175–88.

BREWER, M. S., BROOK, M. & POWELL, D. 1979. Dating of the tectono-metamorphic history of the southwestern Moine, Scotland. *In*: HARRIS, A. L., HOLLAND, C. H. & LEAKE, B. E. (eds). *The Caledonides of the British Isles—Reviewed*. Spec. Publ. geol. Soc. London, **8**, 129–37.

BROOK, M., POWELL, D. & BREWER, M. S. 1976. Grenville age for rocks in the Moine of north-western Scotland. *Nature, London*, **260**, 515–17.

——, —— & —— 1977. Grenville events in Moine rocks of the Northern Highlands, Scotland. *J. geol. Soc. London*, **133**, 489–96.

BROWN, R. L., DALZIEL, I. W. D. & JOHNSON, M. R. W. 1970. A review of the structure and stratigraphy of the Moinian of Ardgour, Moidart and Sunart—Argyll and Inverness-shire. *Scott. J. Geol.* **6**, 309–35.

DALRYMPLE, G. B. & LANPHERE, M. A. 1969. *Potassium-Argon dating. Principles, techniques and application to geochronology*. W. H. Freeman, San Francisco, 258 pp.

DEER, W. A., HOWIE, R. A. & ZUSSMANN, J. 1966. *Rock forming minerals*. Longmans, London, 528 pp.

GILETTI, B. J., MOORBATH, S. & LAMBERT, R. ST. J. 1961. A geochronological study of the metamorphic complexes of the Scottish Highlands. *Q. J. geol. Soc. London*, **117**, 233–72.

JAMES, J. A. 1976. *Correlation of events affecting Moine rocks bordering the Sound of Sleat, Inverness-shire, Scotland*. Thesis, PhD, Univ. of London. (unpubl.).

JOHNSTONE, G. S., SMITH, D. I. & HARRIS, A. L. 1969. Moinian assemblage of Scotland. *In*: KAY, M. (ed.). *North Atlantic geology and continental drift—a symposium*. Mem. Am. Ass. Petrol. Geol. **12**, 159–80.

KENNEDY, W. Q. 1949. Zones of progressive regional metamorphism in the Moine Schists of the Western Highlands of Scotland. *Geol. Mag.* **86**, 43–56.

LAMBERT, R. ST. J. 1969. Isotopic studies relating to the Precambrian history of the Moinian of Scotland. *Proc. geol. Soc. London*, **1652**, 243–5.

LAMBERT, R. ST. J., WINCHESTER, J. A. & HOLLAND, J. G. 1979. Time, space and intensity relationships of the Precambrian and lower Palaeozoic metamorphisms of the Scottish Highlands. *In*: HARRIS, A. L., HOLLAND, C.

Appendix 7.

H. & LEAKE, B. E. (eds) *The Caledonides of the British Isles—Reviewed*. Spec. Publ. geol. Soc. London, **8**, 363–7.

LEAKE, B. E. 1968. Zoned garnets from the Galway granite and its aplite. *Earth planet. Sci. Lett.* **3**, 311–16.

LONG, L. E. & LAMBERT, R. ST. J. 1963. Rb-Sr isotopic ages from the Moine series. *In*: JOHNSON, M. R. W. & STEWART, F. H. (eds). *The British Caledonides*. Oliver & Boyd, Edinburgh, 217–47.

MACQUEEN, J. A. & POWELL, D. 1977. Relationships between deformation and garnet growth in Moine (Precambrian) rocks of western Scotland. *Bull. geol. Soc. Am.* **88**, 235–40.

PIASECKI, M. A. J. & VAN BREEMEN, O. 1979. A Morarian age for the 'younger Moines' of central and western Scotland. *Nature, London*, **278**, 734–6.

——, —— & WRIGHT, A. E. 1981. Late Precambrian geology of Scotland, England and Wales. *Mem. Can. Soc. Petrol. Geol. Alberta*, **7**, 57–94.

PIDGEON, R. T. & JOHNSON, M. R. W. 1974. A comparison of zircon U-Pb and whole rock Rb-Sr systems in three phases of the Carn Chuinneag granite, northern Scotland. *Earth planet. Sci. Lett.* **2**, 105–12.

POWELL, D. 1974. Stratigraphy and structure of the western Moine and the problem of Moine orogenesis. *J. geol. Soc. London*, **130**, 575–93.

—— & TREAGUS, J. E. 1970. Rotational fabrics in metamorphic minerals. *Mineralog. Mag. London*, **37**, 801–14.

—— & MACQUEEN, J. A. 1976. Relationships between garnet shape, rotational inclusion fabrics and strain in some Moine metamorphic rocks of Skye, Scotland. *Tectonophysics*, **35**, 391–402.

——, BAIRD, A. W., CHARNLEY, N. R. & JORDAN, P. J. 1981. The metamorphic environment of the Sgurr Beag Slide; a major crustal displacement zone in Proterozoic, Moine rocks of Scotland. *J. geol. Soc. London*, **138**, 661–73.

RATHBONE, P. A. & HARRIS, A. L. 1979. Basement-cover relationships at Lewisian inliers in the Moine rocks. *In*: HARRIS, A. L., HOLLAND, C. H. & LEAKE, B. E. (eds). *The Caledonides of the British Isles—Reviewed*. Spec. Publ. geol. Soc. London, **8**, 101–7.

RAMSAY, J. G. 1960. The deformation of early linear structures in areas of repeated folding. *J. Geol.* **68**, 75–93.

—— 1963. Structure and metamorphism of the Moine and Lewisian rocks of the North-West Caledonides. *In*: JOHNSON, M. R. W. & STEWART, F. H. (eds). *The British Caledonides*. Oliver & Boyd, Edinburgh & London, 143–70.

ROBERTS, A. M. & HARRIS, A. L. 1983. The Loch Quoich Line—a limit of early Palaeozoic crustal reworking in the Moine of the Northern Highlands of Scotland. *J. geol. Soc. London*, **140**, in press.

SHEPHERD, J. 1973. The structure and structural dating of the Carn Chuinneag intrusion, Ross-shire. *Scott. J. Geol.* **9**, 63.

TANNER, P. W. G. 1970. The Sgurr Beag Slide—a major tectonic break within the Moinian of the Western Highlands of Scotland. *Q. J. geol. Soc. London*, **126**, 435–63.

—— 1976. Progressive Regional Metamorphism of Thin Calcareous Bands from the Moinian Rocks of N. W. Scotland. *J. Petrol.* **17**, 100–34.

—— & MILLER, R. G. 1980. Geochemical evidence for loss of Na and K from Moinian calc-silicate pods during prograde metamorphism. *Geol. Mag.* **117**, 267–75.

TOBISCH, P. T., FLEUTY, M. J., MERH, S. S., MUKHOPADHYAY, D. & RAMSAY, J. G. 1970. Deformational and metamorphic history of Moinian and Lewisian rocks between Strathconon and Glen Affric. *Scott. J. Geol.* **6**, 243–65.

VAN BREEMEN, O., PIDGEON, R. T. & JOHNSON, M. R. W. 1974. Precambrian and Palaeozoic pegmatites in the Moines of northern Scotland. *J. geol. Soc. London*, **130**, 493–508.

——, HALLIDAY, A. N., JOHNSON, M. R. W. & BOWES, D. R. 1978. Crustal additions in late Precambrian times. *In*: BOWES, D. R. & LEAKE, B. E. (eds). *Crustal Evolution in Northwestern Britain and Adjacent Regions*. Geol. J. Spec. Issue, **10**, 82–106.

——, AFTALION, M., PANKHURST, R. J. & RICHARDSON, S. W. 1979. Age of the Glen Dessary Syenite, Inverness-shire: diachronous Palaeozoic metamorphism across the Great Glen. *Scott. J. Geol.* **15**, 49–62.

WINCHESTER, J. A. 1974. The zonal pattern of regional metamorphism in the Scottish Caledonides. *J. geol. Soc. London*, **130**, 508–24.

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Discussion of the structural setting and tectonic significance of the Glen Dessary Syenite, Inverness-shire

Journal, Vol. 141, Part 6, 1984, pp. 1033–42

MR A. W. BAIRD writes: Roberts, Smith & Harris (1984) have produced a significant piece of structural mapping in Glen Dessary but I cannot agree with their interpretation of its regional significance. I propose an alternative which does not involve the unlikely situation of regional scale vertical extension, and is more in keeping with the structural history obtained from studies nearby (e.g. Baird 1982; Powell *et al.* 1981).

Roberts *et al.* describe a series of D_3 major folds with sub-vertical axial plane traces. The D_3 Glen Dessary Synform has an intensely curvilinear hinge line with a sub-vertical extension (X) direction within its fold axial plane. All the D_3 major folds east of the Glen Dessary Synform have sub-horizontal fold hinge lines. They attribute the change of hinge line orientation and development of curvilinear folds to a rapid increase in the amount of D_3 strain westwards, and they link this strain variation to the generation of the 'steep belt', with the Loch Quoich line representing the eastern limit of intense D_3 strain and upright reworking.

The post-D_3 Glen Dessary Syenite was intruded at 456 ± 5 Ma. (van Breemen *et al.* 1979). Elsewhere in the Western Moine schists, D_2 is regarded as a Precambrian (Grenville) tectonic event (Brook *et al.* 1976; Brook *et al.* 1977; Brewer *et al.* 1979; Powell *et al.* 1983). Therefore Roberts *et al.* argue that the production of the steep belt during D_3 deformation must represent the first Lower Palaeozoic structural reworking of the Moine rocks.

However, further southwest, at Loch Eilt, the Sgurr Beag Slide is a regional D_3 structure which has been folded by major tight to isoclinal D_4 folds with sub-vertical axial planar traces and sub-horizontal hinge lines (Baird 1982; Powell *et al.* 1981). Work now shows that prior to D_4 deformation the Sgurr Beag Slide was a flat structure over which the rocks of the Glenfinnan and Loch Eil Divisions were transported to the northwest.

Moving downwards towards the Sgurr Beag Slide, the Glenfinnan and Loch Eil Division rocks show a progressive increase in D_3 strain: D_3 fold interlimb angles decrease and D_3 hinge lines rotate within their fold axial planes towards the extension (X) direction (Baird 1982).

Beyond the eastern limit of steep belt, upright reworking the Loch Eil Division rocks still preserve sub-recumbent major D_3 structures (Druim Beag Synform).

I interpret the Glen Dessary Synform as having formed as a flat-lying D_3 curvilinear fold with its extension (X) direction sub-horizontal towards the northwest, which has been reorientated by D_4 deformation to produce the steep belt. Consequently, one or more of the major folds with sub-horizontal hinge lines east of the Glen Dessary Synform must belong to the D_4 phase of deformation.

It is relevant to note that the suite of microdiorites in the Loch Eil–Loch Eilt area was intruded after the formation of the Sgurr Beag Slide (D_3). The amount of deformation within the sheets is related to the intensity of D_4 steep-belt deformation.

In areas such as Loch Eilt and Glen Dessary, where two phases of intense Caledonian deformation (D_3 and D_4) have produced tight to isoclinal folds, markers within the local deformation sequence, such as the Sgurr Beag Slide and the microdiorite suite, may provide the best means of establishing the complete local structural chronology.

DRS A. M. ROBERTS, D. I. SMITH, A. L. HARRIS and R. E. HOLDSWORTH reply: The main difference between our interpretation of the structural setting and tectonic significance of the Glen Dessary complex and that of Baird involves the age of the structures that comprise the regional steep belt. Baird believes that two sets of major upright folds can be recognized within the steep belt, and regards the earlier of these as being coeval with the Sgurr Beag Slide (ductile thrust) and initially recumbent (D_3 *sensu* Baird). He believes that these 'D_3' structures were reorientated into their present steep attitude by a later set of folds (D_4 *sensu* Baird). However, our regional mapping and detailed fabric studies of a large area between Loch Shiel and Loch Affric, including the Glen Dessary area, show that all the major upright folds within the steep belt formed with their present orientation, and can be referred to a single episode of progressive deformation (D_3 *sensu* Roberts *et al.* 1984). This set of folds overprints the Sgurr Beag ductile thrust in Morar and Ardnamurchan, and also overprints two sets of previously recumbent structures, probably pre-Caledonian in age (see Roberts & Harris 1983,

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p. 891), which retain their original attitude in the 'flat belt' to the east of the Loch Quoich Line (Roberts & Harris 1983, Holdsworth & Roberts 1984).

The field-based evidence outlined below is relevant to this discussion.

1. If the upright folds within the steep belt can be referred to two separate episodes of folding (D_3 and D_4 *sensu* Baird), interference between these two sets of folds should be common. Baird (1982, fig. 4) was unable to demonstrate that any of his D_3 fold axial traces were refolded about D_4 fold axial traces and, similarly, we have recognized no interference patterns of this type. Interference within the steep belt is exclusively between steep belt folds (our D_3) and the earlier D_1 and D_2 structures which we recognize unmodified in the flat belt to the east, and can trace westwards into the steep belt where they become overprinted by our D_3.

2. Baird (1982, p. 652) has pointed out that his minor D_3 and D_4 folds show similar styles and geometry and concluded, perhaps significantly, that 'they are impossible to distinguish in the field'. However, by contrast, the existence of two sets of major folds (say the D_3 and D_4 of Baird) within the steep belt would make mapped fold vergence patterns very irregular and unpredictable on all scales; systematic map patterns of fold vergence would not be expected. The reality is that throughout the regional steep belt, and in the flat belt to the east, careful mapping, and an awareness of the curvilinear nature of the fold hinges, produces a systematic pattern of minor and major fold vergence related to a single set of NW–SE trending, regionally persistent, upright folds. In particular, we should mention that detailed mapping of fold vergence to the east of the Glen Dessary Synform shows patterns consistent with a single set of upright D_3 folds, and that there is no evidence to support Baird's suggestion that one of the major structures of that area is late (viz. D_4).

3. No evidence has yet been presented by Baird to support the correlation of his 'D_3' Druim Beag Synform with his major D_3 folds farther West. This synform has been described by Baird as a major, sub-recumbent structure, and as such it begs a correlation with the (possibly Precambrian) D_2 structures described from the Glen Garry–Loch Quoich and Glen Dessary areas (Holdsworth & Roberts 1984; Roberts & Harris 1983; Roberts *et al.* 1984). These structures have a N–S extension direction, and hence cannot have been produced during WNW-directed Caledonian ductile thrusting.

An alternative explanation has been put forward by R. A. Strachan (1985 and see following discussion). Strachan has mapped the Druim Beag Synform not as a recumbent structure, but as an upright structure, of only local importance, unrelated both to the formation of the regional steep-belt and the Sgurr Beag ductile thrust.

If either of these explanations is correct then Baird's suggestion that the Sgurr Beag ductile thrust and the Druim Beag Synform formed coevally is incorrect.

4. Baird's model involves passive rotation of an initially sub-recumbent Glen Dessary Synform into its present upright attidude during later folding. Such passive rotation would not only reorientate Baird's D_3 folds, for which evidence of interference patterns is notably absent, but also the earlier sub-recumbent D_1 and D_2 folds. Passive rotation of D_1 and D_2 folds on the limbs of steep belt folds has indeed been recorded by Roberts *et al.* (1984). However, if the hinge zone of the Glen Dessary synform is traced northwards from the Glen Dessary area, where it is markedly curvilinear, into the Loch Quoich area, where it has been called the Gleouraich Synform (Roberts & Harris 1983), the large passive rotations of D_1 and D_2 structures required by Baird's model are demonstrably absent. Instead, in the synform hinge zone at Loch Quoich early structures are still commonly sub-recumbent or gently inclined, and are similar to those in the hinge zone of the nearby Spidean Mialach Antiform, and further east, to those in the flat-belt east of the Loch Quoich Line (Roberts & Harris 1983; Holdsworth & Roberts 1984). It is therefore apparent that the hinge zone of the Gleouraich–Glen Dessary Synform was generated as an upright structure.

In the light of the above points we suggest an alternative model for the structural evolution of the regional steep belt. In agreement with Baird, we envisage the Sgurr Beag ductile thrust to be the earliest Phanerozoic structure in the region. However, outside the *c.* 1 km-wide slide zone we do not recognize any structures, and certainly not major folds, associated with the ductile thrust. It is more likely that, following the emplacement of the Sgurr Beag Nappe (Glenfinnan and Loch Eil divisions) onto the Morar Division by the Sgurr Beag ductile thrust, the only structures and fabrics within the nappe itself were the still flat-lying, probably pre-Caledonian, D_1 and D_2 structures referred to above.

It is possible to speculate further on the regional deformation pattern which led to the development of the steep belt structures in both the hanging wall and footwall of the Sgurr Beag thrust. It is probable that movement on the Sgurr Beag ductile thrust was followed by displacements on the underlying thrusts—the Knoydart ductile thrust and the Moine Thrust successively—in a foreland propagating sequence (Barr 1983). Ductile thrusting was succeeded, and may have been partly overlapped, by a major phase of regional shortening which produced the upright folds (our D_3) forming the regional steep belt. These folds fold the earlier Sgurr Beag and Knoydart ductile thrusts. It is possible that sticking on the Moine Thrust itself was responsible for the formation of these upright folds (Rathbone *et al.* 1983). We envisage that these folds propagated eastwards as far as the Loch

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Quoich Line (Roberts & Harris 1983), the rocks to the east of which were scarcely affected.

Baird is concerned about the development of regional, vertical extension. However, significant shortening, as seen in the steep belt, above a sub-horizontal detachment, such as the Moine Thrust, would be expected to produce just such vertical extension (cf. Coward & Siddans 1979). We therefore do not share Baird's concern about the sub-vertical extension described by Roberts *et al.* (1984). Such sub-vertical extension would bring up hotter, deeper level rocks in those areas where sub-horizontal shortening was greatest, i.e. the steep belt. This may explain the presence of the sillimanite 'overprint' in the steep belt and, given that this 'exhumation' occurred subsequent to ductile thrusting, may explain the younger cooling ages of these rocks relative to those further west (see Powell 1983).

DR R. A. STRACHAN provides the further comments: Roberts & Harris (1983), Holdsworth & Roberts (1984), Roberts *et al.* (1984) and Strachan (1985) demonstrate that the Glenfinnan and Loch Eil divisions have undergone a common Precambrian structural history involving two phases of recumbent folding. These workers suggest that subsequent Caledonian reworking may be viewed in terms of (a) displacement along the Sgurr Beag Slide and (b) upright folding. One particularly important set of upright folds may be recognized throughout W Inverness-shire between Glenfinnan and Loch Cluanie, and this corresponds to the major NNE–SSW trending folds such as the Glen Dessary synform described by Roberts & Harris (1983) and Roberts *et al.* (1984). None of these authors recognize any set of folds which might be directly associated with movement along the Sgurr Beag Slide. Baird, in contrast, postulates that the Glen Dessary synform and, by inference, many others of the folds described by Roberts & Harris (1983) and Roberts *et al.* (1984), formed as recumbent structures coeval with thrust-type movement along the Sgurr Beag Slide. Subsequent upright folding is inferred to have rotated these postulated recumbent folds into their present highly inclined attitudes. Baird specifically indicates that evidence for these recumbent folds is present in the Loch Eil division, and cities the Druim Beag synform as an example.

At issue is the regional extent and nature of those folds described as 'D_3' by Baird (1982; this discussion) and Strachan (1985). Recent work in the Loch Eil area demonstrates that the Glenfinnan and Loch Eil divisions have undergone five major phases of folding (D_1–D_5) (Strachan 1985). This area is therefore more structurally complex than either the Glen Dessary or Loch Quoich areas where it is evident that the two divisions have only been affected by four phases of folding (Roberts & Harris 1983; Roberts *et al.* 1984). The apparent discrepancy between these two structural schemes in terms of the numbers of fold phases present is best resolved by the suggestion that the 'D_3' structures of the Loch Eil area are regionally impersistent and die out northwards (Strachan 1985, table 3). The fourth phase folds recognized at Loch Eil would then appear to correlate directly with the upright third phase folds described by Roberts & Harris (1983) and Roberts *et al.* (1984) from Loch Quoich and Glen Dessary. I have examined all these areas in some detail and conclude that there is little likelihood that the 'D_3' folds recognized at Loch Eil can be extrapolated as far north as Baird argues.

In the Loch Eil area, the 'D_3' folds, including the Druim Beag synform, have upright to sub-vertical axial planes and are tight–open in style. The D_3 folds have a regionally arcuate trend about N–S to NNE–SSW trending D_4 folds. The level of D_4 strain within the Loch Eil division is generally low: D_4 folds are gentle–open in style, and D_3 and D_4 fold axes typically intersect at high angles. A rapid westward increase in the level of D_4 strains is marked by the progressive tightening of D_4 folds which define the steep belt in this area. East of the Loch Quoich Line, where D_4 strains are low, it can be seen that D_3 folds are likely to have had an original NW–SE or WNW–ESE trend (Strachan 1985) which is markedly oblique to both the Sgurr Beag Slide and the steep belt. The field evidence indicates that all these D_3 folds, including the Druim Beag synform, were generated with upright to sub-vertical axial planes. Baird's suggestion that the D_3 folds within the Loch Eil division were originally recumbent implies that their axial planes have been subsequently rotated into an upright attitude during D_4 folding. This necessitates that levels of D_4 strain within the area east of the Loch Quoich Line are comparable with those to the west within the steep belt, which is clearly not the case.

In conclusion, Baird's model is not supported by the field evidence. Regional considerations indicate that the D_3 folds are restricted to the area south of Glen Dessary. Furthermore, these structures did not evolve as recumbent folds associated with movement along the Sgurr Beag Slide, but rather as a series of upright structures oblique to the main Caledonian trend within the Western Highlands. It is likely that these folds either represent an early phase of NNE–SSW directed Caledonian compression, or that they relate to the Precambrian event. In view of these criticisms of Baird's comments, the work of Roberts & Harris (1983) and Roberts *et al.* (1984; this discussion) represents a more plausible model for the Caledonian structural evolution of the Sgurr Beag Nappe. All the indications are that the major folds described by these workers from the Glen Dessary and Loch Quoich areas were generated as upright folds, possibly above a major decoupling zone which lies at some depth below the present erosion surface in the Western Highlands.

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DR D. BARR comments: Baird draws attention to two features of the Glen Dessary Synform which he considers to be incompatible with a 'steep belt' age (D_3 of Roberts *et al.* 1984): (a) its intense curvilinearity and the inferred high ductile strain and (b) the unlikelihood of regional-scale vertical extension. Both these features are compatible with and indeed predictable from a model in which the Loch Quoich Line represents an eastern limit of intense D_3 strain, developed above a basal décollement as inferred by Roberts *et al.* (this discussion). Differential WNW–ESE shortening (equivalent on a regional scale to pure shear) gives rise to compatibility problems, which require the existence of a component of simple shear in the zones separating domains of contrasting pure shear. The magnitude of this vertical simple shear component is controlled by the depth to decollement and by the horizontal gradient in the pure shear component. Work currently in progress (with A. M. Roberts) applies this model to the generation of the steep belt and the Loch Quoich Line. At the type locality of the Loch Quoich Line, D_3 reworking declines rapidly eastwards, large, sub-vertical simple shear strains are generated, and F_3 folds are curvilinear (cf. Roberts & Harris 1983, fig. 3). East of Glen Dessary (Roberts *et al.* 1984, fig. 3) there is a gradual transition between the steep belt and the flat belt, the simple shear component is small, and F_3 axes are sub-horizontal. The Glen Dessary Synform is one of a series of augen of low D_3 shortening which crop out within the steep belt. The associated steep gradient in the pure shear component generates large simple shear strains and curvilinear F_3 folds. The magnitude of this simple shear component dies out downwards to zero at the basal décollement, so its sub-vertical X-direction does not create space problems at depth.

Curvilinearity of 'steep belt' folds about a sub-vertical X-direction is not restricted to the Glen Dessary area, and at several localities the Sgurr Beag Slide itself passes around later sheath folds, so demonstrating their post-D_3 age in Baird's chronology. At Kinlochourn [NG 951 063] an antiformal sheath fold of a wavelength of *c.* 500 m is associated with extension lineations plunging at *c.* 80° towards the south (Barr 1983). The Sgurr Beag Slide crops out on the western limb of the complementary synform [NG 945 069], and the core of the sheath fold is occupied by non-migmatitic semipelites identical to the Morar Division rocks which lie west of the Sgurr Beag Slide. The envelope of the sheath fold consists of highly deformed, migmatitic psammites which form the basal member of the local Glenfinnan Division succession. Systematic variations in minor fold vergence confirm that the sheath fold belongs to the set of upright structures which defines the steep belt. The rocks in the core of the sheath fold are interpreted as representing a Morar Division inlier, and so the sheath fold post-dates the Sgurr Beag Slide. A comparable, but synformal, structure almost certainly occurs at the head of Glen Dessary (B.G.S. sheet 62W, Loch Quoich); [NM 90 93], where an isolated elliptical outcrop of Glenfinnan Division lithologies lies west of the Sgurr Beag Slide. Further north, at Sgùman Còinntich [NG 97 29], a 4×1 km outlier of Glenfinnan Division pelites occupies the core of an essentially isoclinal sheath fold within the Morar Division (Barr 1983; subsequent work by R. E. Holdsworth).

The occurrence of both sheath-like, steeply plunging and cylindrical, gently plunging folds, all of which post-date the Sgurr Beag Slide, indicates that deformation within the steep belt was extremely heterogeneous, and so the contrast in structural styles at Glen Dessary is not unusual. Baird's interpretation requires that one of the gently plunging folds east of the Glen Dessary Synform represents its complementary antiform, since the stratigraphic sequence as a whole is right way up. He thus assigns the strain heterogeneity to an earlier tectonic event for which there is no evidence in the Glen Dessary area, and introduces an unnecessary complication into a model based on the demonstrable heterogeneity of upright Caledonian reworking.

DR O. T. TOBISCH writes: The interpretation by Roberts *et al.* (1984) of the Glen Dessary synform as a sheath fold having formed by one continuous deformation is a reasonable one. They suggest, however, that the elongate eyed structures I described from the Glen Cannich area 50 km to the northeast are 'likely' to be "major upright "sheath folds" rather than representing a basin-and-dome interference pattern as interpreted by me (1966).

When analysing the field data and interpreting the genesis of the Cannich structures, I considered the possibility that these structures formed by one continuous deformation in which there had been more 'upward' movement at certain points along the fold axes relative to other points and that this differential movement produced curved fold axes resulting in the observed elongate eyes. A mechanism to produce such structures had already been proposed by Ramsay (1962), which might be considered to be the forerunner of the 'sheath fold' concept developed later (Cobbold & Quinquis 1980). I had to reject this interpretation for the Cannich structures because the bulk of the evidence was more readily interpreted as a fold interference pattern.

This evidence includes the facts that (1) as outlined earlier (Tobisch, 1966), the eyed structures which make up the major Cannich folds are elongate entirely in the direction of F_2; within the hinges of at least three of these F_2 major folds, one finds isoclinal minor folds as well as lineations consisting of both striping and quartzofeldspathic rodding which pre-date the F_2 folds. On a mesoscopic scale, there are a number of localities where one finds isoclinal F_2 minor folds refolding earlier (F_1) isoclinal minor folds, which in some cases show quartzofeldspathic rodding lying parallel to the F_1 axes. It is clear, therefore, that the F_3 folds which form the elongate eyes postdate an earlier period of deformation and metamorphism of some considerable intensity. In addition, these eyed structures occur over a large area, and show a regularity of pattern when the effect of the later deformations (esp. F_3) are removed (see Tobisch 1966, figs 9 and 12). There are also clear hints that just to the west of the area being considered (see Tobisch 1967, fig. 2) the basin-and-dome pattern may continue over an even larger area than originally outlined (Tobisch 1966). To me, the above observations support the interference pattern interpretation much more convincingly than attempting to explain the major fold pattern and other structural/metamorphic features by one continuous process (e.g. sheath fold genesis).

Whereas I do not completely rule out the possibility that the Cannich eyed structures may have formed by one continuous deformation, the evidence is more fully accounted for by interpreting the structural pattern as forming by fold interference of two successive deformations, however closely or widely separated in time.

Drs A. M. Roberts, D. I. Smith, A. L. Harris and R. E. Holdsworth reply: Tobisch disputes our suggestion that the structures mapped by him in the Glen Cannich area are probably 'sheath fold' structures, produced during a single period of upright folding. He does so on the grounds that structures earlier than his D_3/Cannich generation (? our D_3) are recognizable within the hinge zones of his now curvilinear Cannich generation folds. From our own recent observations in Glen Cannich we agree that the Cannich folds are certainly not the earliest structures present; however, we fail to see how this necessitates that the major, upright, curvilinear folds should be the product of fold interference.

In the Loch Quoich–Glen Dessary area (Roberts & Harris 1983; Roberts *et al.* 1984) the upright 'sheath folds' are also not primary structures. They were generated during our D_3 deformation and post-date two earlier sets of originally recumbent structures, at least one of which is also characterized by markedly curvilinear folds (Holdsworth & Roberts 1984). We have recognized numerous interference structures between these early folds and the upright D_3 folds. On a large scale, however, only type 3 'hook' interference patterns are produced (e.g. Roberts & Harris 1983, figs 2 and 4). This interference, as outlined by Roberts *et al.* (1984) was not responsible for the highly curvilinear nature of major D_3 folds such as the Glen Dessary Synform (Roberts *et al.* 1984) and the Spidean Mialach Antiform (Roberts & Harris 1983).

To produce folds with the geometries described by Tobisch (1966), by the process of fold interference, a regionally developed set of NW–SE trending, upright folds of pre-Cannich age must exist. Tobisch (1966) and Tobisch *et al.* (1970) could not detect such a set of folds, and had to "hypothesize" that they might exist. The only pre-Cannich structures recognized by Tobisch *et al.* (1970) were minor structures, described by Tobisch above, of very limited areal extent (Tobisch *et al.* 1970, fig. 3), which are too insignificant to induce marked curvilearity in major, later folds.

We acknowledge that, in the Glen Cannich area, local interference between Cannich and pre-Cannich generation structures certainly exists, but still prefer the interpretation that the major, upright, curvilinear folds were produced during a single period of folding. In accordance with D. Barr (in discussion of Baird above) we believe the curvilinearity may be the result of regional-scale, heterogeneous shortening, with accompanying simple shear, above a gently inclined detachment horizon.

References

Baird, A. W. 1982. The Sgurr Beag Slide within Moine rocks at Loch Eilt, Inverness-shire. *J. geol. Soc. London*, **139**, 647–53.

Barr, D. 1983. *Genesis and structural relations of Moine migmatites*. Unpubl. PhD thesis, Univ. Liverpool.

van Breemen, O., Aftalion, M., Pankhurst, R. J. & Richardson, S. W. 1979. Age of the Glen Dessary Syenite, Inverness-shire: diachronous Palaeozoic metamorphism across the Great Glen. *Scott. J. Geol.* **15**, 49–62.

Brewer, M. S., Brook, M. & Powell, D. 1979. Dating of the tectono-metamorphic history of the southwestern Moine, Scotland. *In*: Harris, A. L., Holland, C. H. & Leake, B. E. (eds) *The Caledonides of the British Isles—Reviewed*. Spec. Publ. geol. Soc. London, **8**, 129–37.

Brook, M., Powell, D. & Brewer, M. S. 1976. Grenville age for rocks in the Moine of north-western Scotland. *Nature. London*, **260**, 515–7.

——, —— & —— 1977. Grenville events in Moine rocks of the Northern Highlands, Scotland. *J. geol. Soc. London*. **133**, 489–96.

Cobbold, P. R. & Quinquis, H. 1980. Development of sheath folds in shear regimes. *J. structural Geol.* **2**, 119–26.

Coward, M. P. & Siddans, A. W. B. 1979. The tectonic evolution of the Welsh Caledonides. *In*: Harris, A. L., Holland, C. H. & Leake, B. E. (eds) *The Caledonides of the British Isles—Reviewed*. Spec. Publ. geol. Soc. London, **8**, 101–8.

Holdsworth, R. E. & Roberts, A. M. 1984. Early curvilinear fold structures and strain in the Moine of the

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Glen Garry region, Inverness-shire. *J. geol. Soc. London*, **141**, 327–38.

Powell, D. 1983. Discussion on a model for the deep structure of the Moine Thrust Zone. *J. geol. Soc. London*, **140**, 519–25.

——, Brook, M. & Baird, A. W. 1983. Structural dating of a Precambrian pegmatite in Moine rocks of northern Scotland and its bearing on the status of the 'Morarian Orogeny'. *J. geol. Soc. London*, **140**, 813–23.

——, Baird, A. W., Charnley, N. C. & Jordan, P. J. 1981. The metamorphic environment of the Sgurr Beag Slide: a major crustal displacement zone in Proterozoic Moine rocks of Scotland. *J. geol. Soc. London*, **138**, 661–73.

Rathbone, P. A., Coward, M. P. & Harris, A. L. 1983. Cover and basement: a contrast in style and fabrics. *In*: Harris, L. D. & Williams, H. (eds) *Tectonics and Geophysics of Mountain Chains*. Mem. geol. Soc. Am. **158**, 213–23.

Ramsay, J. G. 1962. The geometry and mechanics of 'similar' type folds. *J. Geol.* **70**, 309–27.

Roberts, A. M. & Harris, A. L. 1983. The Loch Quoich Line—a limit of early Palaeozoic crustal reworking in the Moine of the Northern Highlands of Scotland. *J. geol. Soc. London*, **140**, 883–92.

——, Smith, D. I. & Harris, A. L. 1984. The structural setting and tectonic significance of the Glen Dessary Syenite, Inverness-shire. *J. geol. Soc. London*, **141**, 1033–42.

Strachan, R. A. 1985. The stratigraphy and structure of the Moine rocks of the Loch Eil Area, West Inverness-shire. *Scott. J. Geol.* **21**, 9–22.

Tobisch, O. T. 1966. A large scale basin and dome pattern resulting from the interference of major folds. *Bull. geol. Soc. Am.* **77**, 393–408.

—— 1967. The influence of early structures on the orientation of late-phase folds in an area of repeated deformation. *J. Geol.* **75**, 554–64.

——, Fleuty, M. J., Merh, S. S., Mukhopadhyay, D. & Ramsay, J. G. 1970. Deformational and metamorphic history of Moinian and Lewisian rocks between Strathconon and Glen Affric. *Scott. J. Geol.* **6**, 243–65.

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LOOSE MATERIAL (MAPS AND SCHEMATIC SECTION).

Map 1) Distribution of lithologies in the area.

Map 2) Structural Map of the Loch Eil Division.

Map 3) Major structures east of the Beinn an Tuim fault.

Map 4) Major structures west of the Beinn an Tuim fault.

Map 5) Major fold axial plane traces in the area.

Map 6) Location of calc-silicate and amphibolite samples.
(and their plagioclase An. content).

Map 7) Microdiorite sheet geometry along the A.830 road.

Map 8) Local topography and sample localities.
(Grid references listed in Appendix 4).

Schematic Cross Section,Skye to Fort William.