Structural and Isotopic studies of Moine rocks: Moine Thrust to the Sgurr Beag Slide.

Thesis submitted for the Degree
of
Doctor of Philosophy

SIMON PETER KELLEY

Department of Geology
Bedford College
University of London

September 1984

CONTENTS

CONTENTS (Continued) Page

CONTENTS (Continued) Page

CONTENTS (Continued)

ACKNOWLEDGEMENTS

My thanks go principally to my supervisor, Dr Derek Powell, whose enthusiasm, help and encouragement in all aspects of the project was greatly appreciated. Thanks also go to Maureen Brook and Dr N.J. Snelling, who supervised my work at the Isotope Unit of the BGS.

Special thanks to Dr Chris Rundle, so often the only person around who could help me when the equipment malfunctioned, and to Dr Bob Pankhurst for discussion of many problems.

I would like to thank my many friends in the Highlands and especially Mrs Cameron and the late Mrs MacLennan for their hospitality.

There are many others who deserve mention: the staff and post-graduates in the Geology Department of Bedford College; staff of the Isotope Unit of the BGS; the technical staff at Bedford College, the Mineralogy Unit of BGS at Grays Inn Road, to whom I am indebted.

My thanks go finally to Mary Burtoft for expertly producing the typescript and not complaining about my spelling.

I gratefully acknowledge the support of this work by the Natural Environment Research Council.

SUMMARY OF THESIS

Structural and Isotopic studies;Moine Thrust to the Sgurr Beag Slide, N.W.Highlands,Scotland.

S Kelley,Bedford College,University of London.

During Caledonian amphibolite grade metamorphism in the Fannich/Ullapool area of N.W.Scotland,high grade migmatitic gneisses were overthrust northwestward along the Sgurr Beag Slide, onto lower grade rocks.A minimum displacement of about 50km is indicated.Features associated with the slide include: the presence of slivers of Lewisian basement along the slide at Achnasheen;the intensification of strain in a 2km zone adjacent to the slide;differences in the early structural and metamorphic histories of the crustal blocks lying above and below the slide;and local inversion of isograds.

The slide zone, and all peak metamorphic fabrics, were reworked by a set of major N-S trending,westward verging asymetrical folds.Associated deformation fabrics intensify towards the Moine Thrust zone,culminating in the development of greenschist facies mylonites.Within 4km of the thrust,folds and fabrics of this generation become sub-parallel to banding and fold axes are rotated towards the movement direction of thrusting (i.e.the W.N.W.).Shear bands are developed in pelitic horizons whilst in psammitic lithologies,strong quartz c-axis girdle fabrics characterise mylonitic rocks.

K-Ar dates on muscovite and biotite pairs from pelites(200-250um grain-size),along an 18km E-W traverse through the area indicate that the whole of the area cooled virtually simultaneously between 424Ma and 421Ma.However uplift possibly as early as 430Ma is indicated close to the thrust.During the same period an excess argon atmosphere was developed in a zone within 4km of the Moine Thrust.

The K-Ar study reveals a relationship between age,reproducibility and grainsize.Fine grain-sizes(125-200um) give reproducible ages with a mean of 423.5Ma(muscovite) and 421.5Ma(biotite).Whereas coarser grain-sizes(250-500um)give unreproducible ages at around 433Ma,significantly older than the finer.Rb-Sr mica pairs of all grain-sizes give ages compatible with the K-Ar ages from the finest grain-sizes.It is suggested that the micas suffered a partial resetting event late in their cooling history.

The major conclusion of the study is that the development of the Moine Thrust zone significantly post-dates peak metamorphism and movement upon the Sgurr Beag Slide.The slide zone is thought to have moved during peak metamorphic conditions around 460Ma whereas the Moine Thrust developed and moved between 430Ma and 420Ma.

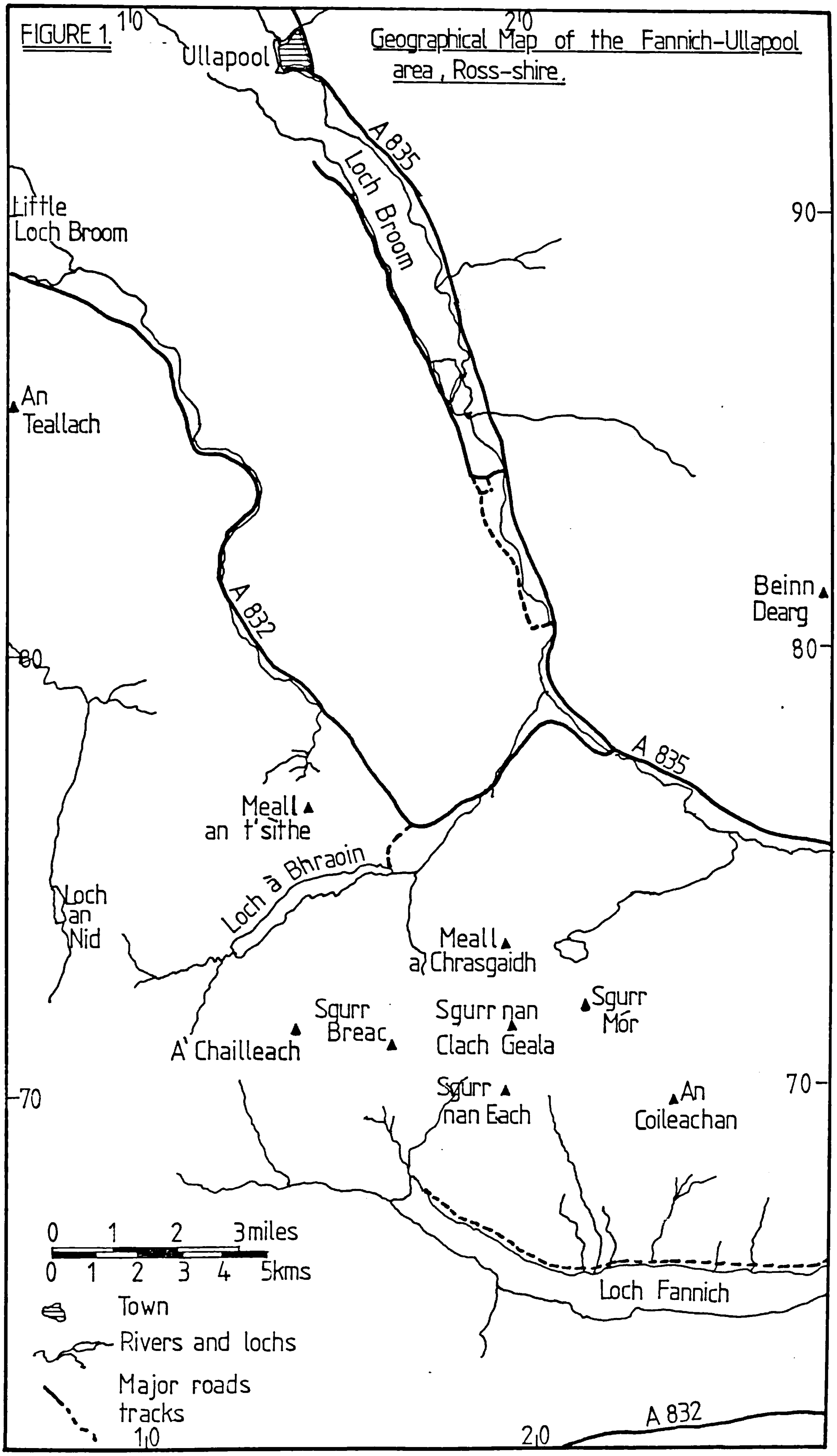


FIGURE 1.
Geographical Map of the Fannich-Ullapool area, Ross-shire.
10
20
Ullapool
A 835
Loch Broom
90
Little Loch Broom
An Teallach
A 832
Beinn Dearg
80
80
A 835
Meall an t'sithe
Loch ā Bhraoin
Loch an Nid
Meall a'Chrasgaidh
Sgurr Mór
Sgurr Breac
Sgurr nan Clach Geala
A' Chailleach
70
70
Sgurr nan Each
An Coileachan
0 1 2 3 miles
0 1 2 3 4 5 kms
Loch Fannich
Town
Rivers and lochs
Major roads
tracks
10
20
A 832

INTRODUCTION

The rocks considered in this survey outcrop in western Ross-shire, Scotland and lie between the A835 to Ullapool, in the north, and the A832 to the Kyle of Lochalsh, in the south (fig.1). Fieldwork was conducted over the seasons 1979-1981 and comprised: One traverse within the inner mobile zone of the orogen, along the northern shore of Loch Fannich (7km), and two main traverses from Braemore to the Moine Thrust at Loch an Nid (13km) and from Braemore along the shores of Loch Broom across the Moine Thrust zone to Rhiroy (14km)(fig.1).

Within the area of study a vertical relief of 2500ft (760m) ranging from a coastal area along the shores of Loch Broom to the mountains of the Fannich forest, provides good three-dimensional control. Degree of exposure varies from 100% at the coast to around 10% in the Fannich area.

Geological Setting and Objectives

The Caledonides of the Scottish Highlands have been interpreted as lying on the northern side of the Iapetus ocean and in this position they became involved in early orogenic activity during collision of a micro continent (Watson & Dunning, 1979) and final orogenic movements during closure.

The constituent rocks include: sediments deformed for the first time during the Grampian orogeny - the Dalradian; a possible cover/ basement sequence of uncertain age in the Central Highlands comprising parts of the Central Highland granulites and the Moine schists of the Northern Highlands to the north of the Great Glen Fault, which probably underwent deformation and metamorphism as early as 1004Ma (Brook et al., 1976).

The Fannich-Ullapool area forms part of the N. Highlands sector and encompasses part of the Caledonian Front - the Moine Thrust zone-and part of the internal metamorphic complex.

No fossiliferous rocks have been found within the Moine schists and thus interpretations of the lithostratigraphy rely on sedimentary 'way up' criteria. The rocks are divided into three assemblages: the Morar, Glenfinnan and Loch Eil Divisions (Johnstone, 1978). The structurally lowest, Morar Division, is overlain by the Glenfinnan Division, and this is in turn overlain by the Loch Eil in a general west to east progression (fig.3).

The boundary between the Morar and Glenfinnan Divisions is now generally recognised to be tectonic - the Sgurr Beag Slide. However, there is some controversy over the boundary between the Glenfinnan and Loch Eil Divisions. The junction is generally coincident with the Quoich line, where Roberts & Harris (1983) propose that the line marks an eastern limit of Caledonian crustal reworking, and that the boundary between the two divisions is conformable. However, Strachan (1982) has suggested that the junction constitutes a tectonic slide separating rocks of the Glenfinnan Division with a pre-Caledonian and Caledonian history, from those of the Loch Eil Division wherein only Caledonian tectonometamorphic events are recorded.

Only one of the divisions, the Morar, is cut by the Moine Thrust zone at the present level of erosion and hence correlation of deformation from the thrust belt to the other divisions is extremely difficult.

In the Fannich area (fig.4) the Glenfinnan Division is preserved in the core of a late syncline, and outcrops along the ridge between Meall an t'sithe and Creag Rainich, where it approaches the Moine Thrust to within 2km. This is the closest approach of Glenfinnan Division rocks to the Moine Thrust zone yet discovered and has afforded the opportunity of investigating the relationships between development of the thrust zone, the structural histories of the Morar and Glenfinnan Divisions and development of the Sgurr Beag Slide zone.

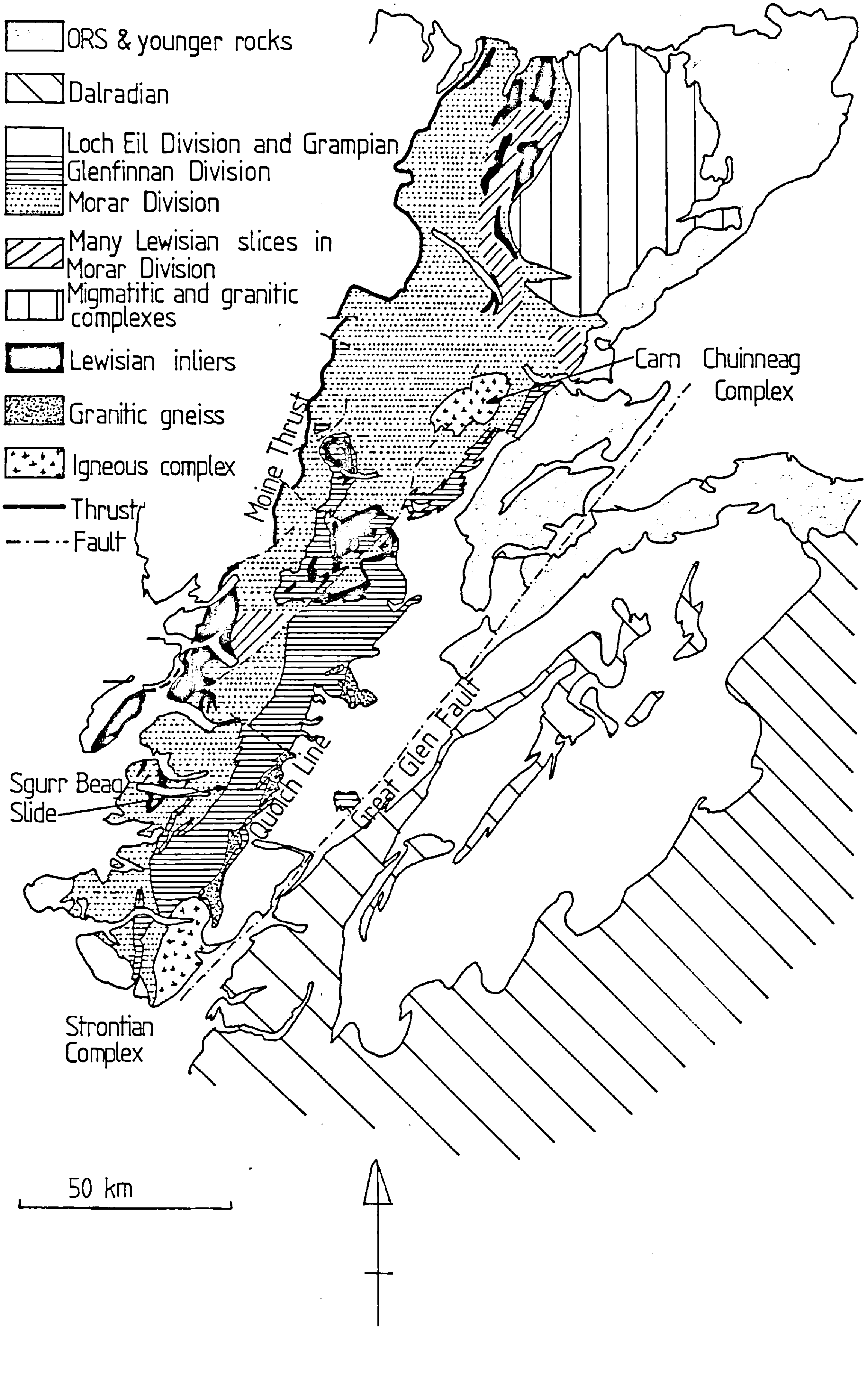
Psammites and pelites of the Morar Division in the field area are similar to those described from elsewhere in the Moine schist outcrop although it is difficult to correlate local lithostratigraphies from area to area. Similarly, the Glenfinnan Division metasediments compare with those described elsewhere with the exception that they do not contain calc-silicates. In the field area Glenfinnan Division rocks are interfolded with highly reworked gneisses which are correlated with the Archaean to Lower Proterozoic Lewisian rocks of the foreland.

The Moine and Lewisian rocks of the Fannich-Ullapool area underwent amphibolite facies metamorphism during the Caledonian orogeny and although the precise age of peak metamorphism is not known, it must post-date 550 ± 10Ma - the age given for intrusion of the Carn Chuinneag granite (Pidgeon & Johnson, 1974) (fig.9).

Elsewhere in the N.Highlands peak Caledonian metamorphic activity has been dated at around 450-460Ma (Van Breemen et al.,1974; Brewer et al. 1979) and thus K-Ar ages for muscovite and biotite of 440 to 420 Ma obtained in this study appear to represent cooling ages related to uplift after peak Caledonian metamorphism.

The main objectives of this study have been to investigate the structural, metamorphic and temporal relationships between development of the Moine Thrust zone and the structural-metamorphic history of the internal metamorphic complex of the N. Highlands.

FIGURE 3. Map illustrating the Geology of the Highlands of Scotland.



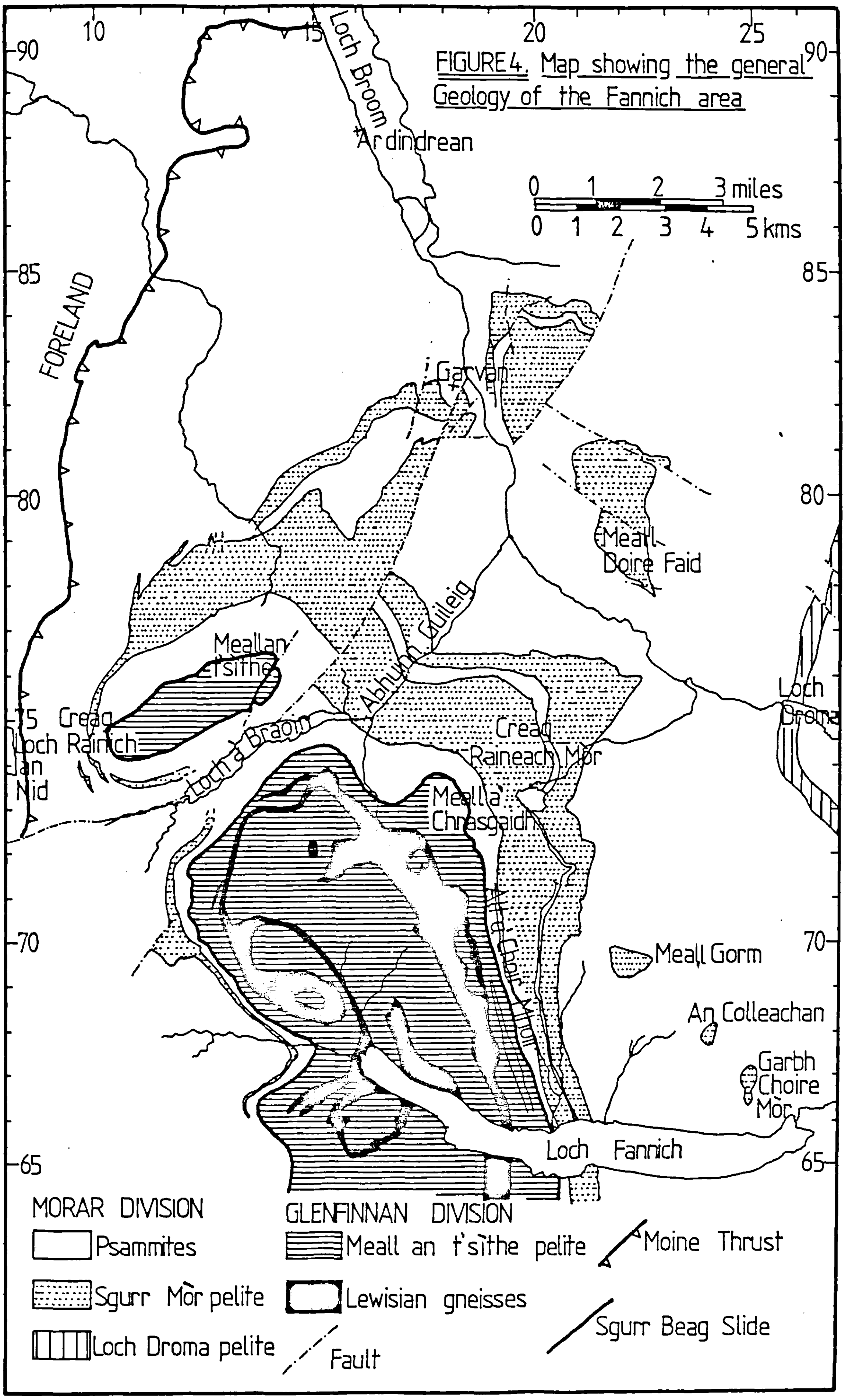
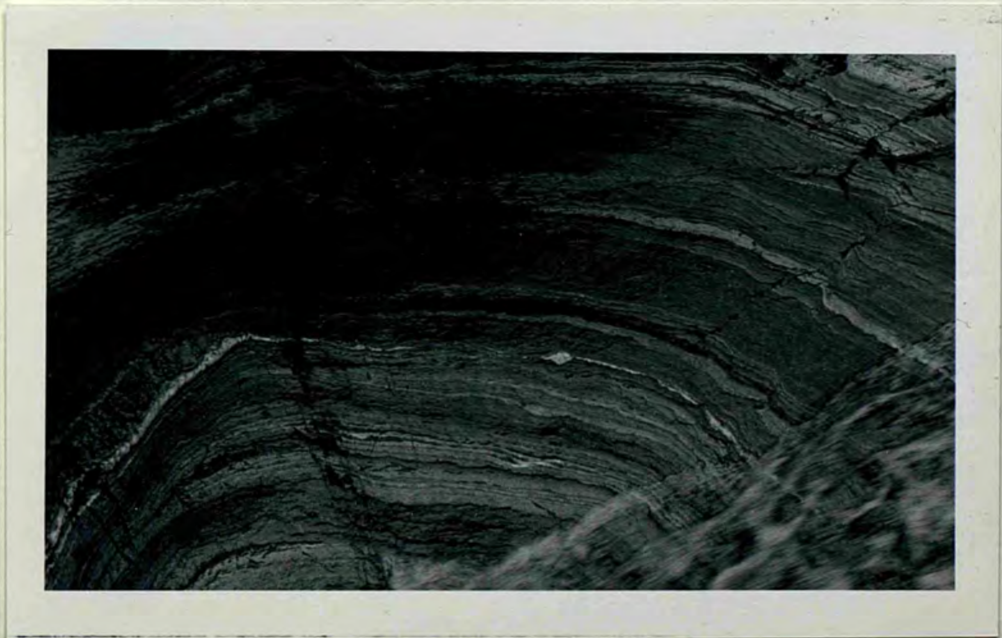


FIGURE 4. Map showing the general Geology of the Fannich area
Loch Broom
Ardindrean
0 1 2 3 miles
0 1 2 3 4 5 kms
FORELAND
Garvan
Meall Doire Faid
Meallan t'sithe
Abhuinn Cuileig
Loch Droma
Creag Loch Rainich
Loch an Nid
Loch a' Braoin
Creag Raineach Mor
Meall a' Chrasgaidh
Allt a' Choire Mhoir
Meall Gorm
An Colleachan
Garbh Choire Mor
Loch Fannich
10
15
20
25
90
85
80
75
70
65
MORAR DIVISION
Psammites
Sgurr Mor pelite
Loch Droma pelite
GLENFINNAN DIVISION
Meall an t'sithe pelite
Lewisian gneisses
Fault
Moine Thrust
Sgurr Beag Slide

FIGURE ~~4~~ 5. a) An exposure of Inverbroom Psammites.
(NH 223775)

b) An exposure of the Sgurr Mor Pelite.
(NH 207 673)

50 cms



c) An exposure of the Meall a'Chrasgaidh Psammite.

(NH 198675)

d) An exposure of the Meall an t'sithe Pelite.

(NH 184697)

10 cms

e) An exposure of the Lewisian Gneisses.

(NH165723)

CHAPTER I. STRATIGRAPHY

St. 1. Previous Research

The results of earlier work concerning the stratigraphy of the study and adjacent areas are summarised in Table 1.

St. 1a. Geological Survey

The earliest stratigraphical succession given for the Fannich-Ullapool area was presented by Gunn, Horne, Peach & Pocock, in the Memoir of the Geological Survey for Sheet 92 (Fannich mountains) published in 1913.

Gunn mapped the area between Loch Broom and the Fannich mountains and reported a 'stratigraphical' succession in the Moine series (Summary of Progress, 1898) in which he recorded the lowest member of the succession as lying against the Moine Thrust, and the highest member as being overlain by coarse acid gneisses which he thought, on lithological grounds, to be Lewisian. On the basis of the distribution of the succession around the north of the area, Gunn inferred that there was a large synclinal fold in Fannich. This was confirmed by Horne in 1898 when he mapped south of the Fannich watershed and found a highly inclined limb of the same fold.

The Moine Series from the Fannich mountains to the Moine Thrust was divided into five groups:

1. Coarse acid gneiss of Lewisian type.
2. Meall an t'sithe rock (muscovite-biotite gneiss)
3. Meall a Chrasgaidh siliceous schists
4. Sgurr Mor biotite schist
5. No name given (flaggy granulitic schists and gneisses).

Gunn & Horne suggested that the apparent anomaly of the high strati graphic position of the Lewisian gneisses could be resolved by:

a) A thrust relationship, at the boundary of the Meall an t'sithe rock and Meall a Chrasgaidh rock - "the first hypothesis". (This was reinforced by the observation of high metamorphic grade present in the Meall an t'sithe and Lewisian relative to the rest of Fannich.)

b) A massive fan-shaped anticlinorium with Lewisian forming the core and therefore basement to the Moine succession - "the second hypothesis".

The first hypothesis seemed to be favoured by Gunn & Horne in 1898 but in the memoir of 1913, the second was employed to construct cross sections of the Fannich range. As described in section SBS.1a, this second hypothesis required the inversion of the Moine Series in the Fannich area.

St. 1b. <u>Rutledge and MacIntyre</u>

Rutledge (1952) and MacIntyre (1954) produced a structural interpretation similar to that of Gunn (1898) (see section SBS.1a) which removed the need for the Fannich series to be totally inverted during deformation (Table 1). Their interpretation did not directly affect the stratigraphy, except that it postulated a tectonic break within the succession to explain the position of the Lewisian gneisses.

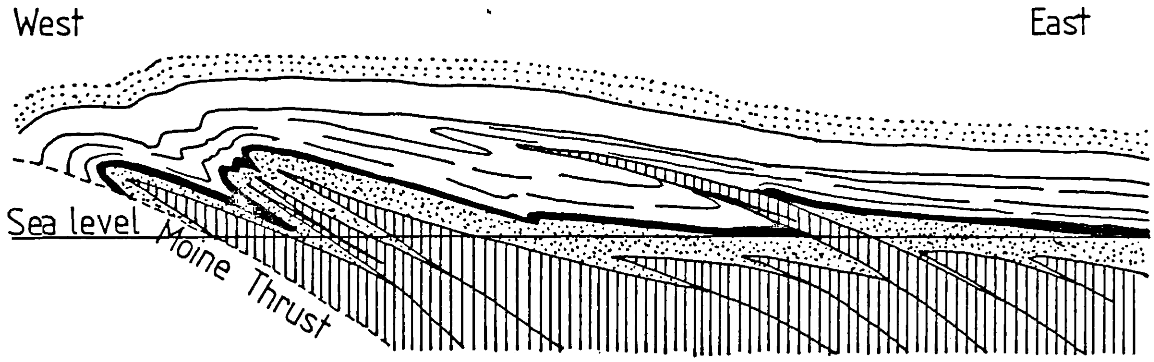
St. 1c. <u>Sutton and Watson</u>

Sutton and Watson mapped the Fannich area in the years 1949-52 and presented a paper on the structure and stratigraphy of the Fannich forest in 1954 (Sutton & Watson, 1954) (Table 1). Mapping the area around Scardroy, they had earlier erected a lithostratigraphic succession using cross-bedding structures as 'way-up' evidence (Sutton & Watson, 1953). The Scardroy succession was then correlated with that of the Fannich

FIGURE 6. Correlation of the areas of Fannich and Scardroy after Sutton and Watson (1954)

	FANNICH	SCARDROY
9. Youngest	Pelitic gneiss with siliceous bands and abundant altered minor basic intrusions	Pelitic gneiss with siliceous bands and abundant minor basic intrusions
8. Fannich gneiss	Dominantely siliceous with calcarious bands and abundant altered minor basic intrusions	Semi-pelitic and siliceous with altered minor basic intrusions
7. Meall an t'sithe pelite	Pelitic gneiss with siliceous bands and abundant altered minor basic intrusions	Pelitic gneiss with abundant altered minor basic intrusions
6. Meall a Chrasgaidh psammite	Uniform siliceous granulite, without current-bedding	Semi pelitic, with hornblendic, quartzitic and calcarious bands
5. Scardroy striped hornblendic group	Not represented	Hornblendic and semi-pelitic, with calcarious bands and altered minor basic intrusions
4. Sgurr Mor pelitic group	Pelitic and semi-pelitic bands, calcarious bands	Not represented
3. Inverbroom semi-pelite	Semi-pelitic with calc-bands near the top; often strongly cross-bedded	semi-pelitic, with hornblendic and calc-bands near top; often current-bedded
2. Loch Droma pelitic group	Pelitic with semi-pelitic bands	
1. Achanalt semi-pelite	Semi-pelitic, often strongly current-bedded	

FIGURE 7.



Lewisian all other lithologies are Moinian

The section is a composite one showing wedges with autochthonous cover proposed by Sutton and Watson for Glenelg in the west and a displaced wedge, pushed into the Moine cover, representing Scardroy. (Sutton and Watson, 1962) mountains (fig.6). The work of Sutton & Watson demonstrated that the acid gneisses lay in the core of a syncline and were apparently younger than the rest of the succession of Fannich. They proposed that the gneisses which had been called Lewisian by the Geological Survey, were an integral part of the Moine succession and they renamed this group the "Fannich gneisses". Sutton and Watson had previously interpreted gneisses, which had been thought to be Lewisian, as Moinian gneisses (Sutton & Watson,1953). In the Scardroy area they interpreted the gneisses which the Survey called Lewisian as altered Moinian calc-silicates or 'Durcha type' rocks. They considered that in Fannich the metamorphic state, igneous intrusions and lithologies could be matched with other areas of the Moine and therefore did not distinguish the Fannich gneisses from the rest of the succession. The gneisses were described as being interbanded with the Meall an t'sithe pelite, whereas the Geological Survey postulated isoclinal folds to explain the outcrop pattern.

Sutton and Watson (1954) considered and rejected the idea that the acid gneisses were thrust into their present position because, like the Geological Survey, they could find no field evidence for thrusting. They therefore concluded that the gneisses were an integral part of the Moine succession which was younger than the lower part of the Meall an t'sithe Pelite.

Subsequently however, Sutton and Watson (1962) (fig. 7) accepted a Lewisian origin for the Fannich gneiss as a result of mapping in Glenelg and Morar (Ramsay, 1957; Sutton & Watson, 1959). Under the new interpretation the stratigraphic succession proposed for Fannich remained the same except that they regarded the Fannich Lewisian as a wedge driven between layers of the Moine succession, thus affirming the presence of a tectonic break in the succession (Table 1).

St.1.d <u>Winchester</u>

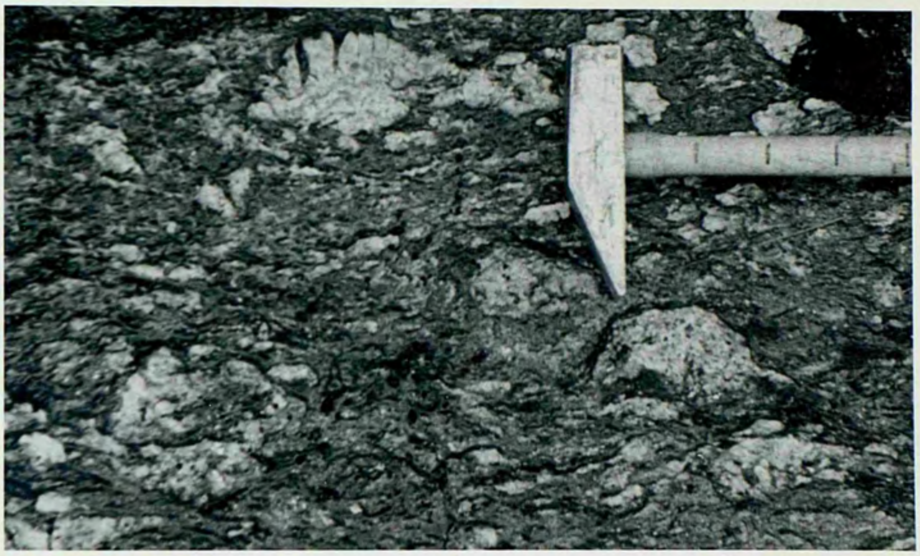
The most recent study of the area was by J.A. Winchester (1970). He mapped essentially the same area as Sutton and Watson (1954) producing refinements to their stratigraphy and extending the known outcrop of the Fannich gneisses, south of Loch Fannich (Winchester, 1973). He concluded that the gneisses were Lewisian on geochemical grounds (Winchester, 1970) (Table 1).

One major difference was that in the Meall an t'sithe Pelite, adjacent to the Lewisian gneisses, SE of Sgurr a Chadha Dheirg, Winchester (1970) described a lithology containing spindle-shaped pods of quartzofeldspathic material that resembled highly deformed pebbles. He then postulated that the lithology might be a basal conglomerate resting on the Lewisian gneisses. He described it as lying in an infold of the Meall an t'sithe Pelite within the Lewisian (Grid ref. NH 187686). (This locality has not been relocated by the present author.) Rocks fitting the description given by Winchester occur in several places within the Meall an t'sithe Pelite, notably in the stream section of Alt a Choire Moir (fig.2) which runs along the tectonic junction representing the Sgurr Beag Slide. In this position an augen gneiss represents a highly deformed equivalent of the migmatitic Meall an t'sithe Pelite (fig.8). Shearing along the Lewisian/Meall an t'sithe Pelite boundary may have resulted in formation of a pseudo-conglomerate too.

Winchester (1970) correlated the two migmatitic pelites lying above and below the Lewisian gneisses and considered them to be the same lithology repeated by isoclinal folding. He further considered, but eventually rejected, the presence of the Sgurr Beag Slide which Tanner <u>et al</u>. (1970) extended through the Fannich area. Tanner <u>et al</u>. (<u>op.cit</u>.) placed the slide at the boundary of the Meall an t'sithe Pelite and Meall a Chrasgaidh

FIGURE 8. Augen gneiss adjacent to the Sgurr Beag Slide formed from migmatites of the Meall an'sithe pelte (Glenfinnan division)

(NH 198 675)



Psammite and explained its presence as a downfold of the Sgurr Beag Slide which rooted to the east of Scardroy. Winchester rejected this idea in 1970, explaining the outcrop pattern in terms of isoclinal folding (see section SBS.1a).

However, in 1973 he subsequently accepted the conclusions of Tanner et al. (1970) seemingly rejecting the major isoclinal folding previously proposed.

Introduction to Lithologies(Table 1 in folder)

During this study evidence was found which supports the conclusion that a tectonic break, probably an extension of the Sgurr Beag Slide, outcrops within the area; therefore it is proposed to consider the stratigraphy in two parts:

1) The Lower succession consisting of the Inverbroom Psammites, Sgurr Mor Pelite, and Meall a Chrasgaidh Psammite (which are correlated with the Morar division. Johnstone et al. (1969), or Glenelg nappe (Tanner et al. (1970)). (fig.5)

2) The Upper succession, consisting of the Meall an t'sithe Pelite (Glenfinnan Division, Johnstone et al. (1969)) and Lewisian gneisses (both of which form part of the Ross-shire nappe, Tanner et al. (1970)). (fig.5).

St. 2 The Lower Succession

Both Inverbroom Psammite and Sgurr Mor Pelite contain well preserved sedimentary structures, indicating the original superposition of the rocks (Table 1). The Meall a Chrasgaidh Psammite is highly deformed and original sedimentary features have been obliterated.

All sedimentary structures seen indicate that this succession is 'right way up' in Fannich and the area to the west, in agreement with the findings of Sutton and Watson (1954) (Table 1). Towards the top of the Sgurr Mor Pelite, close to the boundary with the Meall a Chrasgaidh Psammite, sedimentary structures become tectonically flattened to such an extent that 'way up' criteria are less reliable. In this area the intensity of folding and flattening is related to the presence of the Sgurr Beag Slide.

The metamorphic grade of the rocks comprising the lower succession, from mineralogical evidence in psammitic, pelitic and calc-silicate rocks is almandine amphibolite facies rising toward the Sgurr Beag Slide, and Glenfinnan Division rocks (section SBS 2a,b).

St.2.a <u>The Inverbroom Psammite</u> (fig.5)

This group has the greatest extent of all the lithostratigraphic units. In the far east of the area mapped, the base is not seen but has been described further east (fig.9) by the Geological Survey (1913) and Sutton and Watson (1954). According to these authors it passes by transition into the Loch Droma Pelite. At the eastern end of Loch Fannich (fig.9) the Inverbroom Psammite has a thickness of approximately 2700m.

In the west, the Inverbroom Psammite is progressively cut out against the Moine Thrust, and there is considerable variation in the thickness of metasediments between the thrust and the Sgurr Mor Pelite, which lies above the Inverbroom Psammite. This was noted by Gunn (1913) who stated:

> "It is evident in the area north of Loch an Nid there is no exact parallelism between the line of the Moine Thrust and the strike of the foliation in the schists for as we proceed toward Loch Broom, lower and lower parts of the division successively crop out so that whatever may be the age of these schists, they existed as such prior to the formation of the Moine Thrust."

The present survey demonstrates that above Loch an Nid there lies only 400m of Inverbroom Psammites between the Moine Thrust and the base of the

Sgurr Mor pelite whereas further north, by the shores of Loch Broom, there is in excess of 3000m. It can be clearly demonstrated on the western shore of Loch Broom that the Moine Thrust is oblique to the major litholo-- gical layering within the Inverbroom Psammite. Lithostratigraphic units can be traced successively into the mylonites of the thrust zone (fig.2). (fig 2 is in the folder) Of particular note is a gritty horizon outcropping at Ardindrean (fig.2) (NH155884) which contains graded bedding and small pebbles up to 1cm diameter exhibiting only low relative deformation. This becomes a highly deformed rock with a strong extension lineation, defined by elongation of pebbles and grains of quartz and feldspar, in the mylonites above the Moine Thrust at Creag Cnoeurach (NH120875) only 3km to the west of Ardindrean.

The Inverbroom Psammite incorporates various sedimentary sub-types which occur to a greater or lesser extent in different areas:

a) The majority of the Inverbroom Psammites are interbanded psammites, 10-15 cms thick. These lithologies contain common calc-psammite bands (lenses and pods similar to calc-silicates but with mineralogy consisting of calcite, quartz, plagioclase feldspar and epidote) and also rarer heavy mineral bands. They exhibit well preserved sedimentary structures such as cross-bedding, slump structures, water escape structures and small channel fills. Sedimentary structures are common in some areas but may be totally absent in others.

b) Massive psammite bands are also present, predominantly in the west of the area, along the shores of Loch Broom. They reach up to 2m in thickness, interbedded with less thick psammites. These psammites were not observed to contain calc-psammite or heavy mineral bands and are internally structureless.

c) A series of thick pelite bands is found in the area around Loch a Braoin, within the Inverbroom psammites. These measure up to 1m. in thickness and are interbedded with thinner psammitic bands. They occur stratigraphically towards the top of the Inverbroom psammites. In this area the Sgurr Mor Pelite is wedging out, and it is thought that these pelites represent part of the transition zone at the edge of the Sgurr Mor Pelite.

d) Rarely, in the area to the west of Fannich, in association with the massive psammites, distinctive coarse, gritty horizons occur showing well preserved graded bedding. Only 4-5 such horizons were discovered but they are important as indicators of the non-parallelism of the Moine Thrust and lithological banding. The grains of feldspars and quartz reach up to 1cm in diameter, but they are generally less than 0.5cm in diameter.

The mineral fabrics of the Inverbroom Psammites are strongly affected by the state of strain of the rock and the metamorphic grade at which the last major deformation took place. The psammites, away from areas of high strain, contain approximately equidimensional quartz, oligoclase-andesine plagioclase and subordinate microcline 250-500μ in diameter. The psammites contain around 10% micas, 200-1000μ in length, the majority generally muscovite and trace amounts of garnet, epidote, sphene and opaques.

The semi-pelites and pelites have a similar mineralogy but contain larger proportions of mica and garnet.

St.2.b <u>The Sgurr Mor Pelite</u> (fig.5)

The Inverbroom Psammites grade upwards into the Sgurr Mor Pelite through a sedimentary transition. In the Fannich area the Sgurr Mor Pelite occurs in a vertical belt, running north from Loch Fannich, where it reaches a maximum thickness of approximately 950m. It turns towards the west at Creag Rainich Mor (fig.4) around the axis of the F3 Fannich syncline. The outcrop becomes more complicated further west due to the lower angle of dip and variable topography. It stretches from the summit of Creag Rainich in the west (where it is thinnest, being only 20m thick) to the Broom valley at Garvan in the north-east (NH187820). Several outliers occur at Meall Doire Faid (NH220790), Meall Gorm (NH222695), An Collechan (NH241680) and Garbh Choire Mor (NH249665) (fig.4).

The same pelitic unit is thought to exist south of Loch Fannich (Winchester, 1970; Sutton & Watson, 1954) as a small band thinning rapidly southwards (fig.4).

The lithologies found within the Sgurr Mor Pelite are:

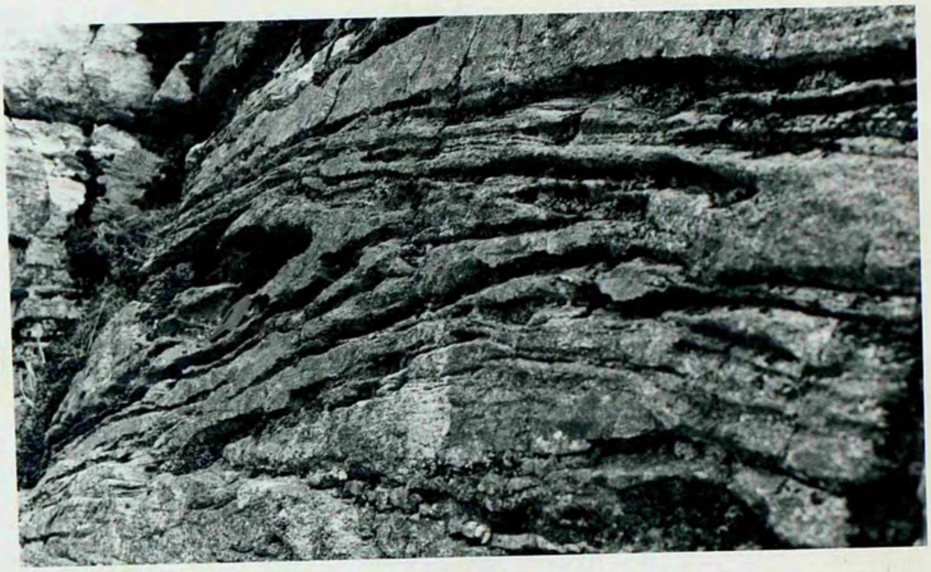
1) Massive pelites with minor interbedded semi-pelites. These generally have little internal structure but do contain calc-silicate bands. Any delicate sedimentary structures originally present, have been destroyed during deformation.

2) Banded semi-pelites and pelites commonly showing sedimentary slump structures, cross-bedding and graded bedding. The beds of this type range from 5cm to 30cm thick and also contain abundant calc-silicates.

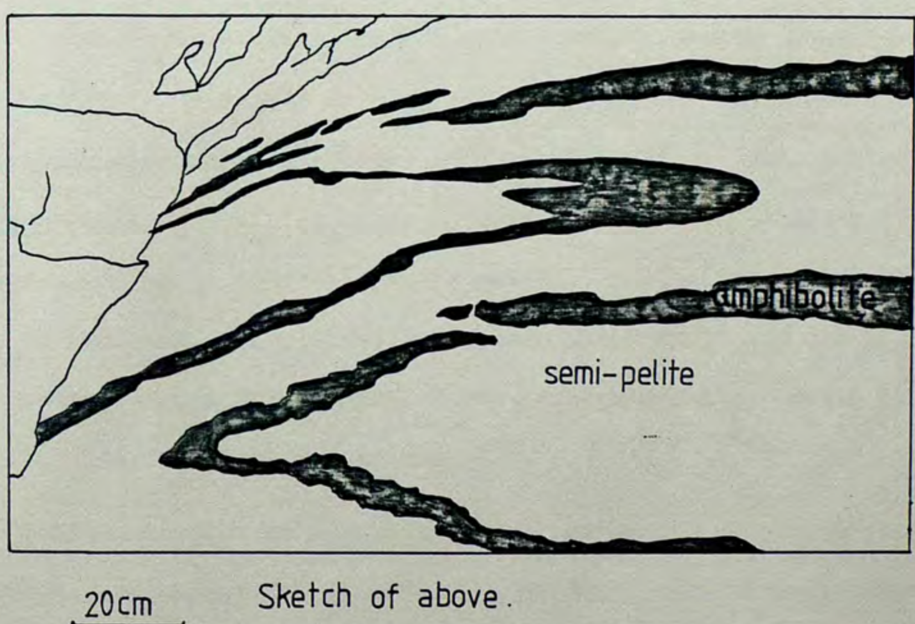
3) Several bands of psammitic rock also occur within the Pelite, by far the most prominent of which is up to 50m thick and divides the pelites of the steep belt north of Loch Fannich (fig.4) and is continuous for some 10km. Several other thinner psammitic bands are interbedded within the

FIGURE 10 .A garnetiferous amphibolite within the Sgurr Mor pelite.It is folded around an F3 fold and can be detected on the photograph by the knobbly weathering.The amphibolite ranges from 1 to 20cm wide.

(NH 144778)



See folder for fig 9



Sketch of above.

pelites. The psammites resemble the Inverbroom Psammites in that they contain sedimentary structures and calc-psammite bands.

The deformation state of the Sgurr Mor Pelite is such that the sedimentary features are distorted and sometimes difficult to interpret. However, in some localities unequivocal 'way up' criteria can be observed. They indicate younging consistently away from the Inverbroom Psammite contact.

The psammitic lithologies contain only calc-psammites but semi-pelites and pelite lithologies contain abundant garnetiferous calc-silicates, which occur as pods and bands up to 10cm thick. These have a varied mineralogy which is sensitive to changes in metamorphic grade. Their petrology and petrogenesis in relation to metamorphism will be described in a subsequent chapter (section SBS.2.b).

Rare strata-bound amphibolites occur towards the top of the Sgurr Mor Pelite and can be seen in the stream section Alt a Choire Mor. They have also been observed north of the summit of Meall an t'sithe (NH148771). Even in areas of relatively low strain, they are parallel to lithological banding and they contain intercalated siliceous lenses which appear to indicate a sedimentary origin perhaps as Ash fall deposits (fig.10). Their chemistry is discussed by Winchester (1976), who concluded that they show calc-alkaline affinities. The amphibolitic horizons range from 5-50cm thick, containing large garnets up to 2cm in diameter. Their mineralogy consists of equidimensional quartz, plagioclase, large biotite laths up to 4mm long, chloritized hornblende laths and trace minerals such as sphene, zircon and opaques. The garnets contain rotated inclusion trails of quartz, feldspar and sphene.

The pelites consist of equidimensional, anhedral quartz and plagioclase (oligoclase-andesine with micas and garnet, trace sphene, epidote, zoisite and opaque minerals). Biotite is usually predominant over muscovite. All original sedimentary microstructures were destroyed in the pelites, now only metamorphic fabrics which are highly susceptible to deformation relative to more psammitic lithologies are seen.

St.2.c The Meall a Chrasgaidh Psammite (fig.5)

The Sgurr Mor Pelite grades into the Meall a Chrasgaidh Psammite over a 5m sedimentary transition. The psammite outcrops on the north shore of Loch Fannich, structurally above the Sgurr Mor Pelite, occupying a belt 100-300m wide between this and the Meall an t'sithe Pelite. It outcrops from the shores of Loch Fannich northward to Meall a Chrasgaidh, then to Loch a Broain and west to the summit of Creag Rainich (NH607752), where it is at its thinnest.

The psammite is highly strained and any sedimentary structures which may have been present have been obliterated. It exhibits almost total parallelism of lithological banding (Section SBS.1.c) throughout its outcrop. The lithologies consist of bands of psammite, 1cm to 10cm thick, interbanded with occasional semi-pelites. The bands have a platey appearance and weathered sections sometimes split parallel to the banding. Rare calc-psammite horizons have been found.

In the stream section west of Alt a Choire Mor (NH198675) thin, dark horizons, approximately 2cm thick, contain hornblende, sphene and epidote but no micas. Such bands were found by Sutton and Watson (1954) in the psammites of Scardroy, which they correlated with the Meall a Chrasgaidh Psammite.

These rocks may represent the relics of sheared Lewisian pods such as those reported to exist along the length of the Sgurr Beag Slide (Tanner et al., 1970; Rathbone & Harris, 1979). They will be discussed in greater detail in a later chapter (Section SBS.1.b).

The mineralogy of the Meall a Chrasgaidh psammites is very similar to that of the Inverbroom psammites. Quartz and plagioclase (andesine) are found as slightly elongated grains. Both muscovite and biotite are present, and define a strong planar fabric. Garnet is very rare, found occasionally in the hornblendic bands which will be discussed later (SBS.1.b).

St.3 The Upper Succession

The Meall an t'sithe Pelite and Lewisian gneisses display coarser grained textures and apparently higher metamorphic grade than the Lower succession. Metamorphic grade is difficult to estimate due to the lack of calc silicate rocks and only rare, relict index minerals in pelites (Section SBS.3.b).

The deformation state is also high and no sedimentary structures were seen in the migmatites. The polyphase deformation state of the Lewisian gneisses is such that it is difficult to determine whether they have sedimentary or igneous origin. The migmatitic pelites lying above the Lewisian gneisses are lithologically very similar to those underlying it and thus they will be considered together as the same stratigraphic unit.

St.3.a The Meall an t'sithe Pelite (fig.5)

This pelite occupies the larger part of the core to the Fannich synform, resting everywhere upon the Meall a Chrasgaidh Psammite across a sharp contact.

An outlier of Meall an t'sithe Pelite occurs on the summit of Meall an t'sithe (NH141764) (fig.5), where it is flat lying. On the summit of Meall Dubh (NH103748) at the western end of the ridge, it lies only 2km from the Moine Thrust (i.e. 400m above the thrust) and this locality represents the closest approach of the Sgurr Beag Slide to the Moine Thrust in the Northern Highlands. (The boundary between the Meall an t'sithe Pelite and Meall a Chrasgaidh Psammite everywhere in the area is considered to be an extension to the Sgurr Beag Slide.)

Only two lithologies have been distinguished in the Meall an t'sithe Pelite:

a) The main pelitic lithology is a lit par lit migmatite, the lits constituting 20-30% of the whole rock. The migmatitic fabric is parallel to lithological (presumably sedimentary) banding, but no sedimentary features have been seen. Nor have any calc-silicate rocks or heavy mineral bands been noted.

In places, against the tectonic boundary with the Meall a Chrasgaidh psammites the migmatite becomes a highly strained and sheared augen gneiss (see also sections SBS.1.c and SBS.2.a).

b) Psammitic migmatites occur at several horizons within the Meall an t'sithe Pelite; these were named the Sgurr nan Each Psammites by Winchester (1970). The psammites form a series of discontinuous lenses between 0 and 100m above the base of the pelite. Mapping west of Meall an t'sithe has extended the outcrop of psammites recorded by the Geological Survey and other workers. The psammites are coarse migmatites, although the amount of pelitic material present is generally less than 20% they contain the characteristic claret-coloured garnets. No calc-silicates or heavy mineral bands were found within this lithology.

The lower part of the Meall an t'sithe Pelite is between 150 and 250m thick; the upper part in excess of 3000m, because the top is above the present erosion level.

Amphibolite pods and dykes up to 2m wide occur in the pelites and were observed in one exposure cross-cutting the migmatite fabric. They are similar to those found within the Glenfinnan Division rocks throughout their known outcrop and the examples in Fannich exhibit a tholeiitic chemistry in contrast to the sedimentary amphibolites of the Sgurr Mor Pelite (Winchester, 1976).

The mineralogy of the pelites is quartz, andesine-oligoclase, feldspar and muscovite which dominate the lits, biotite and garnet, together with muscovite dominate the pelitic matrix with subordinate sphene, zoisite, tourmaline and opaque minerals.

The amphibolites contain quartz, oligoclase-feldspar, hornblende, garnet, biotite, sphene and opaque minerals.

The pelitic matrix contains numerous claret-coloured garnets which are relatively inclusion-free compared with those of the Morar Division; they are generally sub-hedral to anhedral 0.5-2mm in diameter. The Geological Survey reported that during bad storms, the streams had become loaded with garnet sands. It was noted during this study that many shallow depressions in the outcrops contained a layer of well-rounded garnets, sorted from the lighter minerals.

St.3.b The Lewisian gneisses (fig.5)

The Lewisian gneisses outcrop north of Loch Fannich in a band 100-150m wide, folded by the Fannich synform and stretching from the western end of Loch Fannich to Loch a Broain (fig.5). South of Loch Fannich, Lewisian gneisses occur in two small outcrops (Winchester 1970, 1973); they are continuous with the Lewisian north of the Loch (fig.5).

The boundary between the Lewisian gneisses and the Meall an t'sithe Pelite is sharp wherever seen, but its outcrop is very sinuous. Sutton and Watson (1954) considered this to be due to interbanding of the two lithologies. In the Memoir of 1913, the Geological Survey noted massive isoclinal interfolding of the two lithologies and interpreted the boundary in terms of a single phase of isoclinal folding. Winchester (1973) re-interpreted the outcrop of the boundary between the Lewisian gneisses and migmatites inferring two phases of isoclinal folding. The nature of the boundary is discussed in more detail in Section SBS.3.a.

The gneisses constituting the Lewisian outcrop are mainly siliceous acid gneisses with hornblendic bands. Other lithologies reported in previous studies, but not seen in the small portion mapped during this survey, are calcareous, epidotic and pyroxinic horizons (Geological Survey, 1913; Sutton & Watson, 1954).

The chief characteristic of the Lewisian gneisses is their highly deformed state. A strong $L \geqslant S$ tectonite is defined in all lithologies. The schistosity is parallel to lithological banding and the strong extension lineation is defined by hornblende, feldspar, quartz and micas, depending upon the lithology.

The siliceous acid gneisses make up the largest portion of the Lewisian. They are uniform, strongly banded lithologies, interbanded with semi-pelitic lithologies.

The hornblendic layers are massive, internally structureless bodies with parallel sides and are characterised by strong L tectonite fabrics; subordinate banded hornblende-rich and plagioclase-rich, layered lithologies are also present.

Due to their highly deformed state, no features were found which could positively distinguish between a sedimentary or igneous origin for the Lewisian gneisses.

St.4 Minor Intrusions

Two suites of pegmatites and at least one set of basic minor intrusions were found in the area mapped.

The two suites of pegmatites are structurally important due to their different ages of intrusion. They are pre-Sgurr Beag Slide (fig.110) and post-Slide, pre-mylonites (fig.50) in age respectively. The early suite is larger but rarer and possibly associated with the Carn Chuinneag granite. They are described in more detail in Section GC.5. The later suite and its deformation are described in Section MT.2.c.

Several basic sheets were found, 5-20cm wide within the Inverbroom psammites west of Fannich. They are conformable with bedding and all lay within 12 km of the thrust and were highly sheared. They have been strongly eroded relative to the surrounding psammites and are generally characterised, in the field, by a groove eroded into the psammites. Their highly weathered state made collection of specimens for thin section analysis difficult. The two specimens which were sectioned consisted almost entirely of chlorite, quartz and albite. Rare amphiboles (Hornblende) were found relict in the fabric. It appears that the original igneous or peak metamorphic fabrics were obliterated during formation of the Moine Thrust zone and that the basic sheets were more susceptible to the deformation than the surrounding psammites. It is also noteworthy that the sheets were deformed under greenschist facies conditions during the Moine Thrust related deformation.

St.5 Correlations

St.5.a The Scardroy-Fannich area

Correlation of local successions was first attempted by Sutton and Watson (1954) when they compared the succession they had determined in Scardroy (Sutton & Watson, 1953) with that seen in the Fannich mountains (Sutton & Watson, 1954) (fig.6).

In 1962 Sutton and Watson modified their earlier views on the status of acid gneisses in the core of the Fannich synform (Section St.1.c), in the light of further evidence gained during mapping north of Glenelg and from the work of Ramsay (1957). Sutton and Watson accepted the inliers of Fannich and Scardroy, previously regarded as parts of the Moine succession, as Lewisian and postulated a structural history in which wedges of Lewisian basement were thrust into the Moine succession (fig.7). They considered that the Moine remained relatively unaffected by this process except in the proximity of the "tongues" of Lewisian basement. This explained the paradox of the metasedimentary sequence younging toward inliers in Fannich and Scardroy. The reinterpretation had little effect upon the stratigraphic sequence determined for Fannich in 1954, except that it suggested that a tectonic break must exist in the sequence. Sutton and Watson (1962) also suggested correlation with the south-western Moine areas of Beinn Dronaig, Loch Hourn and Morar.

Tanner et al. (1970) reviewed the work of Sutton and Watson while considering the central Ross-shire inliers. Communication with Sutton and Watson and knowledge of the outcrop of the Sgurr Beag Slide in Kinloch Hourn (Tanner, 1971) made Tanner et al. (op.cit.) place the Scardroy Lewisian sheet immediately above the slide zone in their sequence.

FIGURE 11. Correlation of the areas of Fannich and Scardroy

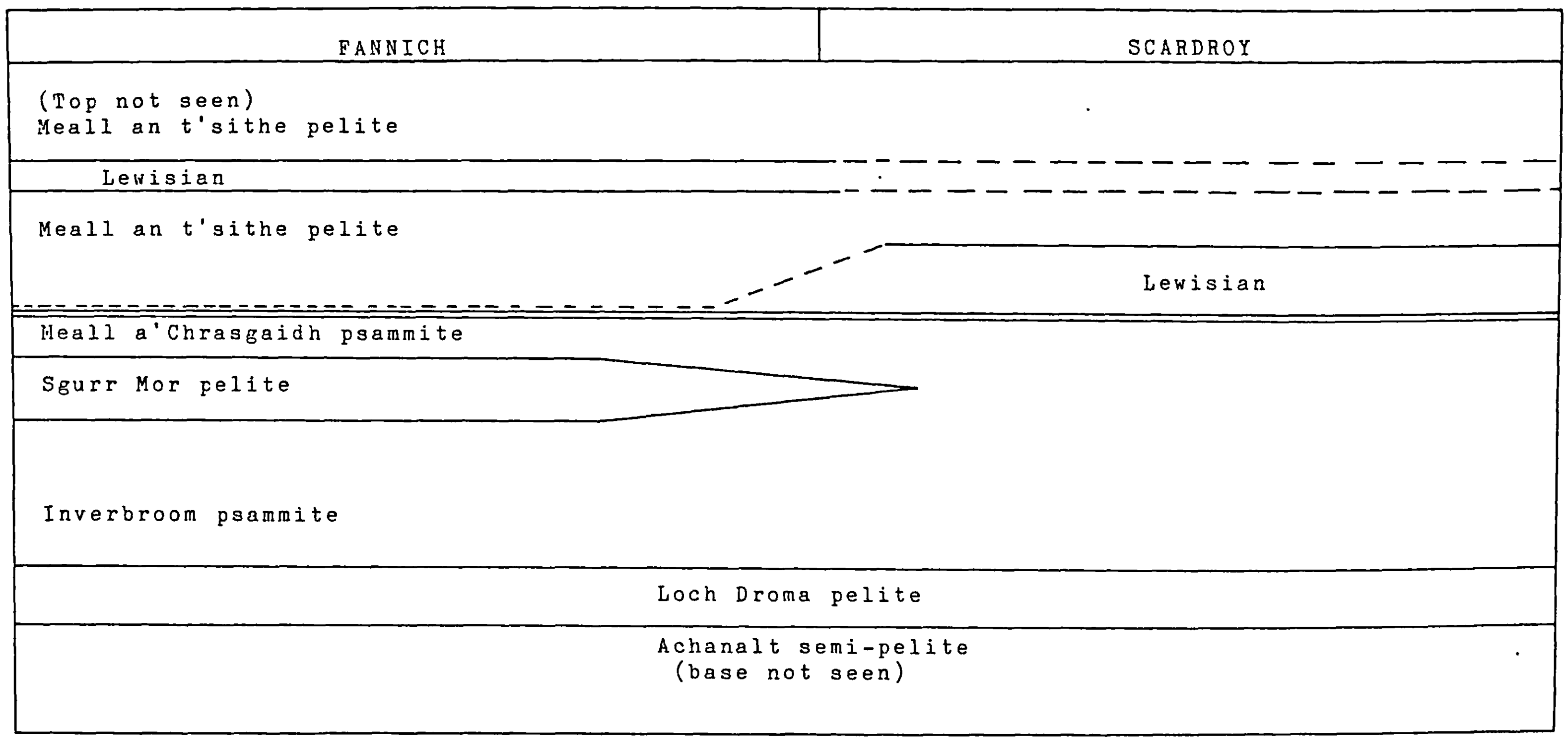
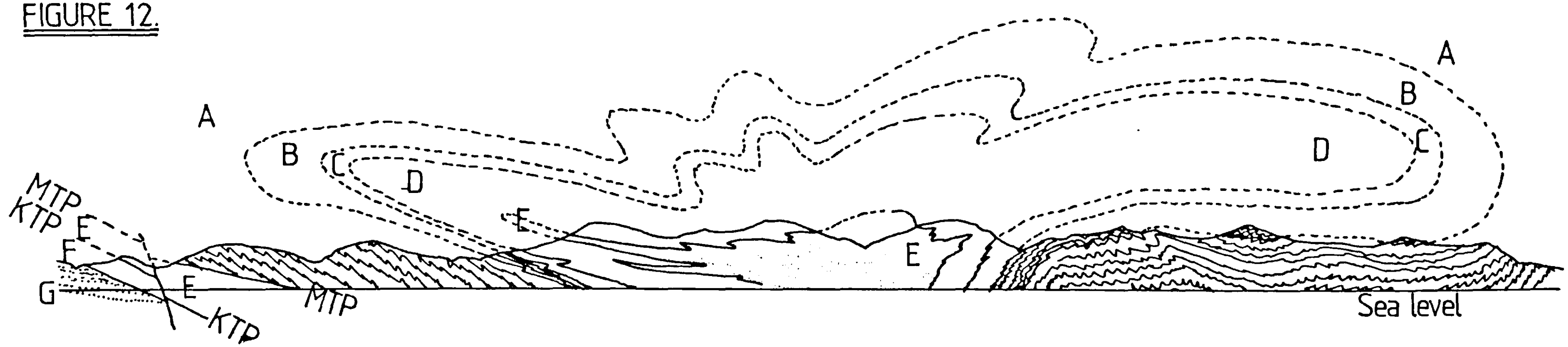


FIGURE 12.



Section from the Moine Thrust, across Fannich to An Coileachan. (The Geological Survey, 1913)

A. Inverbroom psammites B. Sgurr Mor pelite C. Meall a Chrasgaidh psammite D. Meall an t'sithe pelite
E. Lewisian gneisses F. Cambro-Ordovian shelf sequence G. Torridonian sandstones.
M.T.P. Moine Thrust plane K.T.P. Kinlochewe Thrust plane

Although they did not specifically reinterpret the succession determined by Sutton and Watson for Scardroy, this has now proved possible using their interpretation of the central Ross-shire Lewisian inliers (fig.11). Beneath the Sgurr Beag Slide in Scardroy, the sequence is similar to that in Fannich, without the Sgurr Mor Pelite which wedges out. It is therefore convenient to consider the Meall a Chrasgaidh and Inverbroom psammites as one in Scardroy under the title Inverbroom psammites.

Sutton and Watson (1954) considered a psammite (unnamed) above the Scardroy Lewisian sheet to be equivalent to the Meall a Chrasgaidh Psammite seen in Fannich. It is tentatively interpreted here as a highly strained acidic Lewisian layer, since it is described as containing hornblendic and calcic layers by Sutton and Watson (op.cit.). The unnamed psammite also rests within the Glenfinnan Division using the interpretation of Tanner et al. (1970) (fig.11). Reinterpretation of the whole sequence from Fannich to Scardroy is presented in Figure 11.

The Scardroy Lewisian is interpreted (Tanner et al., op.cit.) as a large sheet lying on the Sgurr Beag Slide which rapidly thins towards Fannich. It is retained only in small pods on the slide at Achnasheen and in thin layers within the Meall a Chrasgaidh Psammite of Fannich (Section SBS.1.b) (fig.11). As seen in Figure 11 the Fannich Lewisian outcrop lies at a higher structural level within the Glenfinnan Division migmatites.

St.5.b The Fannich-Carn Chuinneag area

The area of Carn Chuinneag and Inchbae has been much studied since its original mapping by the Geological Survey. Peach et al. (1912) determined a succession west of the Carn Chuinneag granite which was proved to be 'right way up' by Shepherd (1973). The succession is:

Upper Pelite

Upper Psammite

Lower Pelite

Lower Psammite.

The Carn Chuinneag granite sheet lies at the top of this sequence cutting the boundary between the Upper Pelite and Upper Psammite.

The lithology most easily correlated with the Fannich succession is the Lower Pelite. Winchester (1976) reported amphibolites of probable volcanic origin within this, similar to those characteristic of the Sgurr Mor Pelite in the Fannich mountains (Section St.2.b). The two pelites are continuous around a series of F3 folds and are essentially right way up. Below this Pelite, the lower psammite of Carn Chuinneag correlates well with the Inverbroom Psammite and is in structural continuity below the lower Pelite.

The continuity of the Loch Droma Pelite is uncertain but it seems probable that it thins and eventually disappears as it is traced northward (fig.9).

Above the Sgurr Mor elite in Fannich is the Meall a Chrasgaidh Psammite. This psammite is highly deformed and exhibits a high degree of parallelism between early quartz veins, lithological banding and early fold axial planes. It ranges from 100 to 300m width and represents the highest lithology in the Morar Division of the area. The Glenfinnan Division migmatites and the Meall an t'sithe elite directly overlie it (fig.9). Above the equivalent pelite in the Carn Chuinneag area is a highly variable lithology. Shepherd (1973) recorded a total of 9900m of psammite, of which over 3000m are coarse-grained psammites and pebbly beds (Peach _et al._, 1912; Shepherd, 1973). However, the lithologies are laterally discontinuous, indicating that Shepherd probably over-estimated the original sedimentary thickness.

Above the Upper Psammite lies a pelite lithology (the Upper Pelite) and finally, highest in the tectonic sequence lies the Carn Chuinneag granite. In no part of the sequence has any evidence of a tectonic break been reported (fig.9). This creates a problem in interpreting the path of the Sgurr Beag Slide from Garve to Fannich since it has normally been observed to be parallel to lithological banding.

East of the Carn Chuinneag granite the Sgurr Beag Slide is found to outcrop in the Garve area (Wilson, 1975; Rathbone, 1982). The slide outcrops against the equivalent of the Upper Psammite but the thickness of the Psammite is difficult to calculate in the Garve area due to later folding. Based however on the work of Wilson (1975) the Psammite has an approximate thickness of 1000m (fig.47) in the Garve area.

The areas of Fannich and Garve are structurally quite similar but in the area immediately west of Carn Chuinneag (see fig9. in folder) a large section of Morar Division psammites and pelites intruded by the Carn Chuinneag granite, lie above the stratigraphical level at which the Sgurr Beag Slide occurs in both Fannich and Garve (fig.46).

It seems very unlikely that the difference in thickness of psammites between Garve, Fannich and Carn Chuinneag is due totally to differential flattening since the difference would involve at least 90% flattening of the other areas relative to Carn Chuinneag.

Although the Sgurr Beag Slide has normally been found to follow stratigraphical horizons on a larger scale (Tanner, 1971; Johnstone <u>et al.</u>, 1969; Tanner <u>et al.</u>, 1970; Rathbone & Harris, 1979), it has been observed to cross-cut structures (Baird, 1982). It is postulated that

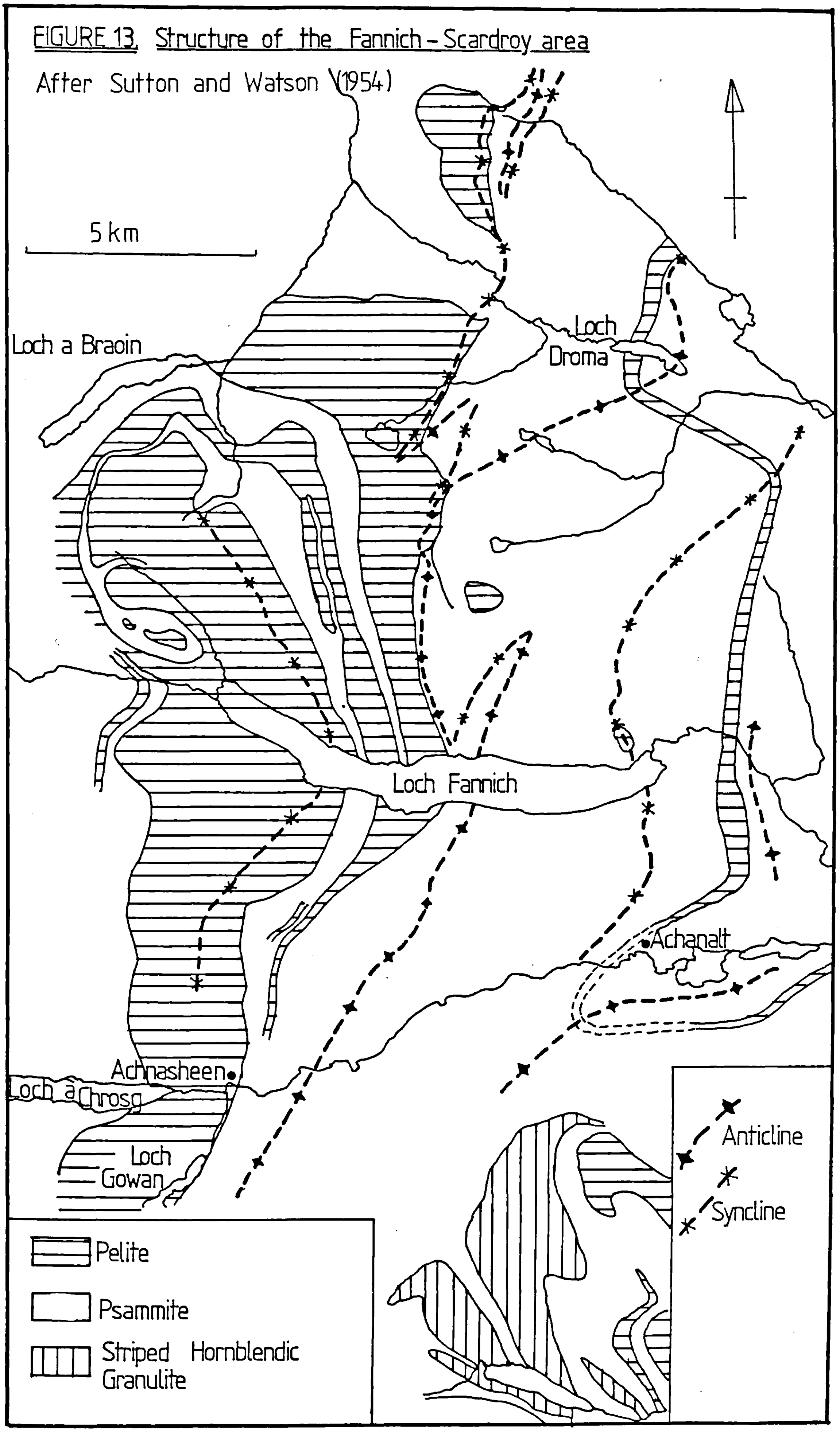


FIGURE 13. Structure of the Fannich-Scardroy area
After Sutton and Watson (1954)
5 km
Loch a Braoin
Loch Droma
Loch Fannich
Achanalt
Achnasheen
Loch a Chrosg
Loch Gowan
Pelite
Psammite
Striped Hornblendic Granulite
Anticline
Syncline

in the Carn Chuinneag area, the Sgurr Beag Slide was forced to deviate from a single stratum plane or horizon by the presence of the granite body, only part of which is preserved. The structural evidence for this conclusion will be discussed in Section SBS.4.

St.5.c Correlation with other areas

Johnstone et al. (1969) produced a three-fold division of the Moine rocks north of the Great Glen. They divided the Moine into Morar, Glenfinnan and Loch Eil Divisions forming belts along the length of the Moine outcrop (fig.3). The classification was not supposed to imply the status of stratigraphic groups since the boundaries between the divisions were thought to be tectonic. Johnstone et al. (op.cit.) used characteristic sets of lithologies and textures to define the divisions.

They considered that closer correlation of lithologies over large distances was not possible due to the lack of knowledge of intervening structures and stratigraphy. This precaution seems to be supported by the observations of the present survey.

Correlations of the succession in central Ross-shire with the Moine succession elsewhere have been attempted by several workers. Sutton and Watson (1962) correlated the Inverbroom Psammite, Loch Droma Pelite, Achanalt Semi-Pelite and Meall an t'sithe Pelite with similar horizons in Morar, Loch Hourn, Glenelg (Sutton & Watson, 1962. Text fig.1, p.528). They used the outcrop pattern of Lewisian inliers to establish correlations.

Shepherd (1973) suggested the Upper Pelite (Sgurr Mor Pelite of Fannich) was equivalent to the Striped and Pelitic group (Richey & Kennedy, 1939) and the Ladhar Beinn Pelite (Ramsay & Spring, 1962) on the similarity of position and lithologies.

Wilson (1975) went on to suggest that the Upper Psammite was probably equivalent to the Upper Psammitic group (Richey & Kennedy, 1939) and Upper Morar Psammite (Ramsay & Spring, 1962). He postulated that the Upper Pelite of Carn Chuinneag was above any of the sequence of the Morar Division exposed in the S.W. Moine. Wilson correlated the Beinn Wyvis Psammite from above the Sgurr Beag Slide in Ross-shire with the Reidh Psammite of Knoydart due to its consistent outcrop immediately above the slide. Rathbone (1982) disagreed with this long-range correlation on the grounds that the texture and content of the Reidh Psammite is different from that of the Beinn Wyvis Psammite.

Wilson also correlated the Beinn Wyvis Schist (Glenfinnan Division migmatites) with the Sgurr Beag Pelite (Tanner, 1971) of Kinloch Hourn and the Lochailort Pelite of the Morar area (Powell, 1964).

In the present survey, it has been noted that lithologies of the succession local to the Fannich and Carn Chuinneag area are laterally discontinuous. All the major pelitic lithologies in the Morar Division wedge out within 50km north/south and even less in east/west directions. The internal lithologies of psammite groups change from area to area and locally become pebbly or gritty horizons which are laterally discontinuous. Thus it is difficult to envisage correlation of individual lithologies over more than 25km without reasonable structural and stratigraphical controls from adjoining areas. It seems unlikely that any of the stratigraphic horizons in Fannich would be continuous to Kinloch Hourn or Morar.

CHAPTER II. TECTONO-METAMORPHIC HISTORY

INTRODUCTION

The present structure of the rocks around Loch Fannich and westward to the Moine Thrust was brought about during two major deformations: The first was movement of migmatitic gneisses of the Glenfinnan Division (and associated Lewisian gneisses, highly tectonised during the Caledonian orogeny) north-westward over lower grade rocks of the Morar Division; the second involved transport of both the Glenfinnan and lower grade Morar Division rocks towards the NW, over very low grade and unmetamorphosed rocks of the Cambro-Ordovician shelf sequence, Torridonian sandstones and gneisses of the foreland that were not deformed during the Caledonian orogeny.

In reporting the tectonometamorphic history of the Fannich-Ullapool area, it has proved difficult to use the conventional system of classification, separating the deformation into F_1 to F_x fold phases with their associated fabrics. Since the deformation consists, fundamentally, of a phase of ductile thrusting (sliding) and subsequent ductile to brittle thrusting, the grouping of events has been related to these events.

The first section of the tectono-metamorphic history deals with the deformation and metamorphism associated with the Sgurr Beag Slide zone and all pre-Sgurr Beag Slide history. The second section covers all deformation and metamorphism post-dating the Sgurr Beag Slide, associated with the Moine Thrust and its effects upon the Sgurr Beag Slide fabrics.

The Sgurr Beag Slide was first detected by Tanner (1971) in Kinloch Hourn (fig.3) and was postulated (Tanner et al., 1970) to pass through Fannich along the boundary between the Meall an t'sithe Pelite and the

Meall a Chrasgaidh Psammite (fig.4). However, a tectonic junction had been postulated at this boundary at a much earlier date (Gunn, 1898). No workers, until the present survey, had attempted to confirm the presence of the slide zone in Fannich or to relate the Fannich structures with those in the Moine Thrust zone.

One of the aims of the present survey was to relate movement on the Sgurr Beag Slide (i.e. the inner mobile zone of the orogen), with movement on the Moine Thrust zone (i.e. the external thrust belt of the orogen). The relationship between the thrust belt and the inner mobile zone has been the subject of controversy since the thrust was first recognised. The view of the early workers of the Geological Survey was that thrusting and its associated metamorphism were a marginal phase of the main metamorphism to the east. This view was followed by Bailey (1955) following new evidence from his survey of the tectonics of Skye.

H. H. Read held the opposing view to Bailey; using evidence gained while writing the Geological Survey Memoir for Sutherland (1931), he proposed that the low grade zone adjacent to the Moine Thrust was a "Zone of dislocation - metamorphism", post-dating the main metamorphism which affected Moine schists further east.

Johnson (1957, 1960), Christie (1963) and Barber (1965), proposed a polyphase deformation sequence (D1 - D4) for the Moine Thrust zone. These deformation phases were related to supposedly similar events which had been detected within the Moine rocks of the inner mobile zone. The early deformation in the marginal thrust zone was regarded as syntectonic with the early deformation and metamorphism further east.

Soper (1971) and Soper and Wilkinson (1975) established that all thrusting post-dated the Cambro-Ordovician shelf sequence of the foreland. This led to more complex correlations between the thrust belt and inner mobile zone, since a pre-Cambrian metamorphism had been indicated in the Moine rocks (Long & Lambert, 1963; Powell, 1974).

In the present survey, a tectono-metamorphic history is deduced for the area from Fannich to Ullapool. The proximity of the Sgurr Beag Slide and Moine Thrust allows the relationship between the two movement zones to be determined, without resorting to long-range correlations of deformation phases.

THE SGURR BEAG SLIDE

Several tectonic slides have been described from the Moine rocks in the NW Highlands of Scotland (Tanner, 1971; Powell, 1974; Rathbone & Harris, 1979; Soper & Barber, 1982). The best documented example, the Sgurr Beag Slide can be traced for over 150km, from the area around the Carn Chuinneag granite and Fannich to Ardnamurchan in the south (fig.3).

Readers unfamiliar with the term 'tectonic slide' are referred to the review of the subject by Hutton (1979). Hutton defines the term as follows:

> "A tectonic slide is a fault which forms in metamorphic rocks prior to or during a metamorphic event. It occurs within a zone of coeval penetrative deformation that represents an intensification of a more widespread, often regionally developed, deformation phase. Within this zone of high strain slides may lie along and be sub-parallel to (although they will cross-cut on a large scale) the boundaries of lithological, tectonic and tectono-metamorphic units."

The Sgurr Beag Slide was first recognised by Tanner (1971) in Kinloch Hourn (fig.3) and by workers of the Geological Survey to the north of this area (Johnstone et al., 1969). It is a tectonic break separating two lithologically distinct groups of Moine rocks of differing sedimentary facies, metamorphic grade and possibly history. These have been termed the Morar (underlying the slide) and Glenfinnan (overlying the slide) Divisions (Johnstone et al., 1969).

In the late 19th Century, Gunn (Summary of Progress, 1898) postulated a tectonic break in the Fannich area at the position now recognised as the Sgurr Beag Slide, but his ideas were not included in the Geological Survey Memoir for sheet 92. Gunn, in these early days, had recognised the possible significance of the strong contrast between the Morar and Glenfinnan Divisions in the Fannich area, which elsewhere appear to be less dramatic.

The Sgurr Beag Slide can be recognised in any particular area by a series of criteria: the presence of Lewisian slivers at the slide zone; an increase of shear strain as the boundary between the two divisions is approached; differing pre-slide histories across the slide; and continuity between the outcrop pattern and that of the known Sgurr Beag Slide outcrops elsewhere. In the Fannich area all these criteria are satisfied in relation to the contact zone of the Meall an t'sithe Pelite and Meall a Chrasgaidh Psammite (fig.4).

The Carn Chuinneag granite intrusion lying east of the Fannich area (Shepherd, 1973; Wilson, 1975; Pidgeon & Johnson, 1974) has provided an important key to correlations of the tectonic and metamorphic history of the Ullapool to Garve area. However, during this study it has been noted that previous descriptions of the relationship between the intrusion and the Sgurr Beag Slide are incomplete.

SBS.1 The Sgurr Beag Slide in Fannich

SBS.1.a Previous work in the Fannich area

Much of the previous work on the Fannich area has centred on a discussion of the relationships between the high grade gneisses of the Fannich synform and the surrounding Moine schists.

During the initial mapping of the Fannich mountains the Geological Survey determined a lithostratigraphy from the superposition of rock types in the Loch Broom area and used this to derive the structure of the Fannich area. It was assumed from knowledge of other areas of the Highlands that the Lewisian gneisses formed a basement to the Moine schists (Peach et al., 1913).

In the 'Summary of progress, 1898', Gunn expressed the opinion that the high grade migmatites of the Meall an t'sithe Pelite and the Lewisian gneisses might have been introduced along a displacement horizon over the Meall a Chrasgaidh Psammite thus explaining the anomalous position of the Lewisian and the higher grade of metamorphism within the Meall an t'sithe Pelite, but no other field evidence of a break was found.

Had Gunn not, unfortunately, died prior to the production of the Fannich mountains Memoir in 1913, the concept of the Sgurr Beag Slide might have arisen earlier. Subsequent to his death, an alternative hypothesis, due to Peach, was advanced involving a major "fountain of nappes" (Peach et al., 1913) (fig.12). This in order to reconcile an apparently continuous, complete stratigraphy with the presence of basement at the centre of the structure. A consequence of this interpretation was that it required inversion of the Moine schists over the whole of the area west of Fannich to the Moine Thrust, and east of Fannich to the Strath Conon fault.

Subsequently Rutledge (1952) and MacIntyre (1954) interpreted the whole of the pelitic belt from Fannich to Ben Dronaig as a gigantic 'tectonic inclusion', essentially reverting to the views of Gunn in 1898. Rutledge and MacIntyre however also failed to find any field evidence for a tectonic break at the boundary of their tectonic inclusions other than a metamorphic contrast.

Sutton and Watson (1954) mapped the whole of the Fannich area and much of their interpretation remains valid, though incomplete, in the light of the present work. Sutton and Watson's study, making use of sedimentary 'way up' criteria developed a structural model for the area involving a single generation of major folds (fig.13) deforming an essentially right way up succession. Sutton and Watson described changes in the style of folds along their strike and commented on the large-scale non-cylindricity (Sutton & Watson, op.cit.).

In 1954, Sutton and Watson, influenced perhaps by the views of Read, regarded the acid gneisses within the core of the Fannich syncline as an integral part of the Moine succession rather than as Lewisian basement (as had been proposed by the Geological Survey). Thus there was no need to invoke the tectonic emplacement of the gneiss. However, in 1962 they accepted the presence of the Lewisian and attempted to explain the anomaly of its occurrence in a model which included the central Ross-shire inliers. The model regarded the Lewisian inliers as a series of basement slices pushed into various levels of the previously complete Moine succession (fig.7) (Sutton & Watson, 1962).

In 1970, Tanner et al. related the presence of the central Ross-shire inliers to the development of the Sgurr Beag Slide. They postulated that the Scardroy Lewisian was brought from depth along the Sgurr Beag Slide. Tanner et al. interpreted the Lewisian in Fannich as a pod lying on a slide plane higher than the Sgurr Beag Slide since the latter was thought to run along the boundary between the Meall an t'sithe Pelite and Meall a Chrasgaidh Psammite. At the same time, Winchester (1970) considered the structural sequence of the Fannich mountains to be radically different to that proposed by previous workers. He postulated the existence of two early, large-scale isoclinal fold phases and considered the major folds described by Sutton and Watson (1954) to be third phase structures. The interpretation of pods of a "conglomerate" lithology (section St.3.b) between the Meall an t'sithe Pelite and Lewisian in the core of the Fannich synform as a basal conglomerate, led Winchester to the conclusion that the Lewisian formed a basement to the Moine schists of the area.

He therefore resisted the ideas of Johnstone et al. (1969) and Tanner et al. (1970) who postulated an extension of the Sgurr Beag Slide running at the boundary between the Meall a Chrasgaidh Psammite and Meall an t'sithe Pelite.

In 1970, Winchester showed that the only evidence supporting the presence of the slide was circumstantial. However, as a result of later work in the Freewater Forest, north of the Fannich area, he accepted its presence (Winchester, 1974).

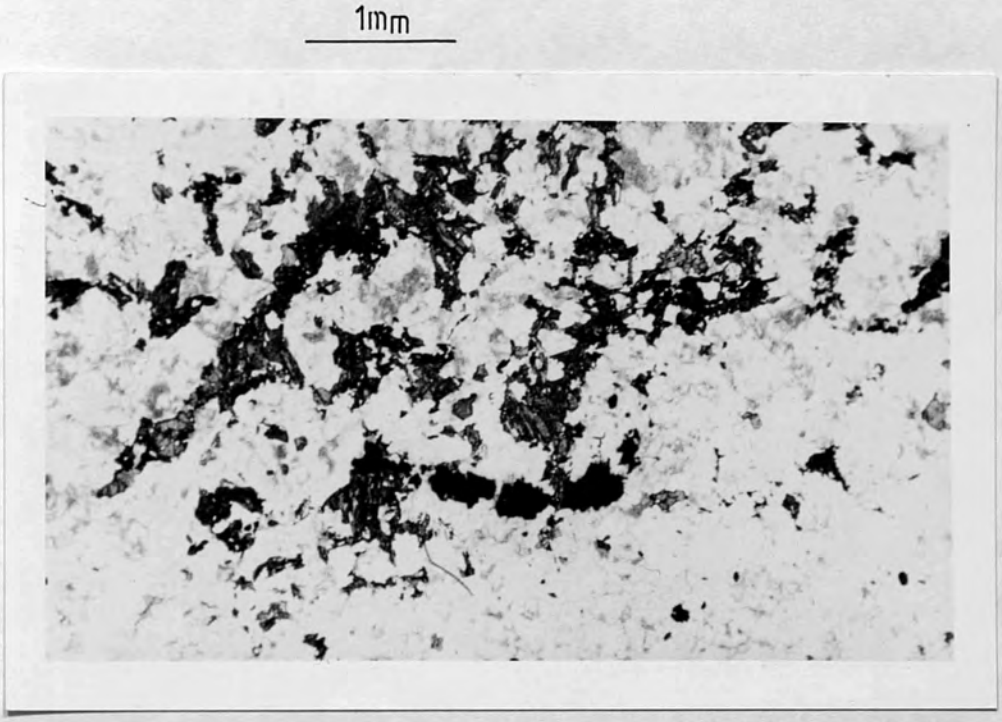
Thus, before the present survey, there was general agreement as to the presence of the Sgurr Beag Slide in Fannich, but the evidence consisted mainly of the apparently anomalous juxtaposition of pelitic migmatites bearing a lithological similarity to migmatites of the Lochailort pelite at Glenfinnan, against lower grade psammites and pelites and the presence of Lewisianoid gneisses at high levels in a Moine succession.

It is accepted here that the acid and basic gneisses within migmatites in Fannich are Lewisian: Geochemical studies by Winchester (1970) in Fannich and at Loch Shin indicate Lewisian parentage and the banded nature of the gneisses is unlike any Moine lithologies seen. Winchester extended the known outcrop of the Lewisian gneisses to the south shore of Loch Fannich (Winchester, 1973).

SBS.1.b. Lewisian slivers along the Sgurr Beag Slide

Hornblendic pods have been recorded with the high strain zone of the Sgurr Beag Slide at Kinloch Hourn (Tanner, 1971) and Garve (Rathbone & Harris, 1979). They have been interpreted as Lewisian inliers and their presence is regarded as primary evidence for the existence of a slide zone, because they occur at high levels in the Moine lithostratigraphic succession. They are accepted herein as pre-Moinian in age.

FIGURE 14. Eroded garnets within the Meall a'Chrasgaidh psammite. The section also contains amphibole, epidote, plagioclase feldspar, sphene and quartz.



In Kinloch Hourn, Tanner (1971) described a banded hornblendic gneiss within the Moine schists which had been boudinaged due to high strain. Tanner showed that the hornblendic boudins and pods exhibited compositional banding and gradational margins. He regarded these criteria as diagnostic of Lewisian parentage since they were not found in the metamorphosed intrusive amphibolites or calc-silicates of the area.

Rathbone and Harris (1979) pointed out the problems of distinguishing the hornblendic pods as Lewisian "inliers" but indicated that in zones of high Caledonian strain, the presence of hornblende might be the only criterion available. They agreed with Tanner (1971) that the banded nature of the hornblendites might be used to determine Lewisian parenthood. A Lewisian inlier at Garve has been described by Rathbone and Harris (1979) and strain within the Morar Division psammites can be seen to increase towards the inlier. It consists of banded acidic and basic gneisses continuous for over 1km along the boundary between Glenfinnan and Morar Divisions.

The Scardroy Lewisian outcrops over 10 x 15km and although highly deformed during post-Sgurr Beag Slide deformation, contains recognisably Lewisian features (Sutton & Watson, 1953, 1962; Tanner <u>et al.</u>, 1970). Tanner <u>et al.</u> (<u>op. cit</u>) have interpreted it as a mass of Lewisian sitting on the Sgurr Beag Slide along the boundary between migmatites of the Glenfinnan Division and psammites of the Morar Division.

BGS Sheet 82 (Loch Carron) shows a Lewisian body over 1.5km long at NH 158586 in Strath Bran, just north of the Achnasheen railway station. The original survey map shows the inlier within a mass of pelite. However Sutton and Watson (1954) distinguish between the migmatitic pelites of the Meall an t'sithe Pelite and psammites of the Sgurr Mor Pelite and the Lewisian inlier marked lies between the two. A further, smaller body of Lewisian (discovered by the author) lies in the river bank just beyond the road junction (NH 157584) again lying at the boundary between migmatitic and unmigmatitic rocks. It consists of a banded gneiss with alternating hornblende-rich and feldspar-rich layers approximately 2cm thick, which lie parallel to the schistosity of the surrounding rocks. The Achnasheen inliers do not represent extensions of the Fannich Lewisian inlier (Winchester, 1973) since they are emplaced at a different structural level in the Moine. They do not show any structures indicative of fold cores, as seen at Glenelg (Ramsay, 1957) but are pods within the highly strained Moine rocks similar to inliers at Garve and Scardroy, which lie along the Sgurr Beag Slide. The tectonic level at which the Achnasheen inliers lie can be traced around major F3 folds and are evidently analogous with the Scardroy Lewisian.

No other large Lewisianoid bodies were found at the slide boundary within the area mapped in detail, although the Achnasheen bodies lie only 5km along strike from Fannich. However, in the Meall a Chrasgaidh Psammite (fig.4), thin layers, approximately 1-2cm thick, are found which are hornblende rich and have the mineralogy: quartz, plagioclase (An 43), hornblende, garnet, epidote, sphene, trace apatite and opaques. Originally large crystals of plagioclase and hornblende are pseudomorphed by clusters of small grains and sub-grains indicating the breakdown of a coarser grained rock during intense deformation. The mineralogy of the layers is similar to rare epidote-rich members of the calc-silicate suite of Fannich (section SBS.2.a. and Winchester, 1974), but whilst garnets are characteristically well formed and up to 1cm diameter in the calc-silicates, in the hornblendic layers of the Meall a Chrasgaidh Psammite, the garnets are skeletal and have an 'eroded' appearance (fig.14). It is therefore concluded that these hornblendic layers may be of Lewisian origin and represent relics of larger Lewisian pods. It is also possible that other Lewisian inliers occur along the slide in the Fannich area but are acidic in nature and are thus extremely difficult to differentiate from the highly strained Morar psammites.

The mechanism for emplacement of the Lewisian inliers at the Morar/ Glenfinnan boundary is uncertain. This problem occurs because the extremely high strain, (Rathbone & Harris, 1979) associated with their emplacement obscures all evidence of the early history of deformation. Two possible mechanisms are:

a) Progressively increasing strain involved in isoclinal interfolding of Lewisian and Moine, followed by shearing and finally slide movement preserving boudined pods of Lewisian fold cores in the zones of high strain.

b) A thrust tectonic-type mechanism (Barton, 1978) where the Lewisian slices represent deep level, far travelled horses detached from the footwall of the slide during movement.

Mechanism (b) is currently popular since the recent interpretation of the Moine Thrust belt in terms of a thin-skinned foreland propagating sequence (Elliott & Johnson, 1980). No final conclusions can be drawn however because of a lack of good evidence in favour of either hypothesis.

The mechanism for incorporation of the larger Fannich Lewisian inlier will be discussed in Section SBS.3.a. since it does not lie on the Sgurr Beag Slide.

Evidence for large displacements upon the Sgurr Beag Slide lies in the absence of any marker which can be traced from the footwall to hanging wall. The only positive evidence is that of the pods of Lewisian on the slide since they must have been stripped from basement, but their source too is unseen. The assumption made therefore is that the distance

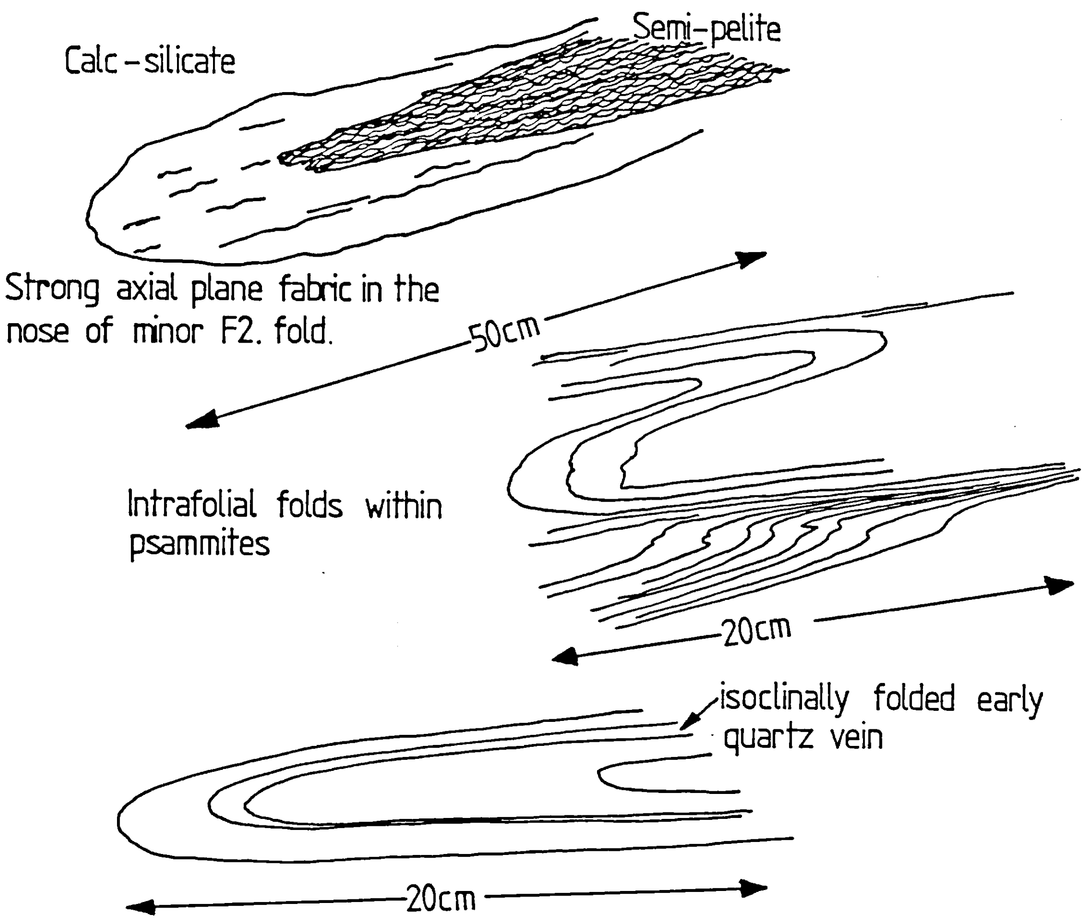


FIGURE 15. F2 folds from the Sgurr Mòr pelite and Meall a'Chrasgaidh psammite

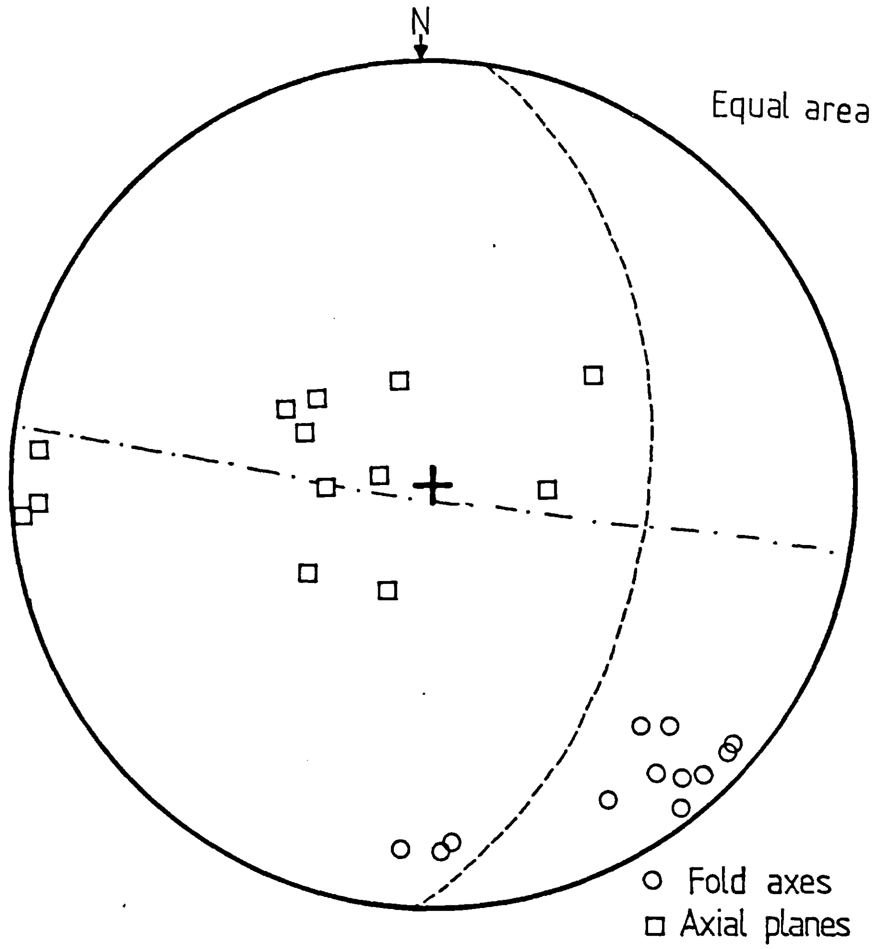


FIGURE 16. Stereonet of minor F2 folds which were deformed during D3. Note F3 axial plane (dashed line) and plane of pre-D3 foliation/bedding in Fannich (dash-dot line).

moved is in excess of the outcrop presently at the surface when measured parallel to the movement direction (fig.3). The movement direction is thought to be parallel to the extension lineation described in Section SBS.1.c. The Sgurr Beag Slide seen adjacent to the Moine Thrust at Loch a Braoin (fig.4) can be traced continuously around post-Sgurr Beag Slide, F3 folds to the Scardroy area before it is lost at the Strath Conon fault. When the effects of F3 are removed, this represents a minimum movement of around 55km.

SBS.1.c. The variation of D2 in relation to the Sgurr Beag Slide

In the Fannich region the lateral equivalent of the Sgurr Beag Slide occurs at the boundary between the Meall an t'sithe Pelite and Meall a Chrasgaidh Psammite, the position predicted by Tanner et al. (1970).

The deformation associated with slide movement is D2 (section SBS.3.a) of the local sequence and the most striking feature of D2 is the concentration of intense deformation close to the slide.

In Kinloch Hourn, Tanner (1971) equated slide movement with development of a series of major and minor folds which possibly outlasted slide movement. In other areas (Powell, 1974; Baird, 1982) the slide deformation has also been associated with the formation of major folds. However, in Fannich this does not seem to be the case - no major F2 folds have been detected in the area. F2 minor folds are generally tight to isoclinal, class 1b or 2 (Ramsay, 1967) (fig.15). They are found almost exclusively within psammitic horizons of the Sgurr Mor Pelite and within the Meall a Chrasgaidh Psammite (fig.4) (section St.2.a,b). Pelitic lithologies may also have contained F2 minor folds but the only evidence found for this was rare, hook shaped calc-silicates with strong axial planar fabrics, relics of original isoclinal minor F2 folds (fig.15).

The minor F2 axes generally plunge at low angles to the S or SE, they are near coaxial with minor F3. However, they do not have any simple geometrical relationship to the major F3 axes because the differing strains on the limbs of F3 folds cause differing amounts of rotation of the earlier F2 axes, (section MT.1.a). The poles to axial planes of minor F2 folds scatter in a broad band (fig.16) roughly at right angles to the axial plane of F3 in the same plane as the foliation due to the Sgurr Beag Slide and the bedding planes of the metasediments (fig.17).

F2 folds are more common close to the outcrop of the slide and become progressively rarer further from the slide and are extremely rare within Morar division rocks further than 1000m structurally below the slide. F2 folds have not been seen refolded by the more common F3 on a minor scale but the plot of axes and axial planes (fig.16) shows that they were folded by the major Fannich synform of F3 age, forming two concentrations of axial plane poles representing F2 folds on the steep and shallow limbs of F3.

Planar fabrics associated with the Sgurr Beag Slide are characteristically parallel to the lithological banding and increase in intensity towards the slide. No cross-cutting shear fabric, predicted by the models of Ramsay and Graham (1970) and Esher and Watterson (1974), was detected oblique to the slide zone, a feature discussed later in this Section.

Since lithologies change from the Inverbroom Psammite (St.2.a), through the Sgurr Mor Pelite (St.2.b), to the Meall a Chrasgaidh Psammite (St.2.c), it is difficult to compare quantitatively the relative intensities of deformation. Hutton (1979) (fig.18) has shown from his own and others work in Donegall, that pelites and psammites react differently to strain intensification close to the slide zones. However, an indication of increasing strain in this study was gained by considering the increase in

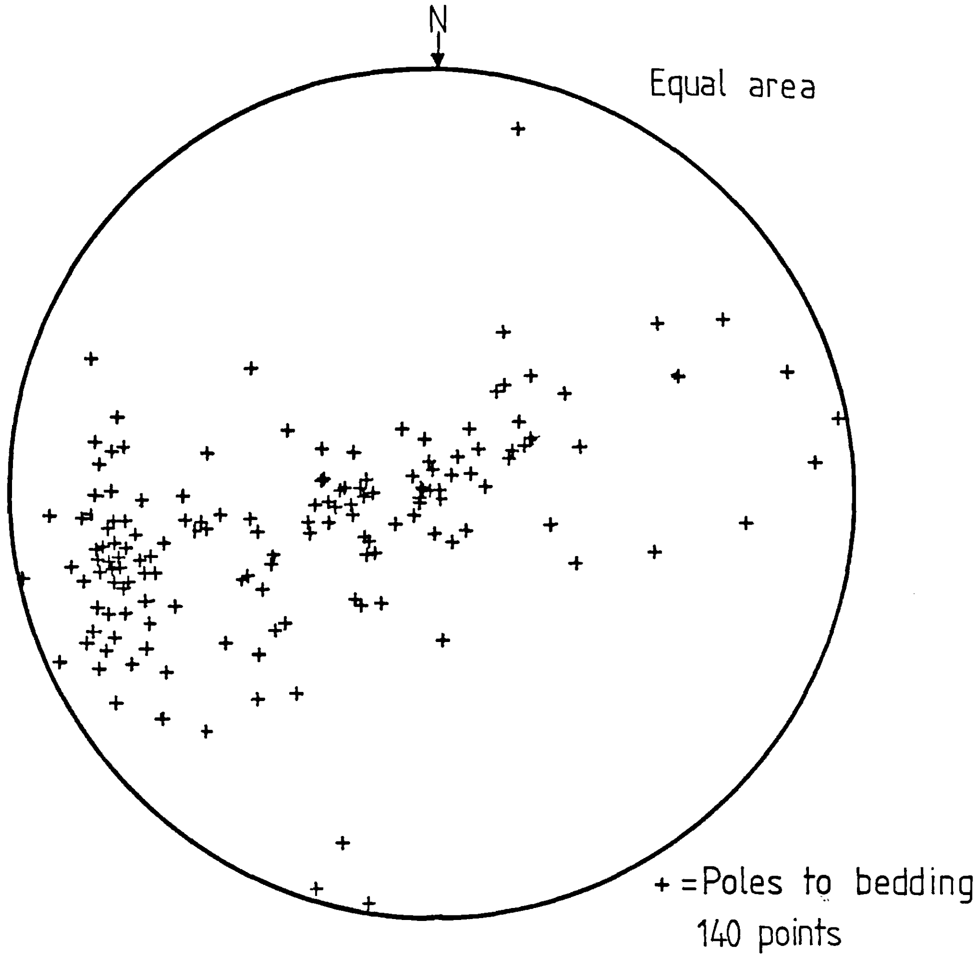


FIGURE 17. Stereonet of Foliation/bedding in the Fannich area

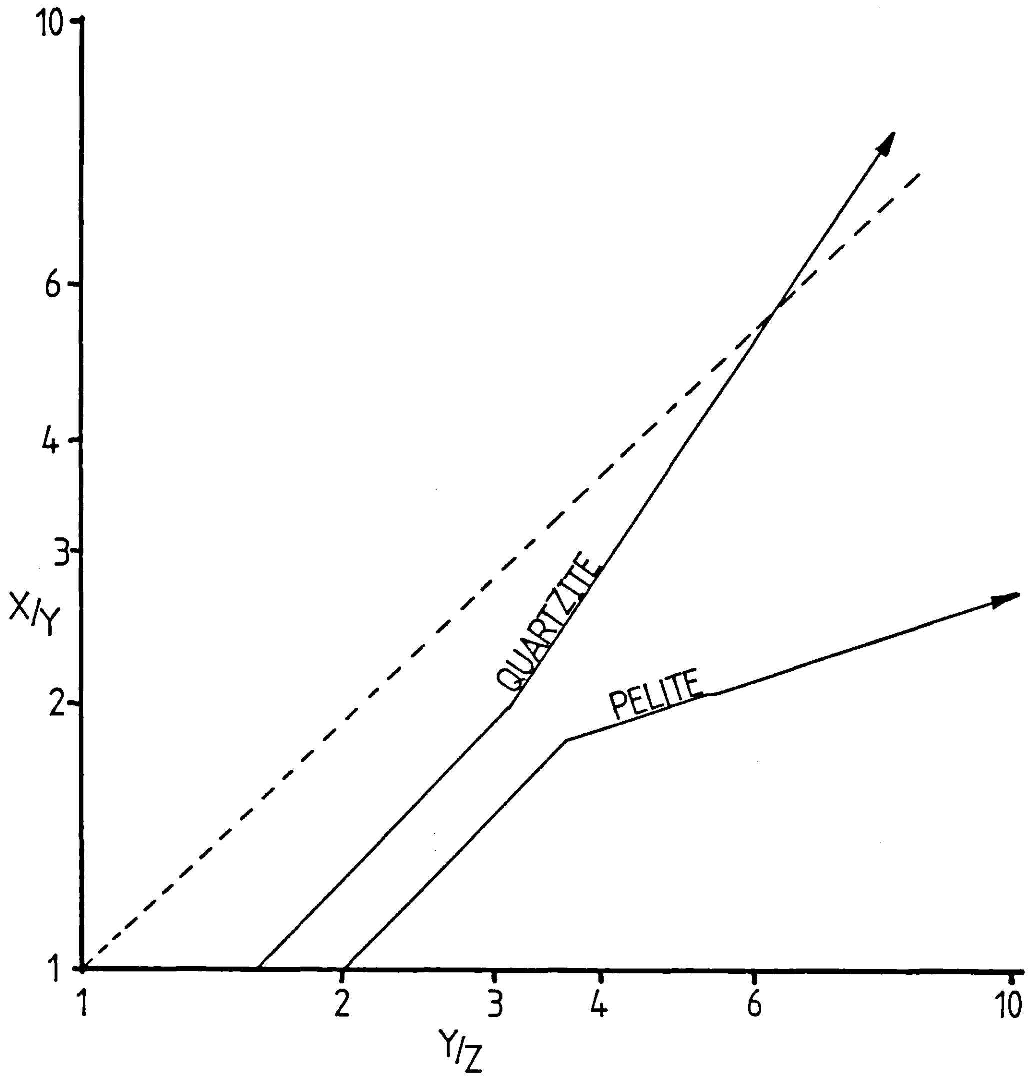


FIGURE 18. Diagram showing the differential deformation of quartzite and pelite in a slide zone, after Hutton (1979)

strength of fabrics in psammitic lithologies only. Pelitic lithologies exhibit a strong alignment of mica cleavage planes parallel to lithological banding but the fabric is susceptible to later deformation and exhibits a ubiquitous D3 crenulation (fig.19).

In areas furthest from the slide, in the Inverbroom Psammites (fig.4), cross-bedding and sedimentary slump features are developed. Although rare, they are relatively undeformed (fig.20). The lithological banding (sedimentary banding) is undulose and varies in thickness from 10cm to 2m (Section St.2.a). Finer banded psammite sequences show planar fabrics defined only by the alignment of mica (001) planes. These are therefore best developed adjacent and within thin pelitic and semi-pelitic bands interbedded with the psammites. No internal planar structures were detected within the thicker psammitic units.

Early quartz veins within the Inverbroom Psammites of Fannich show only low levels of deformation, often remaining at angles greater than 30° from bed parallel. The lack of early highly deformed quartz veins and the presence of undeformed sedimentary structures indicates that the Inverbroom psammites and therefore the whole Morar succession in Fannich probably suffered little deformation prior to the movement of the Sgurr Beag Slide (see Section SBS.3.a).

Coming closer to the slide, within the more psammitic horizons of the Sgurr Mor Pelite (fig.4), sedimentary structures are common though it is often difficult to determine the direction of younging because deformation has reduced the angular discordance of the cross-sets (fig.21). The psammitic bands within the Sgurr Mor Pelite contain a greater proportion of micas than the majority of psammites within the Inverbroom Psammite and hence the planar structures, due to the alignment of (001) cleavages in micas, are more prominent. All early quartz veins show marked deformation,

FIGURE 19. A thin section of a crenulated pelite from the Sgurr Mor pelite.

exhibiting low angles from bed parallelism (typically less than 30°) and much buckling and folding (fig.22).

Within 200m of the Sgurr Beag Slide, psammitic lithologies of the Meall a Chrasgaidh Psammite (fig.23) exhibit an intense planar fabric defined by mica cleavage planes and the elongation of quartz and feldspar. The lithological banding is very fine (typically less than 3cm width) and is platey (fig.23). The banding has become near perfectly parallel and straight, giving it a 'tramline' like appearance. Cross-bedding has not been identified within this psammite. Early quartz veins which were originally up to 90° from parallel in the Inverbroom Psammite are very near parallel to lithological banding (fig.22) varying by less than 5° from parallelism. This increase in parallelism of all disparate planar structures is characteristic of slide zones throughout the Moine (Rathbone & Harris, 1979).

Within the migmatites above the slide, the predominant planar fabric is a shape fabric defined by the quartzofeldspathic lits. The lits and (001) planes of micas in the matrix are parallel to the plane of lithological banding. It is difficult to know to what extent the planar fabrics can be attributed to Sgurr Beag Slide deformation, since formation of the migmatitic fabric pre-dates emplacement of the slide (Section SBS.4.a). Also the migmatites are never seen greater than 600m from the Sgurr Beag Slide in Fannich due to the synclinal nature of the exposure.

The migmatite fabrics do, however, alter in relationship to distance from the slide. Furthest from the slide the lits vary in thickness and length forming a gently undulating shape fabric. Closer to the slide, the fabric becomes less undulose and the lits tend to become elongate. Adjacent to the slide the migmatites develop patches of augen gneiss (fig.8), the quartzofeldspathic lits forming 'eyes' around which the pelitic matrix wraps. In some cases the feldspars within the augened lits have broken under the strain in a similar manner to that reported within feldspathic mylonites in Eriboll (White et al., 1982) (figs. 8,24). The migmatitic fabrics are reworked by the slide deformation and hence the migmatization event must pre-date slide movement.

Linear fabrics also vary in intensity and style in relation to the slide. Within the Inverbroom Psammites furthest from the slide (except in areas adjacent to the Moine Thrust belt), only weak lineations are seen. They are defined by the weak elongation of quartz and feldspar generally only seen on bedding surfaces and within semi-pelitic layers. Within the psammitic lithologies, no structures could be distinguished.

In psammitic parts of the Sgurr Mor Pelite, the lineation is often well developed and easily viewed on bedding surfaces although difficulty is experienced seeing the same lineation within the psammites. Again, the lineation defined by the elongation of quartz and feldspar is particularly well developed in areas with a higher proportion of micas.

Within the Meall a Chrasgaidh Psammite, the lineation has become intense and is obvious on all surfaces, even within very mica-poor lithologies. It is parallel to the lithological banding and constitutes an extension lineation defined by the strongly elongate quartz and feldspar grains. In places the fabric may be defined as an L=S tectonite. The 'S' part defined by mica (001) plane alignment and 'L' by misalignment of micas and elongation of quartz and feldspar (fig.25).

Within all psammites of the Glenfinnan Division and all the Lewisian gneisses, a strong extension lineation is defined by the elongation of quartz and feldspar and alignment of acicular hornblendes. The intensity of these lineations is comparable to that of the lineation within the

FIGURE 20.
Inverbroom psammite sedimentary structures

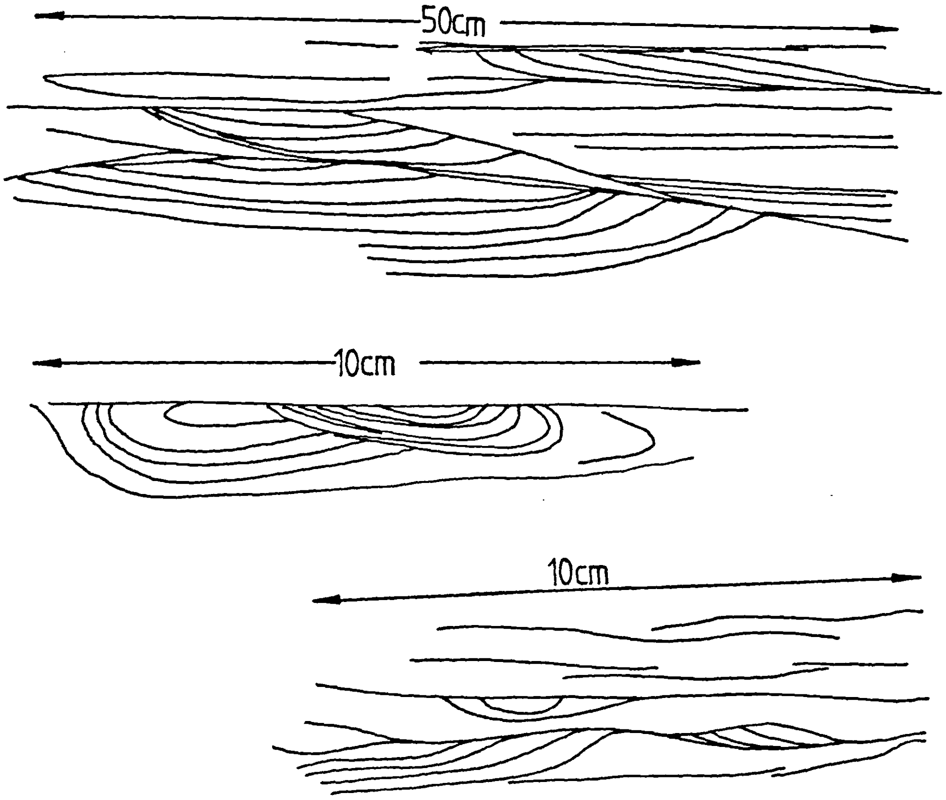


FIGURE 21.
Sgurr Mor pelite sedimentary structures

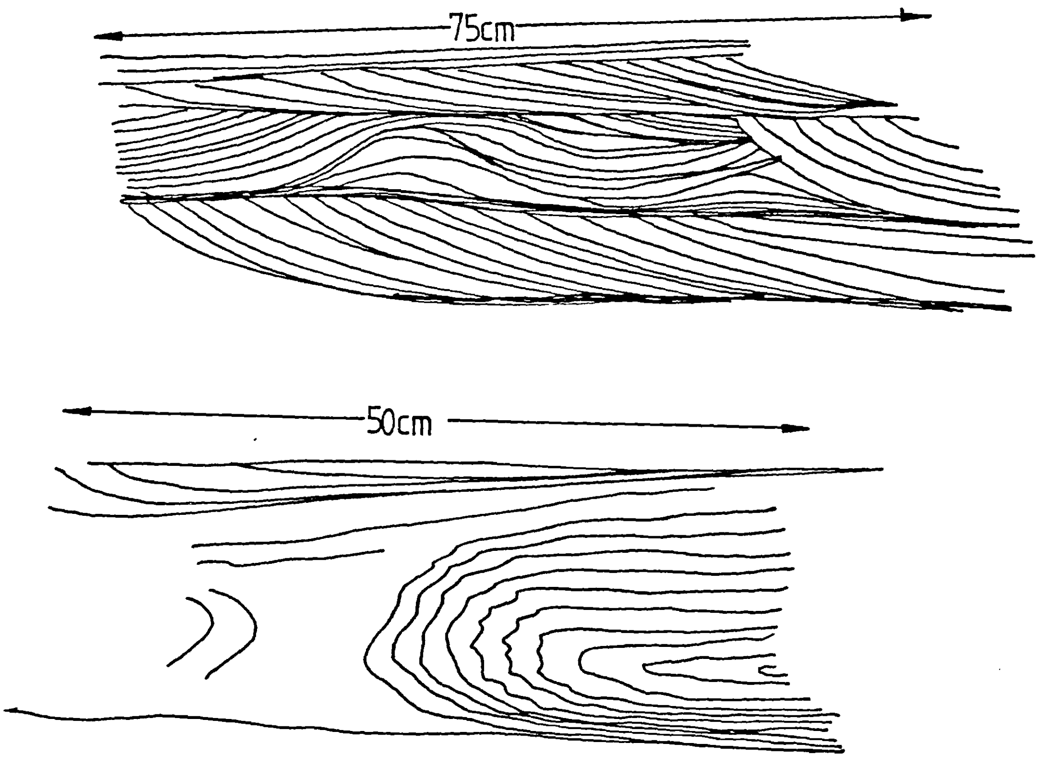


FIGURE 22. Deformation of quartz veins in the Sgurr Mor pelite. See figure 23 for deformed quartz veins in the Meall a'Chrasgaidh psammite.

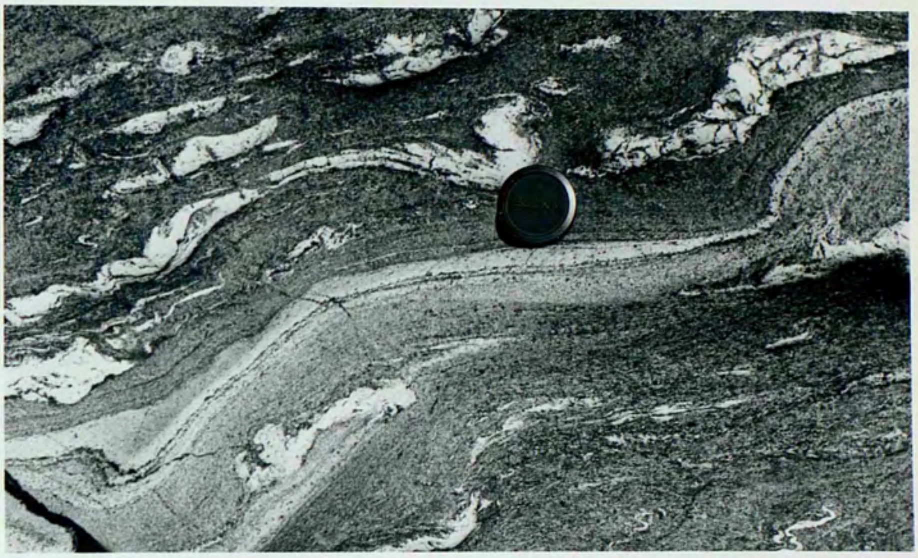
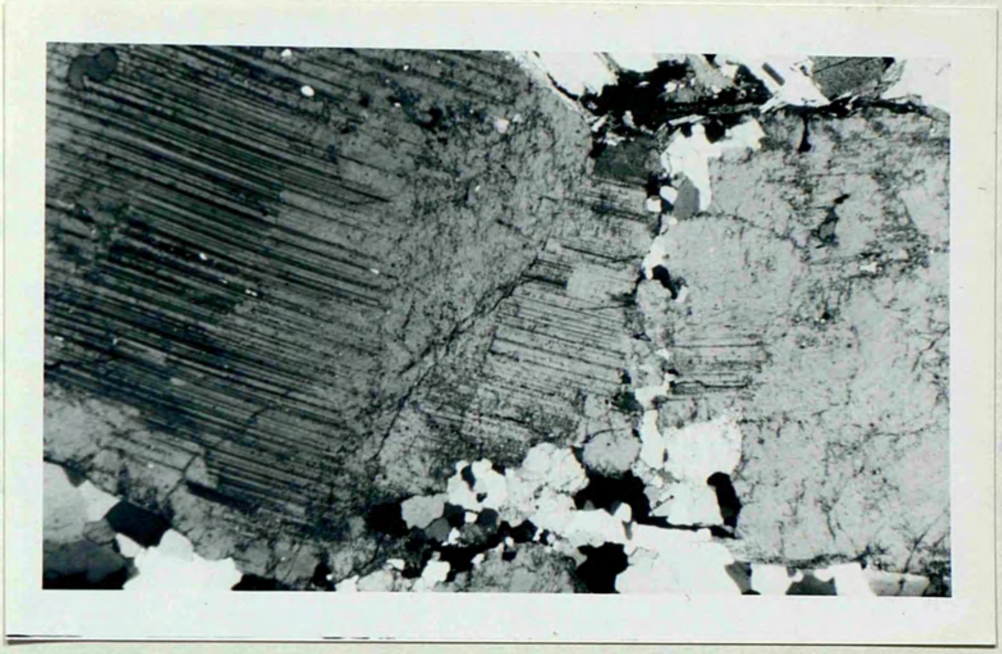


FIGURE 23. Intense planar fabric in the Meall a'Chrasgaidh psammite.

FIGURE 24. A large feldspar lath in the migmatites, deformed and broken during D2.

1 mm



Meall a Chrasgaidh Psammite. However a problem occurs in relating the lineations seen within these lithologies to a single deformation since both the Lewisian and the migmatites have a complex pre-slide history (Section SBS.3.a). It is likely that the lineation seen within the Glenfinnan Division and Lewisian is a composite fabric. The composite lineation is parallel to that generated during Sgurr Beag Slide deformation.

The orientations of L2 are strongly affected by later deformations (Section MT.2.b) but plunge generally S to SE at low angles, closer to south on the vertical F3 limbs, closer to SE on the horizontal limbs (fig.26).

As mentioned earlier, no planar or linear fabrics have been found in the Fannich area, which can be traced progressively rotating into the slide zone, however such fabrics are predicted from the shear zone models of Ramsay and Graham (1970) and Esher and Watterson (1974). Rathbone and Harris (1979) found that where minor folds were common, they could be traced progressively changing orientation into the slide zone. They also found, as in Fannich, that the minor folds are not always common. They noted too that the planar and linear fabrics attributable to slide deformation paralleled the deformation zone rather than forming in what would have been the XY plane of the strain ellipsoid for such a deformation.

This phenomenon may be explained in two ways (fig.27):

a) The zones of weaker Sgurr Beag Slide related fabrics were lost during subsequent recrystallization and deformation, leaving only the near parallel intense fabrics close to the slide.

b) The Sgurr Beag Slide deformation did not conform to a homogeneous Ramsay and Graham type shear zone but involved some inhomogeneous mechanisms.

Hypothesis (a) is difficult to reconcile with the weak fabrics which are found outside the zone of intense deformation. In areas outside the intense deformation zone, fabric paralleled the slide zone rather than forming at angles up to 45°, as predicted by Ramsay and Graham. Another argument against obliteration of weak fabrics is that a good quartz c-axis fabric is found preserved in the most intense zone; such fabrics are easily destroyed (see Section SBS.2.a) (Lister & Williams, 1979).

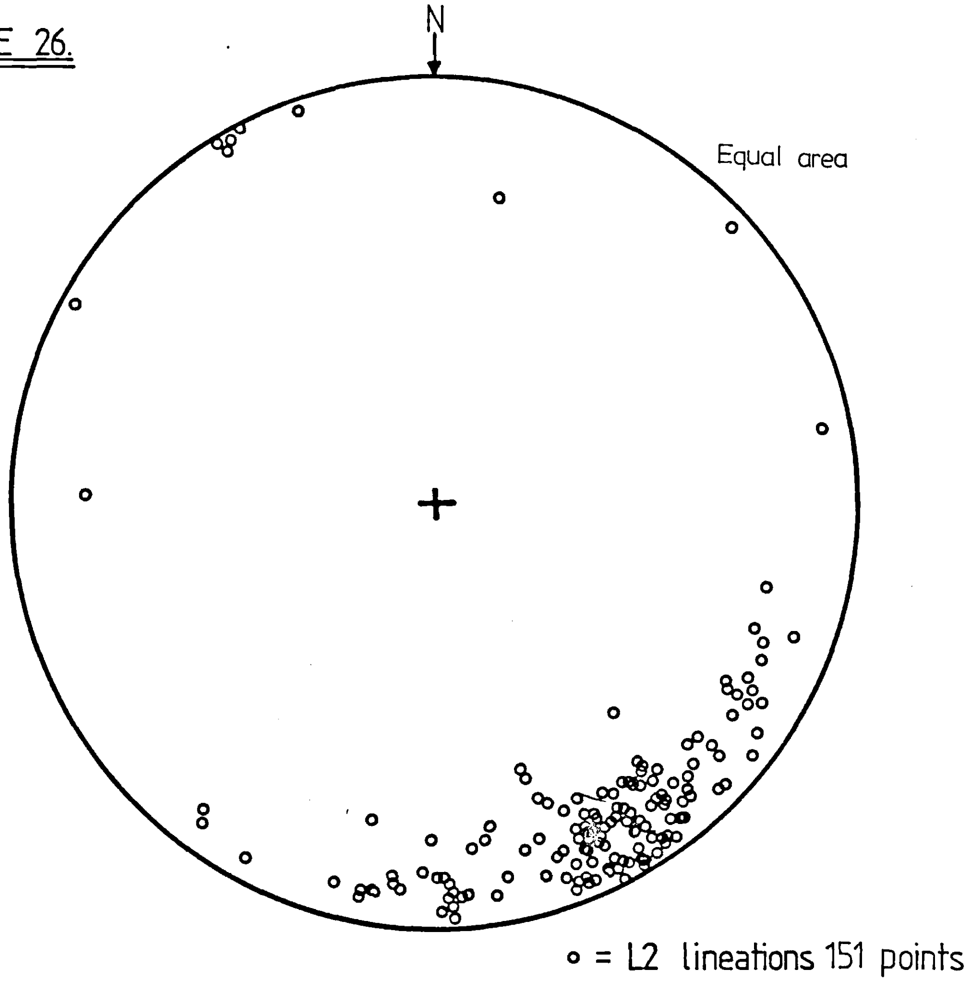
Hypothesis (b), that some type of anisotropic deformation mechanism could explain the lack of oblique shear fabrics is borne out by the work of Berthe et al. (1979). They describe two sets of planes developing in sheared granites within the South American shear zone: "S" surfaces, representing the XY plane of the strain ellipsoid and curving into the shear zone as predicted by Ramsay and Graham (op.cit.); "C" surfaces, small shear planes, parallel to the larger scale shear zone (fig.27). As deformation increases, the two sets of planes intensify and tend towards parallelism with each other at higher strains. The undeformed granite is homogeneous on a hand specimen scale, but if the initial rock had a weak planar fabric, the shears or "C" surfaces would be expected to predominate over "S" surfaces.

Applying this to the Fannich area, some of the deformation which would have led to the formation of a fabric in the less intensely deformed Inverbroom psammites may have been taken up in shearing along bedding planes. This would also explain the stronger planar fabrics and linear fabrics seen on bedding planes but not within beds.

Rathbone and Harris (1979) attempted to quantify increasing strain into the Sgurr Beag Slide zone using criteria such as the increasing deformation of sedimentary structures, early quartz veins and minor folds. A similar attempt in Fannich runs into the problem of interpreting

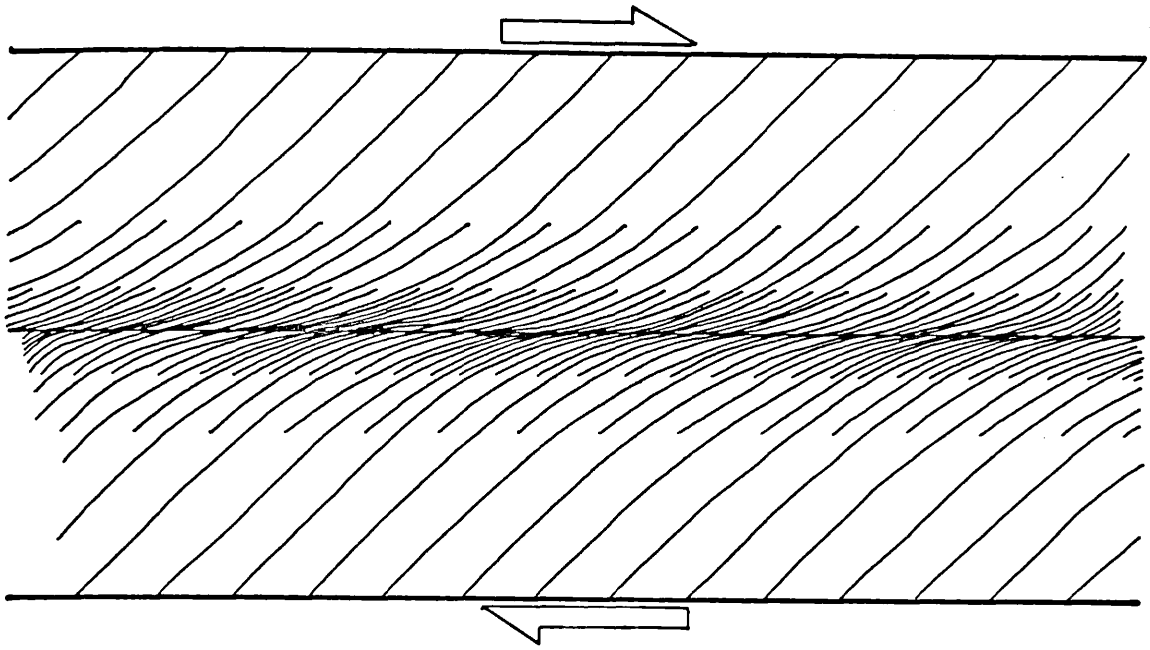
FIGURE 25. Intense stretching lineation within the Meall a'Chrasgaidh psammite. Note the highly deformed quartz vein.

FIGURE 26.

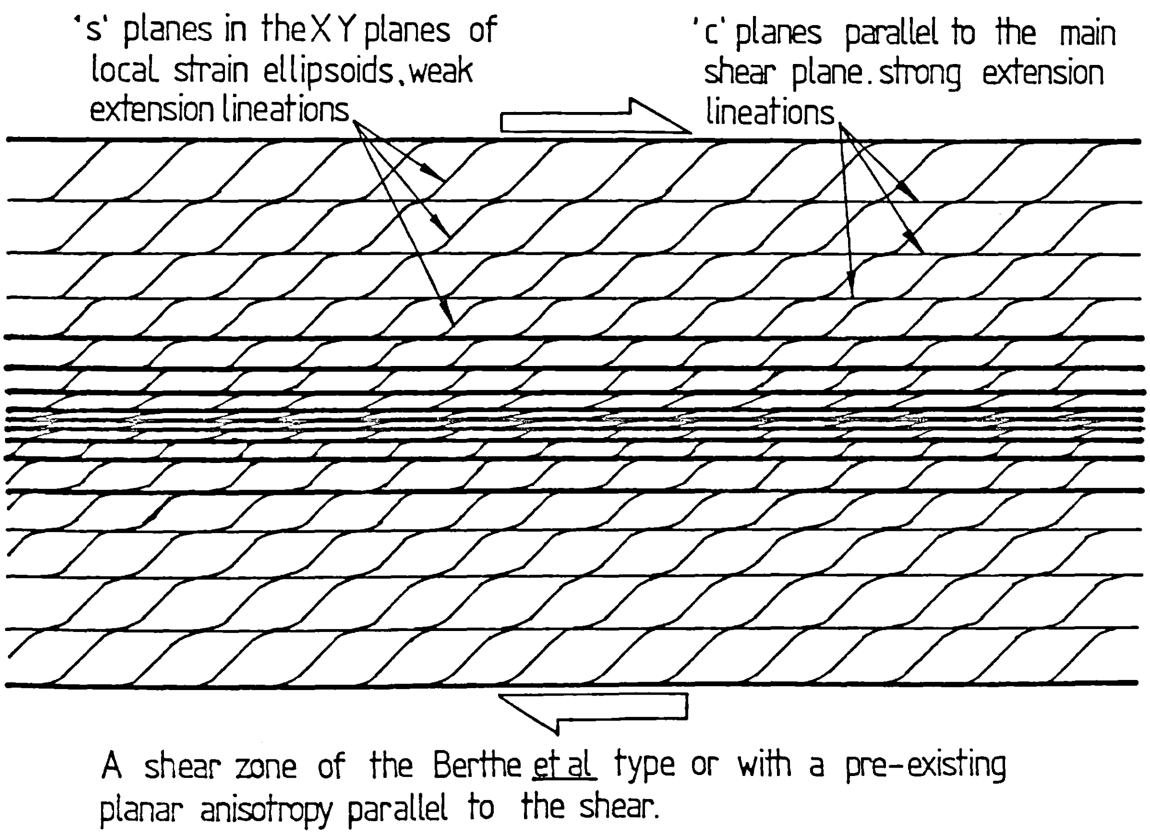


Stereonet of L2 within the Fannich area.

FIGURE 27 Two types of shear zone after Ramsay and Graham (1970) and Berthé et al



A shear zone exhibiting the formation of an extension lineation parallel to the X direction of the strain ellipsoid and its intensification towards the centre of the zone.



A shear zone of the Berthe et al type or with a pre-existing planar anisotropy parallel to the shear.

structures in a variety of lithologies which probably had different deformation paths (Hutton 1979) (fig.18). However, using the criteria of Rathbone and Harris, it can be seen that deformation within the bulk of the Inverbroom Psammites was low enough for good sedimentary structures to survive and they were therefore out of the zone of intense deformation. Within the Sgurr Mor Pelite, sedimentary structures are deformed and planar and linear structures are present along with strongly folded quartz veining, indicating that it was significantly deformed during slide movement. The boundary of the slide zone rests close to the boundary between the two groups, 800m from the Glenfinnan Division boundary. This is a similar value to that recorded by Rathbone and Harris (1979).

Although the intensification of D2 strain does not prove the presence of a slide zone, taken with the presence of Lewisian pips at the boundary (SBS.1.b), the contrast in early deformation histories (SBS.3.a) and metamorphic histories (SBS.3.b) it indicates a structure of the type (b) defined by Hutton (1979); a slide occurring between tectonic units.

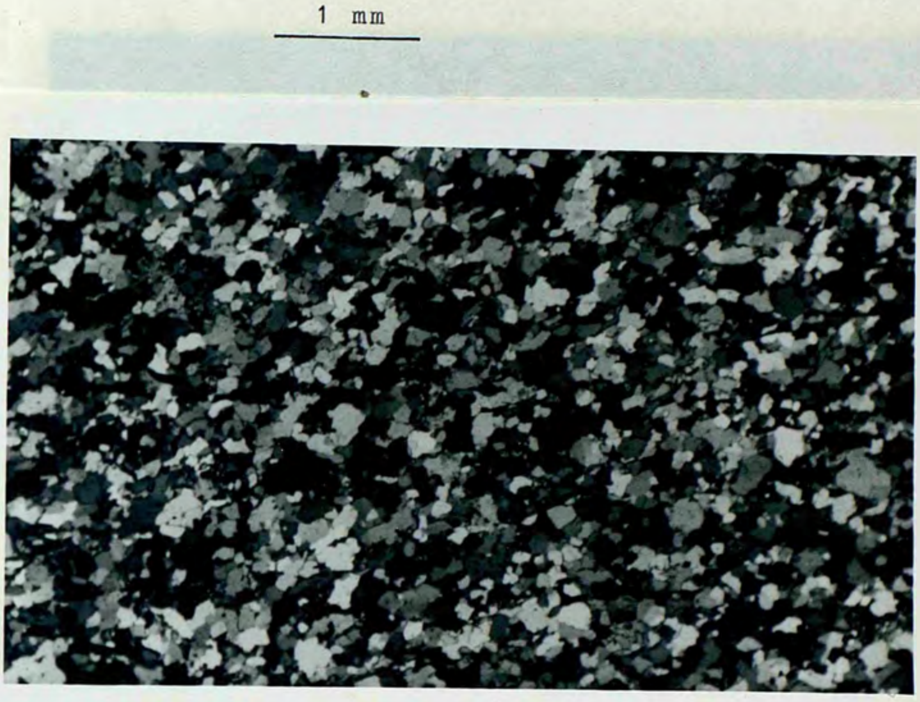
SBS.2 Metamorphic Textures and Microfabrics

SBS.2.a Psammites and pelites

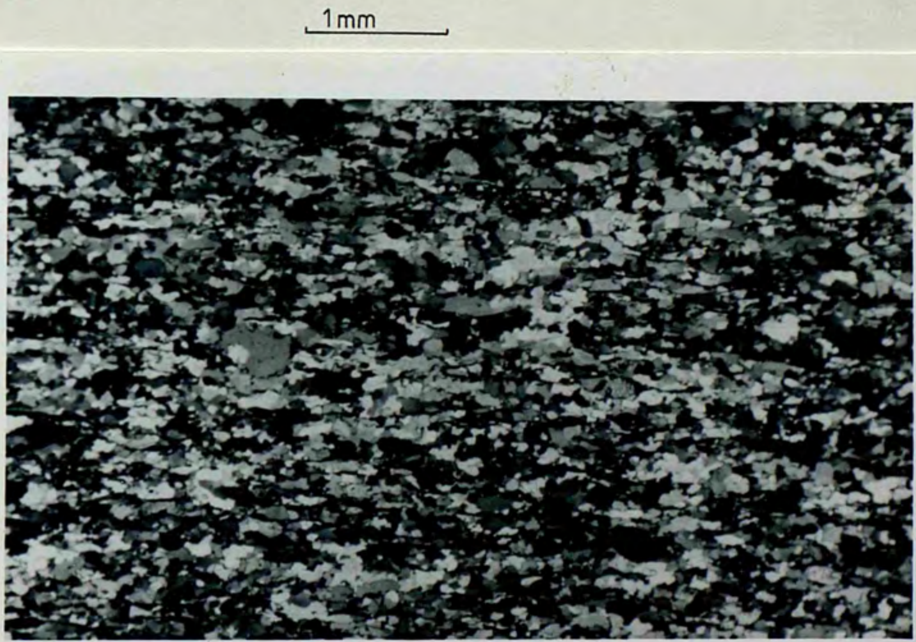
The metamorphic textures and microfabrics of peak metamorphism and movement of the Sgurr Beag Slide are present in the Fannich area. They can be used to reconstruct the environment and style of deformation due to the Sgurr Beag Slide. However, they are affected by later deformation and retrogression. Indeed, the increase in intensity of Moine Thrust microfabrics can be viewed as progressive breakdown of Sgurr Beag Slide related microfabrics (Section MT.5.a,b). It is therefore important to state the problems involved in their interpretation.

FIGURE 28. Progressive change of quartz fabrics into the mylonites.

a) Initial breakdown.



b) Intermediate breakdown.



c) Final stage breakdown.

1mm

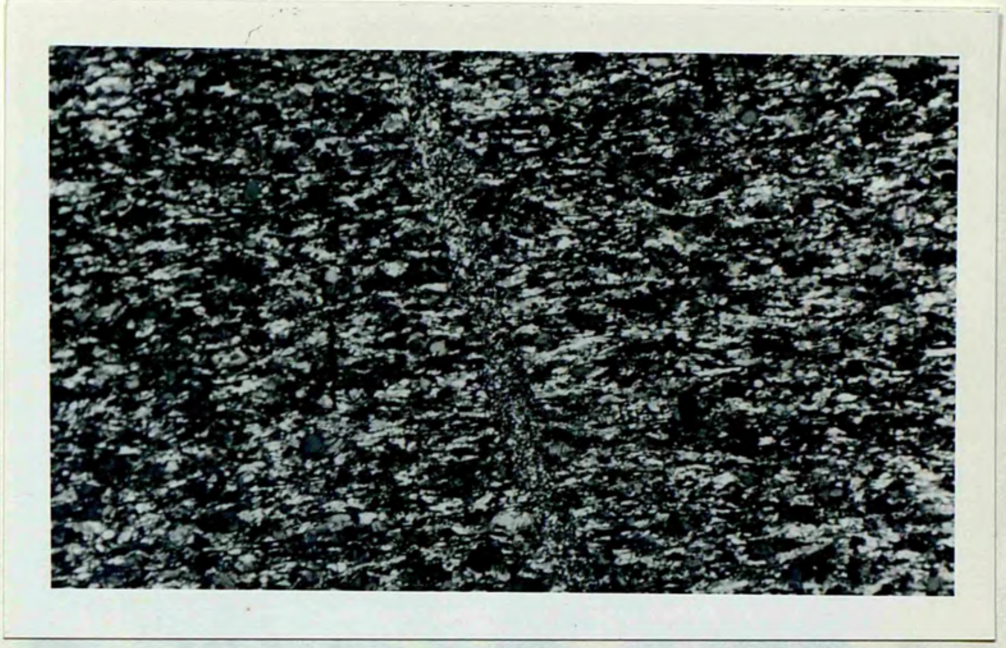


FIGURE 29.A garnet in the Sgurr Mor pelite showing fine grained inclusions and zoning.

1 mm

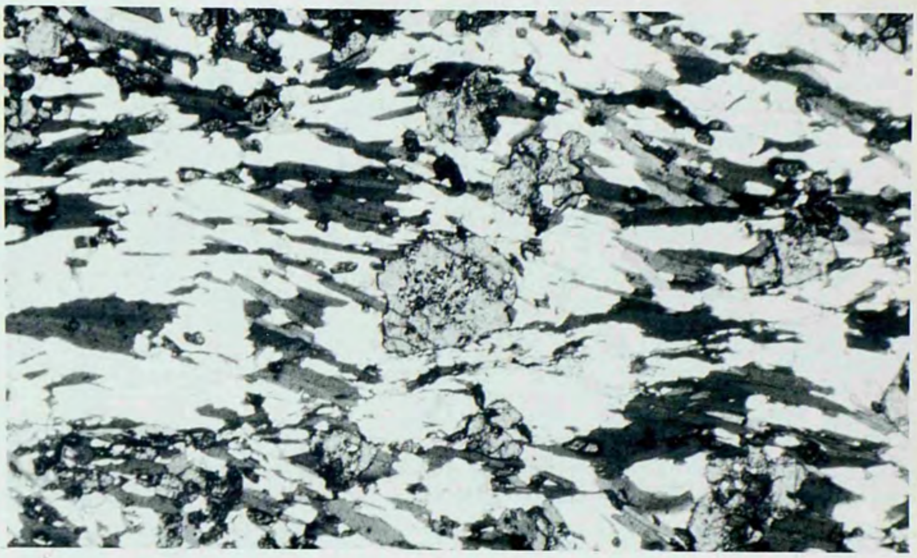
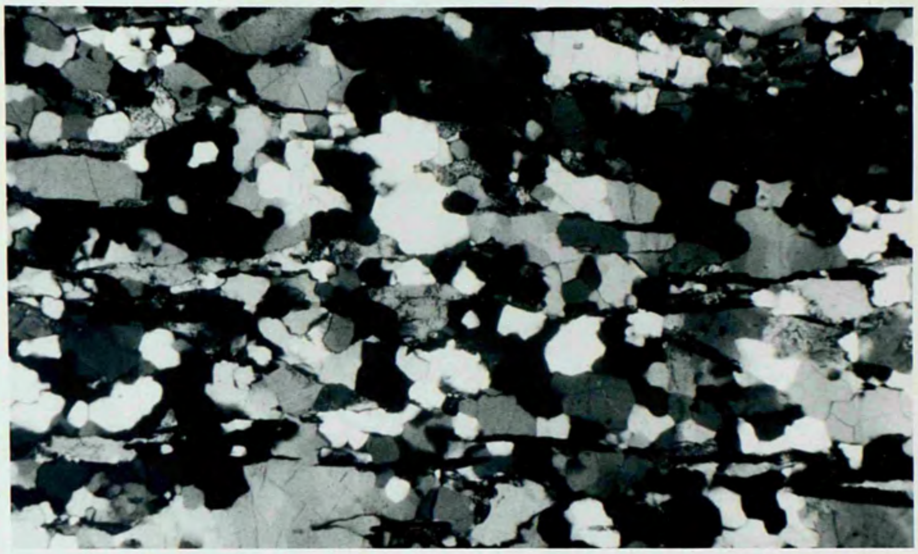


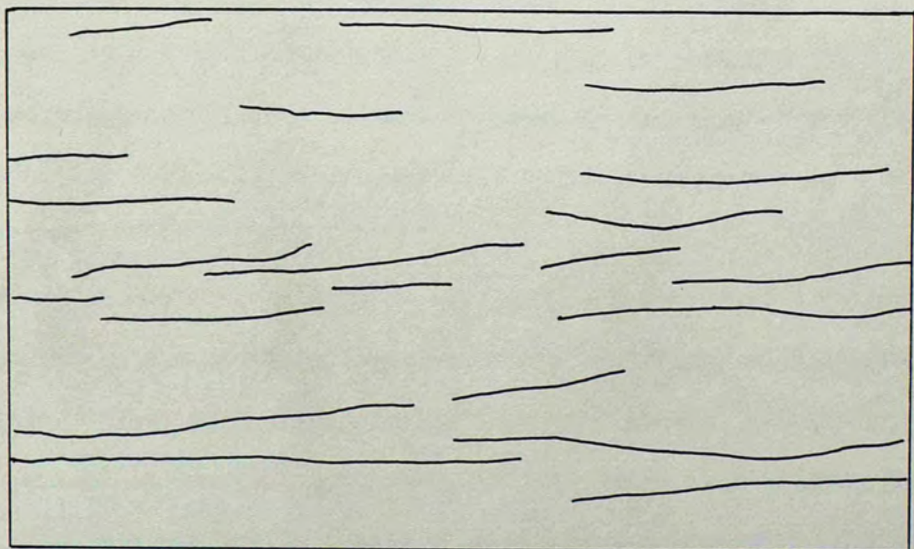
FIGURE 30.A psammite from the Inverbroom psammites showing granoblastic,equilibrium textures.

1mm

FIGURE 31. A highly deformed psammite from within the Meall a'Chrasgaidh psammite, adjacent to the slide.

1 mm





Parallel planes in the above figure defined by micas, straight grain boundaries and strings of small grains.

The psammites and pelites of the Fannich area have a very restricted range of mineralogy. They rarely develop the aluminosilicate minerals used elsewhere as indicators of metamorphic grade. This is also true of other metasedimentary lithologies in the Moine rocks of Scotland (Kennedy, 1949; Powell et al., 1981), their chemistry (Winchester, 1974) making them insensitive as indicators of metamorphic grade.

Microfabrics associated with the Sgurr Beag Slide are more difficult to study than those associated with the Moine Thrust belt for two reasons: firstly, the recrystallization under peak metamorphic conditions after movement on the Sgurr Beag Slide had ceased; secondly, later deformation associated with mylonitization at the Moine Thrust.

The second point is well illustrated by considering Figure 28. This shows microfabrics from the Inverbroom Psammite, first in Fannich, then as they are broken down towards the Moine Thrust. A rock at the same stratigraphical level as the first illustration forms a quartzofeldspathic mylonite just above the thrust.

Microfabrics due to the Sgurr Beag Slide were investigated in the Fannich area, away from the mylonites, using specimens which had suffered least deformation during D3 (i.e. away from major F3 hinge zones).

Psammites

Psammites in the Fannich area develop only a very restricted mineralogy, lacking in most of the aluminosilicate minerals used to determine metamorphic grade. They do however develop almandine garnet, indicating that they reached at least almandine-amphibolite grade conditions. The general mineral composition is quartz + plagioclase + muscovite + biotite $\pm$ apatite + opaques. Quartz and feldspar characteristically comprise around 75% of the total rock. When garnet is present (up to 5%) it shows inclusion trails of finer grain sizes than the matrix and concentric growth zoning (fig.29). Garnets include quartz, feldspar, sphene, apatite and occasionally biotite, but there are no other obvious intermineral relationships. Muscovite and biotite occur in approximately equal proportions of 5-10% characteristically intergrown and inclusion free.

Psammites over 1000m structurally below the Sgurr Beag Slide exhibit microfabrics which are closer to a stable equilibrium state than others in the Fannich-Ullapool area (fig.30). Both quartz and feldspar are internally relatively undeformed, exhibiting only slight strained extinction. Grain sizes range from 50-500μ with a mean of around 250μ and a granoblastic texture is characteristic. Micas 200-750μ long with aspect ratios of 2:1 to 6:1 exhibit a weak alignment parallel to the bedding foliation planes (fig.30) and no detectable internal deformation.

In contrast, closer than 300m to the slide zone psammites of the Meall a Chrasgaidh psammite exhibit a very strong fabric which is obvious in hand specimen. The strong microfabrics (fig.31) correspond to strong planar and linear fabrics seen in the field (Section SBS.1.c).

A strong alignment of all micaceous (001) cleavage planes forms the foliation in the Sgurr Beag Slide zone (fig.31). Micas range in size from 100-1000μ having aspect ratios of 3:1 to 20:1, that is, much higher than those of the low strain psammite. Quartz and feldspar are slightly elongate, up to 750μ long with aspect ratios of up to 4:1. Their boundaries are serrated through processes of sub-grain and new grain formation. As such boundaries are not characteristic of the low strain psammite (fig.30) their formation is attributed to deformation resulting from movement on the Sgurr Beag Slide. Many of the large quartz grains show banded extinction and are broken into several sub-grains, both features being again attributable to slide movement.

It can be seen (fig.31) that the orientation of grain boundaries and thus the shapes of quartz grains in the highly deformed psammite are largely controlled by the alignment of micas. The grain boundaries are predominantly at high angles to adjacent mica surfaces and further, micas are never found totally enclosed by quartz or feldspar grains. The micas "pin" boundaries of the quartz and feldspar grains. Such features imply that post-deformation grain boundary migration was restricted and that quartz and feldspar have been prevented from reaching a larger grain size equidimensional, granoblastic texture. Similar textures are observed in the mylonites above the Moine Thrust, where micas "pinned" the boundaries of quartz and feldspar grains (Section MT.5.a). It was implied, from the discussion in Section MT.5.a, that quartz and feldspar have not undergone massive recrystallization and grainsize increase since mylonitization ceased due to inhibition by the micas. Therefore it seems possible that quartz and feldspar in the Sgurr Beag Slide zone have not undergone massive grain growth and requilibration since the deformation, because the strong alignment and high aspect ratios of the micas, inhibits recovery recrystallization.

C-axis fabrics for quartz were studied for a low strain psammite (SKND27) taken from an outcrop containing practically undeformed cross-bedding structures in the Inverbroom psammites and a high strain psammite (SKND117) taken from the Meall a Chrasgaidh Psammite (Morar Division) less than 1m from the Meall an t'sithe Pelite (Glenfinnan Division) (fig.32).

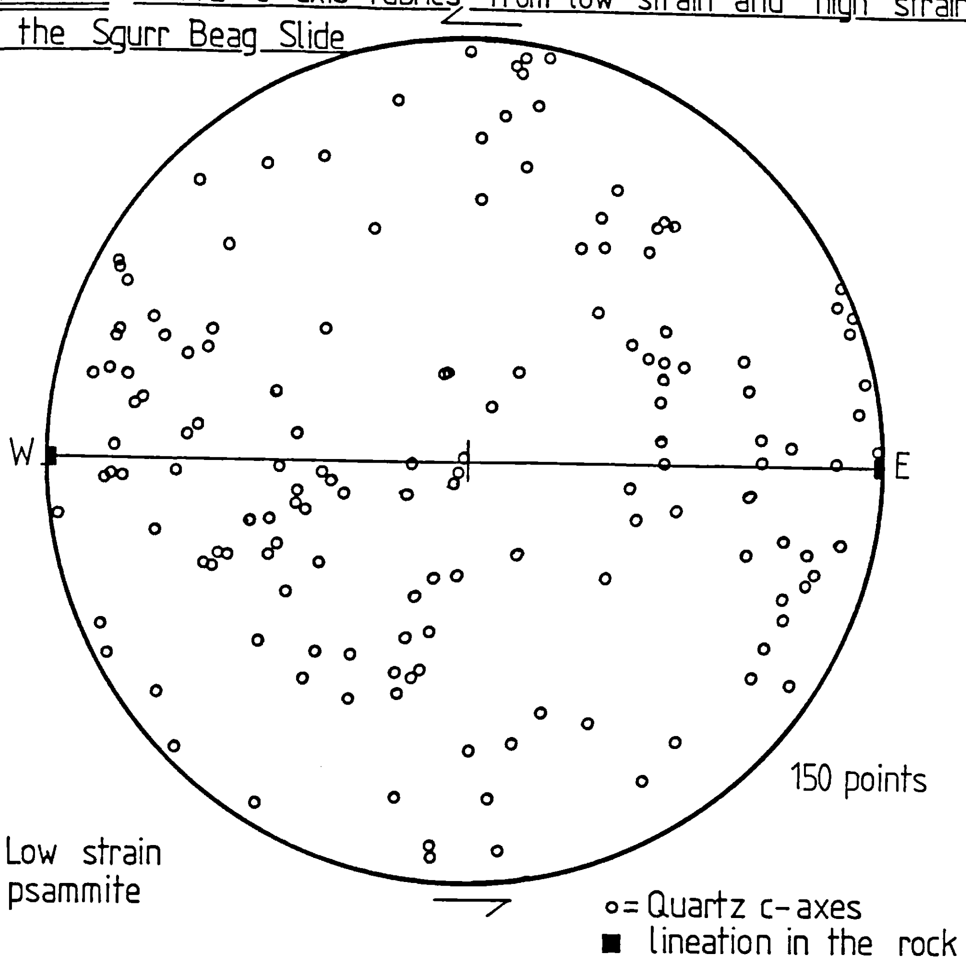
The low strain psammite gave a near random pattern of quartz c-axis orientations (fig.32). The high strain psammite however, exhibits a strongly developed asymmetrical single girdle fabric very similar to those interpreted as forming during shear deformations (Lister & Williams, 1979), those previously reported for Moine mylonites at Eriboll (White et al., 1982) and seen in mylonites west of Fannich (figs.32,87). The high strain sample was taken from a flat lying area of the Sgurr Beag Slide zone west of Fannich but at a distance of over 10km from the Moine Thrust zone in order to minimise post Sgurr Beag Slide deformation. It does not show features related to the mylonite deformation phase, indeed the trend of the mylonite (X) extension direction is oblique to the X direction than can be deduced from the plot. The X direction deduced from the plot coincides with the (L2) extension direction of the Sgurr Beag Slide deformation.

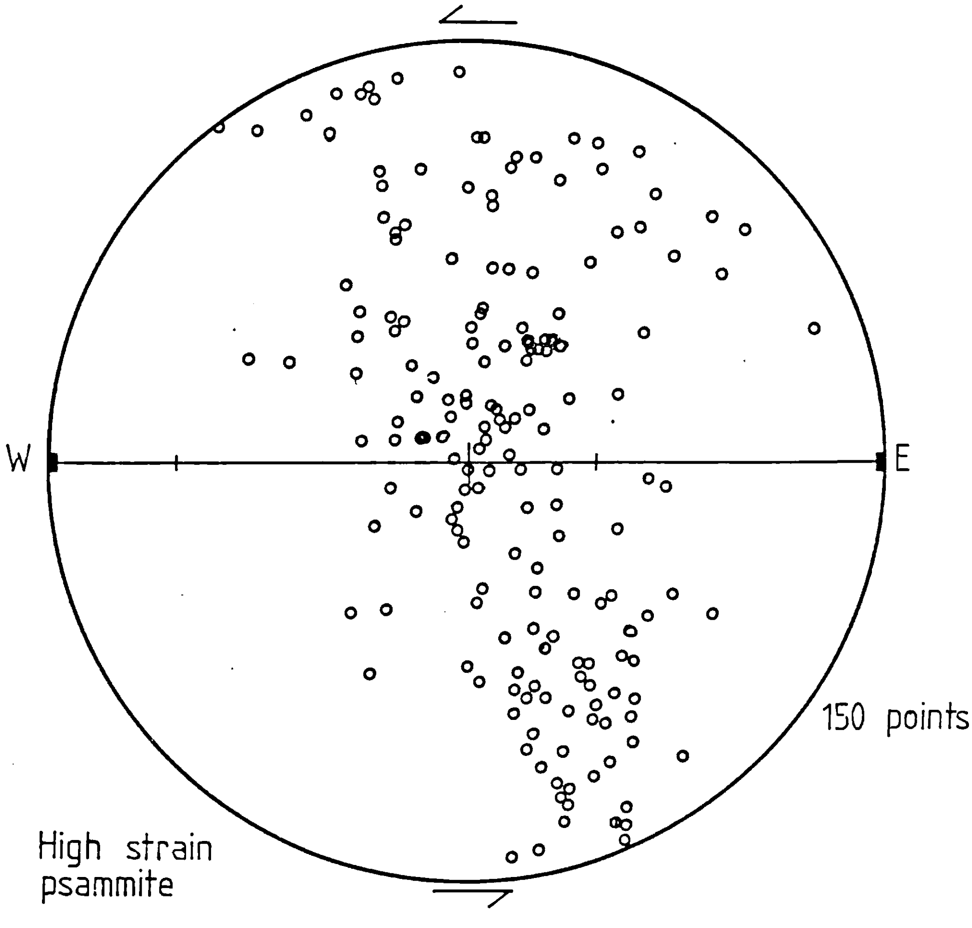
The pattern shows more spread than that obtained from the mylonites west of Fannich (figs.32,87) and this may be due to disruption of the pattern during post-Sgurr Beag Slide deformation.

If the direction of movement deduced for the mylonite quartz c-axis patterns in Eriboll (White et al., op.cit.) and west of Fannich (Section MT.5.a) is correct and generally applicable, that is girdle asymmetry indicates simple shear and is spatially related to the orientations of X, Y and Z strain axes, then it follows that movement upon the Sgurr Beag Slide involved overthrusting towards the NW. Unfortunately however, not all studies of asymmetrical girdles agree on the implied movement direction (Bouchez & Pecher, 1976; Carreras et al., 1977; White et al., 1982).

Most single girdle c-axis fabrics for quartz in the literature are reported for greenschist facies mylonitic rocks; however in this study, such a fabric is developed in psammites which were deformed under amphibolite facies conditions. The apparent absence of similar fabrics relating to the Sgurr Beag Slide elsewhere must be partly a consequence of lack of investigation but also, for example in the SW Moine, to post-sliding recrystallization (Baird, 1982).

FIGURE 32. Quartz c-axis fabrics from low strain and high strain zones of the Sgurr Beag Slide





Pelites

Pelites suffer the same restricted mineralogy as psammites but with proportionally larger amounts of micas and garnet (up to 50% micas). Winchester (1974a,b) has shown that the aluminosilicate minerals are rare because the bulk CaO/Al_2O_3 ratio for pelites is outside the field of alumino-silicate stability at amphibole grade conditions, aluminosilicates rarely developing in CaO-poor lithologies. However, rare occurrences of kyanite have been reported from migmatites of the Glenfinnan Division rocks in the west of the Fannich area (Sutton & Watson, 1954; Winchester, 1970).

The Glenfinnan pelites have a coarse migmatitic lit par lit texture. Quartz and plagioclase feldspars up to 2000μ in diameter, micas up to 3000μ in diameter showing coarse-grained non-directional inclusions and occasional compositional zoning (fig.33).

Morar Division pelites show little evidence even of incipient migmatization (fig.34). Quartz and feldspar are generally 250-750μ in diameter, micas up to 1000μ long with aspect ratios 3:1 to 6:1 and garnets generally less than 500μ in diameter. Inclusions within garnet are fine grained and directional; they also often show compositional zoning (fig.29). From the mineral compositions (above) it can be seen that all the rocks reached at least almandine amphibolite grade conditions.

Garnets are seen within pelites of the Morar Division throughout the traverse from Fannich to the Moine Thrust. They have a similar habit and structural age along the whole traverse. This is an important point since it shows that the whole of the Moine rocks from Fannich to the Moine Thrust was metamorphosed to amphibolite grade conditions during peak metamorphism. Therefore the margin of the Scottish Caledonian orogen at the Moine Thrust is dissimilar to that at the Highland Border Fault.

FIGURE 33.Garnets within the Meall an t'sithe pelite(Glenfinnan division migmatites) 1mm

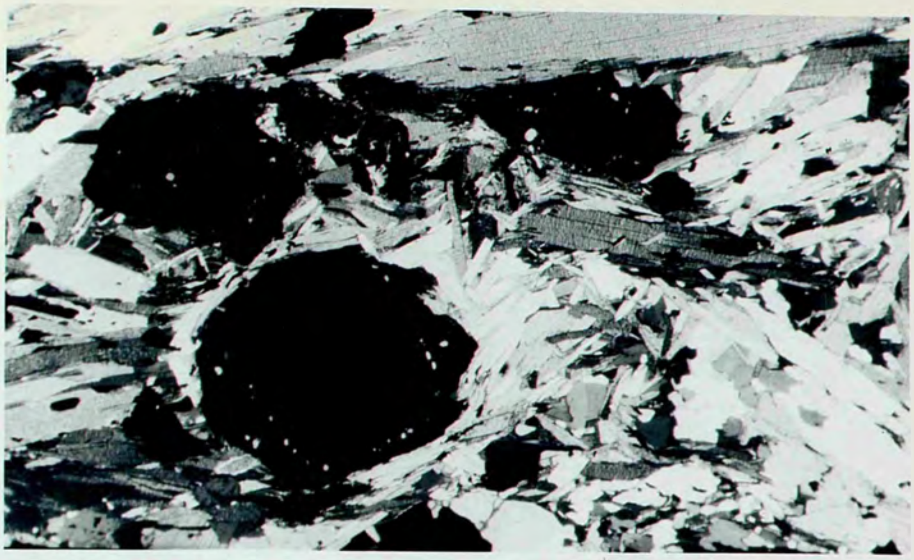


FIGURE 34.An exposure of Morar pelite,no sign of migmatisation(Sgurr Mor pelite)

FIGURE 5. Highly deformed migmatites adjacent to the slide

Migmatites away from the slide are illustrated in figure 4d)

Garnets present within the pelites of Fannich exhibit grainsizes of 100-500μ, they contain inclusions of quartz and feldspar often orientated but not rotational. The inclusions are always fine grained and often contain fine opaque dusty minerals in trails, which are not present in the surrounding rock (fig.29). The garnets can be traced continuously from Fannich to the mylonites and are resistant to the deformation (Section MT.5.a). They retain the fine inclusion trails which, adjacent to the Moine Thrust, can be misinterpreted since they are effectively of similar grainsize to groundmass quartz and feldspar (fig.83). However, they are always helicitic to the mylonite fabrics.

A contrast in metamorphic textures between Glenfinnan Division pelites (Meall an t'sithe Pelite) and those of the Morar Division (Sgurr Mor Pelite) is particularly obvious in Fannich (see above). Both pelites contain dominantly plagioclase feldspar and only trace K-feldspar. The plagioclase composition of both Meall an t'sithe and Sgurr Mor Pelites is in the low andesine range. Winkler (1976) has shown that the formation of migmatites is dependent upon temperature, availability of water, the composition of plagioclase and the ratio quartz : alkali feldspar : plagioclase. If no tectonic break were present between the two divisions in Fannich and they had therefore been metamorphosed under similar conditions, the Morar Division pelites should be significantly less susceptible to migmatisation in order to produce the observed difference in textures.

However, this is not the case. Both divisions contain pelites with low K-feldspar content and plagioclase with compositions in the Andesine range. Therefore, it seems that they were not metamorphosed under similar conditions but were brought together at a later stage.

Neither the relative or absolute ages of metamorphic activity in rocks lying above and below the Sgurr Beag Slide are known. No evidence has been presented here to confute even the suggestion that metamorphism above the slide zone and below it are completely unrelated. However, work from other areas of the Moine may shed light upon the problem. Shepherd (1973) showed that Morar Division rocks did not reach amphibolite grade conditions at any time before the intrusion of the Carn Chuinneag granite (Section SBS.5) at $555^{+}_{-}10$Ma (Pidgeon & Johnson, 1974) and the main amphibolite regional metamorphism is therefore Caledonian in age. Brewer et al. (1979) showed that metasediments in the SW Moine area underwent a major regional metamorphic event at $1004^{+}_{-}28$Ma. The grade of this 'Grenville' event increasing from greenschist in Skye, eastwards, to upper amphibolite conditions in Ardgour. It is therefore tempting to suggest that the Fannich and Carn Chuinneag areas illustrate the supposition of terrains which were separate during the 'Grenville' orogeny and brought together during the Caledonian orogeny.

To the west of Fannich, towards the Moine Thrust, pelites rapidly become deformed in the mylonite related deformation (Section MT.5.a), making it difficult to reconstruct the D2 Sgurr Beag Slide fabrics. Likewise, in Fannich the problem is almost as acute, because in approaching the Sgurr Beag Slide zone from the Morar Division, the core of the F3 Fannich synform is also approached (fig.4). The D2 (i.e. Sgurr Beag Slide) fabrics are strongly crenulated as a result of D3. The pelite fabrics are extremely inhomogenous, reflecting the variation from moderate to intense D3 overprint in different areas. Further, in the zone up to 300m from the slide (the Meall a Chrasgaidh Psammite) only psammitic and semi-pelitic lithologies are present. Thus differences in fabrics related to movement on the slide zone are not seen in Morar division pelitic rocks.

Within the Glenfinnan Division, D3 fabrics are relatively less intense in the Fannich area. This is because the Glenfinnan Division just above the slide zone lies in the less deformed vertical limb between the Fannich synform and Achnasheen antiform. A traverse from migmatites showing long undulose lits to those forming an augen gneiss adjacent to the slide (fig. 5d and 35) illustrates changes which can be attributed to D2.

Away from the slide, it is difficult to separate effects due to the Sgurr Beag Slide (D2) from earlier phases of deformation. Minor F2 folds deform the migmatitic lit par lit fabric and contain an axial planar fabric (Section SBS.3.a & fig.36). The planar fabric shown by micas curves around the migmatitic lits and is parallel to lithological banding. The mica fabric is a composite of S_0? (sedimentary fabric) S_1 and S_2. The preferred orientation is moderately strong but tends to be disrupted where micas are broken and deformed close to garnets (this is possibly due to D3). The garnets showing concentric zones of inclusions are not wrapped by S_2, but they have an 'eroded' appearance - broken zoning within the garnets, irregular shapes and rims of opaque minerals (fig.37) all indicate multi-phase history. Quartz and feldspar remain as large laths 500-2000μ in diameter, showing strained extinction and rare sub-grain formation.

The augen gneisses within the slide zone exhibit stronger mica fabrics and garnets show signs of having been wrapped by micas as a result of D2 deformation. Quartz is found in generally smaller grainsizes, 100-250μ in diameter, exhibiting undulose extinction and sub-grain formation. The grain boundaries are serrated, indicating non-equilibrium conditions but it is unclear how much of this deformation in quartz is due to D3. Feldspars form large strained laths up to 2000μ in diameter which become boudinaged, showing brittle fracture, undulose or patchy extinction and

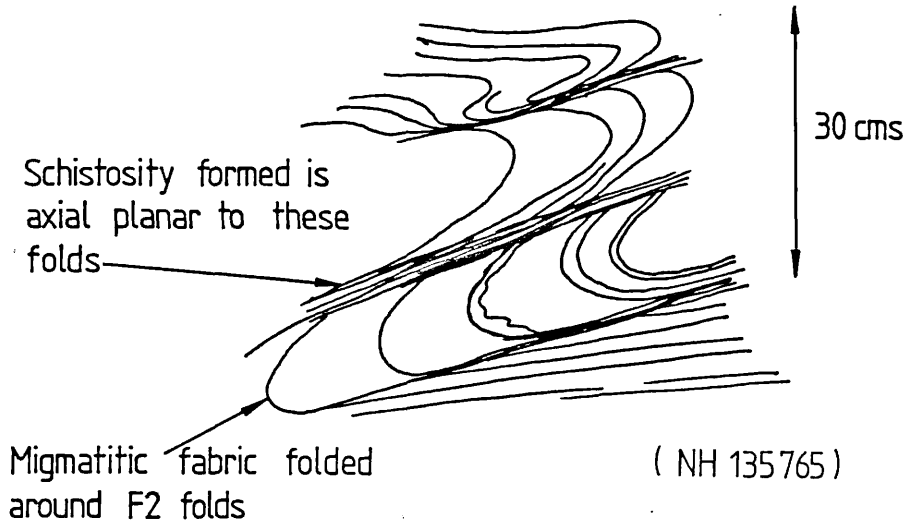
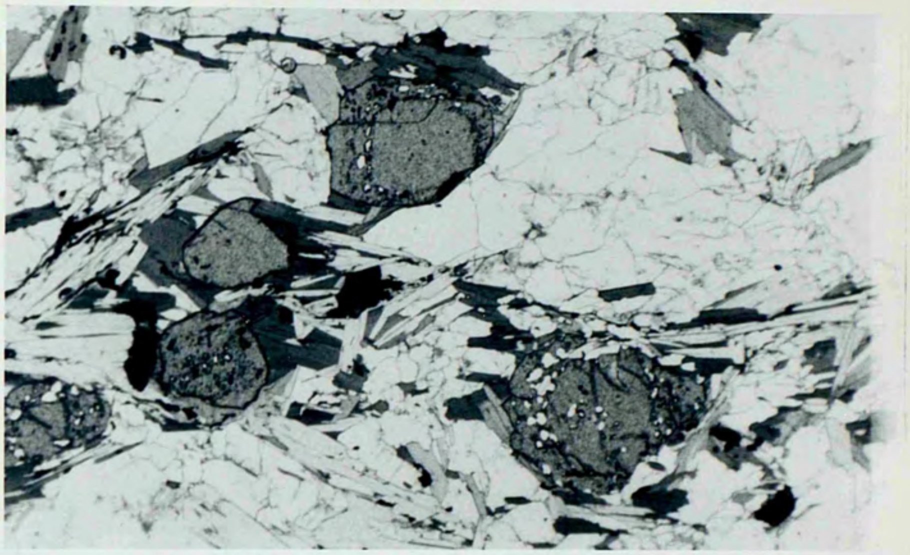


FIGURE 36. F2 minor fold reworking migmatitic fabric

<u>FIGURE 37.</u> Examples of garnets with broken growth zones from migmatites in the slide zone.

1 mm



new grain growth. This is especially strong within quartzofeldspathic augen; it is attributed to D2 deformation and can be observed in the field (fig.8).

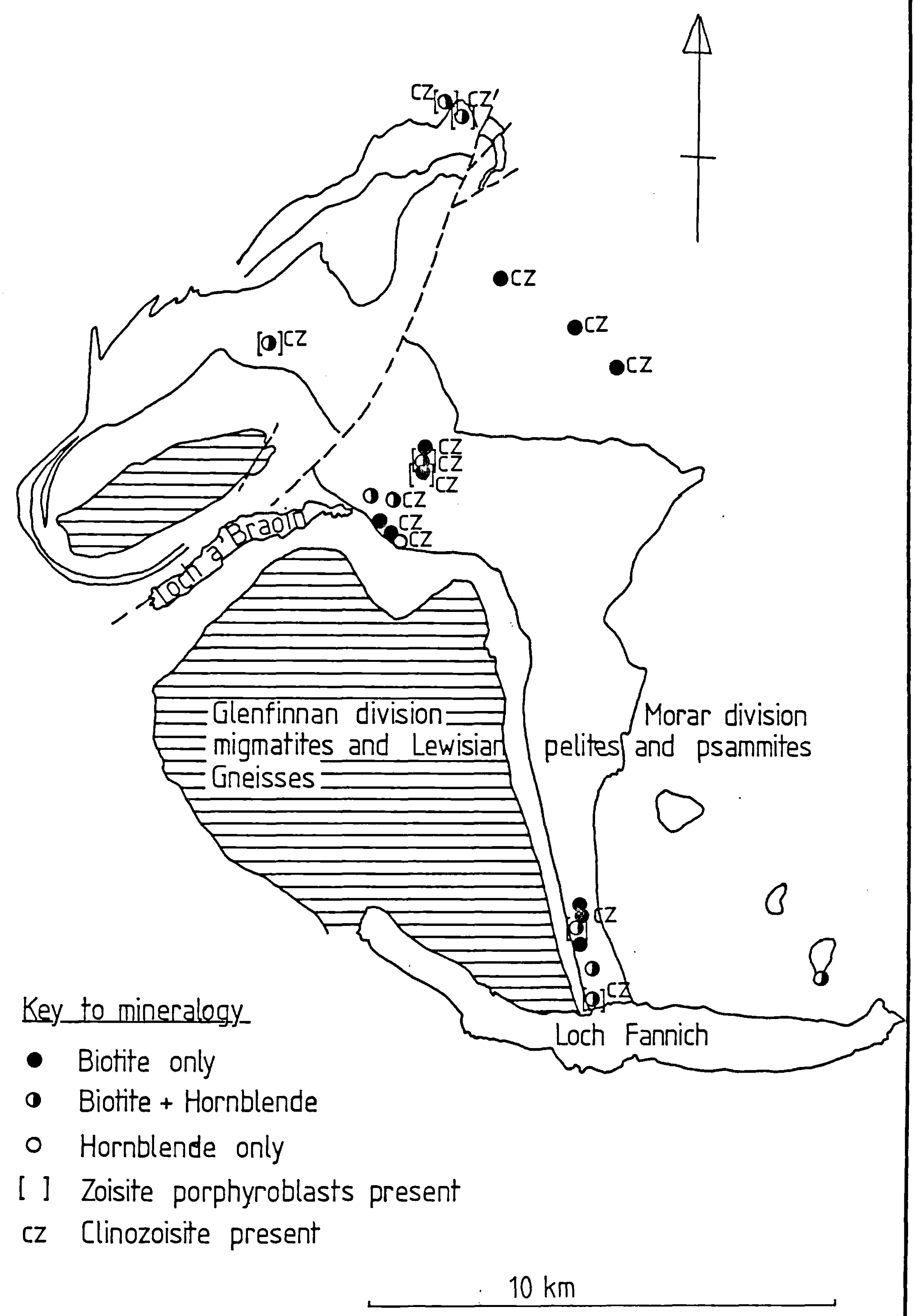
In conclusion pelitic rocks, in general, have been much more strongly affected by D3 and the mylonitization event than psammites. Determining their pre-D3 fabrics is therefore difficult, except in the migmatites of the Glenfinnan Division where features attributable to reworking towards the slide can be related to Sgurr Beag Slide (D2) deformation.

SBS.2.b Metamorphism of the Calc-silicate rocks

Psammites and pelites of the Fannich area are insensitive as indicators of metamorphic grade. However, the calc-silicate bands have a varied mineralogy and chemistry which in Fannich, as elsewhere, has provided evidence for changing metamorphic conditions (Kennedy, 1949; Winchester, 1972, 73, 74; Tanner, 1971; Soper & Brown, 1971; Powell et al., 1981).

Examination of the microtextures and microfabrics of the calc-silicate specimens has revealed a two-phase metamorphic history: Early stage assemblages, with high grade metamorphic textures, exhibiting strong mineral fabrics associated with Sgurr Beag Slide deformation, and later stage assemblages exhibiting near random fabrics post-dating D2 minor folds. In view of this separation, the early textures and fabrics will be discussed in this Section whereas the later textures and fabrics will be discussed in the Moine Thrust Chapter (Section MT.4.b). Both the field relationships and general mineralogy of the calc-silicates will be discussed in this Section.

FIGURE 38. Sample localities of calc-silicates



Occurrence and general mineralogy

Calc-silicates occur as discontinuous bands or lenses sporadically developed within lithologies comprising the Morar Division. They are most common however in more pelitic horizons. Generally 2-5cm thick and laterally continuous for 10cm to 5m. They are characterized by cream/white or pea green colours and distinctive large (up to 5mm) garnet porphyroblasts.

The bands are most common in the Sgurr Mor Pelite, occurring much less frequently in the Inverbroom Psammites wherein they are epidote bearing (i.e. of Arnapol type; Read, 1934). Epidote bearing calc-silicates are also extremely rarely found in the Meal a Chrasgaidh Psammite. No calc-silicates were found above the Sgurr Beag Slide, in Glenfinnan Division rocks of this area either by the author or previous workers. This is in strong contrast to their abundance in the Glenfinnan Division rocks of the south-western Moine (Ramsay & Spring, 1962; Powell, 1964; Tanner, 1971).

A total of 36 calc-silicate samples were collected (fig.38) which show a wide range of mineral assemblages:

a) Quartz + plagioclase + biotite + garnet + zoisite $\pm$ clinozoisite.

b) Quartz + biotite + garnet $\pm$ zoisite $\pm$ clinozoisite.

c) Quartz + plagioclase + biotite + hornblende + garnet $\pm$ zoisite $\pm$ clinozoisite.

d) Quartz + biotite + hornblende + garnet $\pm$ zoisite $\pm$ clinozoisite.

e) Quartz + plagioclase + hornblende + garnet $\pm$ zoisite $\pm$ clinozoisite.

f) Quartz + hornblende + garnet $\pm$ zoisite $\pm$ clinozoisite. (Not found during the present survey but reported by Winchester, 1972).

g) Quartz + plagioclase + epidote + biotite $\pm$ zoisite $\pm$ clinozoisite.

In contrast to calc-silicate rocks described from the SW Moine, those of the Fannich have never been found to contain pyroxene (Kennedy, 1949; Charnley, 1976; Powell et al., 1981) 21 of the 36 calc-silicates collected were free of epidote but only 9 of those contained plagioclase, hence a study such as that of Powell et al. (1981) is impractical in this area. This is possibly due to the generally higher CaO/Al_2O_3 ratios of the Fannich calcs (Winchester, 1970, 72) causing a predominance of the epidote group of minerals.

The chemistry of the calc-silicate samples collected during this study has not been studied since the subject has already been extensively examined by a previous worker in the area (Winchester, 1970, 72, 73, 74,a,b).

Mineralogy of the calc-silicates

Biotite: is found in two distinct relationships, as single grains and small aggregates of grains dispersed throughout the calc-silicate, or as concentrated bands of large biotite laths at the edges of some calc-silicates. They are always fox brown in colour and exhibit slight to total alteration to chlorite.

Hornblende: occurs as sub-hedral to anhedral single grains 250-1000μ long which are rich in quartz and feldspar inclusions and occasionally intergrown with biotite. Alteration to chlorite is common in all sections but replacement by biotite has not been seen (fig.39). Amphibole also occurs as large (up to 2cm long) porphyroblasts in association with large zoisites.

Garnet: is almost always present as large grains, 500-5000μ in diameter. Despite being full if inclusions of quartz, feldspar, sphene and occasionally calcite they generally maintain sub-hedral to anhedral outlines (fig.40).

FIGURE 39. Biotite and hornblende co-existing in one thin section. Both are being altered to chlorite.

1 mm

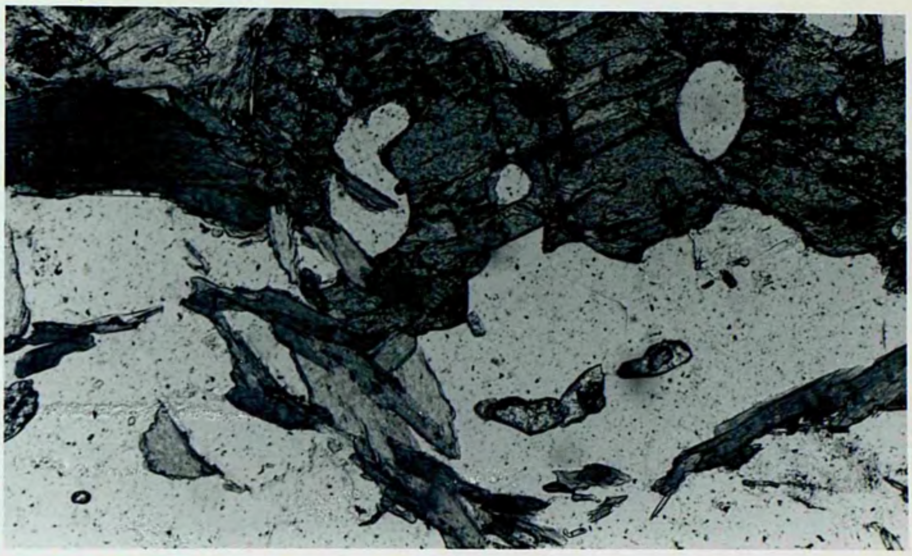


FIGURE40. Sub-hedral garnets within a calc-silicate of the Sgurr Mor pelite.

1 mm

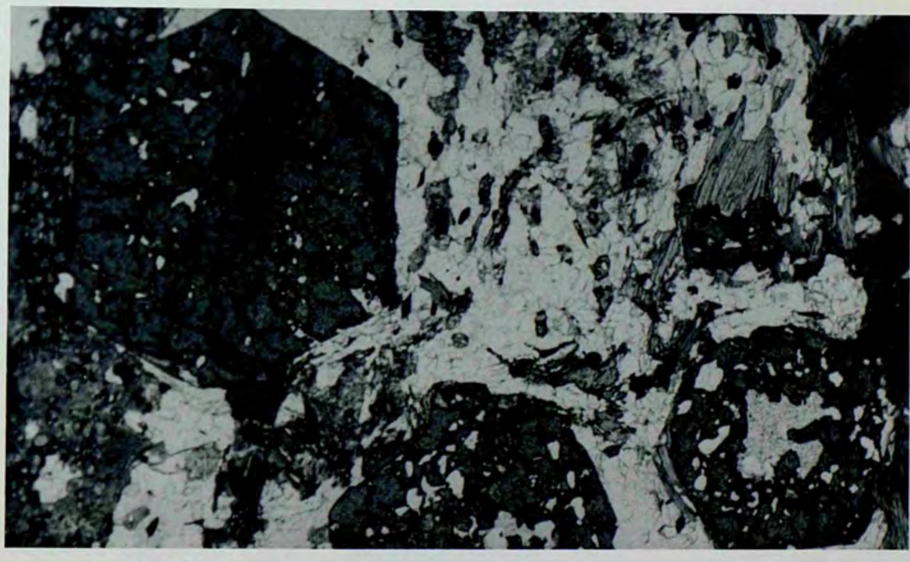


FIGURE 41.Peak metamorphic zoisites in a calc-silicate.Note.The garnet has suffered retrogression.

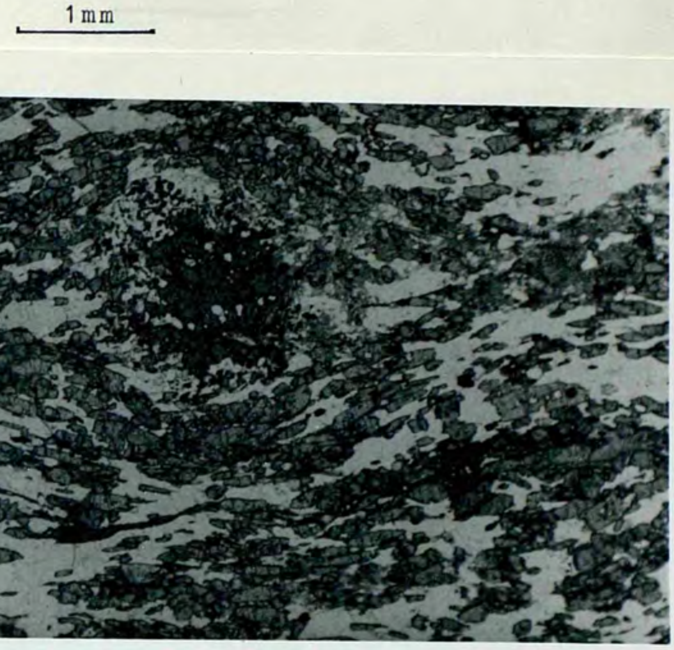
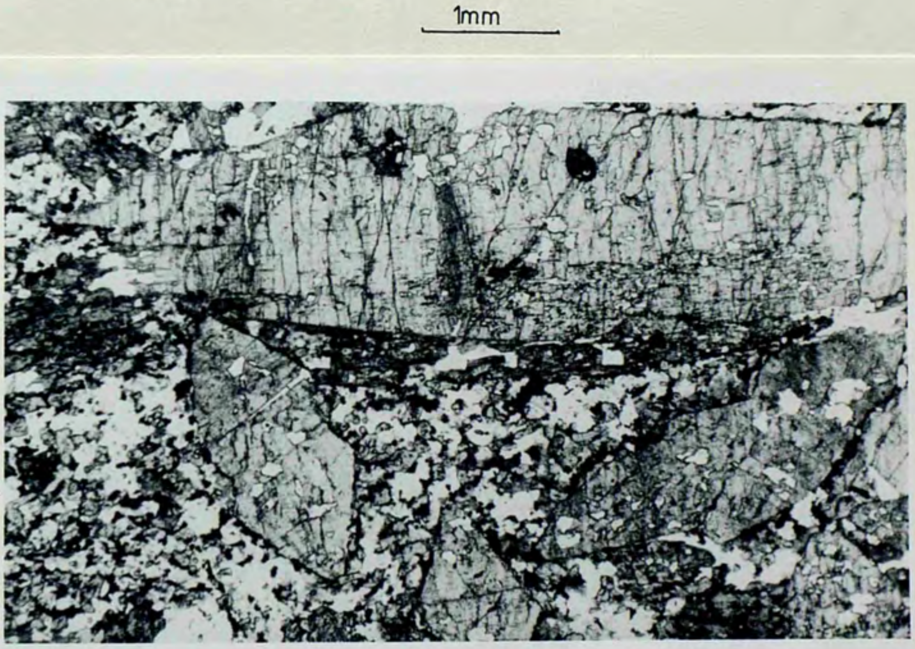


Figure 42.Late zoisites in a calc-silicate.
a)Sub-hedral grains.



b)Late zoisites showing almost skeletal shapes.

1 mm

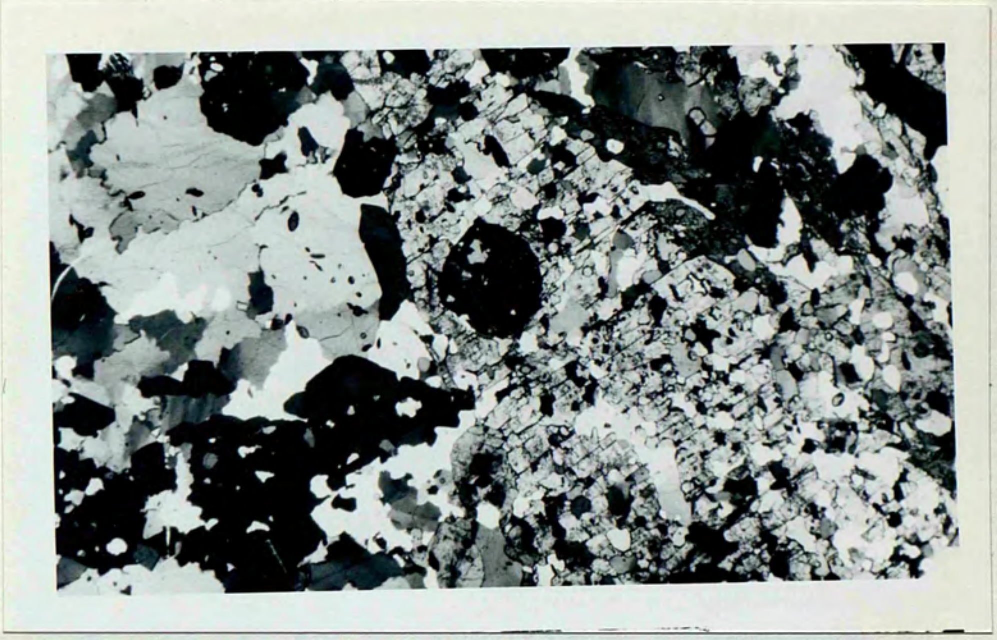
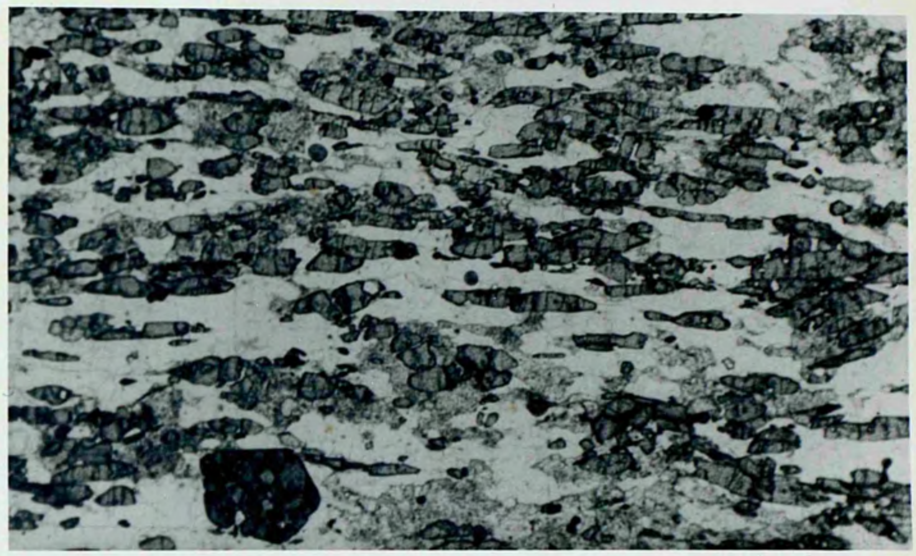


FIGURE 43. a strong extension lineation is seen in many calc-silicates. Here the lineation is defined by the long axes of zoisites.

1 mm



Plagioclase: occurs as anhedral grains 250-500μ in various stages of alteration to sericite and clay mineral aggregates. Optical studies suggest compositions varying from An 39 to An 66 although there is insufficient data to determine whether such variation conforms to that observed by Powell et al. (1981). Plagioclase occurs in less than half the epidote-free calc-silicates in contrast to those seen in Morar.

Quartz: occurs in a similar habit to plagioclase and exhibits near granoblastic textures. This contrasts with adjacent pelites which show a greater degree of grainsize reduction due to the initiation of mylonite related deformation. In some calc-silicates, in the west, quartz shows strained extinction and sub-grain formation at the edges of the grains.

Zoisite: occurs in several different habits; as concentrically zoned stubby grains similar in shape to epidote; as long prismatic grains with subhedral to anhedral shapes and strong preferred c-axis orientations (fig.41); and as late porphyroblasts up to 5 cm long which overgrow peak metamorphic minerals such as garnet, plagioclase, biotite and clinozoisite, but is occasionally intergrown with large hornblende laths (fig.42)

Clinozoisite: is found in two habits: as small, occasionally zoned equigranular grains about 250μ in diameter dispersed throughout the calc-silicate or as large aggregates pseudomorphing garnet.

[Zoisite and clinozoisite were distinguished in this study on the basis of the anomalous blue interference colours of clinozoisite and the inclined extinction of clinozoisite compared with the straight extinction of zoisite.]

Epidote: occurs in calc-silicates which lack or contain little garnet. It is often associated with zoisite and clinozoisite as subhedral to anhedral, stubby grains 250-500μ in length. It is generally found in the calc-silicates of the Inverbroom Psammites.

Early high grade microtextures and microfabrics

The metamorphic microfabrics (deformation related) and microtextures (metamorphic related) are discussed under two headings: Early high grade textures and fabrics (see below) and Late low grade textures and fabrics (Section MT.4.b.iii).

The interpretation of metamorphic history presented here differs from that postulated by Winchester (1970, 72, 74).

Many of the calc-silicates exhibit a strong extension lineation defined by the alignment of elongate grains of feldspar, zoisite and hornblende (fig.43). They also often exhibit a strong planar fabric defined mainly by biotites which (except in the cores of early folds where it is axial planar) is parallel to the lithological (sedimentary/diagenetic?) zoning within the calc-silicates. The linear and planar fabrics parallel L2 and S2 respectively, hence the fabrics are thought to be related to D2, Sgurr Beag Slide deformation.

There is no obvious textural evidence within the calc-silicates to suggest a time sequence for the development of the constituent minerals of the peak metamorphic assemblages. Systematic inclusion relationships are not found except in garnet, but clearly the alignment of minerals took place essentially syntectonic with D2.

The most important mineral relationship in the calc-silicates of the Fannich area is between biotite and hornblende. Kennedy (1949) used the breakdown reaction of biotite to define the upper limit of his lowest prograde metamorphic zone, the biotite-calcite-zoisite zone:

biotite + calcite (zoisite) → hornblende.

Winchester (1972) states that in Fannich, where biotite and hornblende are found co-existing in the same rock, there is always evidence of the breakdown of hornblende:

hornblende + microcline → biotite + Anorthite + quartz.

In the present summary however, biotite and hornblende were found together in 43% of all the calc-silicates collected and in all cases, the textures seen indicated that hornblende and biotite were co-stable at peak metamorphism. The two minerals are occasionally seen intergrown (fig.41) but neither is obviously unstable or replacing the other. In some cases both are seen to be partly replaced by chlorite (see later in this Section for discussion).

Garnets are abundant in the calc-silicates, differing in habit from those found in adjacent pelites. They are generally large, up to 5mm in diameter, and exhibit sub-hedral to euhedral shapes. The garnets contain many inclusions, generally of quartz and feldspar, however these differ from the inclusion fabrics of the pelites in that the inclusions are of larger grainsizes, up to 200 μ diameter. Most inclusion fabrics were not orientated.

Discussion

Although the present survey has involved no chemical analyses of calc-silicate rocks, thin section and field evidence indicates that mineral fabrics and textures have not been adequately explained by previous work.

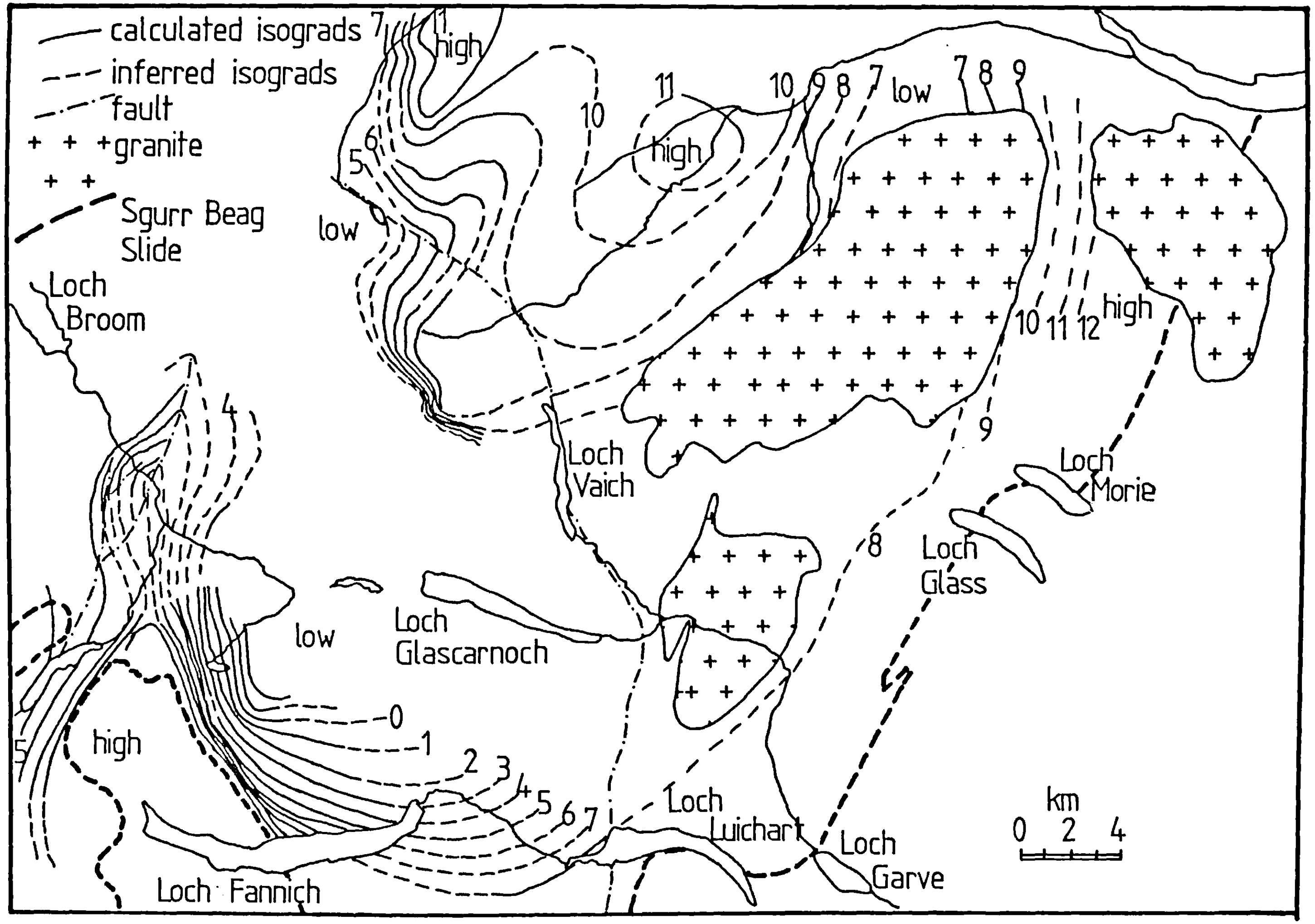


FIGURE 44. Isograds shown by calc-silicates, projected onto a level surface (after Winchester 1974)

Winchester (1970, 72, 74) defined a series of isograds using the transition between hornblende and biotite in calc-silicates for given CaO/Al_2O_3 bulk rock ratios. Hence by plotting the spatial distribution of CaO/Al_2O_3 values of "critical" calc-silicates containing both hornblende and biotite Winchester postulated inverted metamorphic zonation, with grade increasing toward the upper part of the succession (the Meall an t'sithe Pelite). He also postulated that the isograds had been folded by the F3 Fannich synform (fig.44; Winchester, 1974). Winchester's interpretation of the biotite/hornblende relationship as a retrograde replacement of hornblende by biotite meant that the inverted isograds had to be explained by invoking an earlier high grade metamorphic event succeeded by one of a lower grade whose effects were strongest in the lower parts of the succession.

In the present survey it has been noted that in all thin sections containing both hornblende and biotite, neither is obviously replacing the other, suggesting stable co-existence. It is also known from the work of Powell et al. (1981), that the emplacement of Glenfinnan Division migmatites over lower grade Morar rocks along the Sgurr Beag Slide at Lochailort had a profound effect upon metamorphism of calc-silicates in its proximity. The slide caused inversion of metamorphic isograds during emplacement. The isograds plotted by Winchester (1972) increase in grade towards the Sgurr Beag Slide (fig.44). However, no calc-silicate rocks have been found within the Glenfinnan Division migmatites of Fannich in order to confirm the model of Powell et al. (op.cit.).

Although both models (Winchester, 1972; Powell et al.,1981) fit the observed pattern of isograds, it appears that the second model is supported by the textural and microfabric evidence, ie,the strength of D2 fabrics in calcs and the biotite/hornblende relationship.

SBS.3 Pre-Sgurr Beag Slide History

SBS.3.a Pre-Sgurr Beag Slide deformation

The Fannich Area

In contrast with the SW Moine, within the Morar Division of Fannich evidence for pre-slide deformation is sparse. Rocks of the Glenfinnan Division show evidence of a pre-slide deformational history much closer to that seen within Glenfinnan Division rocks of the type area (Baird, 1982).

In Morar Division rocks of Fannich a schistosity defined by the preferred orientation of small muscovite and biotite (001) planes is folded by minor F2 and can be distinguished from S2 in the noses of D2 folds. The early fabric appears to have been planar and bed parallel. It is not possible to determine whether S1 was ubiquitous throughout the area since it is now obscured by the coplanar S2 fabric even where this is weakly developed away from the slide zone.

Only two exposures showing D1 folds have been found within Morar division rocks during this survey. Both are refolded by D2 folds. Other D1 folds may exist but they are extremely difficult to distinguish from early sedimentary fold structures. Any isoclinal fold structures in areas of obvious slumping and cross-bedding, or folds showing intricate fold styles, were regarded as being of sedimentary origin. The presence of axial planar fabrics as evidence of tectonic origin is a difficult criterium to apply here, since all the bed parallel isoclinal folds have an S2 fabric parallel to their axial plane.

One of the identified D1 folds lies within semi-pelite in Abhain Cuileig (NH 165753) and consists of a dome and basin refold pattern (fig.45). The other comprises a hooked, type 3 (Ramsay, 1967) pattern of interbanded semi-pelites and pelites about 2km NE of the other exposure in the Alt Leacach (NH 176768). In this example the F1 axial plane fabric is clearly seen refolded by F2 (fig.45). Both D1 minor folds are isoclinal with axial planes sub-parallel to bedding and hinges trending toward 190°. Precise orientations are difficult to measure due to poor exposure.

Evidence of an early D1 deformation phase is less scarce in the Glenfinnan division of Fannich. The coarse lit par lit migmatites have a strong S1 fabric defined by the alignment of leucosome and melanosome fractions of the migmatitic fabric, sub-parallel to the original lithological boundaries. This early fabric is clearly folded around D2 minor fold closures (fig.36).

The migmatites also become highly strained in the D2 Sgurr Beag Slide zone forming, by disruption of the lits, an augen gneiss (fig.8).

During this survey no folds were found which pre-date the migmatitic fabric. This may be because the Fannich outcrop of Glenfinnan Division rocks is restricted, being only 500m thick, and lies within an area strongly affected by the Sgurr Beag Slide. Earlier fabrics and folds may thus have been obliterated during slide movement.

Large scale isoclinal folds have been reported from the Lewisian rocks in the centre of the Fannich synform (Geol. Survey, 1912) and similar medium and minor scale folds were found during this survey. These had limbs up to 6m and axial planes parallel to the main schistosity within the Lewisian. In the hinges of these folds the schistosity is folded, thus providing evidence of a complex early history relative to the Morar Division rocks of the area. However, because an insufficient area of Lewisian gneisses was mapped, firm correlations with the deformation sequence affecting Moine schists cannot be made and it is not known whether the isoclinal fold structures are pre- or post-deposition of the Moine schists.

<u>FIGURE 45.</u> An F1 minor isocline refolded by a minor F2 fold.

An F2 fold deforming migmatite fabric is illustrated in figure 36.

The Lewisian gneisses

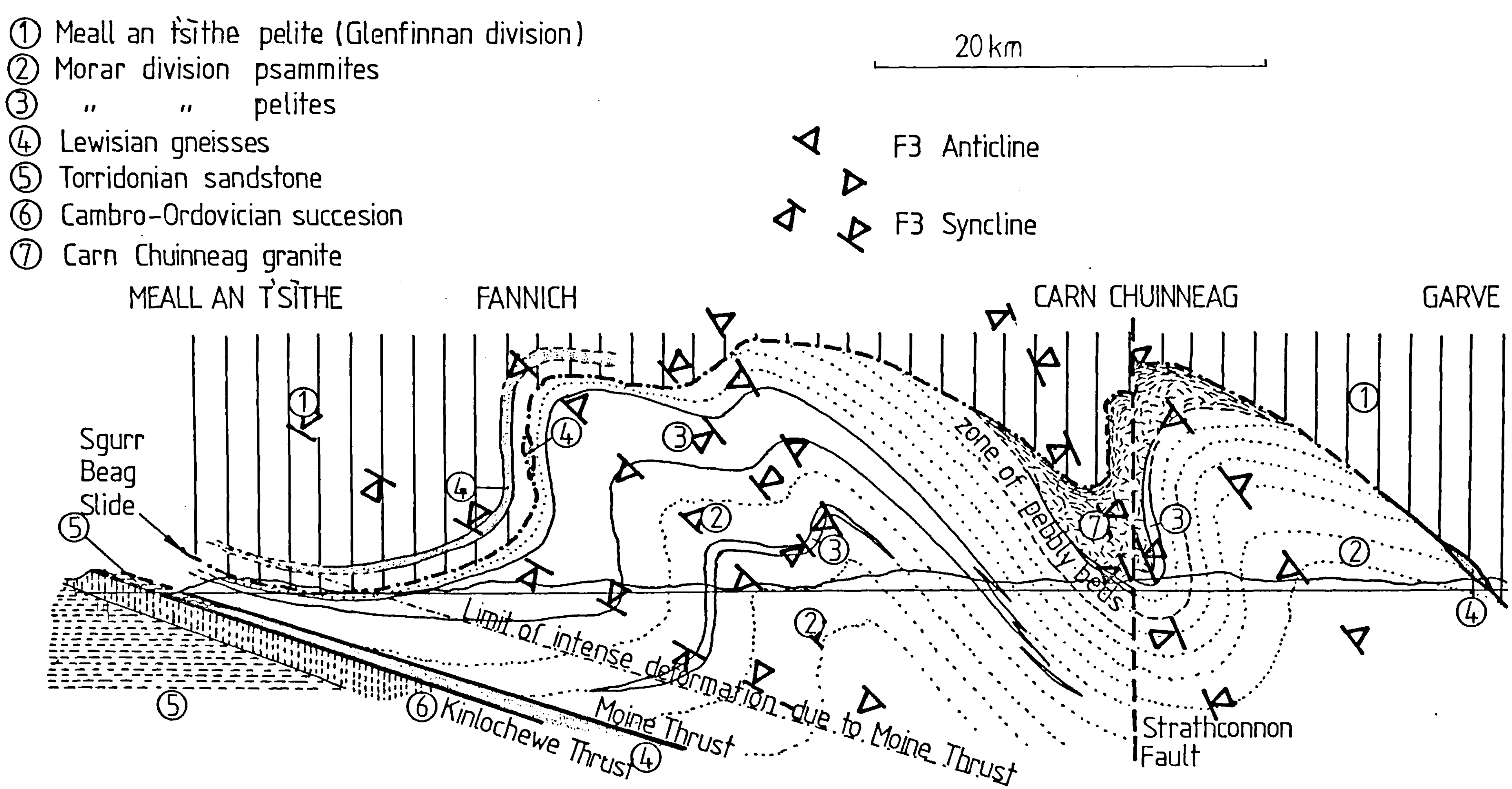
Whilst it seems clear that, as elsewhere, movement of the Sgurr Beag slide effected emplacement of high grade Glenfinnan Division rocks above rocks of the Morar division, the occurrence of small inliers of Lewisian gneiss within the Glenfinnan Division along the whole length of the Fannich-Beinn Dronaig Pelite belt cannot be directly attributed to movement on the slide zone. The inliers occur at a higher structural level than the Sgurr Beag Slide (figs. 3, 11) and have been variously interpreted as: pods lying upon a higher level slide plane (Tanner et al., 1970); the core to a "fountain of nappes" (Geol. Survey, 1912); and an isoclinal fold core (Winchester, 1970).

The idea of a 'fountain of nappes' has already been dismissed (Section SBS.1.a). The possibility of the Fannich Lewisian lying on a higher but similar age slide to the Sgurr Beag Slide is the less likely of the remaining two remaining hypotheses. The boundary between the Lewisian gneiss and Meall an t'sithe Pelite is not a simple planar structure but contains large D2 isoclines which fold it (Winchester, 1970,1973). This indicates either: a) that there has been relatively little movement between the two lithologies, or b) that any slide movement between the two pre-dates movement on the Sgurr Beag Slide.

The Carn Chuinneag area

The areas of Carn Chuinneag and Fannich lie adjacent to each other and no major discontinuity is found between the two. The structural and metamorphic histories are therefore very similar and the well dated Carn Chuinneag granite (555$\pm$10Ma) can provide an important constraint to the absolute time of early deformation and metamorphism for the Fannich area.

FIGURE 46. Cross-section from the Moine Thrust to Garve.



Recent authors (Rathbone, 1982; Wilson & Shepherd, 1979) agree that the earliest and main deformation affecting the Carn Chuinneag granite is of the same age as movement on the Sgurr Beag Slide (Section SBS.5).

De L'Apparant (1935) was the first author to conclude that granite intrusion post-dated an early deformation, describing folded schists from xenoliths within the granite. Subsequently Harker (1970) described folds within the aureole which pre-dated intrusion and the structural analysis of Wilson and Shepherd (1979) recognised one deformation phase prior to intrusion.

As in Fannich, no major D1 folds have been described in the Carn Chuinneag area. In Morar Division rocks around Carn Chuinneag, rare, sporadic minor D1 folds are described (Harker, 1970; Wilson, 1975; Shepherd, 1970,1973). Shepherd also describes medium scale upright, tight to isoclinal D1 folds with N-S axial traces and sub-horizontal axes. All the large D1 folds described by Shepherd and the folds described by Harker (op.cit.) lie within the Upper Pelite (Section St.4.b) adjacent to the granite, within the contact aureole.

The possibility that these early (D1) folds were produced during a forceful granite intrusion and were subsequently flattened during D2 deformation has not been discussed by these authors and thus their origin must remain an open question.

Outside the aureole, Wilson and Shepherd (1979) describe only rare minor D1 folds with a weak bed parallel, axial planar, S1 fabric defined by the alignment of 001 mica planes. The muscovite and biotite grains may originally have taken up that orientation fabric at a lower grade as chlorites; there is no evidence to determine the exact grade of early metamorphism (see Section SBS.3.b). Shepherd also described an extension lineation which he ascribed to the development of the D1 fabric. Wilson (1975) described a pre-D2 lineation in pebbly beds of the Upper Quartzite but accepted that in most places L1 and L2 are inseparable. Wilson (op. cit.) also raised the problem of relating D1 in the Morar Division to D1 in the Glenfinnan Division, a problem that is also encountered in Fannich. Structures recognised as D1 within the Morar Division of Fannich are the same as those described by Wilson and Shepherd with the exception of the D1 folds described by Shepherd from the granite aureole.

D1 deformation in the Glenfinnan Division of the Carn Chuinneag area has been described by Wilson (1975). He states that although no major D1 folds were seen, the quartzofeldspathic segregations are flattened within S1 and elongated parallel to the axes of minor D1 folds. Wilson also showed that these minor D1 folds and lit par lit segregations were refolded during D2 deformation in minor D2 folds.

The early structural history of Glenfinnan Division rocks is therefore very similar in Fannich and Carn Chuinneag. The only difference being that the early D1 structures are better preserved in Carn Chuinneag because the migmatites in Fannich all lie close to the Sgurr Beag Slide (D2) which reworked D1 in both areas.

SBS.3.b Pre-Sgurr Beag Slide metamorphism

The Fannich area

The Moine schists of the NW Highlands of Scotland are in general, poor indicators of metamorphic grade (Section SBS2.a). The difficulty encountered in determining precise metamorphic grade during the Caledonian orogeny is multiplied when trying to discover the metamorphic state of the rocks earlier in their polymetamorphic history. The calc-silicate rocks which are used to determine the grade of more recent events are less useful in determining earlier events due to their susceptibility to changes of grade and lack of relict textures. Polymetamorphism has been proven using structural correlations and internal rotation fabrics of garnets (Powell & MacQueen, 1976; MacQueen & Powell, 1977; Anderson & Olimpio, 1977) and also by detailed interpretation of a 776 ± 15Ma pegmatite which was shown to antedate at least one episode of garnet growth (Powell et al., 1983). However, in all these studies, the evidence indicated only that the rocks had reached at least garnet grade during their early metamorphic history.

The Fannich area

It has not been possible in Fannich to determine the precise grade of early metamorphism, although it is possible to demonstrate that polymetamorphism has occurred. Rocks of the Glenfinnan Division (the Meall an t'sithe Pelite) exhibit a lit par lit gneissic texture which is folded by minor D2 folds. The gneisses are also reworked within the Sgurr Beag Slide zone, forming an augen gneiss. Sutton and Watson (1954) reported one occurrence of relict kyanite from the Meall an t'sithe Pelite which may indicate the early high grade metamorphism. However, the detailed microstructural relationships of the kyanite were not published. The major mineralogy of the migmatitic lits is quartz, andesine and muscovite. The absence of a major k-feldspar phase indicates that the lits are segregations or melts formed at temperatures above 650°C (Winkler & Von Platen, 1961).

The absolute age of the migmatization event cannot be determined from Fannich since no whole rock Rb-Sr dates have been attempted on metasediments of the area. Migmatization must however have occurred at or before 740 ± 30Ma (Van Breemen et al.,1974; see also later in this Section for discussion). No evidence has been found that either supports or refutes the indications of an age of 1004 ± 28Ma for early metamorphism of the Moine schists, as suggested by Brewer et al. (1979).

The nature and full extent of pre-Sgurr Beag Slide metamorphism of the Morar division of Fannich is extremely difficult to determine. An early mica fabric comprising small 200μ aligned muscovite and biotite (recrystallised from early chlorites?) is seen folded by minor D2 folds in psammites, but in calc-silicates and pelites the fabrics are axial planar to minor F2, no relict fabrics have been seen. At most we can therefore deduce that the early metamorphic event may have reached greenschist facies. However, recrystallisation of early micas and chlorites could easily have occurred during later events. The early fabric might alternatively represent a compaction fabric similar to that found in low grade metasediments (Maltman, 1977).

The Carn Chuinneag area

The relationships of the Carn Chuinneag granite and its aureole to the structural and metamorphic history of surrounding Moine rocks are critical to any discussion of the early metamorphic history of the Fannich area since the structural histories of the two areas are closely linked (Section SBS.3.a).

It has long been established that within the aureole of the granite, andalusite (chiastolite) and cordierite were pseudomorphed during subsequent Caledonian regional metamorphism by kyanite, white micas and garnet (Geol.Survey, 1912; Harker, 1954; Shepherd, 1973).

Two hypotheses exist concerning the grade of metamorphism prior to granite intrusion, that it reached garnet grade (Long & Lambert, 1963), or only greenschist facies (Shepherd, 1970,1973). Long and Lambert (1963,

Appendix A) argued that the work of De L'Apparent (1935) (see discussion - Section SBS.3.a) showed that xenoliths within the Carn Chuinneag mass were folded prior to intrusion and that garnet was present prior to growth of hornfels biotite. There has however been controversy over interpretation of the timing of garnet growth within the granite aureole (Geol. Survey, 1912; De L'Apparent, 1935; Tilley, 1935; Harker, 1954; Haugh, 1963; Shepherd 1970,1973). Shepherd made a detailed study of garnet fabrics associated with the granite and demonstrated that inclusion trails within the inner zones of garnets, outside the thermal aureole, have strongly orientated inclusion trails, made up of inequidimensional quartz grains. This he took to indicate the schistose nature of fabrics which pre-dated garnet growth. However, within the granite aureole, Shepherd (op.cit.) showed that the inner zones of garnets only rarely exhibit orientated fabrics or inclusion trails and these are generally weak. More commonly, garnets within the aureole have random internal fabrics and further, garnets found within pseudomorphs of cordierite show similar fabrics to those outside. On this evidence, Shepherd (op.cit.) concluded that garnet growth must post-date the thermal metamorphism and thus intrusion of the granite.

The Glenfinnan Division migmatites of the Carn Chuinneag area exhibit very similar lit par lit textures to those seen in Fannich and they too are reworked by minor D2 folds (Wilson, 1975). Further evidence is afforded by the presence of sericite knots containing relict kyanite (Wilson, op.cit.) indicating probable retrogression of an early high grade fabric. The absolute age of the migmatisation event in the Carn Chuinneag area has been dated at 740 ± 30Ma by Van Breemen et al. (1974). This age was obtained from several Rb-Sr determinations of muscovite books from a pegmatite found concordant with the lit par lit fabric of the migmatites. Van Breemen _et al._ (_op. cit._) interpreted the pegmatite as a syn-metamorphic phenomenon. However, Powell _et al._ (1983) have shown that a similar age pegmatite in Ardnish (Inverness-shire) antedates an early high grade metamorphic event. Thus the date obtained by Van Breeman _et al._, from the Carn Chuinneag area must be regarded as a minimum age of migmatisation.

The early greenschist facies metamorphism of Morar Division rocks in the Carn Chuinneag area can be correlated with metamorphism accompanying D1 in Fannich on several lines of evidence:

a) The proximity of the Carn Chuinneag and Fannich areas and structural continuity around major D3 folds. In other words, the two areas were in a very similar position prior to D3.

b) The similarity of D1 fabrics and minor folds between the two areas. (See Section SBS.3.a).

c) The continuity of the internal fabrics found in garnets. Shepherd (1970) surveyed garnet fabric from Carn Chuinneag towards the Moine Thrust and found very little variation outside the aureole of the granite. Fabrics with the garnets in Fannich and even into the mylonites can be traced continuously to the Carn Chuinneag granite.

The recognition of an early metamorphism at greenschist facies or below in the Morar division rocks of Ross-shire makes a clear distinction between early metamorphism of these rocks and Glenfinnan Division migmatites of the area. The contrast in early metamorphism of the Morar Division from south to north must reflect the waning of pre-Caledonian orogenic activity northward referred to by Powell (1974) and Lambert _et al._ (1979).

SBS.4 A Proposed Relationship between the Carn Chuinneag Granite and the Sgurr Beag Slide

The geological relationships of the Carn Chuinneag granite are critical to understanding the structural and metamorphic development of the Moine schists around Fannich because the date of intrusion of the granite is well established at 555$\pm$10Ma, (Long & Lambert, 1963; Shepherd, 1973; Pidgeon & Johnson, 1979) pre-dating the Caledonian orogeny. The granite lies within 10km of Fannich and aplitic apopheses up to 30cm wide have been recorded in Fannich (Section GC.5).

Previous workers have not considered the path of the Sgurr Beag Slide from its outcrop in Garve (Rathbone & Harris, 1979) to that in Fannich (Section SBS.1) which must pass through the Carn Chuinneag area which lies between the two.

In the earliest survey of the granite and its aureole, the Geological Survey (1912) showed that porphyroblasts within the aureole were pseudomorphed by high grade regional metamorphic minerals and concluded that the granite intrusion pre-dated regional metamorphism.

An excellent review of the early work is presented by Shepherd (1973), whose work provided the first complete structural synthesis of the setting of the Carn Chuinneag granite.

In summary, Shepherd's structural sequence is:

a) D1 deformation (see Section SBS.3.a of this thesis)

b) Intrusion of the granite as a sheet at a high level in the crust. The Moine envelope became distended and underwent contact metamorphism.

c) D2 deformation, after solidification; formation of the main schistosity.

d) F3 open to close folding on N-S axes

e) Open folds on E-W axes.

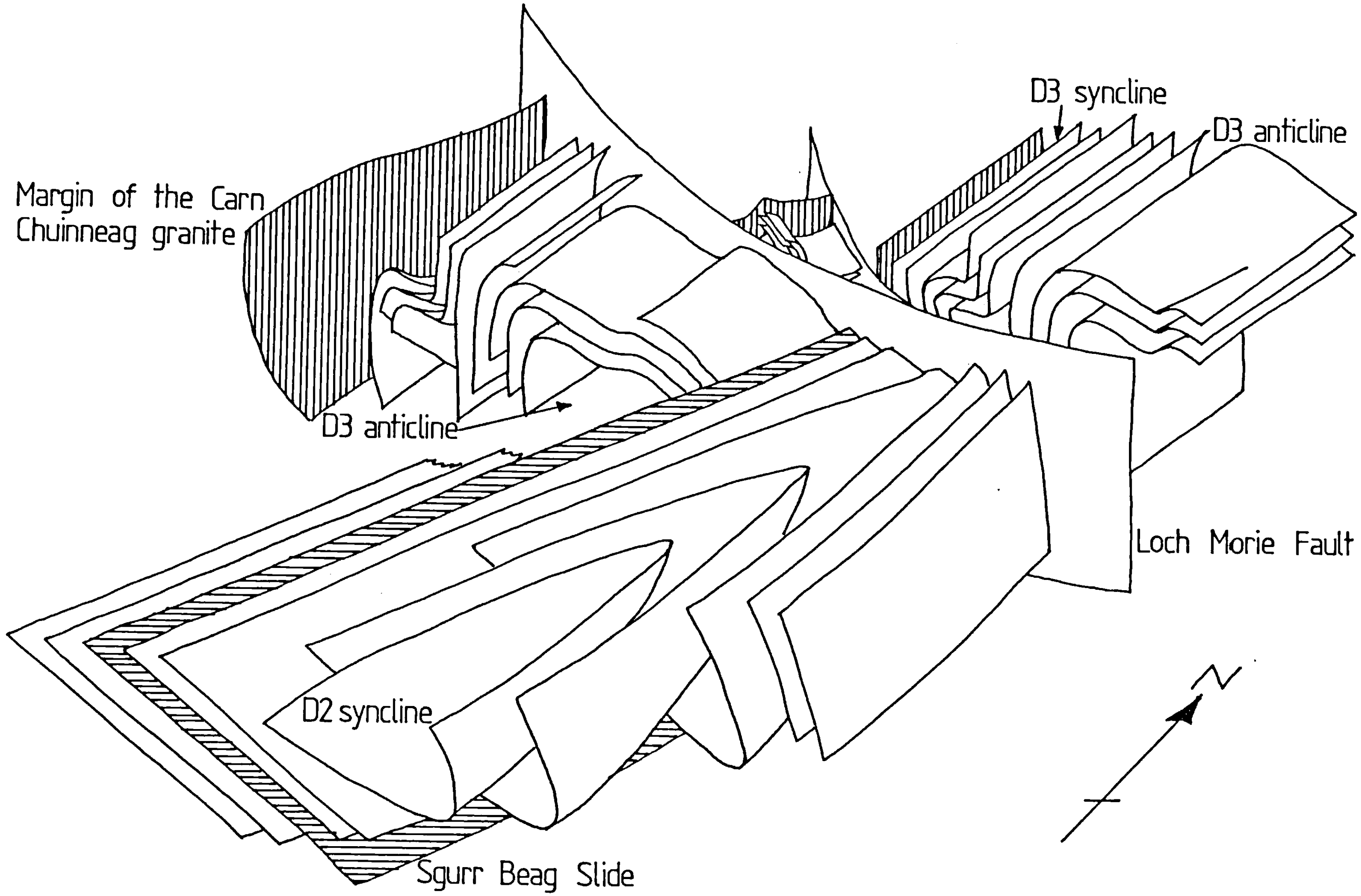


FIGURE 47. Block diagram of the area between Carn Chuinneag and the Sgurr Beag slide (after Wilson, 1975)

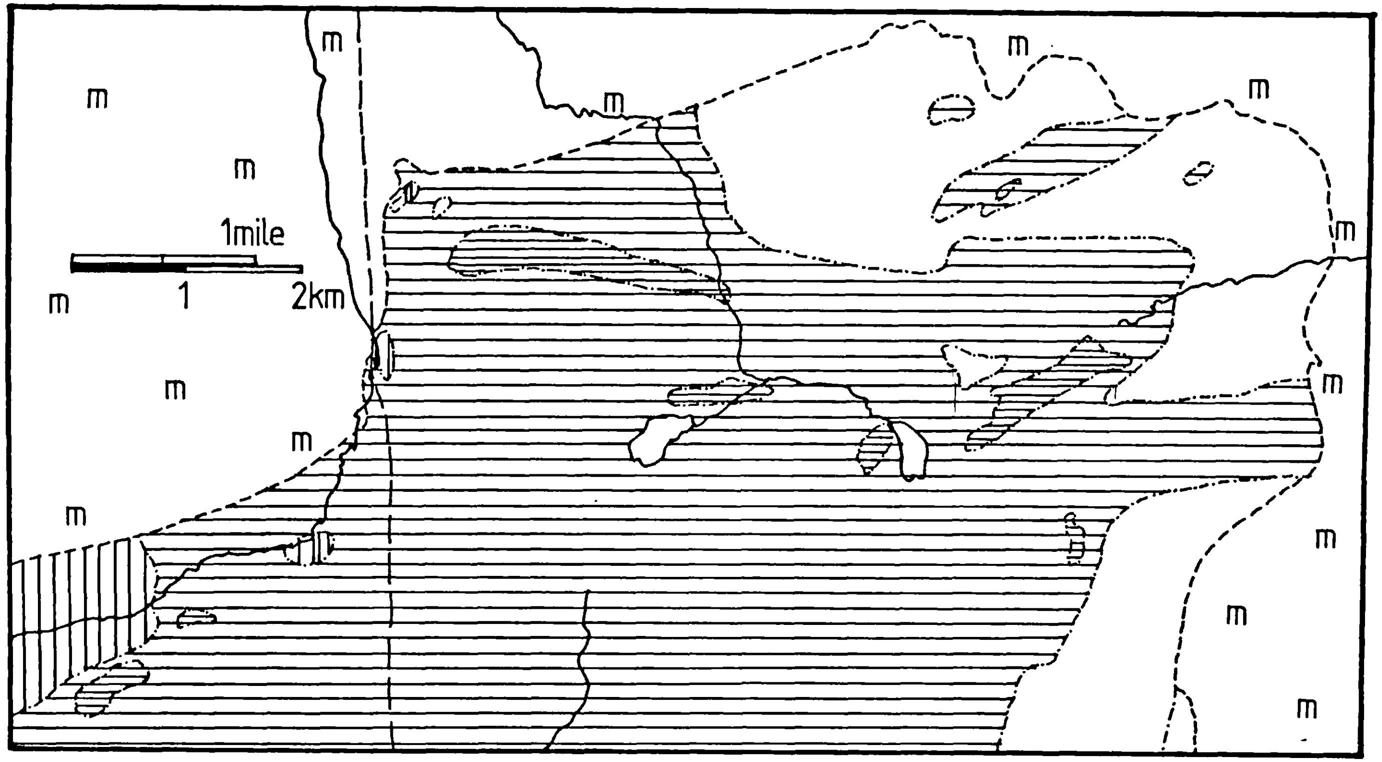


FIGURE 48. showing the intensity of the D2 foliation in the Carn Chuinneag area

(after Shepherd 1973)

Weak gneiss foliation	Weakly foliated amphibolite	m Moine country rocks
Moderate " "	Moderate " "	
Strong " "	Strong " "	

Wilson (1975), also working on the granite area showed that the granite is preserved within an F3 synform in the same way as the Glenfinnan Division rocks in Fannich (fig.4,47). Wilson and Shepherd (1979) combined their work over the whole northern and eastern margins of the granite and determined that its margins are concordant with bedding within the Moine rocks which young consistently towards the granite from all directions. They also showed that the granite foliation is folded by the F3 synform.

Wilson was the first to attempt a correlation between structures in the granite, those now attributed to slide movement, and structures within the Glenfinnan Division rocks to the east. He correlated the main granite deforming event with movement of the slide and D2 deformation in the Glenfinnan Division, a relationship confirmed by Rathbone and Harris (1979). Rathbone (1982) studying the Lewisian tectonic slice lying within the slide zone at Garve (fig.9) recorded breakdown of an early metamorphic fabric within the Morar division increasing towards the slide. He concluded that this process occurred during movement of the slide.

Shepherd (1973) produced a map showing qualitatively the spatial distribution of deformation within the main granite body, improving the Geological Survey's mapping of the occurrence of augen gneiss within the granite (Sheet 82). This distribution can be related to the position of the Sgurr Beag Slide and will be discussed in detail later in this Section.

The granite shows a wide range of internal deformation fabrics (Peach et al., 1912; Shepherd, 1973; reconnaissance by the present author). The granite ranges from areas in which no foliation can be detected to areas showing strong $L<S$ fabrics defined by the alignment of mica cleavage planes and "cake like" augen of K-feldspar. Within the Inchbae granite mass, the fabrics become $L>S$ defined by rod like augen of K-feldspar (Shepherd, op.cit.).

However, since the Inchbae granite lies in the tightest part of the core of a major F3 synform, the D2 (i.e. Sgurr Beag Slide - see below) fabrics may become more strongly modified. The main part of the Carn Chuinneag mass lies in a more open part of the fold.

While producing a stratigraphy for Fannich and attempting to correlate it with other areas, it was noted by the author that an anomaly occurs relating to the position of the Carn Chuinneag granite and its aureole. Because the slide zones at Fannich and Garve juxtapose groups of rocks which can, on lithological and metamorphic similarities, be correlated with the Morar and Glenfinnan Division rocks, it follows that the slide zones in each of the areas represent the Sgurr Beag Slide zone. Thus they are most probably spatially continuous.

In Fannich and the area around Carn Chuinneag, one lithology can be particularly well correlated; the Sgurr Mor Pelite, known as the Lower Pelite in the Carn Chuinneag area. The two pelites are lithologically very similar and more particularly both contain strata-bound amphibolites with calc-alkaline affinities (Section St.2.b). Amphibolites within the Glenfinnan Division of Fannich have been analysed by Winchester (1976) and show tholeiitic trends.

The upper boundary of this pelite may be used as an approximate datum plane. In Fannich the Sgurr Beag Slide lies 300m above this datum plane whilst at Garve it lies less than 500m above. In contrast, in the area around Carn Chuinneag a thickness of up to 3km of uninterrupted Moine metasediments lie above the upper boundary of the pelite and the Carn Chuinneag granite lies at the top of this succession - no higher rocks are exposed.

Over most of its outcrop, it has been shown that the Sgurr Beag Slide closely follows stratigraphic boundaries only rarely showing cross-cutting relationships (Tanner, 1971; Rathbone & Harris, 1979; cf. Baird, 1982). However, this rule does not appear to apply in the Carn Chuinneag area; in order to connect the outlier of Glenfinnan Division and Lewisian rocks that sit above the Sgurr Beag Slide at Fannich with those at Garve, and thus the slide zones, it is necessary to take the slide through or over the Carn Chuinneag granite. In summary, constraints on the course that the Sgurr Beag Slide may follow are:

a) The pelitic rocks of the granite aureole are not migmatitic as are all the pelitic lithologies of the Glenfinnan Division - thus they are regarded as members of the Morar Division.

b) None of the features representative of the Sgurr Beag Slide, such as a high strain zone and the presence of Lewisian slices, are seen in the considerable thickness of rocks lying structurally beneath the granite.

c) The Rb-Sr and U-Pb data (Pidgeon & Johnson, 1974) together with work of Shepherd (1973) shows the granite to have been intruded at 550 ± 10Ma into metasediments which had only undergone one phase of pre-intrusion low grade pre-Caledonian regional metamorphism, at or prior to ca.740Ma (Van Breemen et al., 1974; Brewer et al., 1979; however see Powell et al., 1983).

It therefore appears that the path of the Sgurr Beag Slide must lie through or above the present level of exposure of the Carn Chuinneag granite. The margin of the granite is less than 6km from the outcrop of the Sgurr Beag Slide at Garve (fig.9). Wilson (1975) showed that a major anticline of F3 age exists between the two, refolding the S2 schistosity. However, he did not discuss the consequences of these observations; further, he did not predict the possible path of the Sgurr

Beag Slide west of its present outcrop. Interpretation of the maps and structural data given by Wilson (1975) indicates that if the slide remains concordant with the S2 bedding parallel schistosity as it does elsewhere, and is folded by the F3 anticline, it would outcrop again close to the boundary of the Carn Chuinneag granite (fig.47). At its closest approach to the slide, on the shores of Loch Luichart, highly deformed, large apopheses from the granite crop out only 800m from the slide.

Further evidence of the path is afforded by considering the variable internal deformation of the granite (Shepherd, 1973). In places the granite has no foliation, in others the granite comprises a highly deformed augen gneiss. The most highly deformed parts are the Inchbae mass and central to western parts of the main mass (fig.48). If the effects of major F3 folds are removed, the highest deformation zones in the granite lie above the less deformed and at the highest structural levels. On the basis of the foregoing evidence and extrapolations, it is therefore suggested that the Sgurr Beag Slide passes just above the present outcrop of the Carn Chuinneag granite.

The Sgurr Beag Slide is thought to have been active during peak Caledonian metamorphism at 460-470Ma (Brewer et al., 1979; Powell et al., 1981), that is a time when the Carn Chuinneag granite had already crystallised at a high crustal level. The intense stretching lineations present in the high strain zone of the slide at Garve and Fannich indicate that the Glenfinnan Division rocks moved in approximately NW direction over Morar Division rocks and over the granite sheet.

It is possible that deflection of the Sgurr Beag Slide was caused by the presence of the granite body with the leading edge of the body (the Inchbae mass) and higher levels of the main granite suffering slide related deformation. Those parts of the granite furthest from the slide would be least deformed (fig.48). Grainsizes within the granite are up to 10 times as large as the surrounding metasediments, and it contains no strong lithological banding. It could therefore have formed a large semi-rigid block during deformation, deflecting the path of the Sgurr Beag Slide.

CHAPTER III. THE MOINE THRUST AND ITS EFFECT UPON EARLIER STRUCTURES

History of Research into Folding related to the Moine mylonites

Before considering the present work on the relationship between folding in the Moine nappe and the marginal mylonite zone, it is important to understand the controversy arising from previous studies of folding within the Moine mylonites and correlation of these events with those affecting the rest of the Moine nappe.

The earliest structural analysis of the Moine mylonites was published by Johnson (1957, 1960), for the areas of Lochcarron and Coulin Forest. Christie (1960, 1963) proposed a complex sequence of deformation and re-crystallisation for the Assynt area. The structural sequence of all the early surveys broadly agree with that summarised by Barber (1965) for the thrust zone around Lochcarron and Lochalsh. The deformation was thought to comprise two parts: early mylonitization and, later, cataclastic deformation. The deformation was considered to be polyphased, each phase being followed by static recrystallisation. The major metamorphic event was thought to follow the second phase of isoclinal folding. The sequence of deformation was as follows:

1. Mylonitization - early granulitization along narrow bands with early isoclinal folding.
2. Isoclinal folding on ESE axes and formation of a penetrative lineation
3. Widespread metamorphism and static recrystallisation.
4. A symetrical folding on N-S axes.
5. Monoclinal and kink folds of various axial trends.
6. Brittle thrusting and cataclasis.

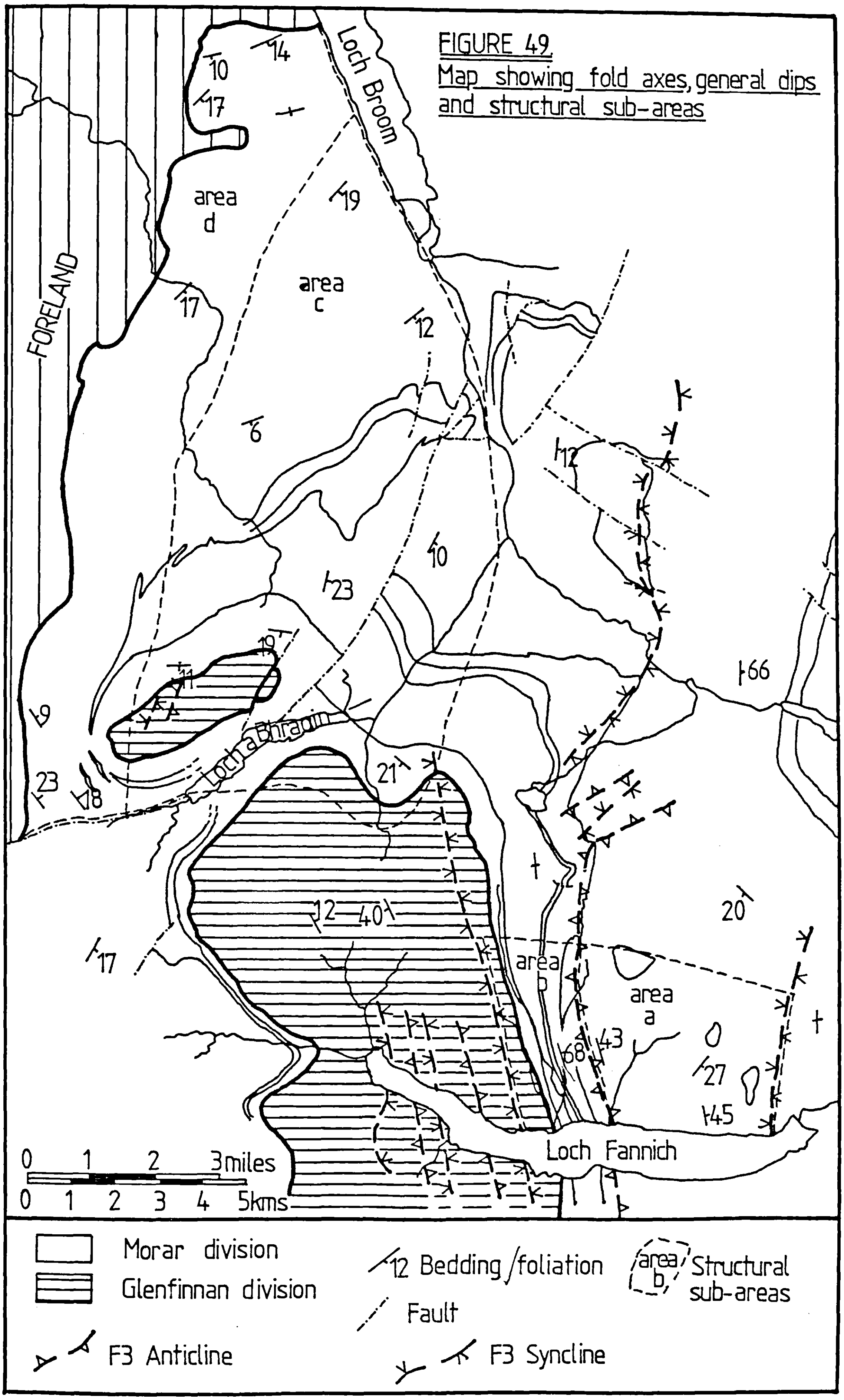


FIGURE 49
Map showing fold axes, general dips and structural sub-areas
Loch Broom
FORELAND
area d
area c
area b
area a
Loch a Bhraoin
Loch Fannich
0 1 2 3 miles
0 1 2 3 4 5 kms
Morar division
Glenfinnan division
12 Bedding/foliation
Fault
area b Structural sub-areas
F3 Anticline
F3 Syncline

Initially, Johnson (1957) proposed that the Moine nappe moved continuously NW. However, later he altered his conclusions and stated that the various structures indicated a more complex, discontinuous movement pattern (Johnson, 1960b).

Christie (1963) proposed that the early phase of ESE trending isoclines and parallel penetrative lineation indicated a component of movement along the strike of the thrust zone in a SSW direction. All authors working on the Moine Thrust zone agreed upon the NW direction of brittle thrust movement despite disagreement on movements during ductile deformation.

Barber (1965) made no assumptions about the direction of early movement but said that the late brittle movement post-dated and was not necessarily related to the early mylonites.

Johnson (1965) stated that the axial planes of the folds and the Moine Thrust are essentially parallel, and that therefore the regional stress pattern remained fairly constant throughout the time of formation of these structures. He suggested that the isoclinal folds described by Christie (1963) were modified in shape and possibly axial trend by later movements.

Bryant and Reed (1969) working in the southern Appalachians discovered a structural sequence in a thrust zone comprising early isoclines with a penetrative lineation parallel to the direction of movement and a later set of a symetrical folds perpendicular to the direction of movement. Bryant and Reed believed that the tight to isoclinal folds had developed from more open folds. They postulated that the open folds they saw were flattened and tightened while the axes rotated towards the direction of movement in progressive simple shear.

The evidence Bryant and Reed cited consisted mainly of the lack of any distinctive gap between the end members, isoclinal folds with axes transverse to the thrust belt and open folds with longitudinal axes. They described a gradation between open and isoclinal folds which was related to the orientation of their axes.

Bryant and Reed postulated that the same process occurred in thrust belts of the Scottish and Norwegian Caledonides. They pointed out that the data of Johnson (1960,a) and Christie (1963), showed weak and partial girdles between the maxima of the two preferred axial orientations for early isoclinal and late open folds.

Soper and Wilkinson (1975), mapped the Moine Thrust at Eriboll on the north coast of Scotland. They produced a structural sequence essentially similar to that of early workers in the Moine nappe. They recognised a D1-D4 sequence: D1, pre-dating the thermal maximum and involving the generation of the mylonitic fabric; D2, associated with the thermal peak and characterised by a widely developed extension lineation; D3, involving the production of crenulation fabrics; D4, 'brittle', often conjugate structures.

All deformation phases involved folding either isoclinal in the cases of D1 and D2 (these were separated by their relationship to the mylonitic fabric) or late a symetrical folding in D3 and conjugate or monoclinal kinks in D4.

Soper and Wilkinson followed discontinuous repeated movements of the Moine nappe towards the NW. They explained the orientations of D1 isoclines as due to passive rotation during later penetrative deformation but did not develop the theme to explain the ESE direction of D2 fold axes.

Soper and Wilkinson seem not to have been aware of the hypotheses of Bryant and Reed although they attempted to prove a simple shear mechanism rather than pure shear (Johnson, 1967). They attempted to correlate the four phases of deformation in the marginal mylonites with equivalent deformations with the rest of the Moine nappe.

On the basis of structural dating of the Vagastie suite of minor intrusions, Soper (1971) suggested a correlation of the earliest mylonitization with D1 events throughout the Moine nappe. This suite was thought to include the deformed Strath Vagastie granite which together with the other intrusives were dated by Soper as pre-D2 and correlated with the Carn Chuinneag granite. Soper stated that the minor intrusions post-dated one episode of mylonitization and therefore concluded that the mylonites were of D1 age.

The earliest mylonitization was known to affect the highest member of the Cambro-Ordovician succession of the foreland, the Eilean Dubh member of the Durness formation. The uppermost member has been shown to be of Arenig age (Dewey et al., 1970; Cowie et al., 1982) or possibly early Llanvirn (Higgins, 1967).

Early dating of the Carn Chuinneag intrusion (Long, 1964; Long & Lambert, 1963) by Rb-Sr whole rock and mineral isochrons, assigned it an intrusion age of 530 ± 10Ma. In consequence according to Soper's (1971) correlation, the earliest events in the Moine Thrust zone had to have occurred before 530Ma, because they pre-dated the minor intrusions correlated with Carn Chuinneag yet post-dated the Arenig or even possibly early Llanvirn.

Soper and Wilkinson (1975) revised Soper's earlier structural correlations since later data from Rb-Sr whole rock and U-Pb zircon analyses suggested an age of 550 ± 10Ma for intrusion of the Carn Chuinneag granite (Pidgeon & Johnson, 1974). The revised age seemed to preclude a correlation of early thrust zone events with earliest events in the Moine nappe since no known time-scale places the Arenig as far back as 560Ma. Johnson and Shepherd (1970) and Shepherd (1973) postulated a D2 age for early mylonitization. Soper and Wilkinson accepted the D2 correlation.

Unfortunately, recent work by Pidgeon and Aftalion (1978) dating the Strath Vagastie intrusion using the U-Pb method, obtained a date of 405 ± 11Ma and Ferguson (1978) showed that the Strath Vagastie granite post-dates the regional D2 deformation, considering it to be a 'Late granite' with the internal deformation witnessed by a schistosity and augen texture belonging to a later event. Hence even the correlations implied by Johnson and Shepherd (op.cit.) are suspect.

McClay and Coward (1982) from studies by Dyan, Madjwick and themselves, produced a new synthesis of events for the marginal mylonites although presenting little evidence. The dominant mylonite formation, D1 of Soper and Wilkinson (1975), was thought to represent a distinct episode of mylonitization at deeper levels. D2 and D3 were regarded as: "Stages in a sequence of essentially simple shear deformation in which the fold axes initiated normal to the transport direction and then were subsequently rotated into the movement direction". They regarded D4 as composed of box and kink folds having axes generally plunging SE, representing shortening almost parallel to the strike of the thrust zone. McClay and Coward (op.cit.) attributed D4 structures to local differential movement between thrust sheets.

This represents a radical change from the ideas of Soper and Wilkinson (1975) and a return to the ideas of Bryant and Reed (1969). Unfortunately the publication of McClay and Coward (1982) contains no evidence for their sequence nor any discussion of the previous work by Soper and Wilkinson (1975) and Bryant and Reed (1969). Their new synthesis can therefore only be regarded as a working hypothesis for further discussion.

MT.1 D3 Deformation in Fannich

MT.1.a F3 Fold Analysis

D3 folding is widespread in both Morar and Glenfinnan Division rocks north of Loch Fannich and the area towards the Moine Thrust zone. It is important to understand the form of F3 in Fannich in order to correlate this deformation phase into the Moine Thrust zone.

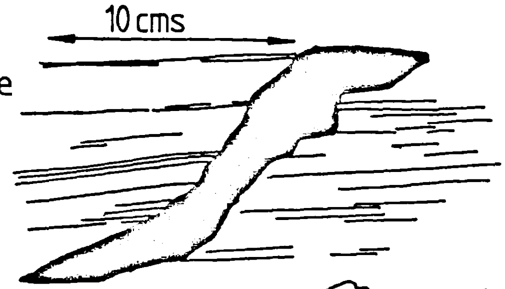
Two major F3 synclines and one major anticline have been mapped in the Fannich area: the Fannich syncline, An Cabar syncline and Loch Droma anticline (fig.49). Two smaller anticline/syncline pairs fold the gently inclined limbs of the larger folds. The major folds are a symetrical and overturned westward. Steep inverted limbs dip between the vertical and about 60°E, the shallow limbs dip at 30°-40°E (fig.49).

Two sets of pegmatites are folded by F3; an early set (Section GC.5) pre-dating D2 and developing a strong S2 schistosity; and a later, more widespread set consisting of quartz, large pink k-feldspar laths and traces of muscovite, apatite and plagioclase (Section MT.2.c). These post-date D2 fabrics (fig.50) but pre-date D3 (fig.50). The history of deformation in the post-D2 pegmatites is of particular importance in establishing a correlation of events from Fannich to the Moine Thrust zone mylonites (Section MT.2.a).

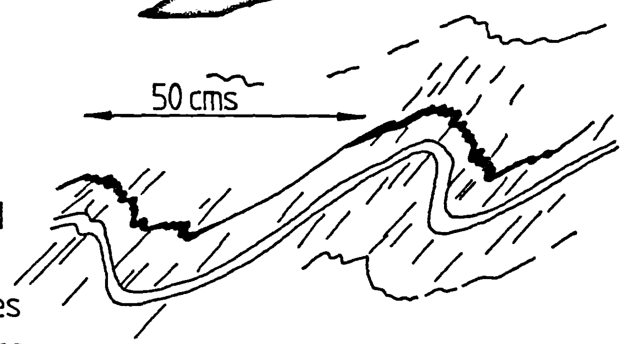
In Fannich, S3 is defined by a strong axial planar crenulation cleavage in pelitic bands and a weak fabric in the hinge zones of folds in more psammitic bands. Associated lineations are weakly developed

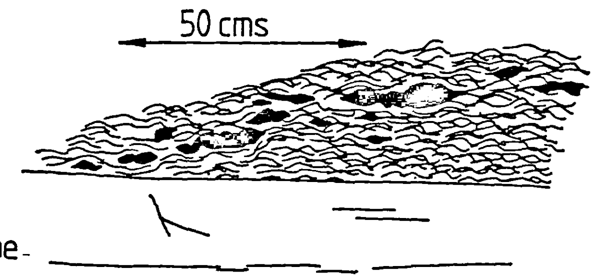
FIGURE 50. Post-D2 pegmatites

Pegmatite cross-cutting the intense D2 fabric of the Meall à Chrasgaidh psammite.

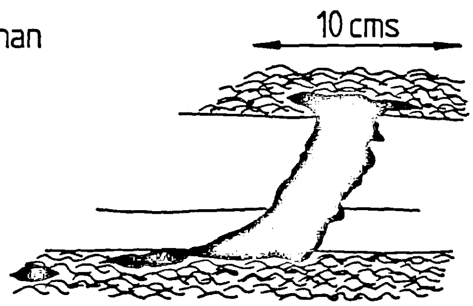


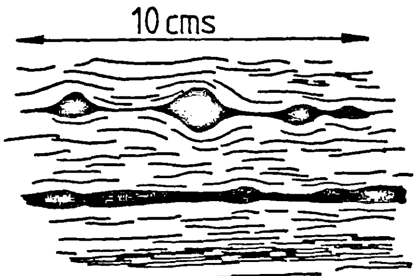
Post-D2 pegmatite folded by an F3 fold, within psammites and semi-pelites of the Inverbroom psammites.





The pegmatites become deformed to much lesser extent within psammites than pelites





boudinaged pegmatite within semi-pelitic mylonites

The contrast between psammites and pelites is also seen in the Fannich area.

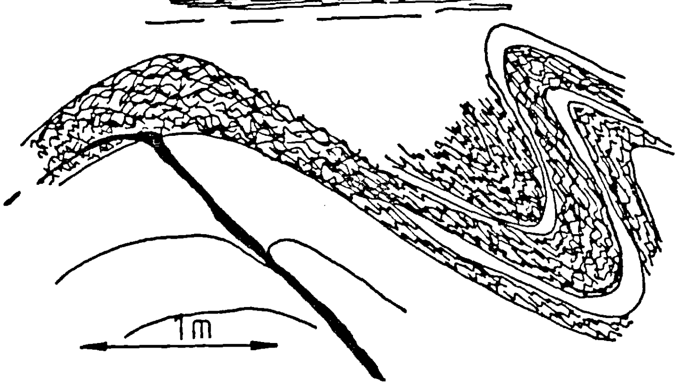
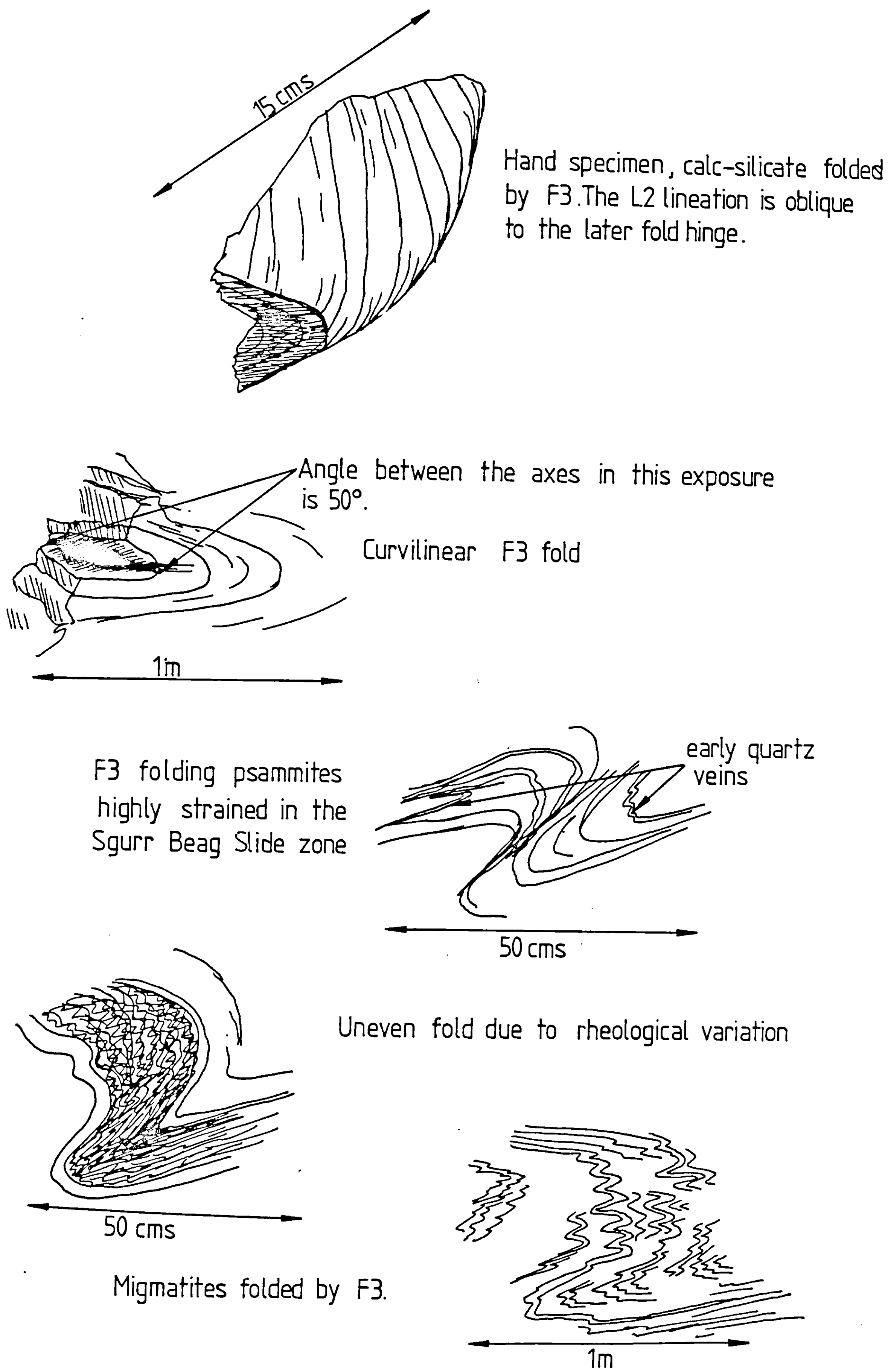


FIGURE 51. Examples of minor F3 folds



and comprise crenulation lineations or intersection lineations in the hinge zone of F3 folds.

Minor folds are generally close, becoming occasionally tight within pelitic bands (fig.51). They fall into classes 1b and 1c within psammites and 1c within pelites (Ramsay, 1967).

The majority of minor folds are concentrated within the vertical limbs of major D3 folds where their axial directions show only slight scatter (fig.52). On the gently inclined limbs very marked non-cylindricity is seen with swings of up to 80° in axial direction within 2m of hinge length.

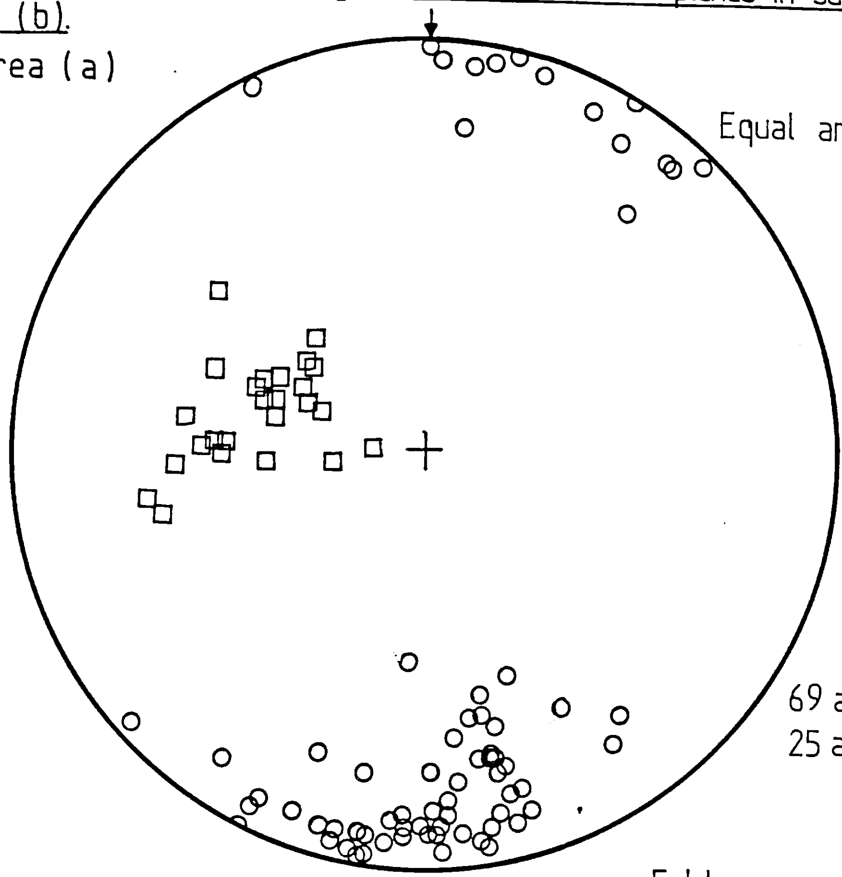
The axial planes of all minor folds interpreted as F3 in the Fannich area fall in a single cluster, 010/30°E (fig.52), implying that they were all part of the same fold episode. D3 lineations are always sub-parallel to F3 fold-axes and therefore show a similar pattern of distribution (fig.53). F3 fold axes are sub-parallel to D2 lineations in some areas (most especially sub-area b) and L2 masks the later, weak lineations. However, west of Fannich some D3 folds exhibit folded D2 lineations oblique to their axes.(fig 51)

As mentioned above, on gently inclined limbs of major folds, minor folds show marked non-cylindricity. This is further shown on the sub-area scale (figs.52,55). Consider the plot of data from sub-area (a) (fig.52), the major fold has a mean axial plane attitude of 010/30°E and whilst the minor fold axes show a concentration of fold plunges at 10°→175, there is evidently a spread of plunges extending along the great circle of the axial plane.

Similarly, the distribution of data for sub-area (b) (fig.52) shows a cluster of axial plane poles about 010/30°E. However, the minor fold axes describe a variable trend within the axial plane from 30°→105 to 5°→030.

FIGURE 52. Stereonets showing F3 folds and axial planes in sub-areas (a) and (b).

Sub area (a)

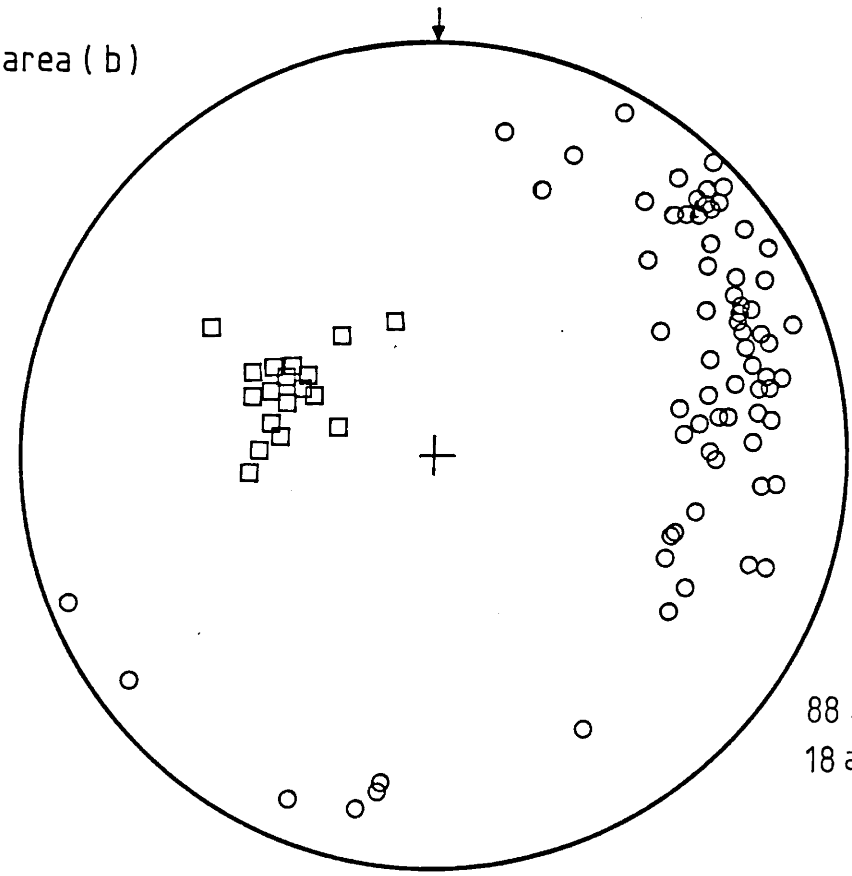


Equal area nets

69 axes
25 axial plane poles

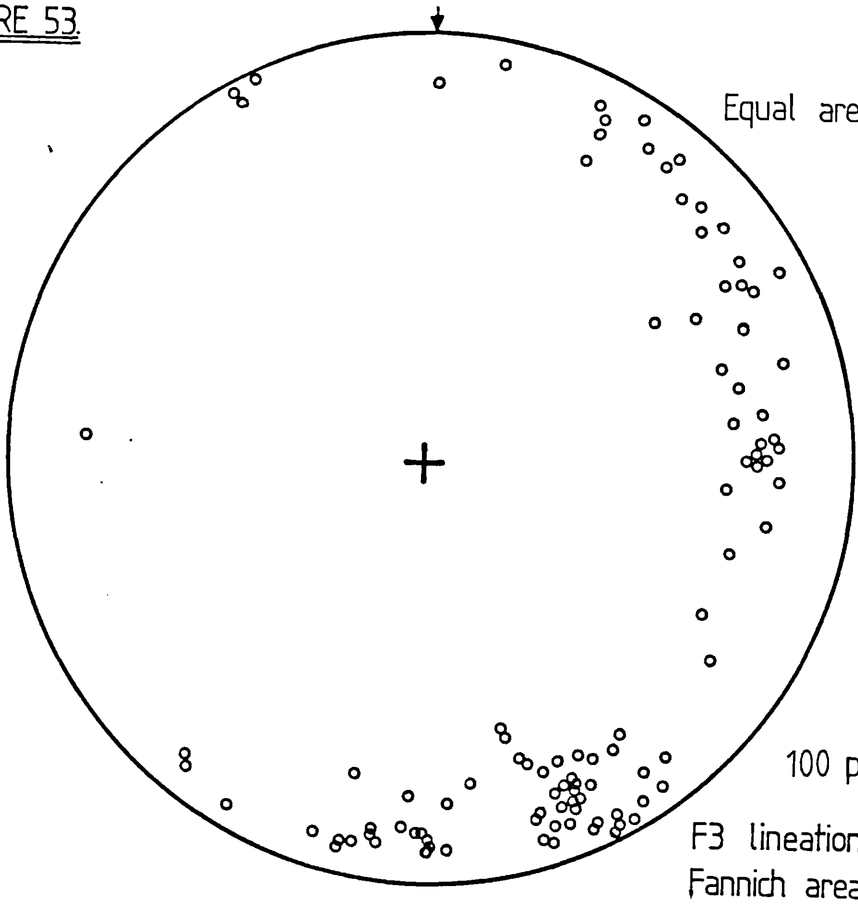
○ = Fold axes
□ = Axial planes

Sub area (b)



88 axes
18 axial plane poles

FIGURE 53.

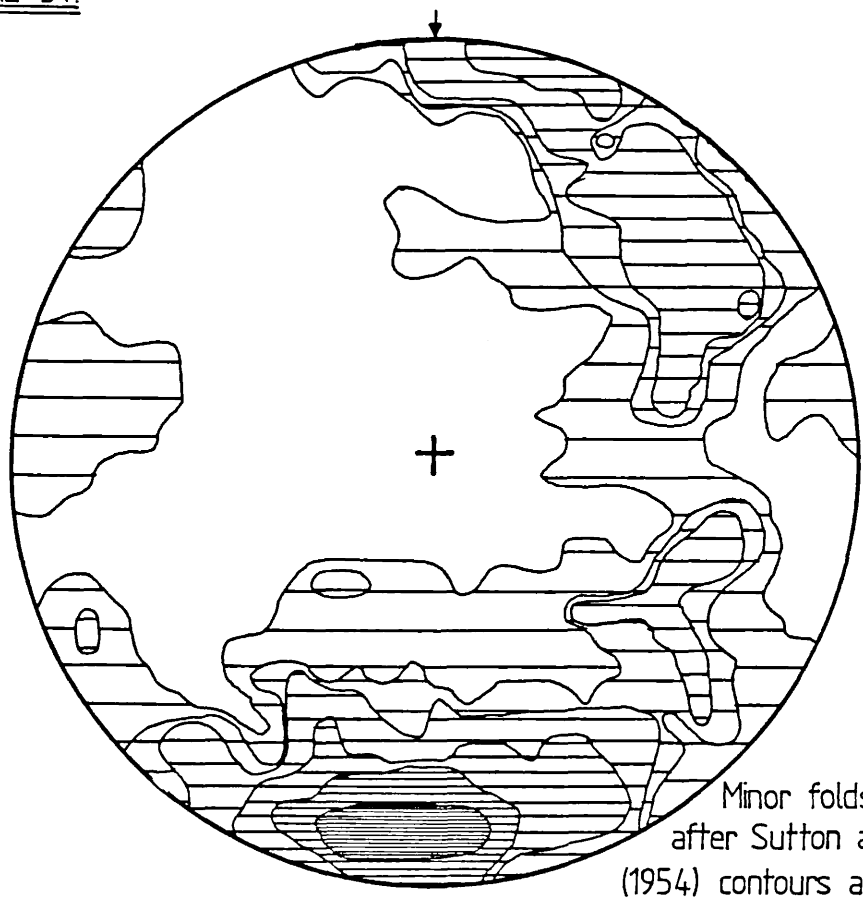


Equal area nets

100 points

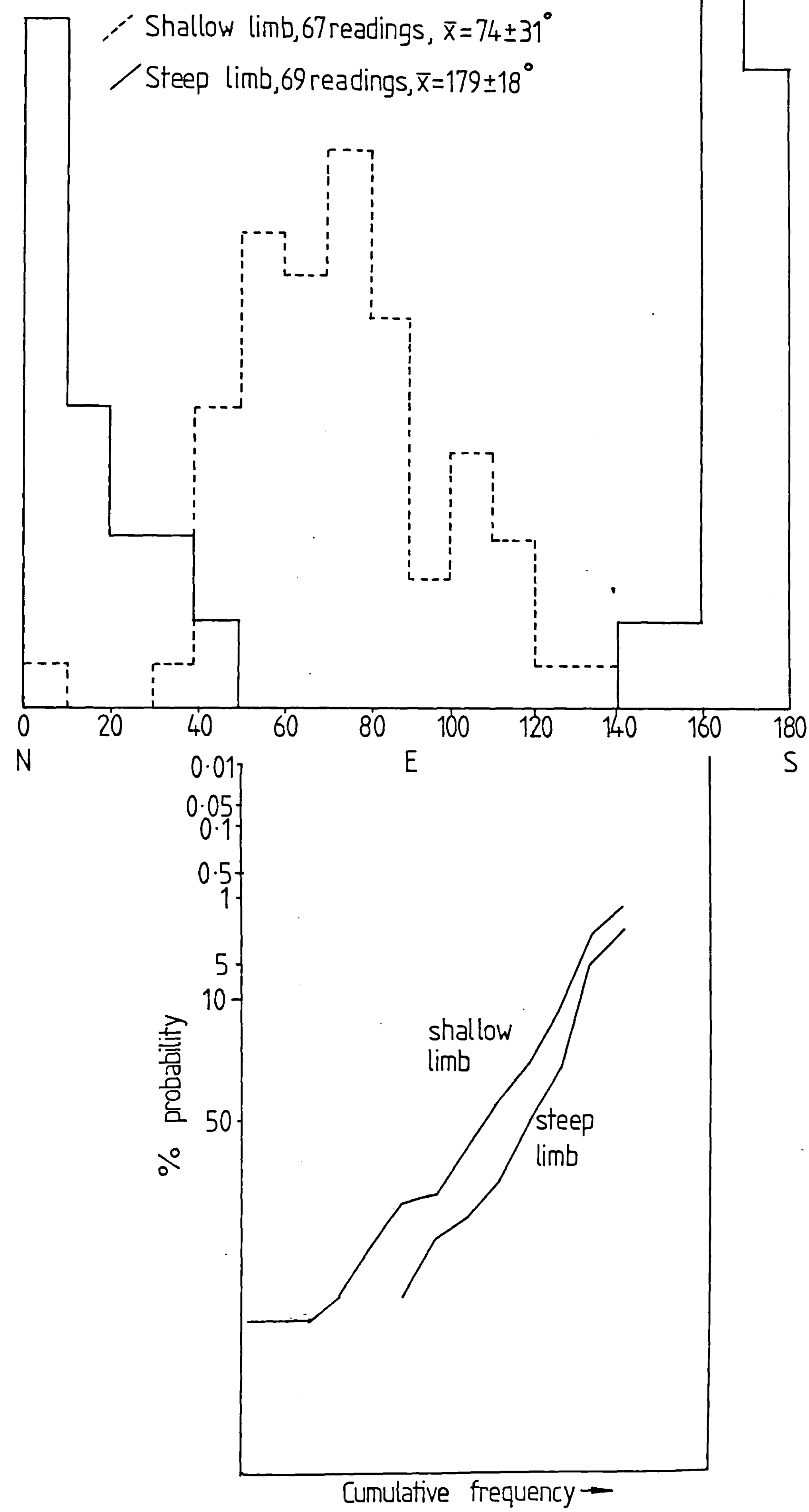
F3 lineations in the Fannich area.

FIGURE 54.



Minor folds, after Sutton and Watson (1954) contours at 0·2, 0·6, 1, 2, 4, 8 %.

FIGURE 55. Frequency distribution of F3 across the Fannich synform and cumulative frequency plot



The difference in attitude of fold plunges between sub-areas (a) and (b) might be explained by refolding, but there is no evidence of such refolding on a major or minor scale. Further, in both areas the folds post-date the second pegmatite suite, they are both associated with a similar crenulation of the S2 fabric, have similar styles being close to tight, class 1b or 1c (Ramsay, op.cit.) and have co-planar axial planes.

Sutton and Watson (1954, figs. 9B and C) reported a similar non-cylindrical pattern for the whole Fannich area, south of Loch Fannich and as far east as the Grudie river (fig.54). However, they did not distinguish between steep and shallow limbs and obtained a complete 180° spread of fold axis orientations within the axial plane.

A diagram showing the frequency distribution of azimuths for F3 fold axes illustrates the contrast between minor folds on the two limbs of a major F3 fold (fig.55). The standard deviation of the minor fold distribution on the steep limb is only 18° about a mean of 179° but the standard deviation is 31° about a mean of 74° on the shallow limb. Both limbs exhibit near gaussian distributions about the mean. However, insufficient measurements were obtained to perform a χ^2 significance test. When the data are plotted on normal/probability paper as a cumulative frequency, both distributions show near straight lines and marked variation from the line at the ends (fig.55).

This deviation from the straight line at low probabilities is again attributable to the lack of data. Further discussion will be devoted to the problem of the near gaussian distribution later in this section.

The distribution of fold axes across this major F3 closure poses a problem because all minor folds might be expected to fall in a gaussian distribution about the main trend for the major folds as has been shown in Dalradian rocks (Roberts & Sanderson, 1974). Minor folds measured on the steep limb conform to this model because the axes of the main Fannich syncline and Loch Droma anticline trend approximately N-S. Minor fold axes on the shallow limb display anomalous axial orientations, greater variability and are non-cylindrical on the exposure scale. The frequency distribution plot (fig.55) for the shallow limb shows two distinct peaks, one at 070 and a minor peak at 110.

Borradaile (1972) described a similar anomalous distribution of axial orientations and non-cylindrical folds in Dalradian rocks in Argyll - shire. He postulated that anomalous fold orientations were caused by rotation during constrictive strain $(00 < k < 1)$. Roberts and Sanderson (1974) examined anomalous fold axis distributions across the Tay nappe using the model developed by Sanderson (1973) to predict the distribution of fold axes during constrictive strain. They showed that two types of fold axis distribution occurred:

a) A unimodal distribution of fold axes within the axial plane, with a mean nearly perpendicular to the stretching direction.

b) A bimodal distribution of fold axes within the axial plane orientated about the stretching direction and associated with strongly non-cylindrical folds.

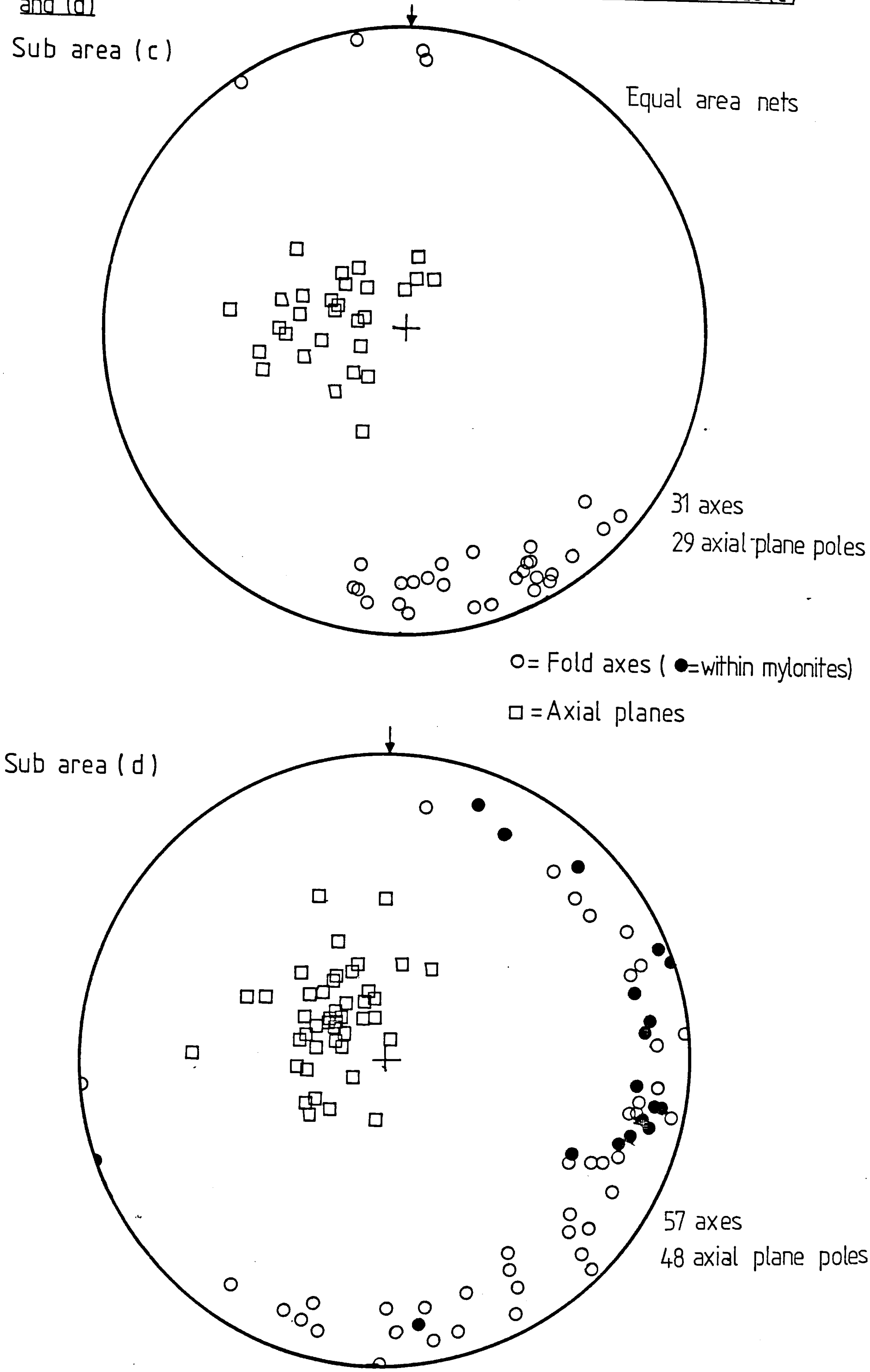
No stretching lineation was found in association with D3 deformation in Fannich. However, the distribution of fold axes on the shallow limb (fig.55) indicates a stretching direction between 090 and 100 using the model of Sanderson (1973). This coincides in direction with the stretching lineation developed during intense deformation further west in the Moine Thrust mylonites. D3 deformation in Fannich is thought to be of similar or slightly earlier age than the mylonites (Section MT.2.d) and it is therefore conceivable that the deformation causing the distribution of minor folds in the shallow limb is associated with the formation of mylonites.

However, this interpretation is not considered proven on the data presented because there are several problems; the orientation of the strain ellipsoid for the Fannich area has not been independently determined; although the distribution of fold azimuths on the gently inclined limb shows a drop at the stretching direction deduced, it is not statistically different from a gaussian distribution because there are too few data points to prove significance. In the work of Sanderson (1973) and Roberts and Sanderson (1974), distributions of fold axes are known to have rotated towards the X direction because independent determination of the strain ellipsoid was possible. However, the distributions presented by these authors cannot be proved, in many cases, to be statistically different from a single gaussian distribution using significance tests such as χ^2 and normal/probability plots. The model of Sanderson (1973) fits a bimodal, theoretically derived curve, to these plots on the basis of their anomalous orientations but never proves the curves to be statistically non-gaussian.

No attempt was made to produce quantitative results for strain from the data because it is possible that the field observations contain a bias and no independent measurement of the strain could be used to confirm an analysis based purely on the fold distributions.

It is concluded that the minor D3 fold axis distributions are caused by differing strains on the steep and gentle limbs of the major D3 folds in Fannich. The steep limbs suffered strains where $X/Y \simeq 1$ and fold axes remained close to the Y axis of the strain ellipsoid, having nucleated there. The shallow limb suffered constrictive or rotational strain and fold axes rotated towards the X axis of the strain ellipsoid during progressive strain.

FIGURE 56. Stereonets of F3 folds and axial planes from sub-areas (c) and (d)



MT.1.b <u>The effect of D3 upon the Sgurr Beag Slide zone</u>

The major part of this project has been to determine the structural relationship between the Sgurr Beag Slide and Moine Thrust zone, both in terms of their nature and movement histories. The effects of D3 on structures and fabrics associated with the Sgurr Beag Slide illustrate the relative ages of D3 in Fannich and the slide zone.

Immediately below the slide lies an extremely platey, parallel-banded psammite lithology, the Meall a Chrasgaidh Psammite (Section SBS.1.c). The banding is attributed to the formation of, and movement on, the slide zone. Minor D3 folds deform the platey banding of the Meall a Chrasgaidh Psammites (fig.51) and thus fabrics in the rocks showing the highest D2 strain pre-date D3.

The strong D2 extension lineation seen within the Meall a Chrasgaidh Psammite, defined by the alignment of elongate quartz and feldspar grains, lies sub-parallel to the axes of F3. In several exposures it becomes slightly oblique to the axes and is seen folded around F3 (fig.51).

The major fabric of D3 in the pelites of the Fannich area is a strong crenulation of an earlier foliation, which is associated with the deformation and movement of the Sgurr Beag Slide described in Section SBS.1.c. It is defined by a strong alignment of micas within the high strain zone of the slide, becoming weaker further away. In all areas it becomes reworked by D3.

During movement on the Sgurr Beag Slide, the boundaries would have been relatively planar and parallel (Cobbold,1977). From Figure 49 it can be seen that the slide zone has been folded around major F3 folds. It is noteworthy, as Winchester (1974) showed, that the metamorphic zonation indicated by calc-silicates is also folded around major F3 folds of the area (Section SBS.3.b).

FIGURE 57. An F3 fold deforming the migmatite fabric on the summit of Meall an t'sithe.

This is taken to indicate that D3 post-dates the peak metamorphic temperatures and pressures responsible for the mineralogy of the calc-silicates. A widespread suite of post-D2 (Sgurr Beag Slide age) pegmatites are folded by minor D3 folds (Section MT.2.c) indicating a possible time lapse between the Sgurr Beag Slide which moved during peak metamorphism, displacing contemporary isotherms (Powell et al.,1981), and D3 to allow for relaxation of metamorphic grade and intrusion of the post-D2 pegmatite suite.

MT.2 The Relationship between D3 in Fannich and Mylonites above the Moine Thrust

MT.2.a Analysis of F3 folds between Fannich and the Moine Thrust

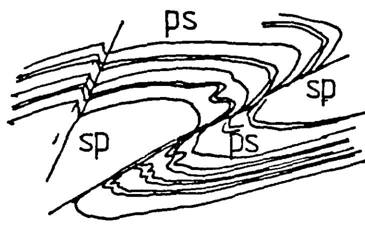
No major fold axes have been found between the Fannich syncline and the Moine Thrust. Minor folds exhibit a NW vergence apart from a small area on the ridge between Meall an t'sithe and Creag Rainich (fig.4,49). Here a medium scale anticline/syncline pair is present similar to that on the gently inclined F3 limb (sub-area b) described in Fannich.

The post-D2 pegmatite suite is ubiquitous throughout the area and always deformed by F3. The gradation in style and intensity of pegmatite deformation is characteristic of increasing strain towards the mylonites. The changes are important and will therefore be discussed later in detail (Section MT.2.c).

Sub-area (c) (fig.49) lies west of the major F3 Fannich syncline and therefore on the gently inclined limb of the major F3 fold. Minor folds are relatively rare and only 27 measurements were taken of axes throughout sub-area (c) (fig.56). The distribution of fold axes displays patterns which are similar to those in Fannich (cf. fig.52), they spread within the axial plane from approximately $20^{\circ}\rightarrow 190$ to $15^{\circ}\rightarrow 130$. The spread

FIGURE 58. Variation of folds (F3) in sub-area d.

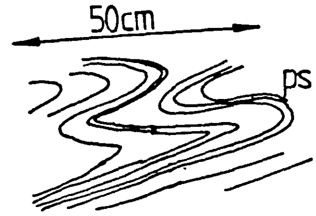
a)



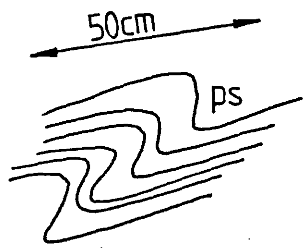
Large folds in sub-area (d) generally have complex forms and exhibit rare axial planar shears.

Many folds vary in style within a single exposure, dependant upon the rock type.

b)

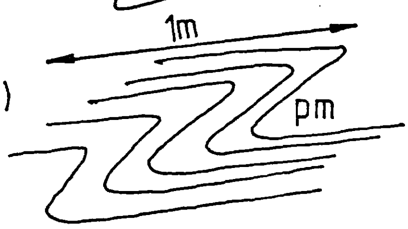


c)



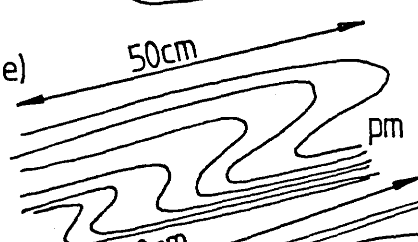
the least deformed F3 folds in sub-area (d) are close, type 1b or 1c.

d)



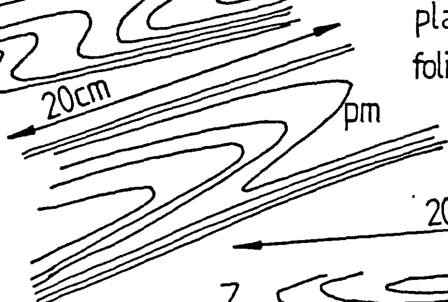
Chevron folds form in some lithologies.

e)



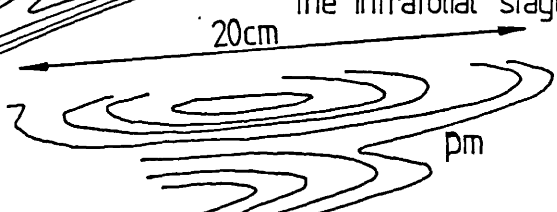
As the folds become more deformed they become tightened and their axial planes approach parallelism with the foliation, becoming intrafolial.

f)

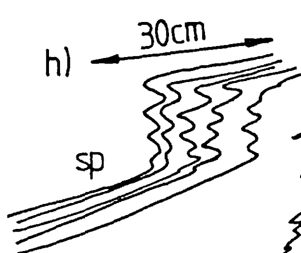


sheath folds commonly accompany the intrafolial stage.

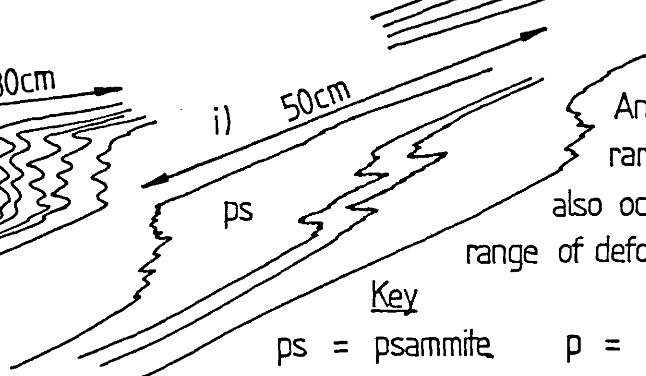
g)



h)



i)



Another, though much rarer "shear" type folding also occurs and exhibits a range of deformation states.

Key

ps = psammite p = pelite

sp = semi-pelite pm = psammitic mylonite might have been expected to correspond more closely to that of sub-area (a) but the small range in distribution may be attributable to lack of data in sub-area (c).

Sub- area (c) differs from both areas defined in Fannich. The axial planes are more flat lying, dipping approximately 20° east whereas in Fannich the mean F3 axial plane dip is 35° east (cf. fig.52). Within sub-area (c) the style of F3 is generally tight rather than close and the associated planar fabrics are stronger than in Fannich, especially in pelites (Sections MT.2.d, MT.5.a). F3 folds are seen to strongly crenulate the S1/S2 fabric within the migmatites on the summit of Meall an t'sithe (fig.57). The folds are seen to fold L2, though this lies sub-parallel to their axes.

Sub-area (d) lies adjacent to the outcrop of the thrust and the data plotted (fig.56) includes folds within the mylonites. A difficulty arises in defining the age of folds in sub-area (d); it is possible that many folds post-date the D3 event in Fannich and this problem will be discussed later in this section. For the moment, F3 folds are defined as: all close to isoclinal folds which fold the post-D2 pegmatites. Only 2 folds were found to pre-date the pegmatites and those were spatially associated with outcrop of the Sgurr Beag Slide at Meall Dubh, they are therefore regarded as D2 in age. The F3 folds vary from close to isoclinal (fig.58) but there is no simple gradation from close folds above the mylonites to isoclinal within them. A complete range of inter-limb angles and axial orientations is present in all parts of sub-area (d). However, folds above the mylonites tend to be close to tight with axes at various orientations to the movement direction of the over-thrust block defined by the extension lineation within the mylonites.

In sub-area (d) the mean axial plane of D3 folds lies within 10° of the mean mylonite foliation (fig. 56). This is lower angle than both Fannich (sub-areas a & b), F3 minor fold axial planes (fig. 52) and those of sub-area (c) to the east (fig. 56). A gradual decrease in the angle between F3 axial planes and the foliation plane is seen toward the thrust zone.

F3 folds are much more common in sub-area (d) than in sub-area (c). A progressive increase in intensity of D3 fabrics such as extension crenulation and associated grain size reduction indicates increasing strain towards the mylonites. The change is considered important in relating D3 in Fannich to the mylonites and is discussed in Section MT. 2. d.

As mentioned above, the spatial distribution of axes and axial planes in sub-area (d) shows a different pattern to that seen in sub-areas (a), (b) and (c) in that the plunges in fold axes vary in orientation through 180° within the F3 axial plane (fig. 56). It could be argued that if enough data were collected, the pattern in Fannich would also exhibit a complete spectrum of 180°. Sutton and Watson (1954) measured over 380 minor fold axes in Fannich and their data (Sutton & Watson, 1954, fig. 9B) shows a complete 180° spread. However, there is a very strong concentration of data sub-parallel to 180, that is the Y axis of the strain ellipsoid deduced earlier (Section MT. 1. a).

Sub-area (d) differs from the Fannich area in that there is evidence of high shear strains, syn to post-D3, witnessed by deformation of post-D2 pegmatites (Section MT. 2. c) and the increased intensity of D3 fabrics (Sections MT. 2. d and MT. 5. a). Increasing parallelism between axial planes, general banding and quartz veins (Sections MT. 3. b) also indicates high strain during or after D3.

The concentration of readings on Sutton and Watson's (1954) plot indicates that the majority of folds in the Fannich area underwent little rotation subsequent to their nucleation, a conclusion supported by evidence of lower strain in Fannich, especially within the steep limbs.

There is an approximate correspondence between the orientation of the axes of F3 folds and their distance above the thrust (fig. 59). Over 250m above the thrust the range of axial orientations is small. It becomes very wide as the thrust is approached, suggesting nucleation and rotation becomes dominant (see later in this Section) and then shows a concentration parallel to the movement direction within the mylonites.

As mentioned earlier, there seems to be an approximate correspondence between fold tightness and axial orientation, the relationship is by no means as strong however as indicated by Bryant and Reed (1969), so the plot of sub-area (d) (fig. 56) is strongly biased towards the area above the mylonites. Thus the concentration of axial orientations of folds within the mylonites is swamped. However, in Figure 56 the folds measured within the mylonites have been indicated and their distribution shows the strong concentration towards 100.

The interpretation proposed to account for the spatial distribution and variation in style of D3 folds close to the mylonites is similar to that of Bryant and Reed (1969) and accords with the mechanism described by Esher and Watterson (1974) for all folds nucleated in shear zones. They showed that folds continuously nucleate at high angles to the tectonic movement direction (near to the Y axis of the strain ellipsoid) and progressively rotate towards the X axis of the strain ellipsoid which in turn tends toward the movement direction under very high shear strains (fig. 60). At any time during the deformation, within the shear zone, there are relatively immature folds which have axes nearly normal to the X axis of the strain ellipsoid and mature folds which are much tighter and have undergone considerable shear strain since nucleation; these have axes sub-parallel to the X-direction of the strain ellipsoid. All intermediate types between these two end members are present.

The model of Esher and Watterson (op. cit.) can be used to explain the distributions found in the Moine mylonites. Because of the fine-grained and planar nature of the mylonites, once the mylonite zone had initiated, it underwent deformation at much higher strain rates than the surrounding rocks (Etheridge & Wilkie, 1978). The process of nucleation and rotation of folds would therefore occur faster within the mylonites than in rocks outside. From this relationship it can be predicted that within the mylonites there should be a greater bias towards isoclinal folds with axes sub-parallel to the movement direction.

The spatial distribution of folds described in relation to other thrust zones is similar to that in the Moine mylonites. The best description comes from the Woodroffe Thrust of central Australia (Bell, 1977). Other examples from Scandinavia (Williams, 1977) and the European Alps (Pfiffner, 1981) are also known to exist. All these examples show 180° distributions of fold axes in association with mylonite belts.

A problem with the mechanism of nucleation and rotation of folds, however, is that folds which nucleate exactly parallel to the Y axis of the strain ellipsoid do not rotate during simple shear. However, Ramsay (1967) and Treagus and Treagus (1981) have shown that if the original lithological layering was oblique to the axes of the strain ellipsoid during deformation, folds would initiate oblique to the Y axis. This is important since it has been shown that the pre-mylonite foliation is oblique to the mylonite foliation just south of Ullapool (Section St.2.a).

Sanderson (1973) used the progressive rotation of folds to obtain estimates of strain values (Section MT.1.a). He demonstrated that if fold axes initiated at angles of only 10° from the Y axis of the strain ellipsoid, rotation would be swift, forming a single concentration close to the X axis. Thus the obliquities seen south of Ullapool are sufficient to explain the occurrence of rotated minor folds within the mylonite.

A problem arises in relating the relative ages of F3 folds in Fannich and the deformation associated with mylonitization in sub-area (d). There seem to be two plausible explanations of the relationship. They are as follows:

a) The mylonites above the Moine Thrust formed at the same time as F3 in Fannich and represent a zone of higher strain.

b) The mylonites and related folding post-date D3 in Fannich and overprint the earlier deformation increasingly towards the thrust zone causing flattening and rotation of earlier F3 folds close to the mylonites.

The problem arises because both models would explain deformation of post-D2 pegmatites, the increasing tightness and rotation of F3 folds into the mylonite zone, and the increased number of folds in the mylonites and adjacent areas. A full discussion is however given in relation to the microfabrics of the area (Section MT.5.a,b) and the relative timing of D3 in Fannich and the formation of the mylonites above the thrust (Section MT.2.d).

In conclusion, it is apparent from the present study that the earliest folds associated with mylonitization deform a set of pegmatites which post-date the Sgurr Beag Slide and are found throughout a continuous traverse to Fannich. The mylonite-related folds form by nucleation normal to the transport direction and rotate towards it during progressive shear strain. The mylonite formation therefore entirely post-dates the Sgurr Beag Slide and formed at the same time as, or later than, D3 of the Fannich area. It has not been possible to determine the exact relationship between D3 and the mylonites on the evidence obtained from the analysis of fold structures.

MT.2.b <u>Analysis of lineations from Fannich to the Moine Thrust zone</u>

Two separate lineations have been recognised in the area west of Fannich. One, spatially associated with the high strain areas of the Sgurr Beag Slide zone, the other a penetrative lineation spatially associated with the mylonites above the Moine Thrust.

The intensity of the lineation associated with the Sgurr Beag Slide is inversely proportional to the distance from the slide (Section SBS.1.b). It is a stretching lineation, defined by crystallographic and long axis alignment of quartz, feldspar and mica grains within the general Sgurr Beag Slide related foliation seen in psammitic rocks. It is also seen in shape fabrics developed in the lits of the migmatitic Glenfinnan division rocks.

The lineation is strong around the shores of Loch a Broain (fig.4), close to the outcrop of the slide. Along the shores of Loch Broom however, the lineation becomes weak and in many of the thick psammite units, no lineation is visible.

The Sgurr Beag Slide related lineation is sub-parallel to F3 in sub-area (c) but is sometimes seen folded around F3 (fig.51).

The mylonite-related lineation is spatially restricted to the area adjacent to the Moine Thrust. It occurs sporadically up to 4 km from the thrust and becomes dominant about 2km from the thrust. It is defined

FIGURE 59 Variation of fold axes with distance above thrust

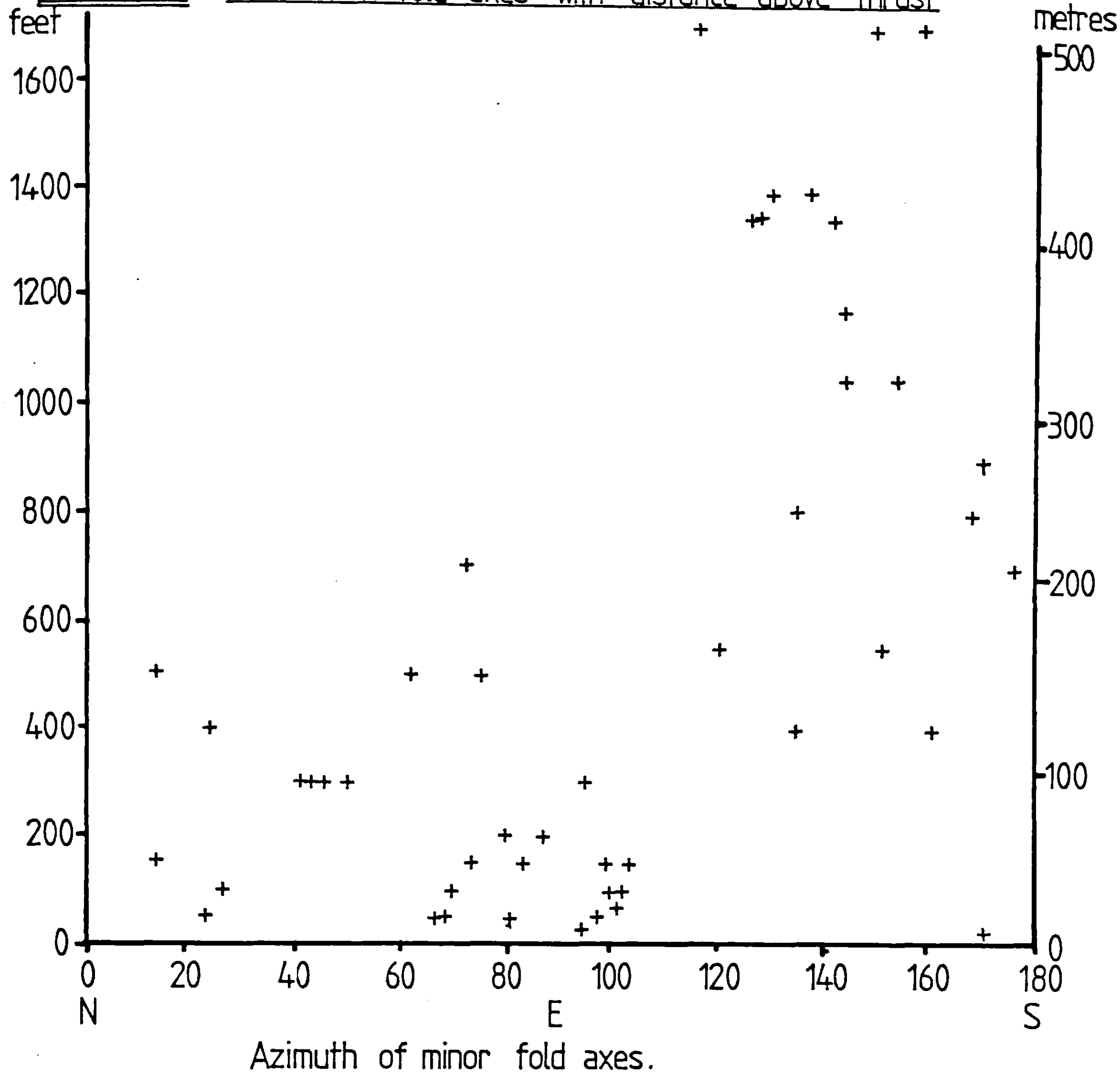
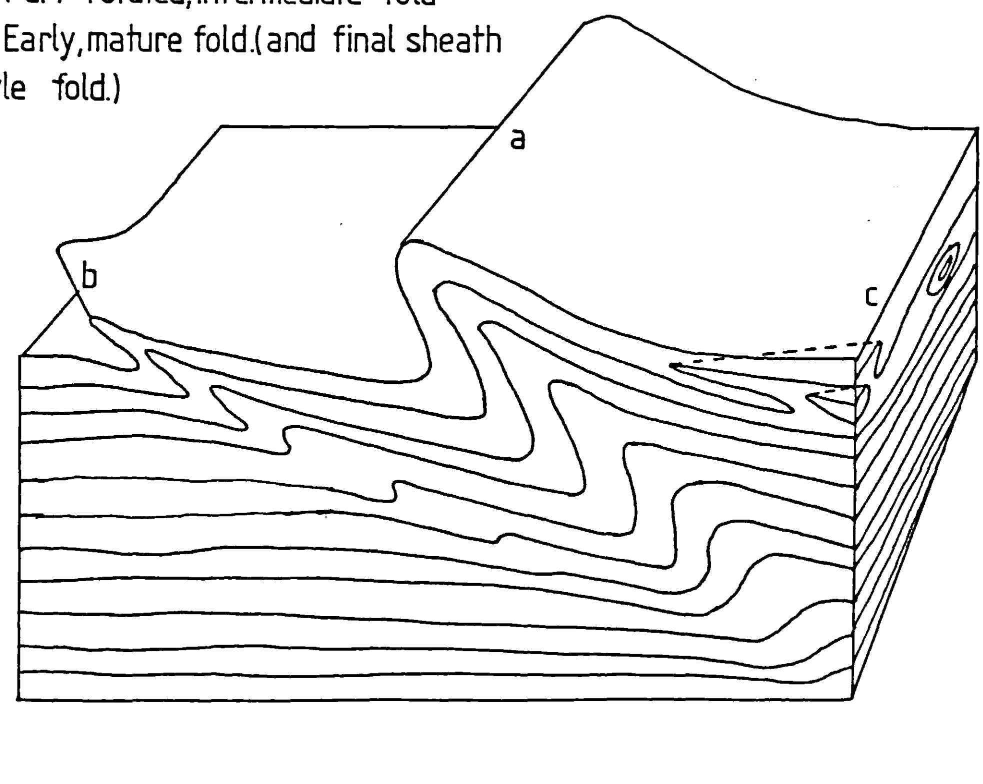


FIGURE 60. Diagramatic representation of fold rotation

a) Late-immature fold.

b) Part-rotated, intermediate fold

c) Early, mature fold.(and final sheath style fold.)



by the strong alignment of elongate quartz and feldspar grains in psammites and pegmatites within the plane of mylonite foliation.

The lineations of sub-area (c) are dominated by the Sgurr Beag Slide (fig.61); the data has a strong concentration trending $10^{\circ} \rightarrow 155$ with a spread extending along the great circle representing the general foliation.

Lineations plotted from sub-area (d) show a greater spread, with a maxima trending $10^{\circ} \rightarrow 100$ but a tail extending towards that deduced for the Sgurr Beag Slide in sub-area (c) (fig.61). The increased scatter of data other than the tail on this plot is due to later folding in the area of the mylonites.

The distribution of lineations in the two areas gives the impression of an early Sgurr Beag Slide related lineation being gradually, passively rotated with the incoming of mylonitic rocks so that it comes to parallel the tectonic movement direction of the mylonites. However, this is not the case. Both lineations can be distinguished, where strongly developed, in a single exposure (fig.62). This affords an important clue to the inhomogeneous style of deformation during mylonite formation.

As strain increases towards the mylonites, up to 4 km from the thrust, the older foliation/bedding planes become disrupted so that pods of relict Sgurr Beag Slide foliation, bounded by platey shear zones with a new mylonitic fabric are developed (figs.62,69). The geometry of the shear zones resembles that of shear bands (Platt and Visers, 1980) (Section MT.2.e). Within the pods, an older Sgurr Beag Slide related lineation is preserved with the early planar fabric. At the outer edges of the pod, in the platey high strain zone, the new mylonitic lineation is dominant trending to 100°.

FIGURE 61. Stereonets of Sgurr Beag slide related and Moine Thrust mylonite related lineations

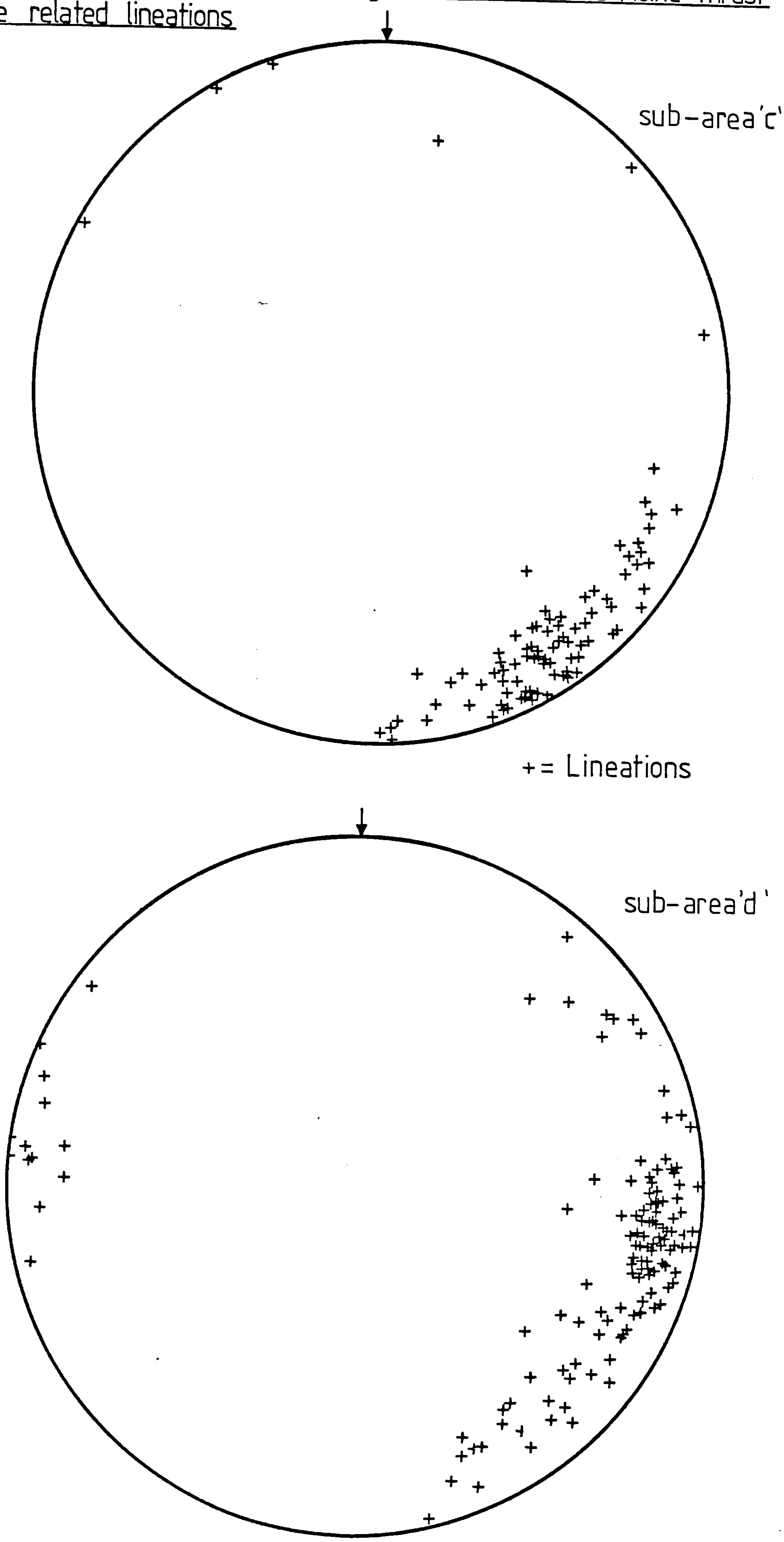


FIGURE 62. Both types of lineation are found within a single exposure

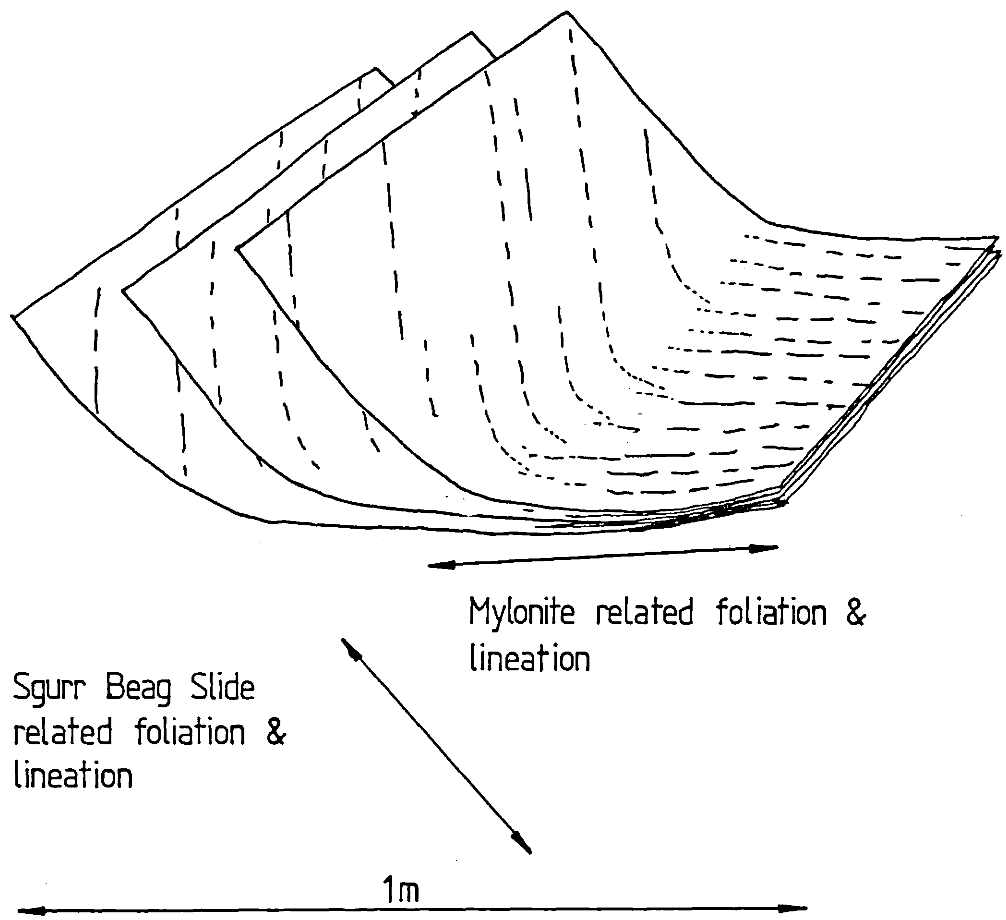
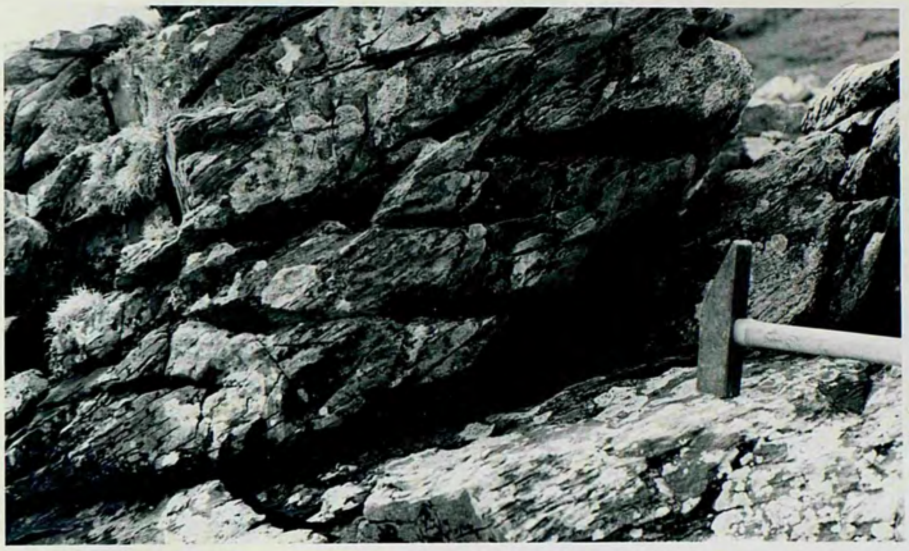
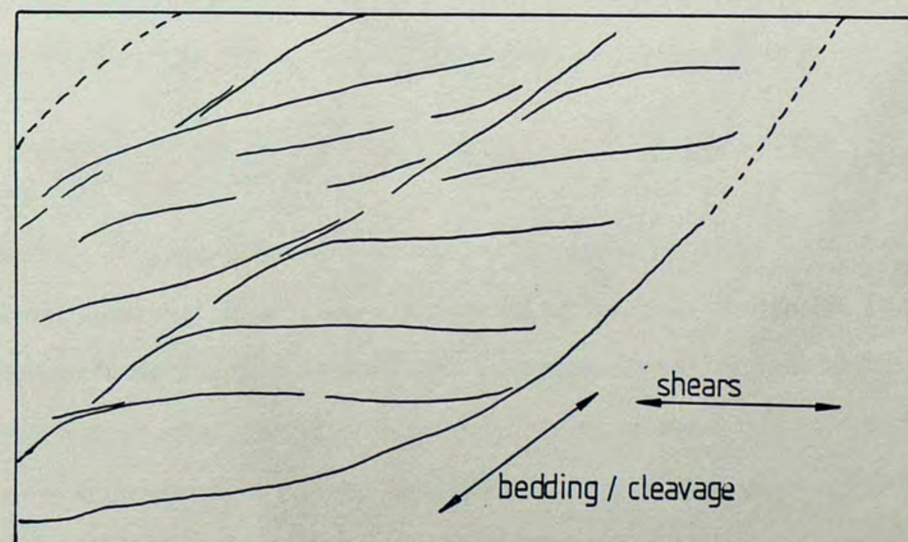


FIGURE 63. A boudin between shear bands in the Inverbroon psammites on the shore of Loch Broom 1km from the Moine Thrust. It is in the process of being broken into smaller boudins.



Flattened boudins on the shore of Loch Broom, see figure 70.



Sketch of above showing detail.

At early stages in the formation of the shear bands, the inner pod remains at low strain and the old lineation suffers at most, bulk rotation. As strain increases, the pod becomes flattened and sometimes breaks down into many smaller pods (fig.63). Eventually the pod becomes so deformed that the internal and external fabrics are sub-parallel (fig.70).

Approaching the thrust zone the pods are generally progressively more deformed so that the early fabric is completely overprinted within 1km of the thrust. Occasionally, however, the early fabric survives within deformed pods and lenses in the mylonites. One such pod exists only 75m above the Moine Thrust west of Creag Rainich (fig.4). It is over 50m long and approximately 5m wide, preserving only partially deformed D2 fabrics and a strong Sgurr Beag Slide related lineation.

The change in azimuth and style of lineations therefore reflects qualitatively the increase in strain associated with thrusting from Fannich to the Moine Thrust zone. The early D2 lineation, related to movement of the Sgurr Beag Slide is rotated toward the X direction of the mylonite-related strain ellipsoid (fig.64). During the deformation, the early metamorphic textures which defined the planar and linear fabrics are broken down to form the new mylonitic foliation and lineation.

MT.2.c Analysis of the relationship between post-D2 pegmatites and the Moine Thrust

Two suites of pegmatites are found in the area; a rare early pre-D2 set comprising quartz + plagioclase + muscovite + garnet + opaques is strongly deformed during D2 and exhibits a strong S2 schistosity defined by the alignment of mica (Section SBS.1.b); a later set post-dates D2 and is ubiquitous throughout the area, occurring in all lithologies. These generally range between 1 and 5cm thick but occasionally reach 10cm, stretching for several metres in length when undeformed. Their most distinctive feature in the field is the presence of large pink K-feldspar laths up to 2cm in diameter. The rest of their mineralogy is up to 50% quartz with subsidiary amounts of plagioclase and trace muscovite, apatite, epidote and opaques. The pegmatites are often associated with pyrite which occurs both within them and in the wall rock. It is thought that the pyrite is late stage since it shows little or no deformation features.

In the most highly strained zones of the Sgurr Beag Slide, pegmatites of the later suite cross-cut the platey psammitic banding S2 at a high angle and are undeformed (fig.50). However, the pegmatites pre-date D3 and are folded in minor D3 folds (fig.50). The style of deformation of the later suite of pegmatites provides an indication of the intensity of D3 and post-D3 deformation. In sub-area (a), the feldspars form augen separated by stringers of quartz and the pegmatites are folded in pelites (fig.50). They show relatively low deformation in the psammites but are deformed in minor F3 fold noses. In sub-area (b), pegmatites are flattened towards S0/S2 foliation planes.

In sub-area (c) there is a strong contrast between the deformation of pegmatites; in psammites they remain at high angles to the bedding/ foliation, often exhibiting small crush zones; in pelites the pegmatites become highly folded and boudinaged (fig.50). In several examples (fig 50) pegmatites can be traced passing from psammites where they show low strain to pelitic bands where they become highly strained. The relative displacement of the pegmatites within the pelites indicates a shearing parallel to the lithological banding towards the NW. An apparent shear strain of $\gamma = 8$ was measured across a pelite 50cm wide in a small quarry over 8km from the thrust. Such shear strains as far as 8km from the Moine Thrust imply much higher strains close to the thrust, as strain can be shown increasing towards the Moine Thrust zone (Section MT.2.a).

Towards the west of sub-area (c) and in sub-area (d) as the deformation of pegmatites increases in intensity in pelitic rocks, they also become strongly deformed in psammitic lithologies. In sub-area (d), the pegmatites show tight - isoclinal folding (fig.50) and approach parallelism with the mylonitic foliation. They exhibit pinch and swell structures within psammites. Some thick pegmatites show internal deformation, leaving lenses of relatively undeformed material surrounded by a highly deformed matrix (fig.65).

The pegmatites illustrate the inhomogeneous nature of deformation by sudden changes in their orientation to foliation within single psammitic units as well as across psammite/pelite boundaries. They are folded by the early tight to isoclinal folds formed during mylonitization.

Within the mylonites, pegmatites are nearly parallel to the foliation phase. They seem more rare but this may be attributed to the breakdown of K-feldspar grains which were used to identify them further east, making them difficult to detect in the mylonites. Many more highly deformed quartz veins are seen in sub-area (d).

It can be concluded that a widespread suite of pegmatites was intruded after the cessation of movement on the Sgurr Beag Slide. The pegmatites were deformed during the D3 deformation in Fannich and during mylonite formation; they record the rapid increase of strain towards the mylonites and the inhomogeneous nature of the deformation depending upon lithology.

FIGURE 65.Internal deformation within a post-D2 pegmatite.Note the large feldspar remaining in the central zone.

See folder for fig 64

MT.2.d The Relationship between D3 fabrics in Fannich and the mylonites above the Moine Thrust

A distinction is made in this Section between "correlation": in the sense of deformation phases seen within the mylonites above the Moine Thrust correlating directly with those within the inner mobile zone; and a relationship between the deformation phases of the two areas. The fabrics developed during D3 can be traced westward towards the mylonites and they increase in intensity westward. It is difficult to establish an absolute correlation with the mylonites although a relationship between the two zones is evident. The possible alternatives will be delineated and discussed.

Any attempt to make a structural correlation from the mylonite zone at the Moine Thrust to structures within the Moine nappe is extremely difficult because the formation of a zone of mylonites at the Moine Thrust would result in massive stress relief within the overlying Moine nappe. As strain increases, the grainsize of the rocks is reduced and aligned, faster deformation mechanisms therefore become dominant (Section MT.5.b and Etheridge & Wilkie, 1979; White, 1979). This causes relative softening in the mylonite zone, especially in rocks containing phyllosilicates (White, 1979). Thus the main movements on the mylonites probably correlate with a static period in the deformation history of the Moine nappe, the nappe acting as a passive overriding block. Any correlations attempted would therefore represent a correlation between those events within the Moine nappe which are associated with the build-up of stress prior to the initiation of mylonitization and the oldest structures preserved within the mylonites.

D3 fabrics in Fannich have been described elsewhere (Section MT1a,b). In the area west of Fannich, around Loch a Braoin and the shores of Loch

Broom (fig.4), minor folds are rarer than in Fannich. It might be surmised that the strain in this area is therefore lower than in Fannich but the fold hinges of F3 minor folds contain penetrative fabrics in psammites as opposed to the incipient fabrics found in Fannich. The pelites, which show crenulations in Fannich exhibit transposed penetrative fabrics in the hinges of minor folds west of Fannich.

Microfabrics indicate the initiation of breakdown within psammites and an extension crenulation cleavage is associated with progressive breakdown of grainsizes in pelites (Sections MT.5.a & MT.5.b).

The extension crenulation cleavage is also seen in the hinge zone of the major F3 Achnasheen anticline (fig.49). The presence of the extension crenulation in the same orientation, both in the F3 fold hinge and towards the mylonites, indicates a relationship between the two.

In the core of the anticline, exposed in cliffs on the west side of Corie Beag (fig.2), mesoscopic vertical and horizontal limbs ($\lambda \doteqdot 100$m) are exposed, parasitic to the major fold. Vertical limbs exhibit a compressional crenulation whereas the shallow lying limbs exhibit an extensional crenulation similar to that seen in the west. The shallow-lying limbs seem to have undergone higher relative deformation, (the pelites having undergone more extensive grainsize reduction than the vertical limbs and quartz veins being more sheared) in the same way as that seen between the relative deformation of the major F3 limbs (Section MT.1.a). Thus compressional and extensional cleavage development is related to F3 fold limbs. The development of an extension crenulation is an important mechanism in the breakdown of early fabrics as the mylonite zone is approached, a feature which will be discussed further in Section MT.5.a.

In Section MT.1.a it was shown that the hinges of F3 minor folds on shallow limbs of the major D3 folds have undergone rotation relative to folds on the steep limb. Folds in the shallow limb of the Fannich syncline, west of Fannich also exhibit orientations at an angle to the regional N-S trend of the F3 major folds.

The extension direction deduced from the rotation of folds in Fannich coincides closely with the extension lineation which can be seen in the mylonites (figs.55,61), thus suggesting that both formed under a similarly oriented stress field.

The possible alternative relationships between the mylonites and D3 deformation in Fannich are as follows:

a) The mylonites pre-date D3 in Fannich.

b) The D3 event in Fannich is equivalent to the earliest part of the mylonite deformation phase.

c) The D3 event in Fannich pre-dated mylonitization and fabrics associated with the mylonites were superimposed upon D3 fabrics in Fannich.

Relationship (a) is unlikely on present evidence, which is fully discussed in Section MT.2.a. Additionally, if the mylonites pre-dated the major D3 folding, they should be folded by F3 and hence reappear within the central Fannich area on the eastern side of the Fannich syncline.

It is more difficult to decide between relationships (b) and (c), superficially both relationships could explain the rotation of folds close to the mylonites and within the low lying limb of the Achnasheen anticline and also the extensional crenulation within the hinge of the major D3 folds. The two events may be related or the mylonite forming event overprint the earlier D3 folds and fabrics, preferentially on the shallow dipping limbs.

In Section MT.1.a it was shown that the hinges of F3 minor folds on shallow limbs of the major D3 folds have undergone rotation relative to folds on the steep limb. Folds in the shallow limb of the Fannich syncline, west of Fannich also exhibit orientations at an angle to the regional N-S trend of the F3 major folds.

The extension direction deduced from the rotation of folds in Fannich coincides closely with the extension lineation which can be seen in the mylonites (figs. 55,61), thus suggesting that both formed under a similarly oriented stress field.

The possible alternative relationships between the mylonites and D3 deformation in Fannich are as follows:

a) The mylonites pre-date D3 in Fannich.

b) The D3 event in Fannich is equivalent to the earliest part of the mylonite deformation phase.

c) The D3 event in Fannich pre-dated mylonitization and fabrics associated with the mylonites were superimposed upon D3 fabrics in Fannich.

Relationship (a) is unlikely on present evidence, which is fully discussed in Section MT.2.a. Additionally, if the mylonites pre-dated the major D3 folding, they should be folded by F3 and hence reappear within the central Fannich area on the eastern side of the Fannich syncline.

It is more difficult to decide between relationships (b) and (c), superficially both relationships could explain the rotation of folds close to the mylonites and within the low lying limb of the Achnasheen anticline and also the extensional crenulation within the hinge of the major D3 folds. The two events may be related or the mylonite forming event overprint the earlier D3 folds and fabrics, preferentially on the shallow dipping limbs.

Relationship (b) is preferred because it is difficult to explain why a later deformation should affect only the shallow lying and not the steep limbs of F3 major folds. Further, relationship (b) also explains the extensional and compressional cleavages and their relationship to F3 folding. Relationship (c) requires that the compressional crenulation cleavage formed during D3 within the steep limbs of the major folds and that during D3 the shallow limbs were unaffected. The extensional circulation would then, following relationship (c) have to form within F3 shallow limbs after the folds had formed, superimposing D3 but not affecting the crenulations in the steep limbs.

Thus, it appears most likely that D3 deformation in Fannich occurred as part of the initiation of mylonitization at the Moine Thrust zone.

MT.2.e Mechanism for the formation of extensional structures within the rocks west of Fannich

In the area adjacent to the Moine mylonites on the gently inclined western limb of the Fannich syncline, structures are observed which indicate extension sub-parallel to the existing foliation. These structures are present in both pelites and psammites. In pelites (Section MT5.a), they form small shear bands oblique to the early foliation. Relict foliation is seen, partially deformed, in lenses or eyes 0.25-2mm long. Microfabric evidence indicates a progression: shear band formation, flattening and re-deformation. In psammites, similar structures occur within 2km of the thrust belt but on a larger scale, the shears are up to 10cm wide and structures of this size are generally known as 'Foliation boudinage' (Platt & Visers, 1980); they have also been found related to extension in glacier ice (Hambrey & Milnes, 1975).

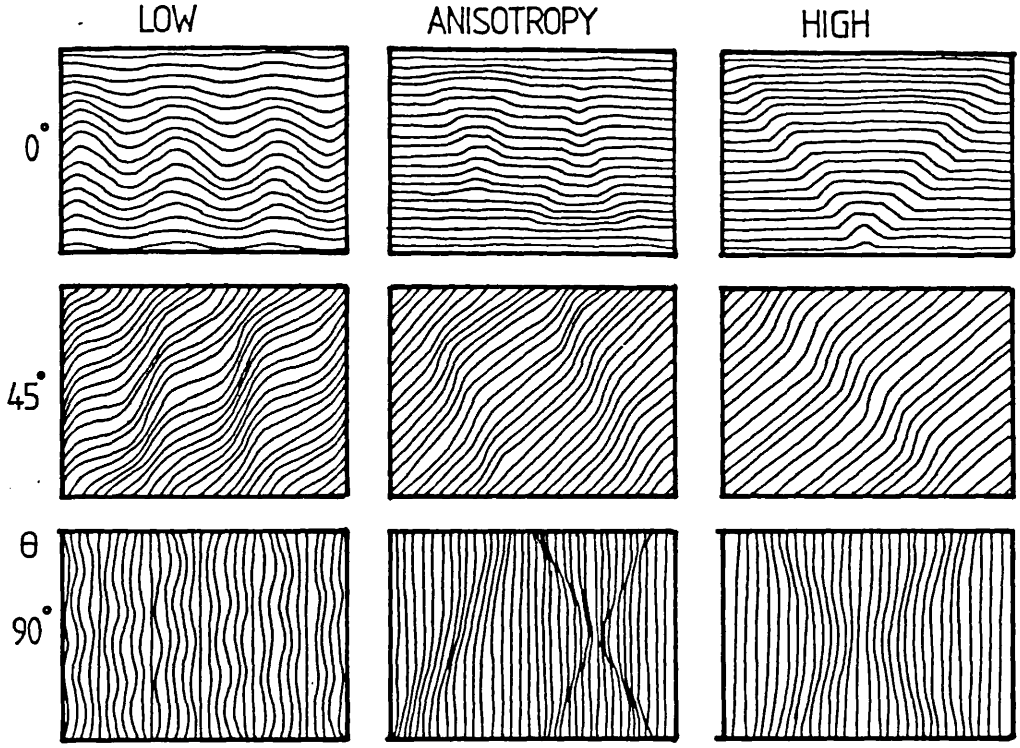
Extensional structures on several scales have been reported from mylonite zones by Berthe et al. (1979,a,b); White (1979), Platt and Visers (1980), White et al. (1980) and Gapias and White (1982).

Early theoretical work by Biot (1965) and work by Cobbold et al. (1971), Cosgrove (1976) and Means and Williams (1972) has illustrated that when rocks containing a planar anisotropy are deformed, they produce a series of structures dependent upon the orientation of the major strain axes and the degree of anisotropy (fig.66). In general, compression parallel to the plane of anisotropy produces kink and buckle folding; compression perpendicular to the layering produces boudinage or pinch and swell structures (fig.66). If the major compression axis is intermediate, shears are produced at an angle to the original foliation (fig.66).

Platt and Visers (1980) describe crenulation structures seen in the field which represent extension near parallel to layering and are related to the shears predicted in the theoretical work. Platt and Visers show that their extension crenulations were produced as a result of weaknesses formed by the structures predicted by Biot and Cobbold et al. (op.cit.). The crenulations or shear bands are shown to be associated with grainsize reduction and dynamic recrystallisation, but as Platt and Visers note, no mineralogical changes are observed associated with the shear bands. From this they deduce that the deformation occurs without any volume change within the shear bands. This is also a feature of the shear bands in pelites west of Fannich (Section MT.5.a).

In order for the shear bands to form, there must be a weakening process so that deformation becomes localised within the bands. The grainsize within shear bands west of Fannich are always smaller than the grainsizes within the relict lenses, indicating that deformation and dynamic recrystallisation have reduced the grainsizes within the shear bands. This produced a marked softening in the shears and therefore higher strain rates (White, 1979,b; Etheridge & Wilkie, 1979) so that deformation can be accomodated. The shear bands develop because bulk deformation

FIGURE 66.



Some possible modes of expression of "buckling instabilities" in materials with different anisotropy and at different angles (θ) to the maximum compression direction. The anisotropy was defined by multiple parallel layers of two rheologically different materials. (after Cobbold _et al_ 1971) cannot accomodate the imposed strain rate. Gapias and White (1982) estimated that strain rates within the shear bands were 5 times faster than those for bulk deformation.

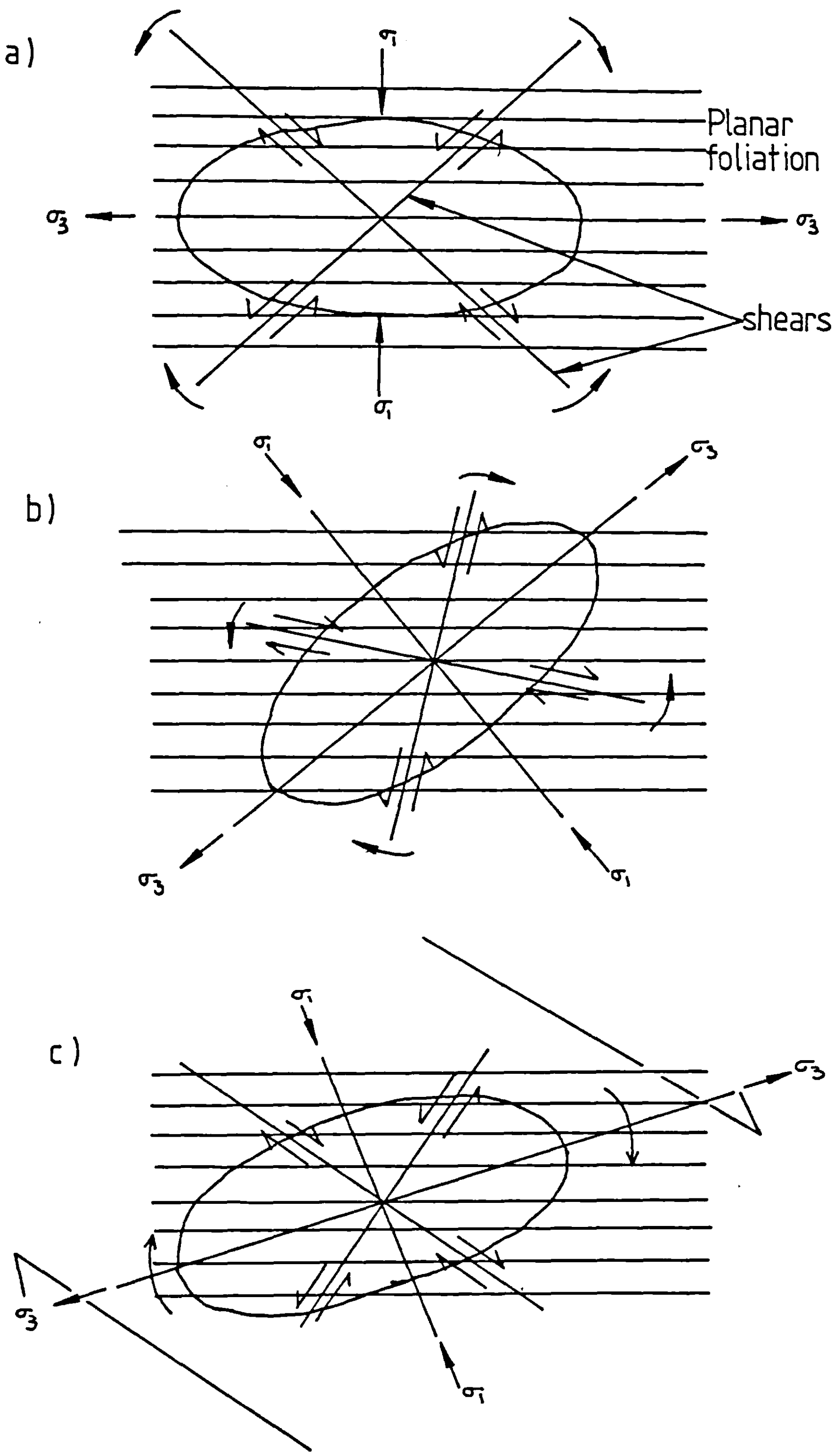
Platt and Visers divide the structures into three types: symmetrical-coaxial, a symetrical-coaxial and non-coaxial (fig.67). The first type, symmetrical-coaxial, produces two conjugate sets of shear bands which accommodate the strain although one set is often dominant. The second type, a symetrical-coaxial, produces one dominant set of shear bands at a high angle to the banding. This set is dominant because the sense of shear on the high angle set is consistent with continued deformation as it rotates toward the major extension direction (fig.67). The low angle set is weaker for a symetrical-coaxial deformation because the sense of shear across them is wrong for continued deformation.

The third type, non-coaxial, produces one set of shear bands, a low angle set at 45°-n (fig.67) from the foliation. The high angle set rotates towards the extension direction and rapidly become inactive.

The present study has revealed structures west of Fannich similar to those described by Platt and Visers but they exhibit some features not previously described in the literature. The structures closely resembling those in the area west of Fannich are described by Platt and Visers in the Betic movement zone of the Alpajarride nappe in southern Spain. In this movement zone, they are associated with a flat-lying mylonite and thrust belt associated with retrogression of amphibolite facies rocks to greenschist facies.

Platt and Visers measured the angles between the shears and an enveloping surface in order to gain a mean value of the angle. From work on the pelite microfabrics and psammite structures, the range of

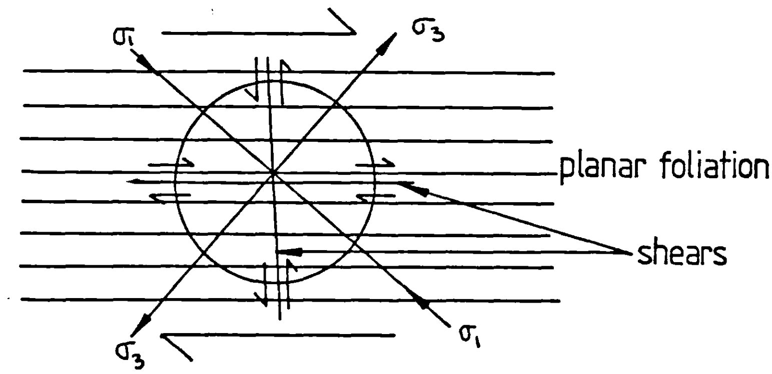
FIGURE 67. Three types of shear band (after Platt and Visers 1980)



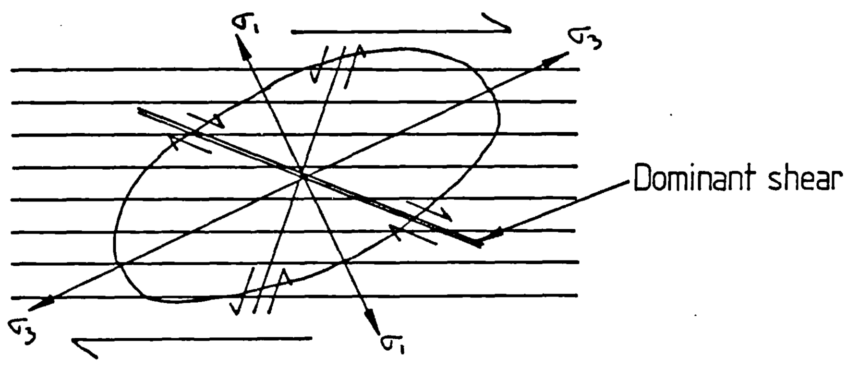
a) Coaxial deformation, with σ_1 parallel to the foliation.
b) Coaxial deformation, with σ_1 at an angle to the foliation.
c) Non-coaxial deformation σ_3 rotating towards parallelism with the shear direction.

FIGURE 68.

Simple application of the Platt and Vissers non-coaxial model



Generalised shear band west of Fannich



angles seen west of Fannich is very similar to that described by Platt and Visers (Section MT.5.a). However, west of Fannich, the whole range of angles may be found evenly distributed in one thin section while in others, only a narrow range is present; there is also a general decrease in angle towards the thrust.

The structures described by Platt and Visers (op.cit.) occur exclusively within mylonites and are related to late mylonite movement, pre-dating all post-mylonite folding. West of Fannich however, the shear bands are found in pelites throughout the area, even 14km east of the Moine Thrust in the core of the Fannich synform. They initially formed before the mylonites but continued to form during the process of mylonitization, comprising an important mechanism of grainsize reduction. Some represent a late phase and are very similar to structures described by Platt and Visers (op.cit.) in the Betic movement zone and by Peach et al. (1907) in the "oyster shell" rocks from Eriboll.

The structures west of Fannich do not fit Platt and Visers' symmetrical coaxial type (fig.67) because the shear bands show consistently only one sense of movement, towards the west. The a symetrical-coaxial type does not account for field observations west of Fannich either (fig.67). The shear bands seen have the wrong orientation and sense of shear for the deformation exhibited by the displacement of quartz veins and pegmatites across pelitic layers (fig.50).

The non-coaxial type seems to fit the structures west of Fannich best. One set of shear bands has developed at a low angle to the pre-existing foliation. However, in the example presented by Platt and Visers to illustrate 'non-coaxial' deformation, shear bands are shown propagating parallel to the boundaries of the imposed shear zone (fig.67). This solution assumes that shear bands begin to propagate immediately as deformation starts. If this model is applied directly to the pelite bands of Fannich (fig.68), it predicts shear band propagation sub-parallel to the early schistosity. However, as Platt and Visers point out in a separate part of their paper (Platt & Visers, 1980, p.400), the shears propagate along planes which are the product of pinch and swell structures caused by the early deformation (fig.64). This implies significant deformation prior to the initiation of shear bands. If the initial strain is taken up in homogeneous strain and shear parallel to the layers, the direction of maximum extension rotates towards parallelism with the early schistosity (fig.68) and shear bands then begin to form.

The highest mean angles measured between shear bands and an enveloping surface (the boundaries of the pelite horizon) west of Fannich are about 25° (Section MT.5.a). Assuming that the shear bands at these angles are immature, this indicates a maximum extension direction at about 20° from layer parallel. A shear strain of about $\gamma=3$ is required to produce such an extension direction prior to the formation of shear bands for this model. Although, intuitively, this seems a large strain, pegmatites within pelites over 8km from the thrust exhibit shear strains of $\gamma=8$ and these contain microfabrics showing only intermediate stage breakdown. All pegmatites seen in the flat belt west of Fannich show high shear strain where they cut pelites (fig.50).

As deformation progresses, the shear bands rotate towards the extension direction and become inactive because the shear couple across them decreases. In the pelites, shear bands which have become inactive are overprinted by the new high angle shears. The structures west of Fannich differ from structures previously described because they are continually formed, flattened and redeformed (Section MT.5.a).

An equivalent of shear bands (foliation boudinage) is seen in psammites adjacent to the Moine Thrust; however these structures show no evidence of the polyphase shear band formation. It is thought that the initial formation of the large scale shears disrupts the anisotropic nature of the psammites to such an extent that no further shears initiate. The shears become flattened passively towards the foliation plane of the mylonites.

In conclusion, the shear bands formed in psammites and pelites west of Fannich differ from shear bands previously described in the literature in several ways:

a) They pre-date the mylonitization event with which they are associated.

b) They occur on several scales in different lithologies.

c) They show a greater variation in angles between the shears and the major shear plane and a relationship between strain and this angle.

d) They initiated above the biotite isograd (Section MT.5.a).

Despite these differences, the mechanism of formation is essentially the same as previously reported examples. It is an inhomogeneous deformation related to a planar anisotropy in the rock. White (1979,a) has suggested that shear band development is associated with temperature drop during deformation. Structures west of Fannich developed in rocks which were deformed and recrystallised above almandine grade. The shear bands therefore initiated after a temperature drop, since garnet is not recrystallised during early shear band formation and is broken down to chlorite in the later shear bands. It is thought that the instability of the coarse-grained anisotropic D2 (Sgurr Beag Slide) fabrics, under the high strain rates imposed during mylonitization, were the main cause of the initiation of shear bands.

MT.3 Structures associated with the Mylonites

MT.3.a Foliation boudinage in psammites

In an area adjacent to the mylonites about 2km wide, the early Sgurr Beag Slide related foliation (S2) in psammites is disrupted by the mylonitization process. Initial deformation associated with mylonites is inhomogeneous forming small shear zones which cross-cut the early foliation and occasionally simple non-rotational boudinage (figs.62,69).

These structures are best developed between the outcrop of the Sgurr Beag Slide on Meall Dubh and the Moine Thrust above Loch an Nid (fig.4). The ~~early~~ S2 foliation is strong in this area and the development of such good examples of the later fabric seems to be associated with the strength of the earlier fabric.

The term "foliation boudinage" was defined by Milnes (see below) and refers to the feature called "internal boudinage" by Cobbold et al. (1971). Hambrey and Milnes (1975) illustrated the existence of foliation boudinage in glaciers from the Swiss Alps. They showed that boudinage had arisen in strongly foliated ice which was lithologically isotropic but structurally anisotropic. The length of boudins recorded by Hambrey and Milnes (1975) was generally around 5m but occasionally more. They noted that the laminations were the same both inside and outside the layer showing boudinage, in other words the formation was not dominated by lithologies. They also noted the strange orientation and shear senses across the boudinage considering the movement direction of the ice. Both these features are apparent in the structures within strongly foliated psammites adjacent to the Moine Thrust.

The structures adjacent to the mylonites consist of lenses or pods of gently folded D2 foliation ranging from 50cm to 5m long, separated by

FIGURE 69. Rotational and non-rotational boudinage in the psammites above the Moine Thrust.

a) Rotational

b) Non-rotational.

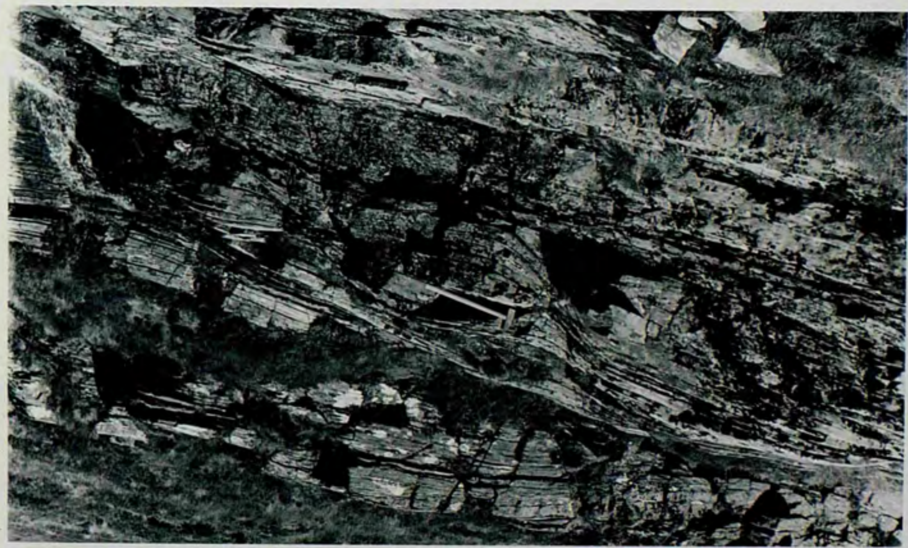
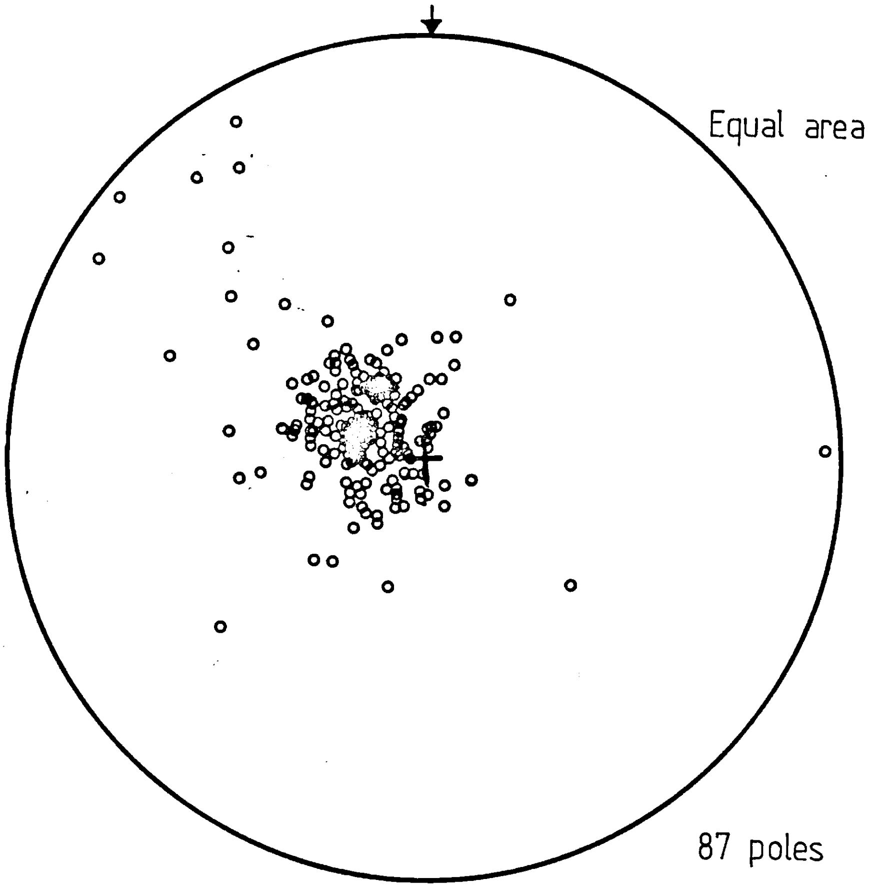


FIGURE 70. Low angle shear bands in psammites close to the Moine Thrust along the shores of Loch Broom.

FIGURE 71. Stereonet of Mylonite Foliation



o = Poles to mylonite foliation

oblique bands of platey mylonitic rock. The D2 fabric is deformed in these bands which exhibit high shear strain (fig.62); the early fabric and associated lineation are destroyed in these zones (Section MT.5.b). The shear zones cross-cut the early foliation (S2) at angles of 20°-30° in this area and always dip down towards the west (fig.69). The shears are generally about 10cm wide and strike approximately parallel to the mylonites; this means that the long axes of the lenses are parallel to the mylonite belt trending approximately 010°.

Microfabric analysis has shown that a relatively undeformed D2 fabric remains within the pods whereas a fabric identical to that seen in psammitic mylonites above the Moine Thrust is established in the shears (Section MT.5.b).

Further north along the shores of Loch Broom, similar structures are developed but are not as regular as those by Loch a Braóin. They are generally smaller and the shears exhibit lower angles to the early foliation. In many cases they are only recognised because of a slight discordance between adjacent foliations (fig.70).

After initial formation the shears rotate passively towards the XY plane of the strain ellipsoid (Section MT.2.e). In pelites, shears become sub-parallel to the existing foliation and new shears initiate at high angles. However, in psammites the formation of shear bands seems to disrupt the planar fabric of the rock so much that no further shear bands form. As the shears rotate towards the XY plane during progressive strain, the lenses or pods of early foliation become more deformed. They generally deform homogeneously, becoming flattened into the mylonitic fabric (fig.70); this can be seen along the shores of Loch Broom where it is sometimes difficult to see the pods close to the mylonites because the angle between internal and external fabrics is less than 10^{o}. Occasionally they deform by forming similar lenses within the larger ones (fig.63).

The small shear zones in psammites have the same geometry and shear sense as shear bands described on a microscopic scale (Gapias & White, 1982) although some examples up to 15cm long have been reported from the Varoise Massif of the French Alps (Platt and Visers, 1980).

The foliation boudinage within psammites west of Fannich corresponds to the 'non-coaxial' type (Platt & Visers, 1980) indicating overthrusting towards the NW. The shears between the pods contain chlorite, albite and quartz, whereas the relict pods occasionally retain partially chloritized biotite.

The structures west of Fannich are larger than has previously been reported in the literature for rocks (cf. Hambrey & Milnes, 1975). It seems that the scale of the foliation boudinage or shear bands depends upon the scale of anisotropy in the pre-existent foliation of the rocks (Section MT.2.e).

MT.3.b A Comparison of deformation features within the mylonite and Sgurr Beag Slide zones

Several features indicate an increase in the state of strain between Fannich and the Moine Thrust to the west: Analysis of changes in the orientation and style of F3 folds (Section MT.2.a); changes in quartz veins although these may, in part, be syn-deformational veins towards the Moine Thrust; post-D2 pegmatites which are widespread throughout the area (Section MT.2.c), deformation of sedimentary structures and formation of a mylonitic foliation. It is noteworthy that these are very similar features to those used to describe the increase in strain towards the Sgurr Beag Slide in Fannich (Section SBS.a.b) and used by Rathbone and

Harris (1979) in describing the increase in strain adjacent to the Sgurr Beag Slide at Lochailort and Garve. Similarities and distinctions between the sizes and features of the Moine Thrust and Sgurr Beag Slide can thus be discussed.

Rathbone and Harris (1979) erected a series of zones describing the increasing strain towards the Sgurr Beag Slide. The same exercise can be performed on the Moine Thrust west of Fannich in order to compare the width and distribution of strain within the deformation zone.

At outcrop the mylonites dip at approximately 15° towards the east and strike N-S (fig.71). In order to estimate the thickness of zones it has been necessary to assume that the mylonites continue to dip at approximately 15° under the area west of Fannich (a distance of 10km). The brittle Moine Thrust between Ullapool and Loch an Nid has been used as a datum plane. Unfortunately it is probable that the brittle Moine Thrust climbs up and down through a wide zone of mylonites. (A 50m thickness of mylonites is seen above the Moine Thrust west of Fannich, whereas 600m is recorded adjacent to Loch Eriboll (Soper & Wilkinson,1975).)

Four zones of deformation have been delineated (fig.72):

D) Greater than 1000m above the Moine Thrust:

Undulatory sedimentary banding - strongly varying thicknesses of bedding. Sedimentary structures unaffected by the mylonite related deformation. Quartz veins and pegmatites exhibit a great variety of orientations often lying at angles greater than 30° from bedding. Pegmatites and quartz veins show little or no signs of internal deformation.

C) 1000-500m above the thrust:

Sedimentary banding thicknesses are still variable but sedimentary structures are flattened. The angular discordance between lithological banding, pegmatites and deformed quartz veins is less than 20°. Pegmatites begin to show pinch and swell structures in those which were intruded at low angles to bedding.

B) 500-50m above the thrust:

Sedimentary banding and early foliation become disrupted by foliation boudinage (Section MT.3.a). Pegmatites generally show near parallelism and boudinage. An angular discordance between pegmatites, early quartz veins and banding is still visible in the field. There are several phases of quartz veining, the early phases show strong internal deformation witnessed by streaks in the veins.

A) 50-0m above the thrust:

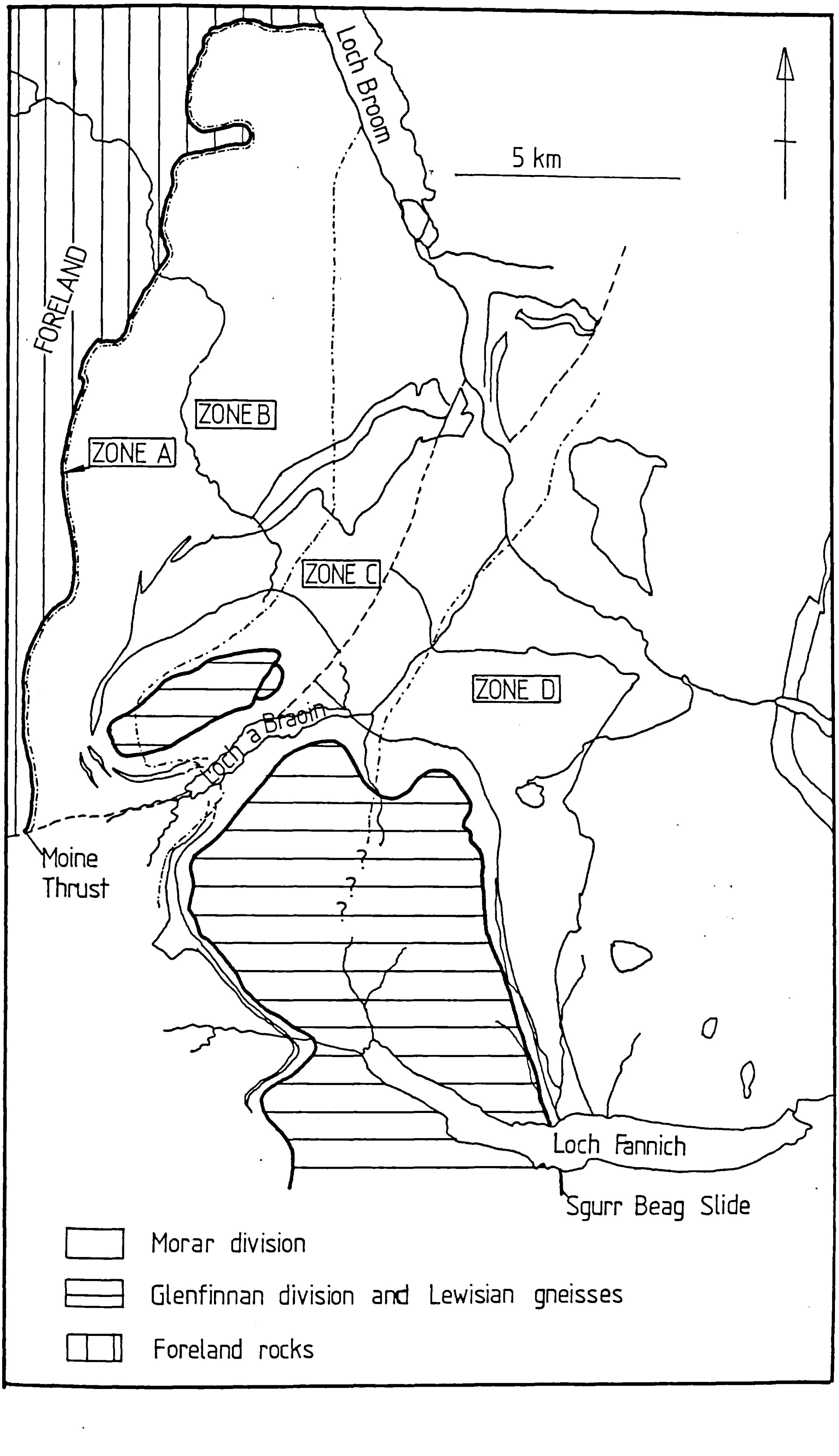
Banding is very fine, platey in places, early sedimentary banding and foliation transposed into a new mylonitic foliation. No angular discordance between pegmatites, early quartz veins and mylonitic foliation can be distinguished in the field. Both early quartz veins and pegmatites show strong internal deformation, pegmatites show strong boudinage and the boudins often become completely separated.

The salient features of both zones show striking similarities:

i) Both show wide zones of deformation (up to 1000m for the Moine Thrust, up to 750m for the Sgurr Beag Slide (Rathbone & Harris, 1979)), but only a narrow zone of intense deformation at the centre of the movement zone.

ii) Both develop strong platey foliations and extension lineations in their most intensely deformed zones. Both, in fact, develop similar quartz

FIGURE 72. Map showing zones of increasing strain west of Fannich



c-axis fabrics in their most intensely deformed zones (Sections MT.5.b, SBS.2.a).

iii) Both zones moved in approximately the same direction, towards the NW, although care must be taken not to use this fact to imply any connection between the genesis of the two.

iv) Both zones involve tightening and rotation of pre to syn-genetic folds (Section MT.2.a; Rathbone & Harris, 1979).

v) Both zones have widths of a similar order of magnitude. Despite the problems of measuring the width of the Sgurr Beag Slide zone, which may be partially obscured by later deformation. Widths of 400-800m were measured in Fannich (Section SBS.1.b) and approximately 750m at Lochailort (Rathbone & Harris, 1979).

However, the major difference between the two zones is the metamorphic environment in which they formed. The Sgurr Beag Slide is thought to have moved at peak metamorphism, under amphibolite conditions (Powell et al., 1981). The Moine Thrust initiated under greenschist facies conditions retrogressing the pre-existing fabrics (Section MT.5.a,b). All fabrics associated with the Moine Thrust in the area between Fannich and the Moine Thrust rework pre-existing Sgurr Beag Slide related fabrics. This gives rise to the characteristic foliation boudinage in psammites and shear bands in pelites.

The metamorphic grade within the mylonites indicates that they formed at higher levels in the crust than the Sgurr Beag Slide. However, the mylonite zone might be expected to widen at depth following the model of Sibson (1977) proposed for the Outer Isles Thrust. Extrapolating the Moine Thrust mylonite zone to depths equivalent to the present outcrop of the Sgurr Beag Slide, implies that the equivalent of the mylonite zone is wider than the Sgurr Beag Slide zone. This is extenuated when mylonites beneath the Moine Thrust are added to the Moine mylonites. However, the Moine Thrust zone has been complicated by later brittle thrusting. As mentioned previously, the least deformed zones of the Sgurr Beag Slide may be obscured by later deformation and thus the true original width of the Sgurr Beag Slide may be wider than the present estimates.

In conclusion, it can be stated that the two zones have a similar "order of magnitude" width and this implies that they represent similar displacements, neglecting brittle movements.

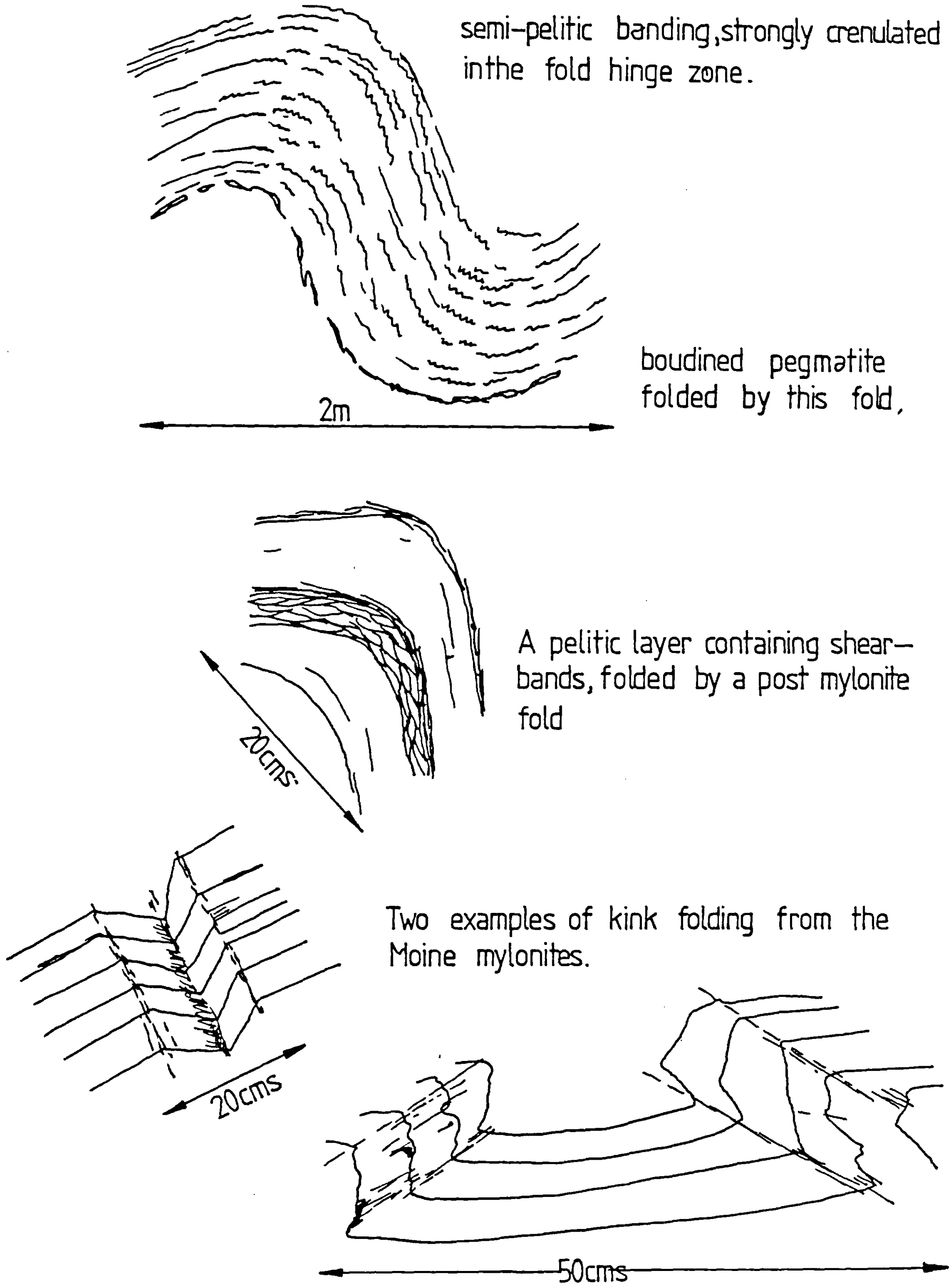
MT.3.c <u>Post-mylonite folding in the Moine Thrust zone</u>

In recent years, confusion has arisen over the relationship between late folding in the thrust zone and the mylonitization event. Early workers (Barber, 1965; Johnson, 1960; Christie, 1963) separated a distinct post-mylonitization event represented by N-S trending a symetrical folds overturned towards the west. They considered that these folds always post-dated the formation of the mylonites and designated them F3 in the thrust zone.

Bryant and Reed (1969) observed that there was a spread of 'F3' axes from their N-S trend, towards the tectonic movement direction of the mylonites. They postulated that the N-S orientated F3 folds were relatively undeformed equivalents of tight to isoclinal folds which had undergone flattening and rotation during the progressive shearing towards the WNW that accompanied mylonite formation.

McClay and Coward (1982) have suggested that the early mylonites formed at deep levels and that the sequence of folding seen (D2 and D3 of their sequence) represents stages in a sequence of simple shear deformation

FIGURE 73. Examples of post-mylonite and kink folds



during which folds nucleated at high angles to the transport direction and rotated towards parallelism during progressive shear.

During the present survey, it was possible to distinguish syn-mylonitic folds from post-mylonitic folds: the early folds have axes trending approximately N-S; these represent unsheared equivalents of the isoclinal ESE-WNW trending folds, which have not undergone extensive reorientation during the mylonitization event (Section MT.2.a). The axial planes of these folds are low angle, dipping 15^{o}-20^{o}E, even where they are found in lower strain rocks to the east. They always exhibit strong fabrics and are small, having wave-lengths of less than 1m. In contrast, a set of folds are distinguished which post-date the mylonite event. Their axial planes strike approximately N-S and dip 45^{o}E. They have a wave-length of 2-5m and are open to close in style, type 1b (Ramsay, 1967) (fig.73). Psammites contain only incipient axial planar fabrics in the fold noses and sometimes suffer brittle fracture cleavages. Pelites contain a crenulation of previous fabrics which is emphasised by pressure solution. The folds deform boudined pegmatites and late extension crenulation cleavages associated with the recrystallisation of chlorite (Section MT.5.a).

The stereographic projection of post-mylonite folding illustrates the observations made by Bryant and Reed (1969) on the spread of axes towards the tectonic movement direction of the thrust zone (fig.74). However, these later folds can be distinguished from the earlier mylonite related folding. They are consistently larger than the mylonite related folds, have higher angle axial planes and exhibit pressure solution and crenulation fabrics.

In previous discussion it has been postulated that the initiation of mylonitization would cause enormous stress relief in the overlying Moine block, allowing the area around Fannich to be regarded as a passively overriding block. However, if the mylonites stuck, stress build-up would occur in the Moine block and result in folds trending approximately at right angles to the over-thrusting direction. These folds might be expected to concentrate within weaker rocks but differ from syn-mylonite folds. It is therefore proposed that the post-mylonite folds of the thrust zone nucleated by such a mechanism.

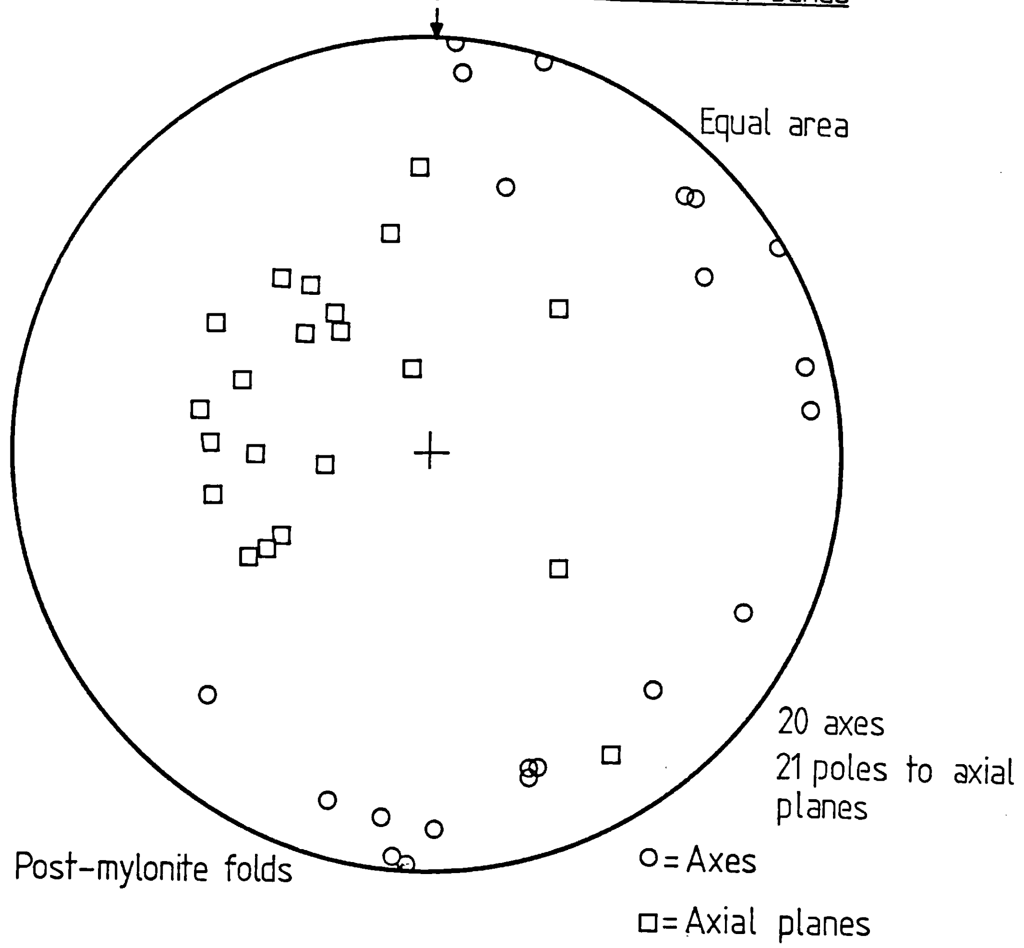
MT.3.d Kink folding and Cataclasis

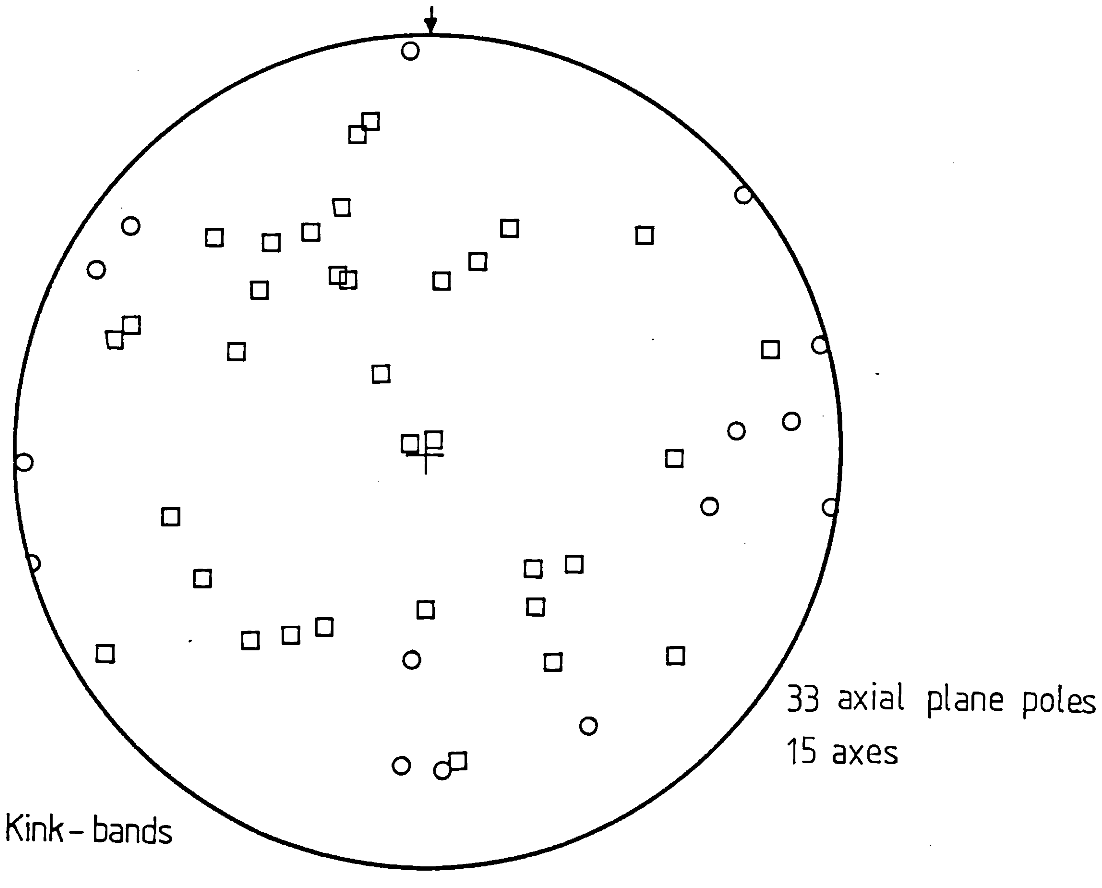
Kink folding is common throughout the mylonites. The interlimb angles range from approximately 120° to 45° and in profile they are generally less than 1m across (fig.73). Both single and conjugate types are common and all seem to be compressional. They show a near random distribution of orientations when plotted on a stereographic projection (fig.74).

Although they do occur throughout the mylonites, they tend to form concentrations within areas of about $50m^2$. It has been postulated (McClay & Coward, 1982) that kink folding may be due to differential movement between thrust sheets. The concentrations of kink folding may be areas of high differential stress. The random distribution of orientations also suggests that the kink folds were not all formed in a single stress system.

The largest concentration within the area studied is the Dundonell structure described by Peach et al. (1907) and Elliott and Johnson (1980). Elliott and Johnson interpret the structure as resulting from a lateral ramp within Cambrian pipe rock. The structure exposes folded Torridonian sandstones, Cambrian quartzites and Fucoid beds at its core. It was used by Elliott and Johnson to illustrate the westward stacking of thrusts

FIGURE 74 Stereonets showing post mylonite folds and kink-bands





towards the foreland and calculate a total movement of 35km by balanced sections.

In the present survey, the structures observed by Elliott and Johnson within the Moine rocks were broadly confirmed but the Dundonell structure can be observed in a new road section on the eastern shore of Loch Broom directly in line with the same structure on the western shore. This appears to deny the existence of a late cross-fault postulated by Peach et al. (1907) which was accepted by Elliott and Johnson (1980).

Cataclasis is spatially associated with the mylonites. It is seen generally in thin bands, near parallel to the mylonite schistosity destroying the early fabric, but also in vertical shear zones 1-5cm wide which may be due to differential movement between thrust sheets (Coward & Kim, 1982).

A larger zone of cataclasis and brittle deformation is associated with the Moine Thrust plane but this varies greatly from a very thin zone where mylonites rest directly upon Torridonian sandstone to a zone of disruption 1m wide. The most striking cataclastic bands are within psammites along the ridge between Meall an t'sithe and Meall Dubh. These bands are between 5mm and 10mm wide and are common in the psammites which were highly strained during movement on the Sgurr Beag Slide. Many of these high strain psammites lie above the mylonites and suffered only partial breakdown during the mylonitization event. Cataclastic banding is much more common here than along the shores of Loch Broom. The probable explanation of this phenomenon is that the highly strained psammites are more brittle than their lower strained equivalents in this tectonic regime, White (1973) postulated that the highly orientated nature

FIGURE 75 Metamorphic sequence in the Moine Thrust zone after Johnson (1965)

	F1	F2	post F2	F3	F4
Lewisian...	Mylonitization in "movement zones" slight recrystallization	Shearing	Growth of quartz, biotite, albite, and epidote	Shear on F3 cleavage; mainly mechanical metamorphism with local growth of quartz quartz	Local cataclasis; biotite chlorite
Torridonian...	Granulitization	Shearing	Growth of sericite, chlorite, (?biotite)	Mechanical shearing	....
Moine......	Granulitization (?mica growth)	Shearing	Growth of quartz, biotite, muscovite, albite (?garnet)	Quartz recrystalizes; locally reoriented	Local cataclasis; biotite chlorite, garnet chlorite

of the grains within a high strain zone made it become more brittle when deformed in a subsequent regime.

In conclusion, kink folding seems attributable to the stresses caused during movement of the brittle thrust over the Cambrian foreland sequence. Its formation may involve stresses produced between thrust sheets during movement. Cataclastic deformation is associated with brittle movement of the Moine Thrust but is also concentrated within rocks which were previously highly strained during movement of the Sgurr Beag Slide or during mylonitization.

MT.4 Metamorphism associated with the Moine Thrust Zone

MT.4.a Previous work

The earliest discussions of metamorphism in the Moine Thrust belt were concerned with the relationship between the thrust belt and the Moine nappe. Peach (1907), Kennedy (1955) and Bailey (1935) considered that the metamorphism seen in the thrust belt represented a low grade marginal phase of the regional metamorphism. Read's (1931,1934) view was that a "dislocation metamorphism" along the thrust zone was a localised effect breaking down fabrics associated with regional metamorphism.

Johnson (1960) argued that the thrust belt metamorphism could be separated into different phases of varying character and intensity. He related them to the structural sequence which had then been worked out (Johnson 1957,1960a) (fig.75).

During the F1 mylonitization, Johnson states that garnets and hornblende crystals were deformed but not retrogressed in Lewisian mylonites. However, he showed that plagioclase was unstable, being sericitized and sausseritized and that hornblende was reduced to rounded grains sheathed by an envelope of comminuted material. He also indicated that in some rocks, a blue/green amphibole was developed (possibly Actinolite).

Johnson split the second phase (thought by him to represent the peak of metamorphism in the thrust zone) into two parts: syn-D2 shearing and breakdown followed by post-D2 recrystallisation. He stated that the second phase of metamorphism varied over the region. In Lewisian rocks, the mylonite belts he considered to be zones of renewed shearing (initial shearing was D1) followed by periods of static growth of quartz, biotite, albite and epidote. In the Torridonian rocks of the Loch Carron area, D2 was thought to induce a schistosity defined by streaking out of quartz, feldspars and thin felts of sericite along the schistosity. In Torridonian of the Coulin area, the schistosity was described consisting of muscovite, biotite and flattened, occasionally helicitic garnets.

Johnson related D2 deformation and metamorphism in the thrust belt to the peak of Caledonian metamorphism in the Moine nappe (but see Section MT.2.a). He therefore made no distinction between Moine rocks adjacent to the thrust belt and those further east except that the degree of crystallinity increased eastwards into the Moine nappe. He interpreted garnet and mica microtextures as syn- to post-D2 of the thrust zone, although there is an indication that he was unsure of the relationships because a question mark has been placed against garnet in his correlation table (Johnson, 1960; table 1). Some of the garnet textures described can be reinterpreted in the light of the present study (Section MT.5.a).

Johnson interpreted the metamorphism associated with the asymmetrical, post-mylonite folding in terms of minor quartz recrystallization. The only strong retrogression was correlated with F4 structures when both garnet and biotite were retrogressed to chlorite.

Barber (1965) agrees with Johnson (1960) as to the structural sequence and the importance of post-D2 recrystallization but then seems to contradict the conclusions of Johnson. He states that the grade of metamorphism during recrystallization was very low and represented by the growth of quartz, chlorite and white mica.

Soper and Wilkinson (1975) describe a strong retrogression of hornblendic Lewisian rocks during mylonitization. A plagioclase-amphibole-epidote (biotite) assemblage is progressively altered to albite-epidote-chlorite in the mylonites.

They also state that almandine garnet occurs in some Moinian localities further east. Soper and Wilkinson interpreted the thrust belt metamorphism as a marginal zone of the regional Caledonian metamorphism, following the ideas of Soper and Brown (1971). They therefore interpreted the growth of almandine as syn- to post-D2 in the thrust zone and located the syn-thrusting almandine isograd only 10km east of the thrust.

MT.4.b Metamorphism associated with the thrust zone: the present survey

The metamorphism which accompanied early shearing on the Moine Thrust zone south of Ullapool was a retrogressive phase increasing in intensity westward. The metamorphic textures associated with the Sgurr Beag Slide and peak metamorphism are broken down increasingly toward the thrust (Section MT.5). Deformation is inhomogeneous but the metamorphic changes can be associated with specific stages in the known sequence of structural and microfabric events (Section MT.5).

The earliest shearing associated with the Moine Thrust zone takes place in pelite horizons, initially in the form of shear bands (Section MT.5.a). Shearing and fabric breakdown initiate in the pelites without significant deformation occurring in adjacent psammites. The intensity of deformation increases westward where psammites become significantly deformed, finally forming a quartz-feldspar mylonite within 300m of the brittle thrust.

Evidence for the ambient metamorphic grade during deformation will be described for pelites, psammites, calc-silicates and other, minor lithologies. The progressive change of metamorphic grade with time will be discussed.

I) Pelites

The initial deformation which can be directly associated with movement on the Moine Thrust zone, takes place within pelites west of Fannich (MT.5.a). It occurs in obliquely inclined micro-shear zones (shear bands) which deform the earlier peak metamorphic fabrics. The mineral grain-sizes are reduced as the rocks pass into the microshears which are 100μ-250μ wide. Original grainsizes in excess of 250μ diameter within the host fabric are reduced to approximately 20μ -50μ within the microshears. Grainsize reduction occurs by simple dynamic recrystallization in quartz, but a more complex system is involved in micas (Wilson & Bell, 1979). The two micas present, muscovite and biotite, react differently during the deformation. Muscovites tend to remain as large bent and kinked grains, stable throughout the deformation process. Biotites tend to recrystallize into very small (20μ) unstrained laths and also become retrogressed to chlorite. The recrystallization of biotite initially is taken to indicate that this deformation occurred above the biotite isograd. Garnets are also present within pelites from Fannich all the way to the Moine mylonites (Section MT.5.a). The garnets are associated with early Caledonian metamorphism (Section SBS.2.b). They remain relatively undeformed during initial shear band deformation but are never seen overgrowing the shear band structures or being recrystallized during grainsize reduction. Many garnets show angular 'broken' outlines and where zoned, the zones are often incomplete (fig.76) and cut by the angular edges. Many are rimmed by opaques, a feature not found further east (fig.76).

These features are held to indicate retrogression in this area. It is therefore concluded that the initial shear band formation took place below almandine grade but above biotite grade.

In more highly sheared pelites, the grainsize has been severely reduced. Muscovites tend to remain as large strained relic grains up to 1mm long. Quartz and feldspars have been reduced to near mylonitic grainsizes of 20μ-50μ. Biotite occurs rarely as large relics associated with muscovite or small grains distributed throughout the quartzofeldspathic groundmass (fig.79). Nearly all of the small, totally recrystallized grains at this intermediate stage are biotite although a few small grains and occasionally larger grains are partially altered to chlorite. Some of the small mica laths which crystallized as undeformed grains early in the shearing process become deformed during later shearing. Garnets are generally rolled (i.e. their internal fabrics bear no relation to the external rock fabrics) and occasionally exhibit strain shadows (fig.83). They generally have rounded or broken shapes and are rimmed by opaques.

The main changes from initial deformation to the more sheared pelites are the general reduction of grainsize, the deformation of larger, early muscovite books and increased retrogression of garnets. The ambient metamorphic grade is similar to that found during the earliest shearing but all the processes initiated during that early phase have progressed further.

Within the mylonites a range of mineralogical associations are found in pelites. Most highly sheared pelites contain biotite and chlorite, some contain only biotite and rare ones only chlorite.

Garnets are found up to 1mm in diameter, always rimmed by opaques and partially or even totally replaced by chlorite. They have also been observed replaced by zoisite + calcite + opaques.

Relics of the large muscovites remain as highly sheared mats (fig.81). Biotites are almost always reduced to mylonitic grainsizes 20μ-100μ and are sometimes completely altered to chlorite although this occurs only in specific areas (see below). Quartz and feldspar are reduced to mylonitic grainsizes although larger relict feldspars do remain.

Two possible explanations of the presence of chlorite seem possible in the mylonites:

a) Mylonitization occurred entirely at temperatures above the biotite isograd and was subsequently partially retrogressed to chlorite.

b) Mylonitization initiated at temperatures above the biotite isograd but during deformation passed into chlorite grade.

The second hypothesis is favoured because the occurrence of chlorite, in rocks containing both biotite and chlorite, is often related to late shear bands (fig.77) (Section MT.5.a). The alteration to chlorite in these cases is associated with a particular, late stage event in the deformation process.

Late cataclastic and "cold working" of the mylonites is associated with the formation of chlorite and epidote.

II) Psammites

Psammites appear more competent than pelites during mylonite formation. This is well illustrated by the displacements of post-D2 pegmatites across pelites and their low deformation state within adjacent psammites (fig.50). It is therefore concluded that psammites remain relatively undeformed during the time when pelites were undergoing initial shearing and shear band formation.

In psammites which have not undergone any mylonite-related deformation, the peak metamorphic mineralogy remains: quartz, feldspar (plagioclase and K-feldspar) muscovite, biotite, garnet plus trace sphene, zoisite and epidote. In psammites which exhibit a significant amount of strain microtextures such as patchy or banded extinction and sub-grain formation in quartz and strained extinction in micas and feldspars, biotite is stable, indicating that the initial deformation took place above the biotite isograd.

However, with the initiation of high strains, in the margins of psammitic 'foliation boudinage' (Section MT.3.a) a mylonitic fabric is initiated and chlorite replaces biotite in the fabric. As in the pelites, the replacement is thought to be syn-tectonic because it is spatially related to the formation of mylonitic bands (Section MT.5.b). As strain increases the grainsizes of the psammites are broken down progressively. Plagioclase in the psammites close to the Sgurr Beag Slide exolves microcline during mylonitization. The psammites become quartz/feldspar/sericite or chlorite mylonites close to the thrust. In the mylonites, quartz has a strong c-axis fabric (fig.87) and chlorite has replaced the majority of the biotite. Relics of biotite remain but are generally partially altered to chlorite. It is therefore concluded that initial deformation in the psammites occurred above the biotite isograd but that the majority of movement on the mylonites occurred at chlorite grade.

In conclusion, it appears that diachroneity can be seen in the retrogression of pelites and psammites mirroring a diachroneity seen in the deformation of the two rock types. Pelites show the greater part of their shearing above the biotite isograd and psammites, in the form of the mylonites, underwent the greater part of their high strain below the biotite isograd. This situation is explained if the pelites were deformed earlier than psammites during a process of progressively decreasing metamorphic grade.

III) Calc-silicates

Calc-silicate rocks of the Fannich area contain two distinct phases of metamorphism: an early peak metamorphic event and a later retrogressive event. The general field relationships, mineralogy and early peak metamorphic textures of the calc-silicates have already been described in Section SBS.2.b.

The description of later stage features in the calc-silicates has been placed within the Moine Thrust chapter because they clearly post-date peak metamorphic textures. However, it is difficult to correlate the late features in calc-silicates with movement on the Moine Thrust zone, they simply post-date all deformation associated with the Sgurr Beag Slide.

The most notable post-peak metamorphic texture is the growth of large porphyroblastic zoisites. These are easily seen in the field and reach up to 5cm in length (fig.42). They grow with random orientations and cross-cut the S0/S2 planar fabrics which are often strong in the calc-silicates (Section SBS.2.b). Their distribution is limited to the calc-silicates although they are seen to grow across the boundaries between

(NH184807)

calc-silicates and pelites. In a single exposure the large zoisites were seen within a calc-silicate folded during F3. The crystals did not show any alignment within the F3 axial plane or any signs of deformation which may have been attributed to folding. Although the evidence is from only one exposure, it is tentatively suggested that the zoisites post-date this F3 fold.

In thin section, the zoisites have been seen enclosing quartz, biotite, garnet (fig.42) and small equidimensional grains of clinozoisite (which may represent altered feldspars), but have not been seen enclosing fresh feldspars. In some cases they show good crystal shapes with relatively few inclusions (fig.42) but there is a complete spectrum ranging to crystals which are strongly poikiloblastic (fig.42).

Large porphyroblasts of hornblende occur in association with the large zoisite laths in two of the calc-silicates analysed. They are intergrown with the zoisite crystals. The amphiboles include quartz, feldspar and garnet. It is concluded that these are also post-peak metamorphic.

The presence of clinozoisite in calc-silicates has been interpreted in other areas as indicating retrogression (Charnley, 1976; Powell et al., 1981) and it is significant in this regard that clinozoisite, whilst uncommon in the Fannich area, becomes ubiquitous towards the west, that is, closer to the Moine Thrust (figs.40,38). However, the presence of clinozoisite included within the porphyroblastic zoisite crystals remains unexplained.

Hornblende, biotite and garnet crystals within calc-silicates undergo retrogression to chlorite. The retrogression is more pervasive within the calc-silicates than in adjacent pelites (figs.39,42). However, the retrogression to chlorite is unlikely to be coeval with the late stage retrogressive event (which formed the large late zoisites in calc-silicates) since it involves breakdown of hornblende to chlorite.

The only analogue with the zoisite event, within pelites, is the early breakdown during the mylonite event which occurs at biotite grade (Section MT.4.b.i). The presence of hornblende porphyroblasts within the calc-silicates therefore presents a problem. However, the calc-silicates containing late hornblendes have not been analysed so their bulk chemistry may hold the clue.

All the K-Ar and Rb-Sr isotope mineral systems are thought to have become closed at around the same time (during early mylonitization, Section GC.3.a). It is therefore tentatively suggested that the evidence, though circumstantial, indicates a late stage retrogressive metamorphic event of greatest intensity west of Fannich.

MT.5 Microfabrics associated with the Moine Thrust zone

The breakdown of peak metamorphic fabrics towards the thrust zone is a very important part of the evidence in understanding the relationship between the Sgurr Beag Slide and Moine Thrust zones. The present survey has produced new evidence for the breakdown of peak metamorphic fabrics and associated metamorphism leading to the Moine mylonites at the thrust zone.

The new interpretation arises from observation and interpretation of fabrics seen in the rocks adjacent to the mylonites and uses the work on deformation mechanisms which has recently seen several major breakthroughs. The interpretations used are based especially upon the work of White (1979), Etheridge and Wilkie (1979), Wilson and Bell (1979 and Platt and Visers (1980).

FIGURE 76. Broken garnets within pelites close to the Moine Thrust. Note that the inclusions within the garnets are the same size as those in figure 29 the grainsize of the groundmass has been reduced in this section.

1mm

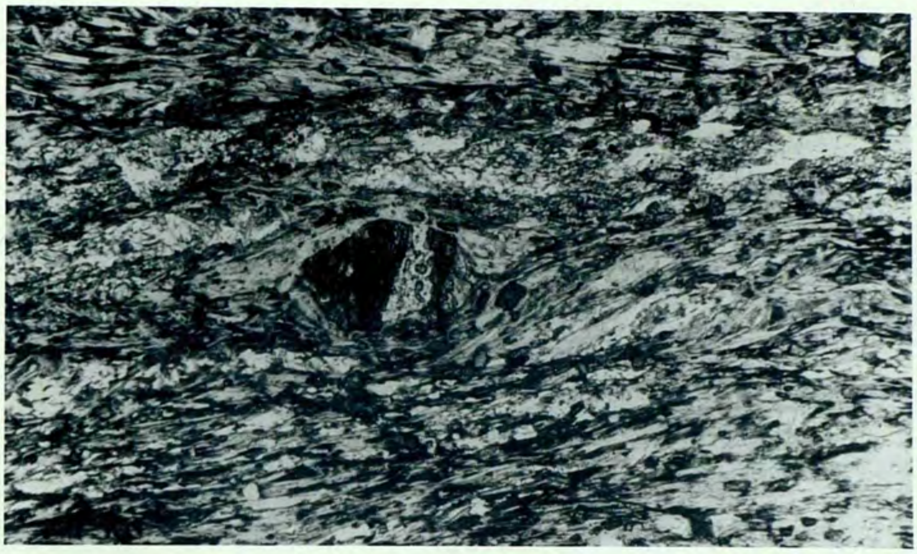


FIGURE 77. Chlorite only is recrytallising within the late shear, biotite remains in the rest of the rock

1mm

See figures ~~76~~, ~~77~~ and ~~78~~ 78, 80, 81 Shear band angles from pelites west of Fannich

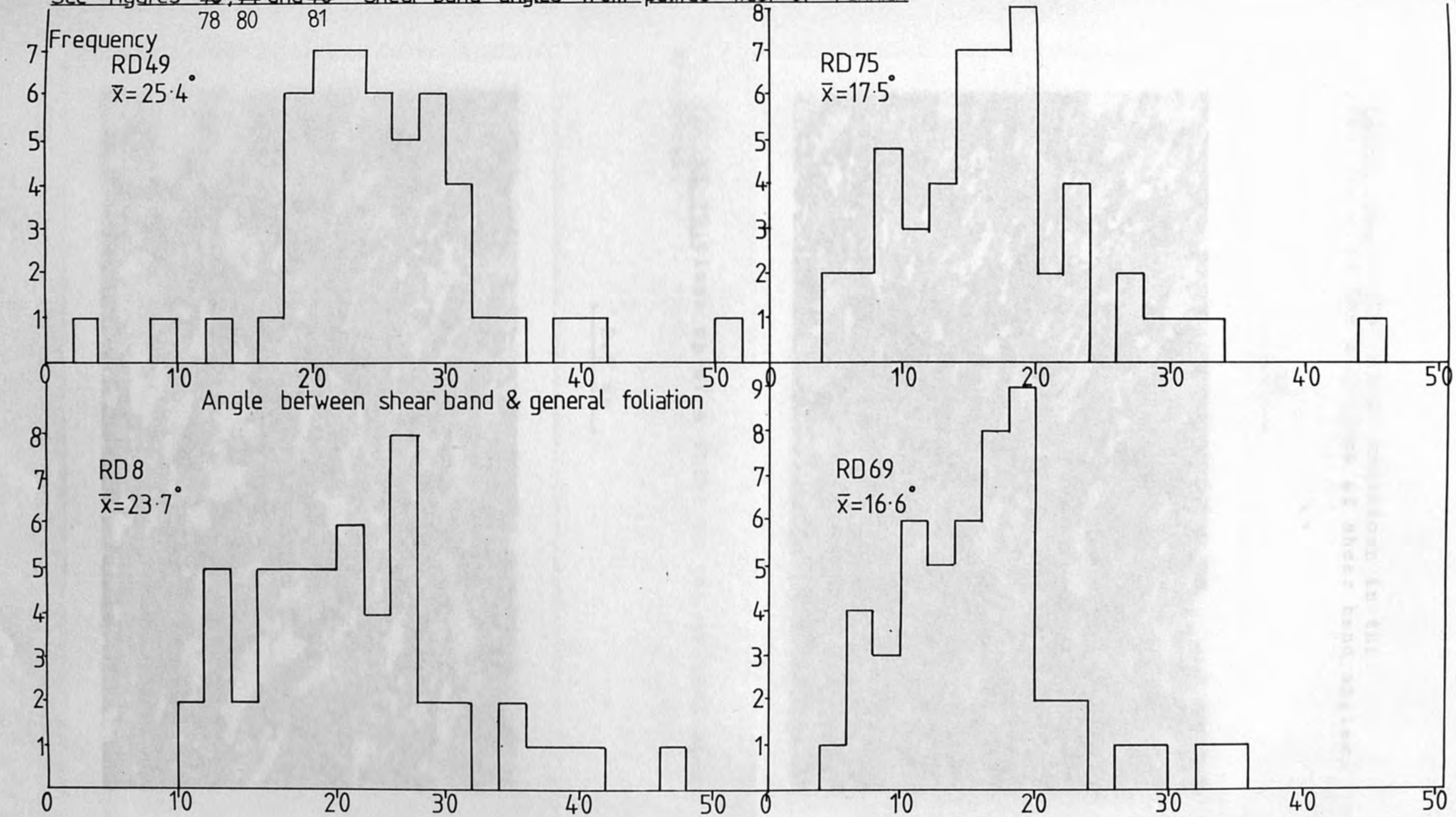


FIGURE 78. Initial stage breakdown in the pelites. See also the histogram of shear band angles.

1 mm

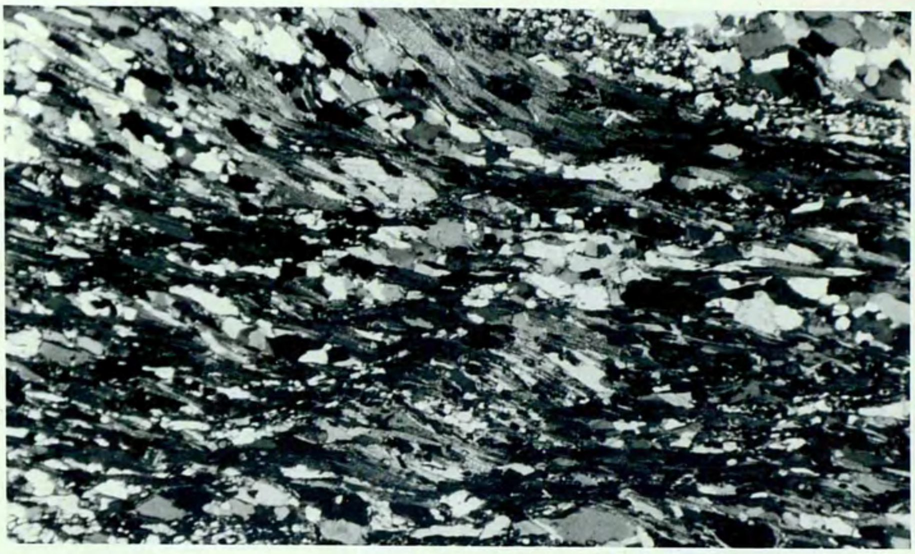


FIGURE 79. Close up of a shear in the initial stage breakdown.

0.5 mm

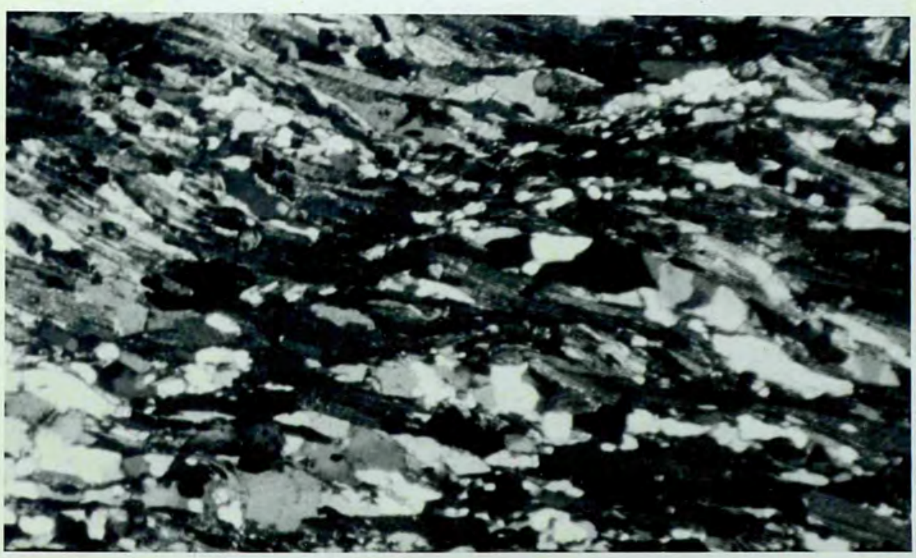


FIGURE 80. Intermediate stage breakdown in pelites. See also the histogram of shear band angles.

1mm

FIGURE 81. Final stage breakdown in pelites. See also the histogram of shear band angles.

1mm

FIGURE 82. A low angle reverse kink in final stage pelites.

1 mm

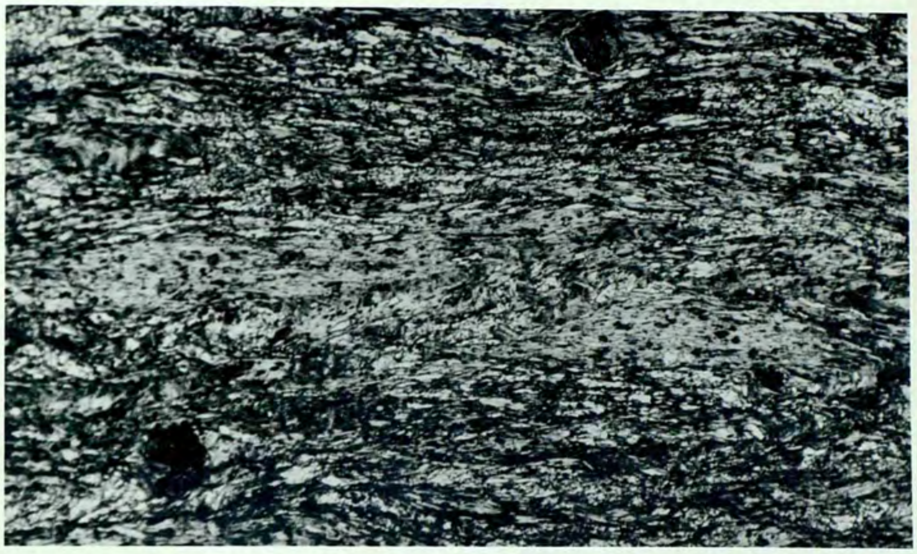
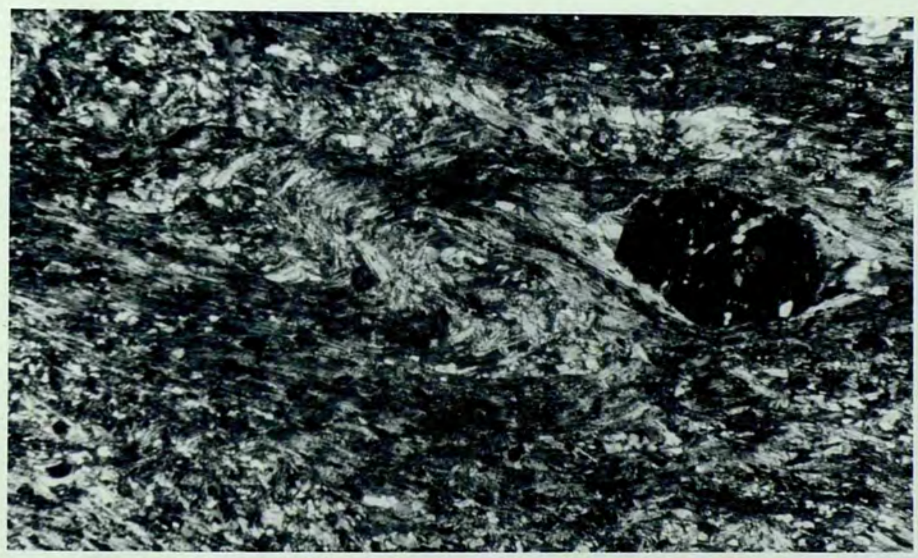


FIGURE 83. A garnet in final stage pelites. Note the pressure shadow and kink zone nucleating on the garnet.

1mm



MT.5.a Microfabrics associated with the mylonites - pelites

This section is concerned with the breakdown of fabrics established in pelites during peak Caledonian metamorphism. The coarse pre-existing fabrics in pelites suffer inhomogeneous deformation and grainsize reduction associated with mylonitization in the Moine Thrust zone.

It has already been shown that significant deformation occurs in pelites retrogressing them to biotite grade prior to the occurrence of significant deformation in psammites (Sections MT.2.a, MT.2.d). The deformation and grainsize reduction occur in two ways:

a) Homogeneous deformation with grainsize reduction

b) Inhomogeneous deformation through the formation of localised zones of high shear strain, oblique to the pre-existing fabric (shear bands). These represent an important process in accommodating large strain deformations at relatively low temperatures (Berthe et al., 1979; Watts & Williams, 1979; Platt & Visers, 1980; Vauchez, 1980; White et al. 1980; Gapias & White, 1982). The mechanism involved in their formation has been discussed in Section MT.2.e.

In order to illustrate the fabrics associated with breakdown, the process has been divided into three stages: Initial, Intermediate and Final. It is understood that the divisions are not real, the three divisions are part of a single process and serve only to illustrate gradual changes.

I) Initial stage (fig.78)

The early fabric of the pelites west of Fannich is strongly planar, caused by the alignment of micas during the Sgurr Beag Slide deformation. Initial deformation of the structures during mylonitization takes place by homogeneous and inhomogeneous mechanisms. Homogeneous deformation is a diffuse, widespread effect. The evidence for its existence comes from strained quartz and micas within pelites which exhibit no grainsize reduction zones or shear bands. It is therefore evident that some of the mylonite related deformation pre-dates the shear bands.

Inhomogeneous deformation takes the form of shear bands (Platt & Visers, 1980). In the early stages of deformation the shear bands within pelites west of Fannich resemble closely those described by Platt and Visers (1980). The mean angle between shear bands and an enveloping surface representing the attitude of the early foliation (fig.68) is 25.4° (fig.78). This compares with an angle of 29.4° quoted by Platt and Visers (1980) for shear bands in a mylonite of the Betic movement zone in southern Spain. The angles range from less than 10° to 50°.

The main characteristics of this fabric are that the early foliation is affected by open asymmetrical microfolds with a wavelength between 1mm and 2cm. One limb is thinner and has undergone shearing; this limb invariably dips towards the west (fig.62). Shearing and grainsize reduction in the shear bands are restricted to a small length of shear band, they tend to curve towards parallelism with the early foliation at their terminations and the zones are typically only as long as the individual microfolds. An anastomosing network is formed but the individual microfolds do not have great lateral extent (fig.78). When viewed on surfaces parallel to the early foliation they show lensoid undulations of the surface, leading early workers to name such structures "Fish scale", "Button schist" and "Oyster shell" textures.

Muscovite and biotite react quite differently during the grainsize breakdown (Wilson & Bell, 1979). Two size fractions are present with differing characteristics:

Type I varies from grains over 4mm long to grains 0.5mm long which show varying amounts of deformation and kinking.

Type II are undeformed and much finer grained (150μ to less than 10μ long).

In areas relatively unaffected by the shear band deformation, micas show type I structures with grains up to 4mm long. They exhibit only a small degree of strained extinction. As the shear bands are approached to within 1mm, both muscovite and biotite show strained extinction, kinks and microfractures. Muscovite tends to retain its large grainsize close to the shear and is sometimes terminated at the boundary of the shear zone by a microfault. Close to the shear, large muscovite grains become segmented and show low angle misalignments along boundaries parallel to 001 (fig.79). A small amount of the muscovite adjacent to shears may form type II grains. Within the shear zones, muscovite occurs as small type I grains up to 100μ long and zones of predominantly muscovite can sometimes be traced across the shears. It also occurs as type II grains within shear bands, these are typically larger than type II biotite grains.

As biotites approach the shears, they begin to show low angle misalignments similar to those seen in muscovites, but they seem to be associated with a reduction of grainsize from 1-2mm long to 250μ or less in length. Also associated with the breakdown into the shears is the growth of small type II biotite grains in areas where the type I biotites are extensively deformed (fig.79). These are generally less than 5μ long and show random orientations, growing across deformed type I biotite grains. Within the shear bands, type II biotites are common, type I biotite grains are very rare. The type II biotites within shears show a random orientation distribution. Layers of predominantly early biotites outside the shears cannot be matched up with a trail of biotites across the shear bands. However, it would be difficult to explain the even distribution and random distribution of orientations of type II biotites within the shear bands using a solid state, host-controlled replacement of grains by a recovery recrystallization mechanism. It is thought that the most likely mechanism for this phenomenon is that proposed by Wilson and Bell (op. cit.). They proposed that the type II grains were nucleated from a metamorphic fluid during deformation. In areas containing muscovite and biotite, type II muscovite is rare indicating that dissolution and recrystallization from metamorphic fluids during deformation was a significantly less important mechanism for muscovite deformation.

Quartz and feldspar grains are 250-500μ in diameter away from the shear bands. Feldspars show relatively little deformation apart from slightly strained extinction. Quartz grains exhibit strained, patchy extinction and sub-grain formation. The quartz grainsize is reduced significantly even within the areas not affected by shear bands. As the shear bands are approached, the grainsize of quartz is quickly reduced to the range 20μ-50μ, similar to that of the quartz/feldspar mylonites above the Moine Thrust. The explanation for the small grainsize within the shears is that small type II muscovites and biotites 'pin' the boundaries of quartz grains (White, 1979) preventing extensive recovery recrystallization. This is confirmed by the presence of larger quartz grains in areas within the shears which are devoid of micas. Feldspars are also reduced in size within the shears but remain as large, strained grains 100μ-250μ in diameter (fig.78).

Zoisites and garnet form hard, resistant grains during the initial deformation. Micas are deformed around them. Garnets show irregular broken outlines and internal zoning within the garnets is often terminated at the broken edges.

II) Intermediate stage (fig.80)

The intermediate stage illustrates partial breakdown of all the early fabrics. The distribution of shear band angles has a mean value of 23.7° in the measured sample (fig.80). The frequency plot shows a skew towards the lower angles. This is characteristic of all intermediate and final stage pelites which were measured. It reflects higher shear strain and hence the rotation of early shear bands towards the XY plane of the strain ellipsoid (Section MT.2.e). This type of distribution has not been previously reported in the literature, the only distribution reported was that of Platt and Visers (1980) who illustrated an even distribution about a mean of 29.4°. The shear bands of intermediate stage form a significantly lower angle with the early foliation than those illustrated by Platt and Visers (op.cit.) and much lower angle than the mean of 35° described by Gapias and White (1982) from a quartzite in the Hercynian belt of Brittany.

As the mean angular difference between the shear bands and early foliation decreases, deformation of the minerals increases. Micas show considerably more deformation features and wide bands of fine-grained mylonitic textures are found. They consist of fine type II biotites, quartz and feldspars, with grainsizes of 10μ-50μ. The fine biotite laths have a random orientation but studies of quartz fabrics within shear bands by Gapias and White (1982) using X-ray texture goniometry indicate a strong preferred c-axis orientation. This indicates that shear band formation caused dynamic recrystallization type processes in quartz. However, the mechanism to explain the present random distribution of biotite laths must be more complicated, as mentioned above.

Large type I biotite laths occur in association with feldspar and muscovite grains which tend to retain sizes of up to 500μ. Muscovite grains remain predominantly type I although some type II are observed at this stage in the fine grained bands. Type I muscovites show much greater deformation. Most muscovites show a segmented appearance; one original grain has broken down into several slightly misaligned grains. Other features present in muscovites are kink bands, in which small type II muscovites can occasionally be seen, extensional shears or microfaults and simple pull-apart structures. All type I muscovites are deformed at this stage.

The breakdown of quartz and feldspar proceeds inhomogeneously. They are reduced to mylonitic dimensions in the shear bands but retain grain-sizes up to 500μ in the intervening microlithons.

Garnets tend to remain within the microlithons. Their resistant nature during the deformation may, in part, maintain the microlithons around them. Some garnets are seen broken and retrogressed (rimmed by opaques) indicating that the deformation took place under greenschist facies conditions.

III) Final stage (fig. 81)

The final stage denotes those pelites in which all the early large grainsizes associated with peak metamorphism have been broken down. The frequency plot of the angles between shear bands and an enveloping surface (the original orientation of the early foliation) shows a greater skew towards low angles than that of the intermediate stage. This distribution is considered to be unevenly weighted towards higher angles since the lowest angles become increasingly difficult to observe under the microscope. The problem stems from the difficulty in distinguishing between genuine low angle relict shear bands and low angle shears between adjacent muscovite grains as they became deformed.

Two samples were measured in order to confirm the results. The means obtained were 17.5° and 16.6°. Shear bands present in the highly sheared pelites are therefore considerably lower angle than any previously reported in the literature. This reflects the different setting of these shear bands relative to those previously reported (Platt & Visers, op.cit.) (Section MT.2.e). The process of shear band formation west of Fannich is associated with grainsize breakdown and mylonitization in the Moine Thrust zone in contrast to previously reported occurrences which post-dated the main mylonitization event.

Pelites classified as 'final stage' develop a texture not seen at earlier stages. Low angle compression crenulation bands break up the highly aligned nature of the pelites. They lie at low angles from 0-15° from the sheared fabric in the opposite attitude to the shear bands (fig.82) but possess the same shear sense as the shear bands. They are up to 100μ wide and continuous for up to 1cm. The origin of these bands is unknown and they have not been described in the literature associated with shear bands. They are never found deformed by extensional shear bands and may represent a later stage in the progressive shear process.

The state of muscovite grains in the final stage pelites is the most important change from the intermediate stage. Type I muscovites are found in lenses of very small, highly sheared grains. They range in size from 20μ to 250μ, showing much higher length/width ratios than those seen at earlier stages of deformation. The muscovites are however found concentrated in lensoid structures as in the initial and intermediate stages. These lenses represent highly sheared relict microlithons. They do not become dispersed through the pelite as minute grains in the way the biotite does. The muscovites form "mats" of sheared relict muscovites showing strong alignment of 001 planes but each 'fibre' in the 'mat' is slightly mis-aligned to all adjacent ones (fig. 82). New type I muscovite grains are occasionally found crystallizing with random orientations or nucleating at kink bands. Biotite remains as grains of type I within the muscovite mats and has similar grainsizes and textures. The majority of biotite is found as minute type II grains randomly orientated within the highly sheared matrix. Some trails of type II biotite grains do however show strong 001 parallel orientation distributions in places. These represent areas where shearing of the pelite occurred during late shear band formation.

Quartz and feldspar exhibit grainsizes less than 50μ. In places quartz grains can be seen completely enclosing fine biotite grains evidently having grown from much smaller grainsizes but in general the micas occur along grain boundaries. The grainsizes of quartz and feldspar are less than those seen in psammitic mylonites of the Moine Thrust zone. The explanation for this phenomenon is thought to be the interference of fine grained phyllosilicates with the recovery of quartz grainsizes (White, 1979; Etheridge & Wilkie, 1979).

Garnets exhibit a wide range of structures. In many pelites they remain as porphyroclasts, rolled during the deformation (i.e. their internal fabrics bear no relationship to the external fabrics seen in the mylonites, fig. 83). Most have opaque rims or are sheathed by chlorite due to retrogression during the deformation (Section MT.4.b) but many are practically unaltered apart from cracking and rounding. They are associated with asymmetrical pressure shadows (fig. 83) and form nucleii for low angle crenulations. The most striking aspect of the garnets is the close resemblance of the internal fabrics of garnets found in mylonitized and unmylonitized pelites.

The preservation of garnets during the biotite grade mylonitization perhaps provides evidence for the sudden and short-lived nature of the event.

Some final stage pelites contain high angle shear bands (fig.77). A rare set of late shear bands are found which do not show significant rotation as seen in the intermediate and final stage pelites. These late shear bands occurred at chlorite grade and are restricted to pelites which have already been deformed to the final stage during mylonitization. The late shear bands generally exhibit angles of around 30° from an enveloping surface. They are very similar to the "normal" type of shear bands described in the literature (Platt & Visers, 1980; Gapias & White, 1982) because they post-date the main shearing movement.

MT.5.b Microfabrics associated with the mylonites - psammites

As previously discussed (Section MT.4.a) early workers interpreted the reduction in grainsize towards the Moine Thrust as either a progressive breakdown of earlier fabrics (Read, 1934) or as fabrics formed at increasingly lower grade towards the thrust during a single phase of Caledonian metamorphism (Barber, 1965; Johnson, 1963; Christie, 1963; Bailey, 1955; Soper & Wilkinson, 1975).

Several advances have been made in the interpretation of deformation fabrics and textures within the last few years. The most important of these is the recognition of the syn-tectonic, recovery-recrystallization mechanism in the deformation of quartz (Etheridge & Bell, 1976; White, 1976).

The deformation fabrics of the Fannich-Ullapool area have been examined optically and the distribution of c-axes for quartz measured in representative specimens using a Universal stage. Psammites adjacent to the Sgurr Beag Slide zone exhibit strong c-axis orientations and alignment of micas.

FIGURE 84. Psammite / pelite boundary, the pelite showing a much higher deformation state. Note the garnet resisting deformation.

1 mm

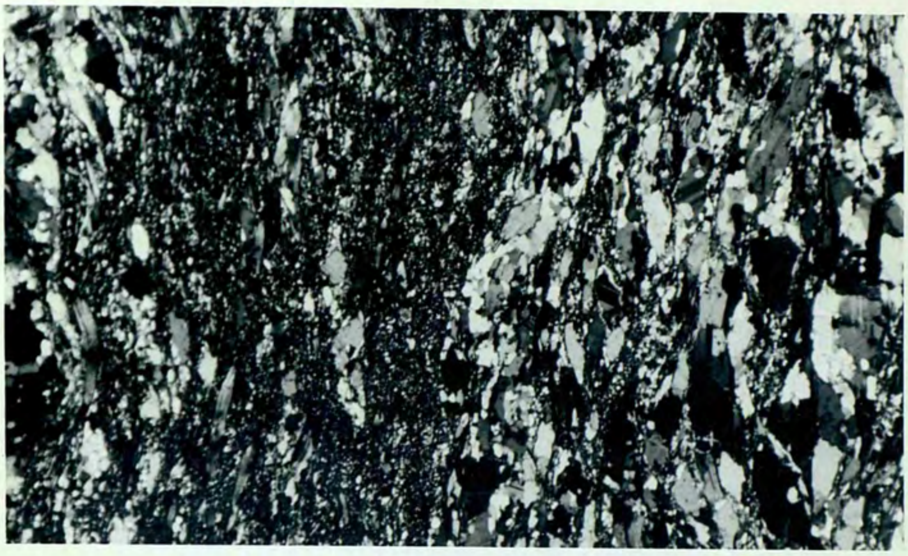
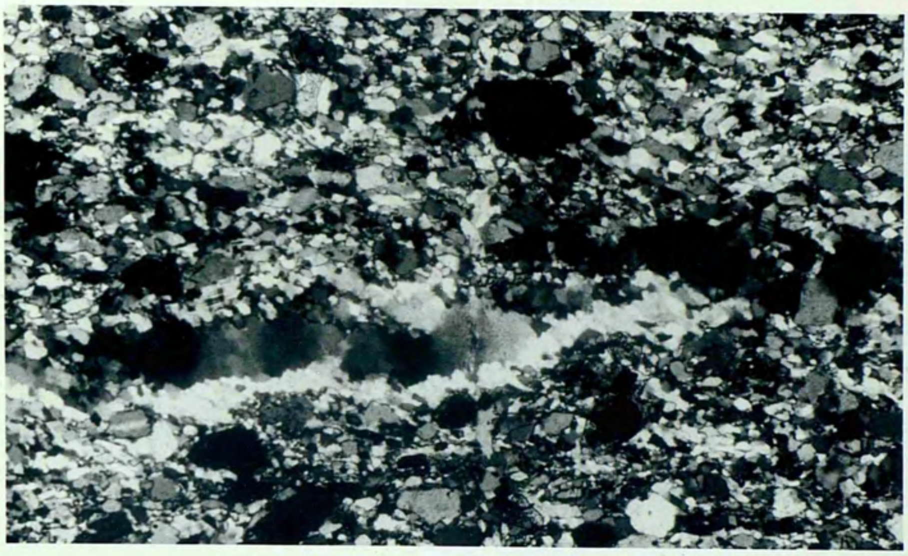


FIGURE 85. A psammite showing only minor grain size reductin due to the Moine mylonites.

1mm

FIGURE 86. A psammite showing intermediate stage breakdown. Note the quartz vein and core-mantle structure.

1 mm



Psammites away from the slide exhibit random c-axis patterns and granoblastic textures (Section SBS.3.a). There is therefore a variation in the fabrics of psammites prior to deformation in the mylonite event. Psammites adjacent to the slide and those away from the slide react in slightly different ways during the initial mylonitization; they develop 'foliation boudinage' to varying extents (Section MT.3.a). No difference is seen between the microfabrics of the mylonites developed close to the outcrop of the Sgurr Beag Slide at Loch a Broain and those developed far from the slide along the shores of Loch Broom.

Psammites are more resistant than pelitic horizons to the mylonite related deformation (fig.84) but it is difficult to define exactly where mylonitization begins in the psammites. Throughout the area west of Fannich, all psammites exhibit at least minor strain of quartz and sub-grain formation (fig.85). In these relatively undeformed psammites quartz and feldspar have grainsizes from 250-500μ and micas have grainsizes up to 1mm long, 250μ wide, although their sizes range down to less than 50μ long. Deformation textures are found only in quartz grains. Quartz exhibits banded or strained extinction, and also minor sub-grain formation; the sub-grains however make up less than 20% of total quartz grains present. The c-axes fabrics obtained from these low strain psammites are near random (fig.32).

At distances of approximately 5-6km from the Moine Thrust, psammites begin to exhibit significant mylonitization (fig.86). Micaceous minerals take on a preferred orientation and are deformed internally. It is extremely difficult to recognise this stage of deformation on an exposure scale, psammites look little deformed and have no measurable lineation in the field.

Feldspar grains show a greater tendency to retain their original grain-sizes. They exhibit patchy extinction, exsolution textures and fractures caused by internal deformation. Quartz exhibits grainsizes generally between 100μ and 250μ but many grains or sub-grains are still visibly parts of larger relict grains. Some of the originally large plates of quartz within veins show large deformed cores mantled by fine sub-grains (fig.86).

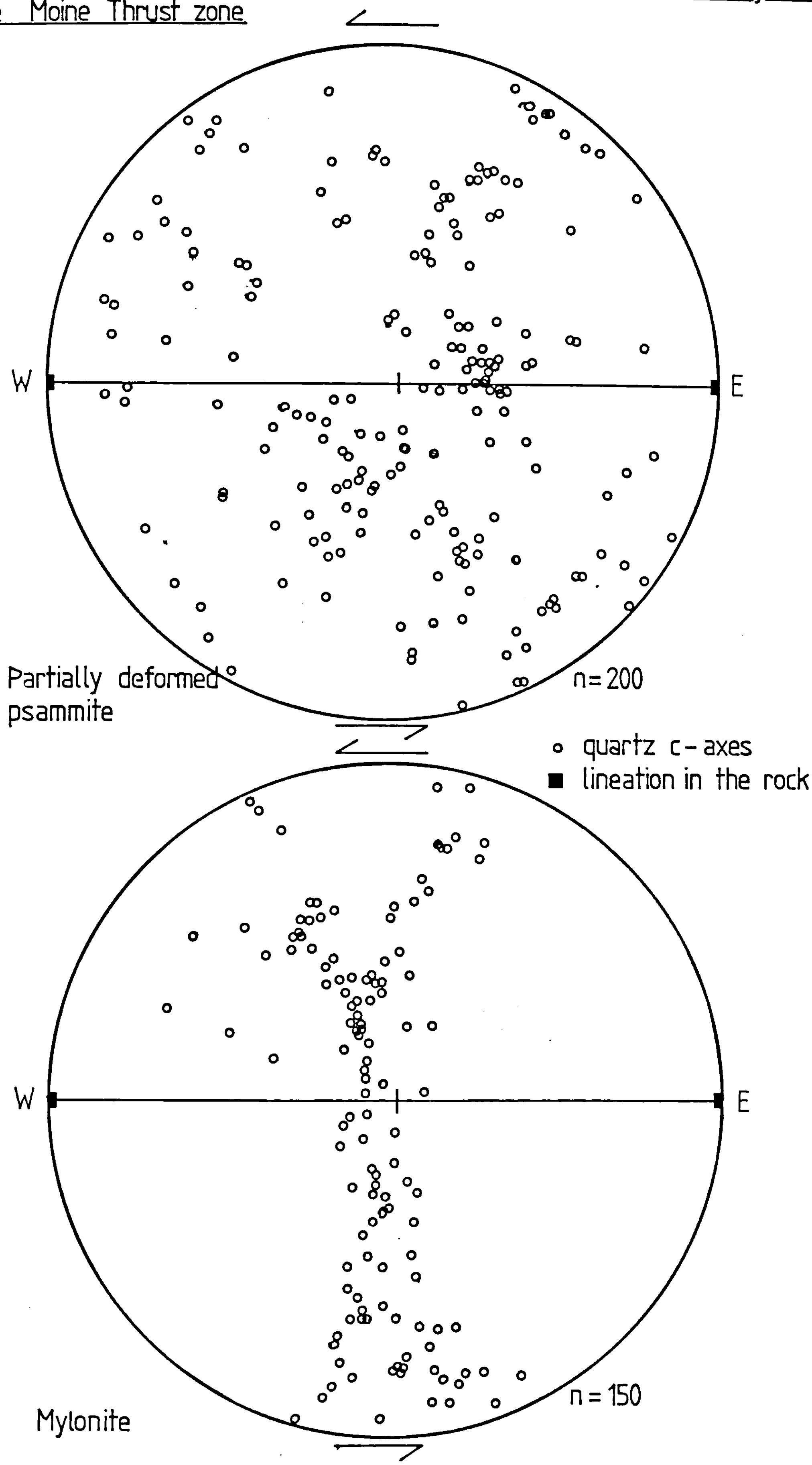
A c-axis fabric determined for a psammite at this stage of deformation showed only a weak fabric which might be interpreted as cross-girdle type (fig.87). However, only one determination was performed on a partially deformed psammite so no clear conclusions can be drawn.

In the intermediate breakdown stage, inhomogeneous deformations of the psammites occur (Section MT.3.a). This takes the form of 'foliation boudinage', especially in the highly strained psammites associated with movement on the Sgurr Beag Slide. The 'foliation boudinage' represents an inhomogeneous deformation on a mesoscopic scale, but the investigation of microfabrics has revealed that the mylonitization associated with these shears is inhomogeneous too.

Within the pods of less deformed rock which retain a Sgurr Beag Slide related lineation, the microfabrics reveal an intermediate type breakdown described above. As the shears are approached, thin mylonitic bands appear 0.5-3mm wide, spaced 5m to 1cm apart. The bands are orientated at angles of 20-30° from the pre-existing foliation in the psammites and are parallel to the main shears in the psammites.

Within the mylonitic bands, the grainsize of quartz and feldspar are reduced to less than 100μ, biotites are retrogressed to chlorite and muscovite is reduced to the 'mats' seen in final stage breakdown of pelites. Rare garnets become broken and retrogressed.

FIGURE 87 C-axis fabrics from partially deformed psammite and mylonite in the Moine Thrust zone



As the shears associated with 'foliation boudinage' are approached, the number and width of the mylonite bands increase until all the early fabric is entirely mylonitized at the centre of the shear.

A similar situation has been described in a granodiorite from a Hercynian shear zone in Brittany (Berthe et al., 1979) and during mylonitization of Lewisian rocks in Eriboll (White et al.,1982). The undeformed igneous rocks of Berthe et al. (op. cit.) became mylonitized not in a simple, homogeneous manner but by the formation and continued movement on millions of parallel shears which Berthe et al. (op. cit.) named C-bands. The rock between the shears was progressively deformed and also formed a foliation due to flattening (S-bands). The S-bands were not seen in the present case.

The final stage of breakdown is the development of a quartzofelds-pathic mylonite. This stage is generally recognised in the field by close parallelism of all early fabrics, pegmatites and quartz veins. It also shows a strong ESE stretching lineation and strongly planar banding.

Quartz grainsizes are generally 25μ to 100μ, feldspars are generally slightly larger, ranging from 25μ to 200μ. Quartz shows strong subgrain formation, subgrains being less than 50μ in diameter. Quartz also exhibits strained extinction and elongation parallel to the foliation.

The 001 planes of micaceous minerals lie closely parallel to the foliation, muscovite grains vary from 50-500μ in length and show textures similar to those seen in the final stage of pelite breakdown. Chlorite occurs in thin 'mats' less than 50μ wide and up to 2mm long. Chlorite is always associated with opaque minerals and is thought to represent the relics of biotite grains which were present in the original psammite although 'mats' may also contain sericite from the breakdown of feldspars.

The quartz grains show strong crystallographic orientations (fig.87). The c-axis patterns observed are very similar in style to those described by White et al. (1982) for Moine mylonites at Eriboll. A single girdle is orientated approximately perpendicular to the lineation (fig.87) with an asymmetry in the direction of shear. However, there seems to be no strong concentration parallel to the Y axis of the strain ellipsoid as seen in Eriboll. The reason for this phenomenon is unknown.

Since 1975 there has been considerable discussion in the literature on the development of crystallographic fabrics in quartz-rich rocks which have been plastically deformed in shear zones. It is not proposed to review all the literature here but a comprehensive review of the problems has been presented by Lister and Williams (1979). There now exists general agreement as to the significance and interpretation of these fabrics in terms of non-coaxial strain.

The orientation of the fabrics are regarded as being controlled by the kinematic framework rather than bulk strain. In other words, for a single shear zone, fabrics will have the same orientation throughout. It has been shown that the reason for the preferred orientations is the preferred alignment of the dominant crystallographic slip plane in quartz parallel to the flow direction. In quartz, the dominant slip seems to be on the crystallographic basal plane, c-axes tend to become orientated perpendicular to the slip direction. This is useful in assigning kinematic significance to the lineation in the Moine mylonites, since the lineation seen is perpendicular to the girdle of c-axis fabrics formed during the process of mylonitization. It is related to recovery-recrystallization and crystallographic reorientation as a result of the deformation.

The lineation seen in the mylonites was developed at the same time as the planar mylonitic fabric being a result of the shearing deformation. It was not superimposed, as suggested by Soper (1971) and Soper and Wilkinson (1975).

The shear sense of the mylonites is known from field observation of the displacement of the pegmatites and the sense of asymmetry seen in the c-axis fabric agrees with those of White et al. (1982) for the mylonites in Eriboll.

However, there seems to be a disagreement in the literature. Some authors interpret patterns similar to these as indicating shear in the opposite sense. It may be that both types of pattern exist, caused by slightly different mechanisms in different shear zones. In areas where no field evidence for shear sense is available, determination of the a-axes in quartz has been used to confirm shear sense.

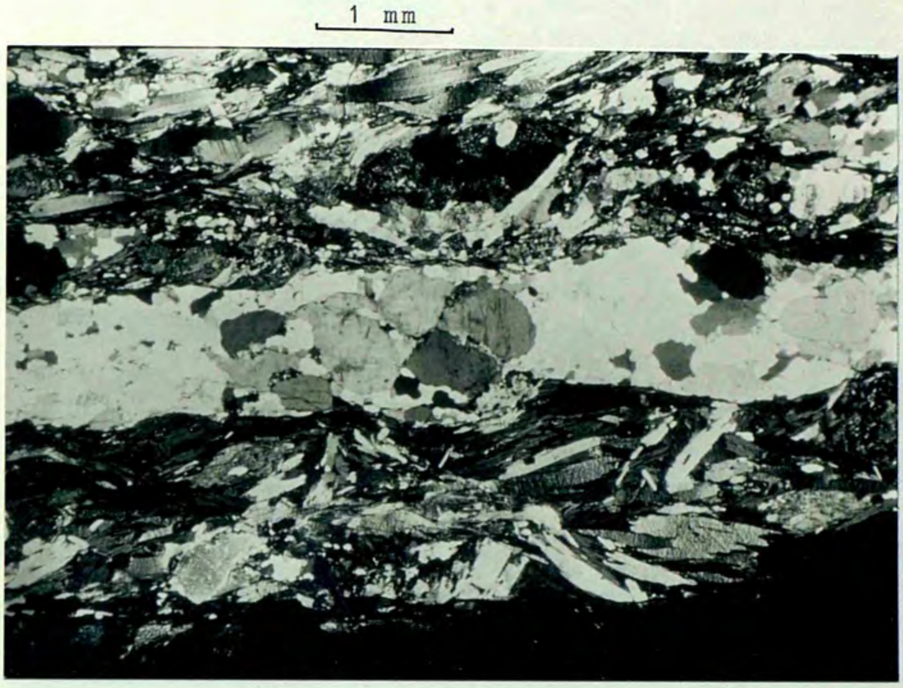
MT.5.c Microfabrics associated with the Moine Thrust Zone - post D-2 pegmatites

The field relationships of these pegmatites have already been discussed in Section MT.2.c. They provide an important indication of post-Sgurr Beag Slide strain as they were intruded and crystallized after the cessation of movement on the Sgurr Beag Slide. The pegmatites exhibit large grainsizes, feldspars up to 2cm in diameter and similar size quartz plates. Both minerals are unstrained in pegmatites taken from areas of low strain, away from the mylonites (fig.88).

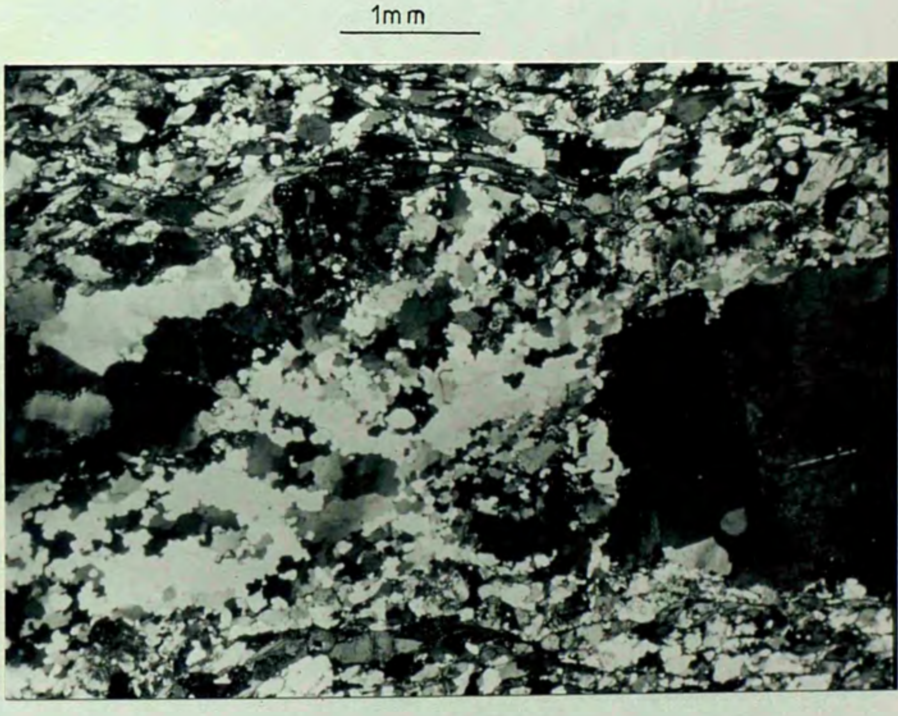
As the mylonite zone is approached the pegmatites begin to show deformation in the field (Section MT.2.c). This occurs both as passive rotation and folding. In thin section quartz takes up the majority of the strain. It shows banded and strained extinction and is deformed

FIGURE 88. Post D2-pegmatites being deformed into the Thrust zone.

a) initial stage.

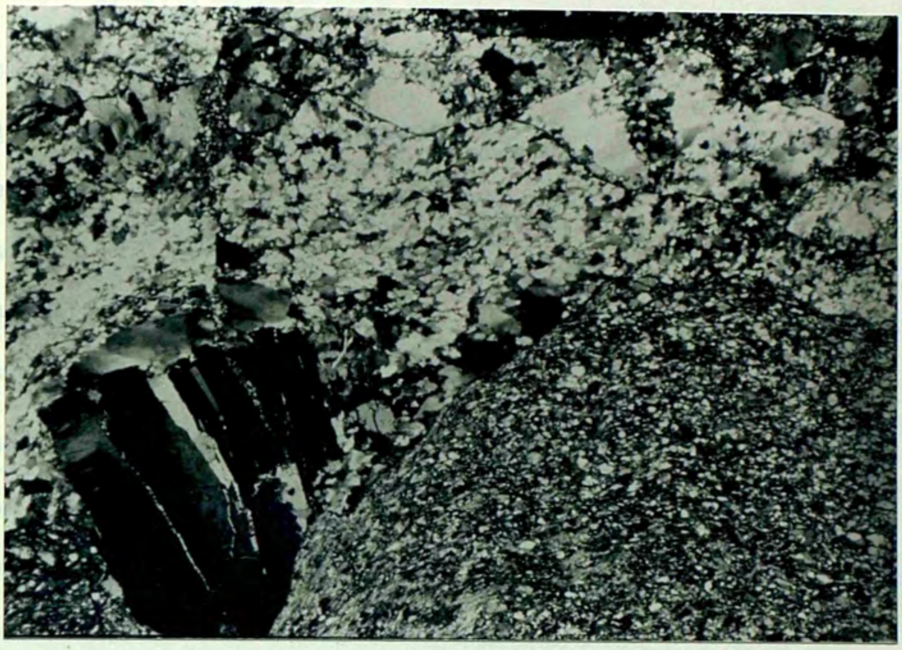


b) Intermediate stage.



c) Final stage, note the breakdown of the feldspar.

1 mm



around the feldspars. The feldspars show relatively little deformation, they become cracked and show minor recrystallization (fig.88).

Close to, and within the mylonites, the pegmatites show intense deformation. Quartz grainsizes have been reduced to mylonitic dimensions and show ribbon textures (fig.88). Feldspars are broken and recrystallized, they show fractures perpendicular to the foliation (fig.88) and are strongly recrystallized adjacent to the cracks. Feldspars also become embayed into the country rock at the edges of the pegmatites due to the reduction in vein width during deformation (fig.88). They act as resistant bodies during the deformation of pegmatites as in the pelites and psammites.

The deformation of the pegmatites and the increasing intensity of mylonitization correlates well with the mylonite formation in psammites and pelites of the area west of Fannich. The pegmatites were mylonitized during the same event which affected other lithologies.

The pegmatites do not correlate with those described within the mylonites of Eriboll by Soper (1971) since these are thought to post-date the main mylonitization event.

MT.5.d <u>D3 microfabrics of the Fannich area</u>

Deformation of the rocks in Fannich during the post-Sgurr Beag Slide event was accompanied by retrogression and occurred at biotite grade. Large grainsizes, 250μ-500μ established during syn- to post-slide metamorphism and crystallization are reduced in the zones of high post-Sgurr Beag Slide strain.

The microfabrics relating to D3 in pelites are crenulations of the earlier D2 foliation (Section SBS.3.b). They are found in the hinge zones of minor F3 folds and are widespread throughout the steeply and shallowly inclined limbs of major folds. The crenulation consists mainly of a compressional type and is accompanied by grainsize reduction of all pre-existing minerals except garnets which resist deformation and deflect the crenulations. Feldspars also tend to retain their original size, showing strained and patchy extinction. Quartz has a wide range of grainsizes from 100μ to 500μ, shows strained and banded extinction and minor sub-grain formation. Micas show strained extinction and re-crystallization at the hinges of crenulations into smaller unstrained grains 100μ-250μ long. Larger grains of both muscovite and biotite of type I (Section MT.5.a) occur as relics.

An important feature of D3 microfabrics in Fannich is the presence of an extension crenulation in the low angle limbs of microscopic folds within the overturned limb of the Fannich syncline. The shear bands are at angles around 25° from an enveloping surface and therefore represent the initial stage of the breakdown defined further west (Section MT.5.a). However, these shear bands differ in that quartz and micas have undergone substantial recrystallization within the shears after their formation (fig.89).

The shear bands are similar to those further west in that biotite has evidently undergone the same grainsize reduction and muscovite has tended to remain larger, occurring with lenses. Quartz has been broken down within the shears. However, widespread recrystallization appears to have annealed the large muscovites as they show very little strained extinction, the small type II biotites are still present but exhibit larger grainsizes than type II biotites further west. Quartz grains are found entirely enclosing the biotites, unlike examples further west where the small type II biotites tended to 'pin' the quartz grain margins.

This kind of "recrystallized shear band" is occasionally found in pelites which only underwent minor, initial stage, deformation further

FIGURE 89. Shear bands fron the Fannich area, kink folded during D3.

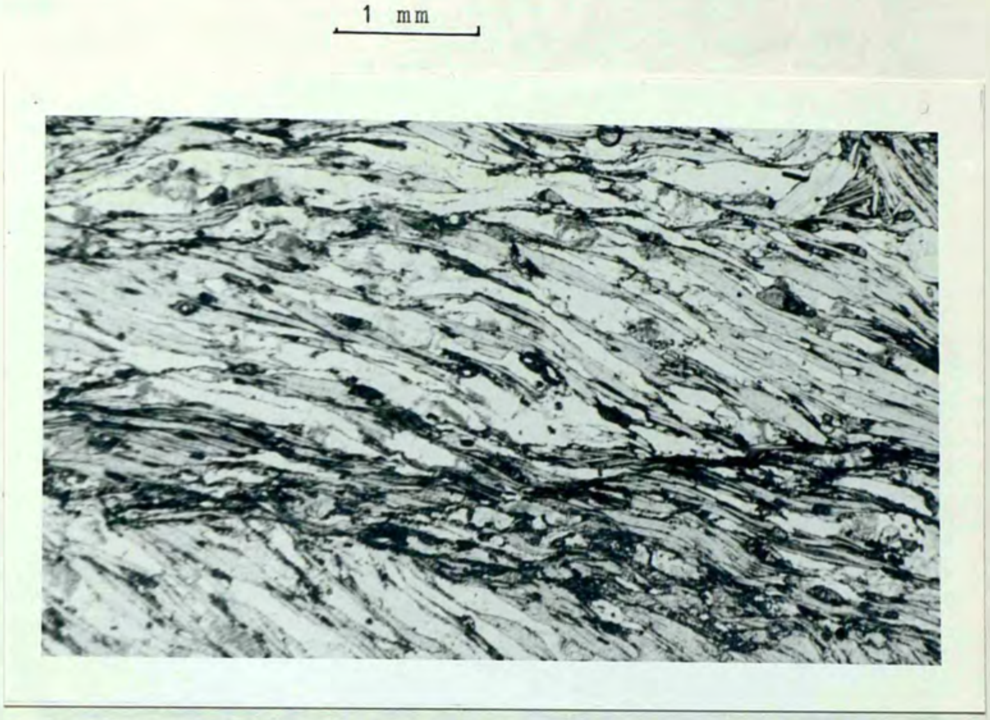


FIGURE 90. A semi-pelitic ultra-mylonite, taken from the base of the Moine mylonites.

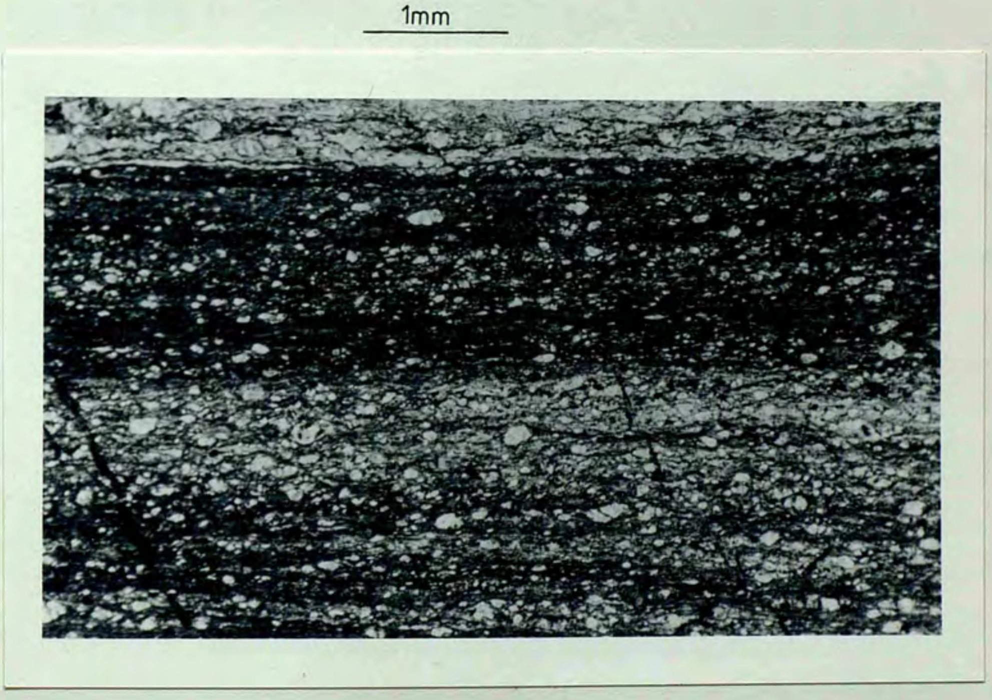
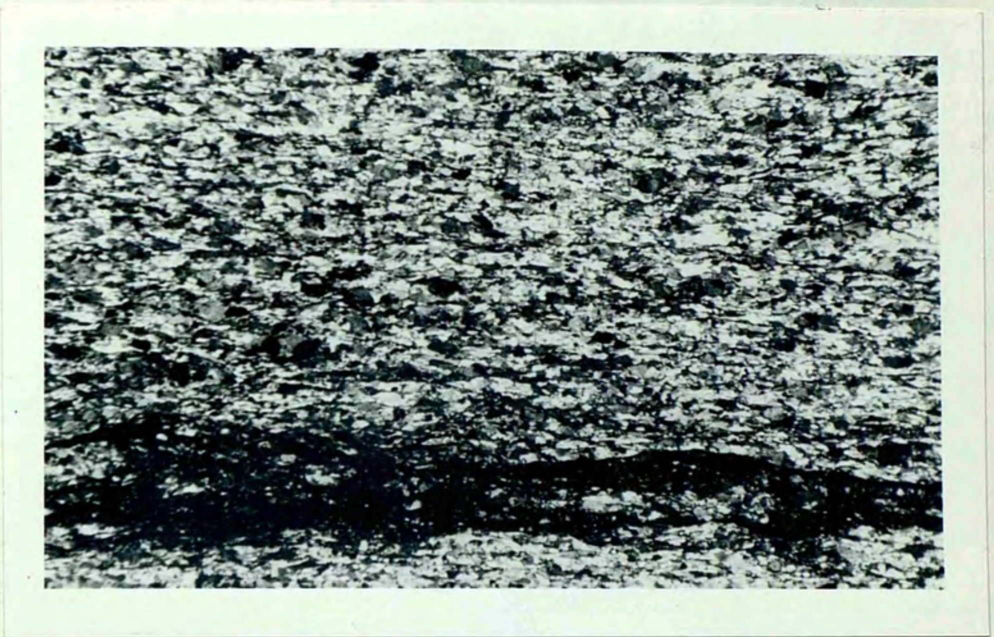


FIGURE 91. Cataclasis due to brittle deformation in the Moine thrust zone.This is significant in that the section was taken from the Meall a'Chrasgaidh psammite adjacent to the Sgurr Beag Slide on the summit of Meall Dubh.

1 mm



west. This type of shear band is thought to have initiated and ceased movement early in the D3 deformation. This is corroborated by the evidence of shear band formation in psammitic and pelitic lithologies towards the thrust; the intense deformation became localised to the thrust zone and the development of shear bands further away ceased. In this situation, the fine grainsizes within the shears and deformed muscovites in Fannich would tend to move towards more stable equilibrium states and produce granoblastic textures by recovery-recrystallization.

Microfabrics within psammites of the Fannich area consist of homogeneous but minor grainsize reduction of quartz and feldspars. Quartz exhibits banded and strained extinction and occasionally sub-grain formation; feldspars show only slightly strained extinction.

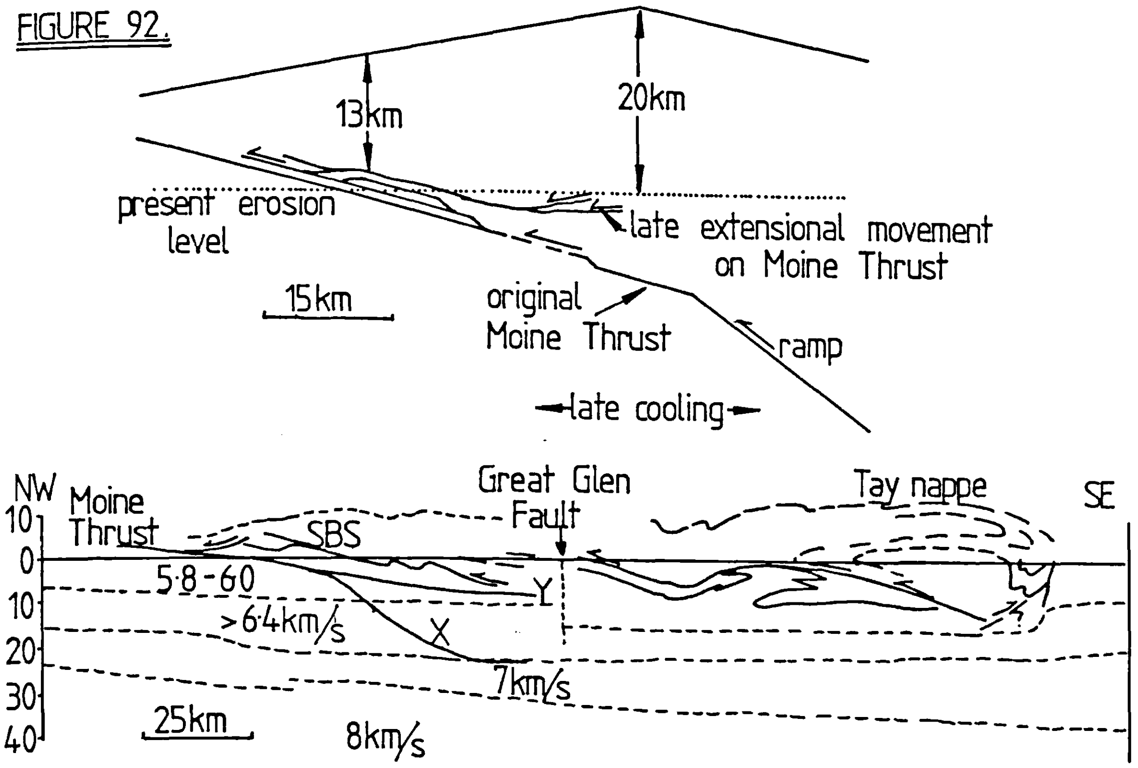
MT.5.e <u>Low temperature mylonites and Cataclasis</u>

Within 10m of the Moine Thrust some rocks exhibit much finer grainsizes than those found throughout the rest of the mylonites (fig.90). Grainsizes become too small to measure optically and the phyllosilicates form cryptocrystalline 'mats'. Only feldspars remain as relict, elongate porphyroclasts up to 250μ by 100μ and these exhibit pull-apart fractures. Originally quartz-rich and mica-rich layers can still be distinguished. These textures are thought to indicate reworking of the earlier mylonites at low temperatures (White <u>et al</u>., 1982) prior to the initiation of the brittle Moine Thrust.

Cataclastic textures are found in many of the rocks west of Fannich from the mylonites to the high strain zone of the Sgurr Beag Slide (Section MT.3.d) (fig.91). In all rocks, the cataclastic zones break down the pre-existing fabric to chlorite grade in sub-horizontal bands. The grainsize of the matrix is below that detectable optically. Angular rotated blocks are commonly found within the matrix, exhibiting a wide range of sizes and containing the earlier fabrics. Muscovite grains occasionally survive as well as single feldspars but quartz is not found outside the composite blocks. Many of the cataclastic bands are haematized.

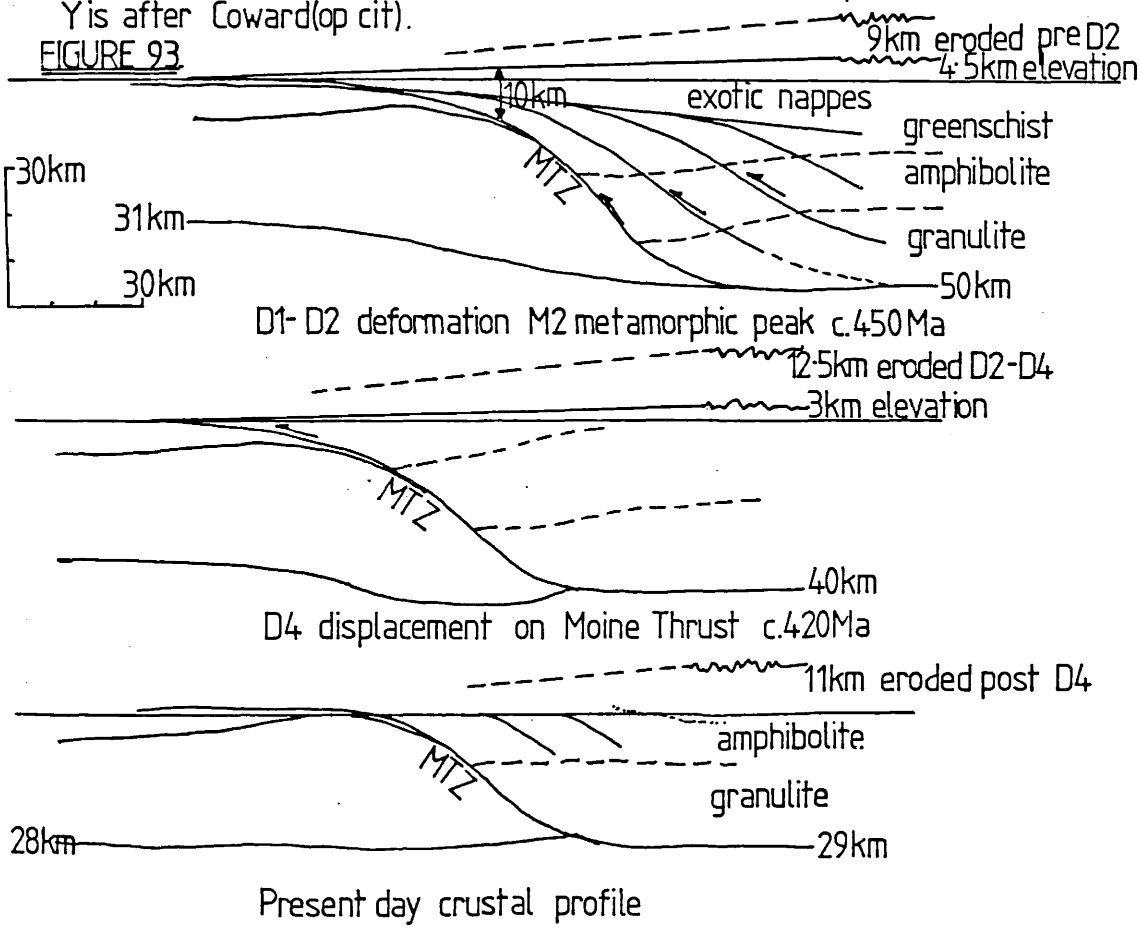
Cataclastic deformation post-dates all the mylonite and Sgurr Beag Slide related fabrics and, as previously mentioned, the intensity of cataclastic bands is spatially related to the high strain zone of the Moine Thrust.

FIGURE 92.



Diagramatic representations of the Moine Thrust after Coward(1983)
The extension of the thrust at depth: X is after Soper and Barber(1982)
Y is after Coward(op cit).

FIGURE 93



Diagramatic representations of the Moine Thrust after Soper and Barber(1982) MS = Meadie slide NS = Naver slide MTZ = Moine Thrust zone

Base of ORS

Discussion of Tectonometamorphic History

Conclusions and discussion concerning the tectonometamorphic history within the area of this study have been covered within relevant individual sections. However, the implications of the conclusions drawn in relation to the history of the N. Highland Caledonides in general and in particular, the development of the Moine Thrust zone have not yet been discussed.

Models for the Moine Thrust zone

Three models have recently been proposed for the crustal geometry of the Moine Thrust zone: The 'thin skinned' model (Coward 1980, 1983; Elliott & Johnson, 1980); the 'crustal duplex' model (Soper & Barber, 1982); and the 'thick skinned' or 'steep thrust' model (Watson & Dunning, 1979; Stewart, 1982).

The 'steep thrust' model postulates that the Moine Thrust steepens at depth, becoming nearly vertical. The evident large-scale horizontal movements on the thrust were explained by Stewart (1982) as gravity sliding off the top of a crustal block uplifted in the east. However, no seismic evidence for this model is seen in the M.O.I.S.T. seismic reflection traverse (Smythe et al., 1982) and it is difficult to envisage 60km of movement on the Moine Thrust (Coward, 1983) and more on lower thrusts, occurring by this mechanism. The model is therefore discarded.

The 'crustal duplex' model of Soper and Barber (1982) postulates that the Moine Thrust zone steepens east of its present outcrop, becoming shallow again at depth, in the lower crust, and joining a floor thrust perhaps at the Moho (fig.93). Soper and Barber (op.cit.), when originally proposing this model, postulated that the Moine Thrust zone at high levels, dipped approximately 3° east, steepening quickly to 30°

before flattening out again at depth. Coward (1983) proposed that the Moine Thrust maintained a shallow trajectory of around 3°, steepening over a ramp structure to a dip of 30° at least 25km and possibly up to 60km east of the present outcrop (fig.91). Coward further suggested that the Moine Thrust flattened out, joining a possible floor thrust at 10-15km. This represents the 'thin skinned' model for the Moine Thrust zone proposed by Coward (1980) and Elliott and Johnson (1980).

The 'crustal duplex' model, presented by Soper and Barber (1982) is difficult to reconcile with data from the present survey, in several respects:

a) They proposed NW movement along the Moine Thrust plane early in the history of the Caledonian orogeny; the present survey indicated that movement along the Moine Thrust occurred only after the cessation of movement along the Sgurr Beag Slide and that a time lapse between the development of the two structures is indicated (Sections MT.1.b, 4.b).

b) Soper and Barber proposed that the main movements upon the Moine Thrust zone occurred during peak Caledonian metamorphism; it has been shown here that all metamorphic fabrics associated with the Moine Thrust zone post-date the Sgurr Beag Slide fabrics and formed at lower grade (Section MT.4).

The presence of the "exotic nappe" postulated by Soper and Barber in order to explain amphibolite facies rocks adjacent to the Moine Thrust zone is removed, since the evidence from the Fannich-Ullapool area indicates that prior to Moine Thrust movement, the rocks adjacent to the thrust had cooled to greenschist facies conditions (Section MT.5.a).

c) If the low grade, late movement model of the Moine Thrust presented in this study is accepted, together with the 60km of movement proposed by Coward (1983), the Soper and Barber model predicts that Granulite facies rocks would have been brought to the surface at the present level of erosion. The rocks adjacent to the thrust zone are of amphibolite or greenschist grade.

The model proposed for the Moine Thrust zone by Coward (1983) fits more closely with data from the Fannich-Ullapool area. Coward postulated that the thrust levelled out at 10-15km since rocks adjacent to the thrust were retrogressed to greenschist facies. However, he also suggested (like Soper & Barber) that the internal zone of the Moines was at amphibolite grade during early thrusting. Thus he assumed that formation and movement on the Sgurr Beag Slide and Moine Thrust zones were essentially coeval with peak metamorphism. However, as previously argued (Sections MT.1.b, 2.a, 2.b, 4.b, 5.a, 5.b) the Moine Thrust post-dates peak metamorphism and movement associated with the Sgurr Beag Slide.

Given this relationship, Coward's (op.cit.) proposed model involving collapse of a thickened crust ahead of the ramp structure in the Moine Thrust zone (fig.92), to provide late extensional movement upon the brittle Moine Thrust is untenable. The increased crustal thickness relating to peak metamorphism had been reduced before movement on the thrust zone. A second mechanism, suggested by Coward (op.cit.) involves the intrusion of the large mass of Caledonian granites east of the thrust zone, causing the bulge and thus gravity sliding; this seems a more acceptable mechanism for the late movement on the Moine Thrust.

Correlation of the structural histories of the Ullapool-Fannich-Carn Chuinneag area with those reported in other areas

Correlation of the deformation history of the Ullapool-Fannich-Carn Chuinneag area with those of other areas of the Moine schists has recently taken on more significance. It has become apparent that different areas of the Moine rocks have undergone radically different tectono-metamorphic histories. Roberts and Harris (1983) postulated that rocks east of the Quoich line underwent polyphase deformation and high grade metamorphism prior to the Caledonian Orogeny with relatively minor re-working during the Caledonian. They contrasted this pattern with rocks west of the Quoich line which underwent polyphase deformation and meta-morphism during the Caledonian Orogeny. Roberts and Harris (op.cit.) proposed the Quoich line as an eastern limit of crustal reworking during the Caledonian. However, it should be noted that this interpretation is not universally accepted. Strachan (1982) preferred to regard the Quoich line as a slide zone and Winchester et al. (1979) regarded it as an unconformity. More recent studies tend to confirm the interpretation of Roberts and Harris (Millar, pers. comm.). If the interpretation of Roberts and Harris is correct, it points to a stark contrast between the area east of the Quoich line and the Fannich-Ullapool-Carn Chuinneag area, west of the Quoich line. The Morar Division rocks of this area show only sparse evidence of possible pre-Caledonian deformation and metamorphism (that is pre-intrusion of the Carn Chuinneag granite (SBS.4.a,b; Shepherd, 1973)).

Correlation of structures from Fannich to Quoich is therefore very difficult and Roberts and Harris did not attempt any fold phase to fold phase correlations. They suggested that the steep belt of the Quoich line formed during the local D3 event was folded by later D4 warping, these two deformation phases representing Caledonian events. It is tentatively suggested that the major D3 folds of Quoich are of the same generation as the major upright D3 folds in Fannich. However, this does not imply that the D1 and D2 structures identified in Fannich and Quoich are identical since D2 in Fannich is Sgurr Beag Slide age and D2 in Quoich is almost certainly pre-Caledonian (Holdsworth & Roberts, 1984).

It seems likely that the Moine Thrust related deformation wanes rapidly eastwards from Ullapool to the Carn Chuinneag area (fig.46) and thus, by analogy, may have little or no expression in the Quoich area.

In northern Ross-shire, the evidence for early deformation in the Morar Division is scarce (Section SBS.3.a) whereas within the Glenfinnan Division rocks there is evidence for an extensive pre-slide history (Section SBS.3).

The apparently simple Morar Division history in northern Ross-shire contrasts with that reported from the SW Moine area (Powell, 1974; Baird, 1982). The Glenfinnan Division rocks of both Ross-shire and the type area of Morar have, however, broadly similar histories with polyphase deformation and migmatization prior to the Sgurr Beag Slide related deformation.

In the Morar division rocks of the SW Moine, polyphase deformation and garnet grade metamorphism are held to have taken place prior to ca.770Ma (Powell et al., 1983). In contrast, the Morar Division in northern Ross-shire has a simpler early tectonometamorphic history, apparently involving at most, low grade metamorphism and rare isoclinal folds (Shepherd, 1970; Wilson, 1975; Wilson & Shepherd, 1979). Further studies will be required before the change between Morar and northern Ross-shire is understood.

That part of the northern Highlands from the SW Moine area to Ullapool and Fannich seems, however, to have undergone a similar history since emplacement of the Glenfinnan rocks along the Sgurr Beag Slide; large asymmetrical N-S trending folds deform the originally relatively flat-lying Sgurr Beag Slide. In northern Ross-shire, however, these folds can be directly related to formation of the mylonites in the Moine Thrust zone, whereas further south such a relationship has not yet been suggested.

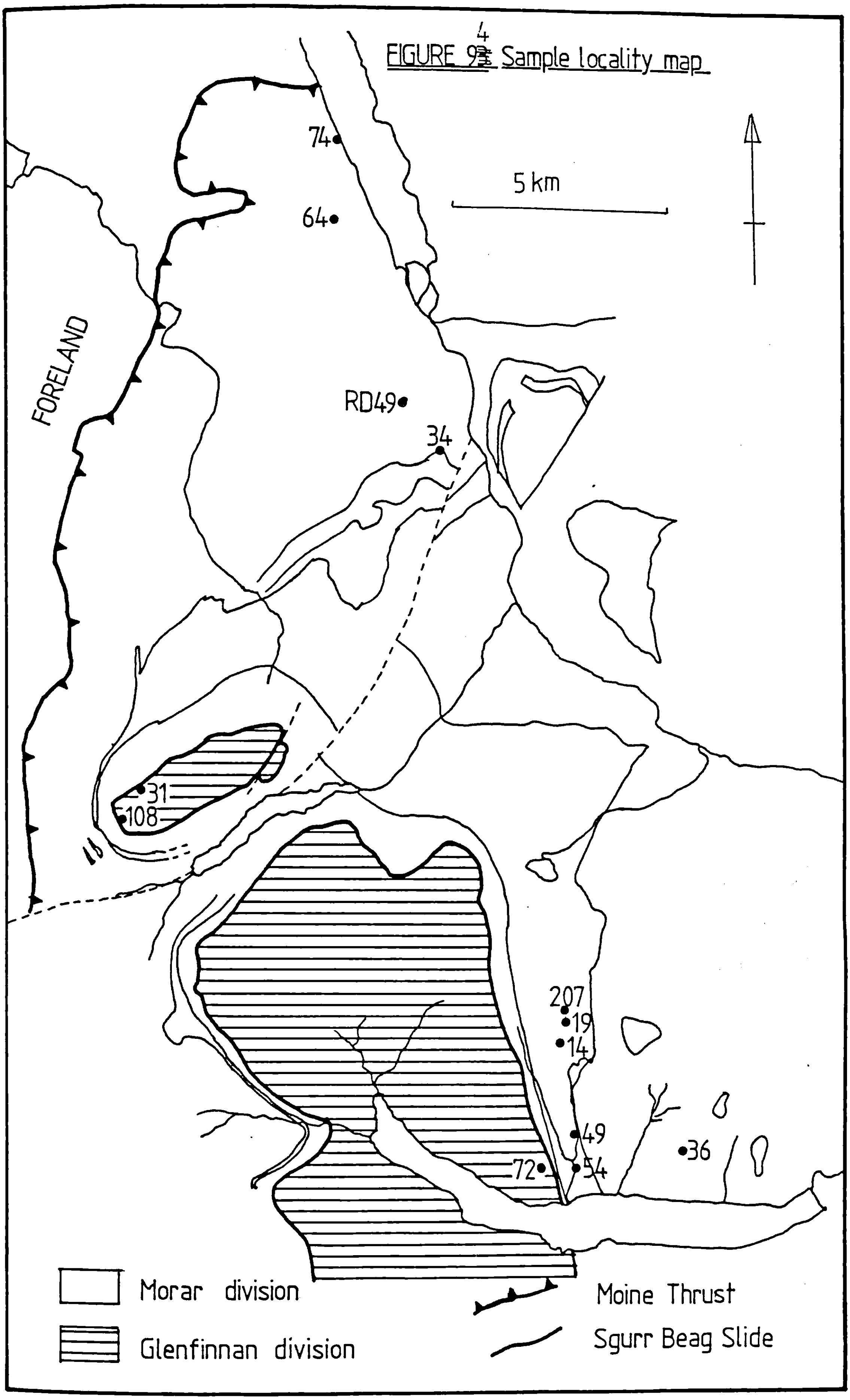


FIGURE ~~9~~ 4 Sample locality map

CHAPTER IV. GEOCHRONOLOGY

Sample Preparation

Samples weighing between 1 and 2kg were collected from pelites and semi-pelites during the fieldwork in the area from Loch Fannich to the Moine Thrust (fig.94). The cleanest and least weathered samples were selected for use in K-Ar analysis.

Samples were initially washed to remove any soil and organic material. They were then reduced to blocks less than 5cm in diameter using a hydraulic splitter and jaw crusher to below 8mm grain size.

At this stage samples were sieved and the finer fractions retained while those which did not pass through a 30 # (500μ) sieve were reduced in a cone grinder and re-sieved. This was continued until over 90% of the sample had been passed through a 500μ sieve.

Several fractions were selected for analysis but not every fraction was recovered from all samples. Those used for potassium and argon analysis were:

30 - 60 # (250-500μ)
60 - 72 # (200-250μ)
72 - 120 # (125-200μ)
120 - 240 # (75-125μ)

The sieved fractions were washed to remove very fine grained particles which adhere to the surfaces of larger grains. Washed, sieved fractions were then passed through a rotating disc, dry magnetic separator (Davis separator) giving a crude separation of biotite + garnet + opaques from quartz + feldspar + muscovite. A heavy liquid separation technique was then used to obtain refined muscovite and biotite concentrations from the first magnetic separations. It is not possible to quote exact specific gravities for the minerals because they varied slightly from sample to sample falling into the general ranges:

Biotite	2.86 - 3.16
Muscovite	2.6 - 3.0

The plagioclase for sample SK42 was obtained by handpicking from the non-magnetic fraction which floated in a liquid with a specific gravity of 2.7

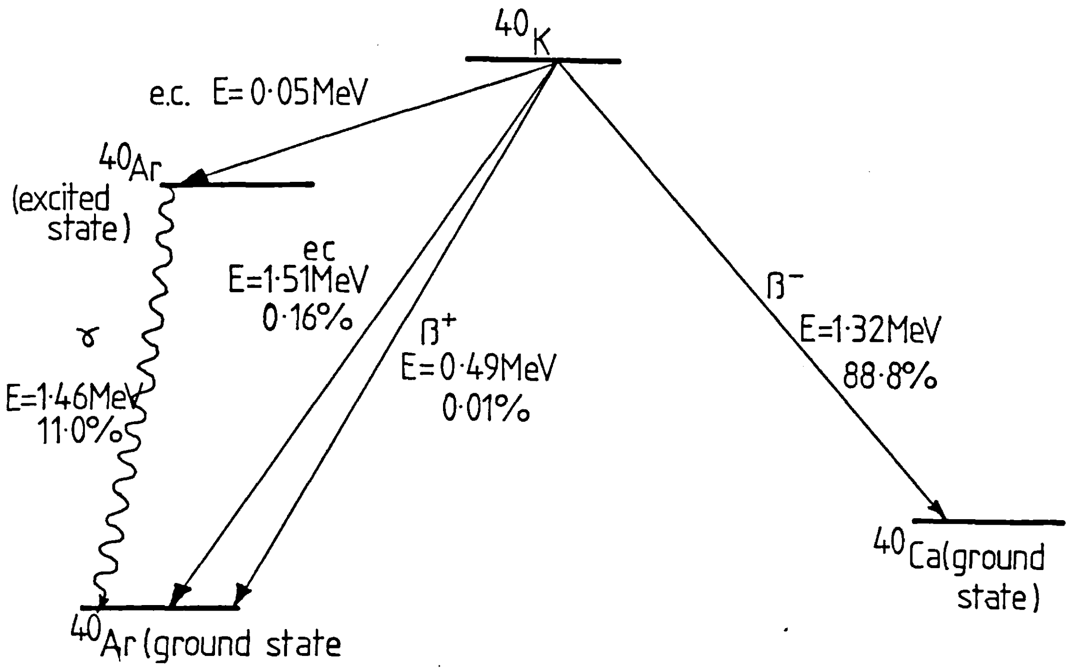
The concentrations of muscovite and biotite were purified further using a Franz Isodynamic separator until they were better than 99% pure. Special care was taken with biotite samples to remove chlorite and partially chloritized biotite. Chlorite has an erratic retention potential for argon and could have an effect upon the final result.

GC.1 The K-Ar Dating Method

GC.1.a Theory

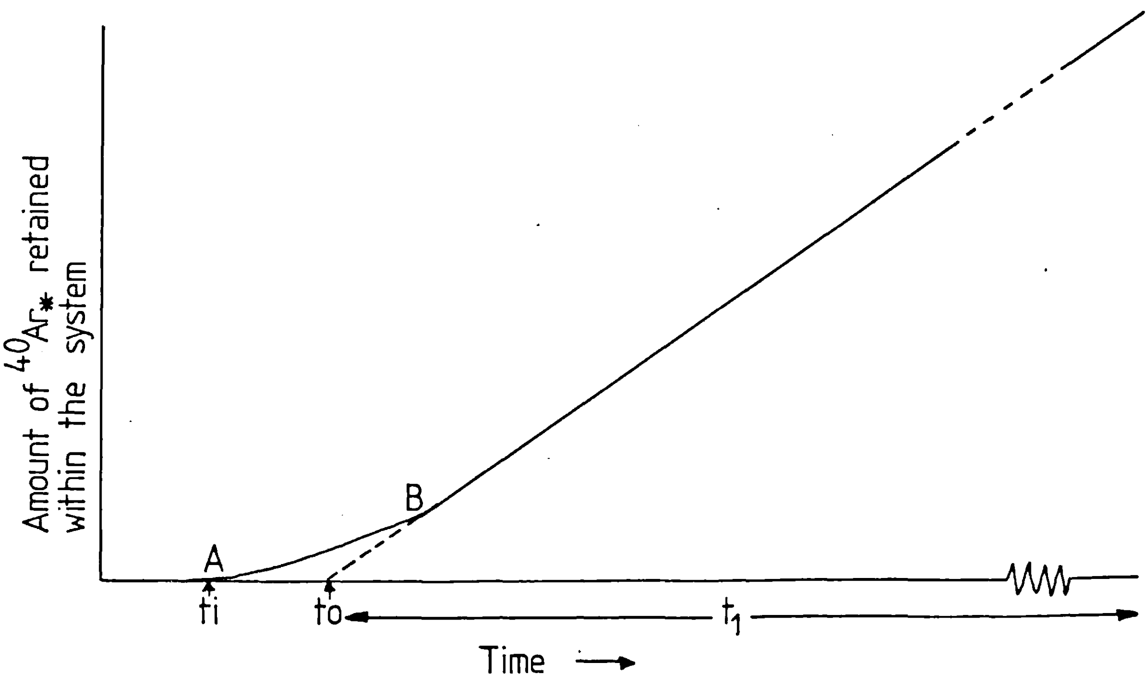
The K-Ar dating method utilises the decay of ^{40}K to ^{40}Ar in natural systems. ^{40}K comprises 0.01167% of the total atomic abundance of potassium in present-day rocks. It undergoes decay by several mechanisms (fig.95); 88.8% decays to ^{40}Ca but this is also the most common isotope of Ca in nature. Several attempts have been made to use the decay of ^{40}K to ^{40}Ca to date rocks, but naturally occurring ^{40}Ca swamps any which is produced by decay of ^{40}K resulting in large errors in any dates obtained. The remaining 11.2% ^{40}K decays to ^{40}Ar by three mechanisms (fig.95): decay by electron capture direct to the ground state, decay by emission of a B^+ particle and, most important, decay by electron capture to an

FIGURE 95. Decay of ^{40}K



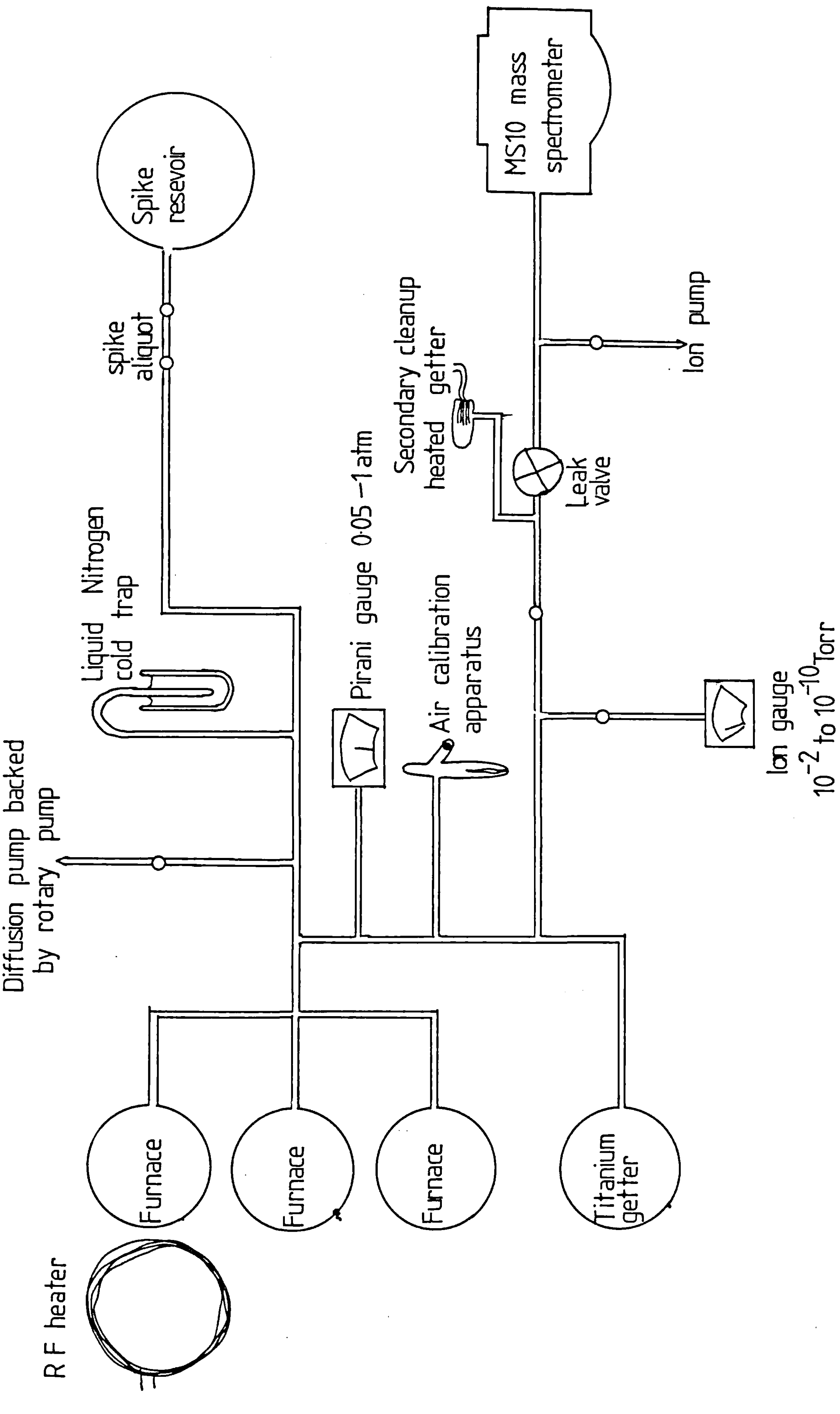
after Dalrymple and Lanphere (1969)

FIGURE 96



The time $t_0 - t_i$ is regarded as instantaneous with respect to the total age, t_1 of the system.

FIGURE 97
Extraction system attached to the MS10 mass spectrometer



excited state of ^{40}Ar and emission of a γ-ray dropping the energy to the ground state.

For the general case of a parent (P) isotope decaying exponentially to a daughter (D) isotope, the equation relating the number of daughter atoms at the present day to the number of parent atoms remaining is:

$$D = P(e^{\lambda t} - 1)$$

where λ = decay constant and t = time since initiation of a closed system.

When two daughter products are involved this becomes:

$$D_1 + D_2 = P(e^{(\lambda_\epsilon + \lambda_\beta)t} - 1)$$

Therefore,

$$^{40}Ar_* + {}^{40}Ca_* = {}^{40}K\ (e^{(\lambda_\epsilon + \lambda_\beta)t} - 1)$$

where λ_ϵ = decay constant for the reaction $^{40}K \rightarrow {}^{40}Ar$

λ_β = " " " " " $^{40}K \rightarrow {}^{40}Ca$

$^{40}Ar_*$ = atoms of radiogenic ^{40}Ar.

$^{40}Ca_*$ = " " " ^{40}Ca.

The ^{40}Ar and ^{40}Ca daughter atoms are produced in the ratio of $\lambda_\epsilon/\lambda_\beta$. For the calculation of K-Ar decay ages, only the decay to ^{40}Ar is used so the equation becomes:

$$^{40}Ar_* = {}^{40}K \frac{\lambda_\epsilon}{\lambda_\epsilon + \lambda_\beta}\left[e^{(\lambda_\epsilon + \lambda_\beta)t} - 1\right]$$

$$t = \frac{1}{\lambda_\epsilon + \lambda_\beta} \log_e\left[\frac{^{40}Ar_*}{^{40}K}\left(\frac{\lambda_\epsilon + \lambda_\beta}{\lambda_\epsilon}\right) + 1\right]$$

substituting the values recommended by Steiger and Jäger (1977)

$\lambda_\epsilon = 0.581 \times 10^{-10}\ yr^{-1}$ $\lambda_\beta = 4.962 \times 10^{-10}\ yr^{-1}$

$$t = \frac{1}{5.543 \times 10^{-10}} \log_e\left[\frac{^{40}Ar_*}{^{40}K}\ \frac{5.543 \times 10^{-10}}{0.581 \times 10^{-10}} + 1\right]$$

In order that the age obtained may be interpreted as the time since closure or "true age", several assumptions must be made. Several lists of the assumptions are available (Dalrymple & Lanphere, 1969; Faure, 1977). They can be summarised as follows:

1. The general physical assumptions about radio-active decay are true, i.e. there is a constant decay rate regardless of physical conditions.

2. The chemical system has remained closed since its closure to the diffusion of potassium and argon.

3. The length of time between initial retention of argon and total retention was short relative to the length of time that the system has been 'closed' to argon diffusion.

4. No fractionation of potassium or argon isotopes has occurred, i.e. the ratio $^{40}K/^{41}K$ is the same for all materials at the present day and no fractionation of argon isotopes has occurred other than through the decay of ^{40}K.

Assumption No.1 is long established and well validated, see Dalrymple and Lanphere for review (Chapter 2, 1969).

Assumption No. 2 is probably the most discussed in geochronological literature since it is more often untrue than the other three. It is important to the present study and will be discussed further in Section GC.4.a).

Assumption No. 3 embodies the blocking concept (fig.96) (Dodson, 1973). The blocking point at which a system becomes closed is regarded as instantaneous with respect to the time since blocking occurred. This is an assumption to simplify the calculation; if this assumption were not made it would be very difficult to define an age for the system. The age could lie anywhere along the curve AB (fig.96). Further, in recent literature, the question of the validity of discrete blocking temperatures has been questioned (Chopin & Maluski, 1980; Desmons et al., 1982, Chopin & Maluski, 1982).

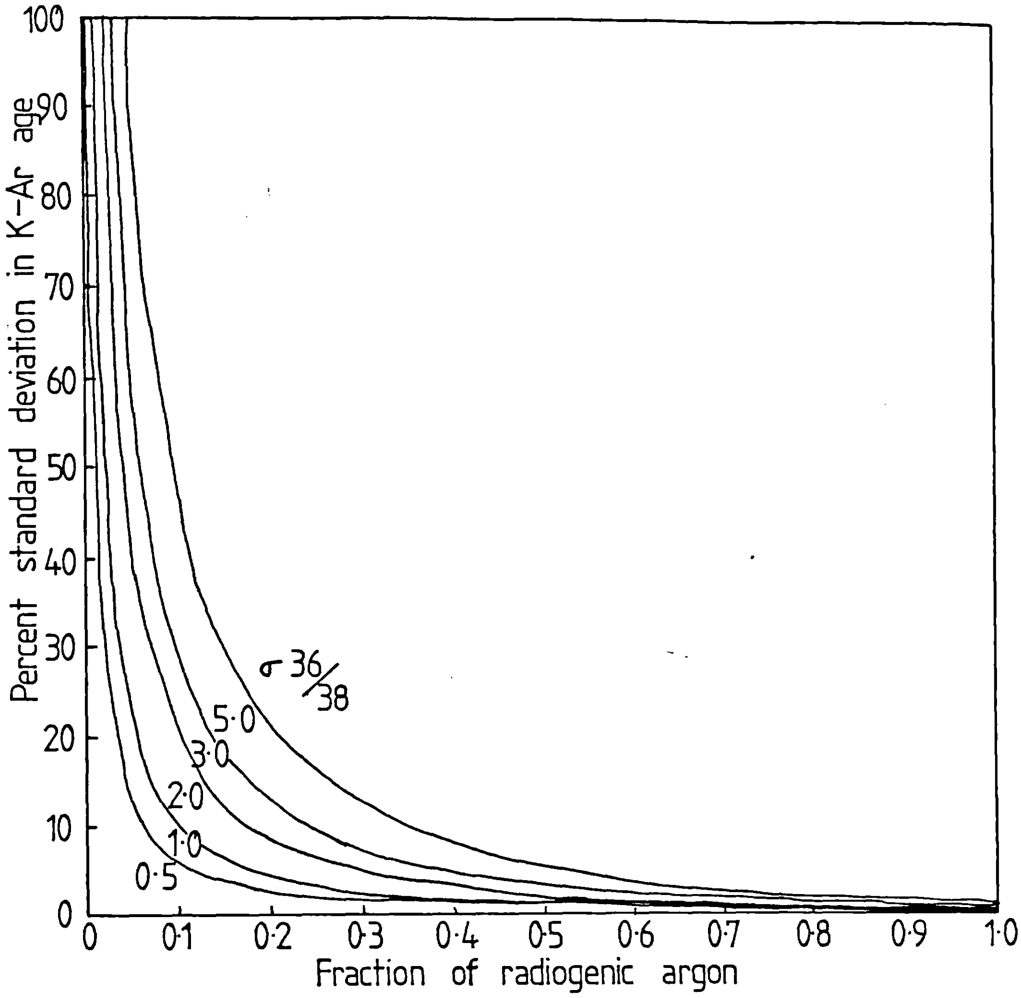
Assumption No. 4 is sometimes found to be invalid for argon isotopes, ^{40}Ar being introduced from outside the system (Brewer, 1969; Roddick et al., 1980; Harrison & McDougall, 1980, 1981; Allen & Stubbs, 1982).

This is also important to the present study and will be discussed in detail in Section GC.3.b. This assumption was also questioned in the 1930s, several workers claiming to have found fractionation of $^{40}K/^{41}K$ in nature. However, more recent work has determined that in the majority of cases, potassium fractionation is negligible (Kendall, 1960). Notable exceptions being meteorites where it has been shown that cosmic bombardment can cause a type of open system behaviour in the production of ^{40}K from ^{40}Ca (Burnett et al., 1960) and at igneous margins where fractionation has been shown to occur over distances of up to 3cm (Verbeek & Schreiner, 1967). The $^{40}K/K_{tot}$ value used during this study is 0.0001167 (Steiger & Jäger, 1977).

GC.1.b Potassium Analysis

Potassium analysis was carried out using an IL543 digital flame photometer with an internal lithium standard. Potassium analysis by flame photometry utilises the excitation of potassium ions in a regulated flame. Potassium is introduced into the flame in an aerosol of aqueous solution and air. The flame causes excitation of outer shell electrons to a higher energy state. As they decay to the ground state, light of a characteristic wavelength is released in the visible spectrum. The intensity of the emission is proportional to the concentration of potassium

FIGURE 98. Diminishing effect of error on 36/38 on the final error



In samples used in the present survey the fraction of radiogenic argon is typically greater than 0·9. The error in measurement of the 36/38 ratio is relatively unimportant to the final error on the age.

After Cox and Dalrymple (1967)

atoms present. Hence the concentration of potassium in the sample can be calculated.

Approximately 0.1g of sample was accurately weighed and digested in a mixture containing $8cm^3$ HF with $2cm^3$ $HClO_4$ on a hot plate. When the samples had evaporated to dryness, the residues were taken up in de-ionized water and made up to $200cm^3$ in a volumetric flask. The final concentration was generally around 25ppm potassium in this solution.

An internal lithium standard solution was used to minimise interference effects. Potassium emits a characteristic spectrum on excitation but interference from other elements in the sample solutions can mask the potassium emission relative to lithium in both standard and sample, minimises the interference effects.

$5cm^3$ of the sample solutions were diluted with $5cm^3$ of 200 ppm lithium standard solution. Standard potassium solution ($\simeq$ 50 ppm) and a blank solution were also diluted in this way. The flame photometer was calibrated using standard and blank solutions, intermediate concentrations were only analysed occasionally since the correlation between response and potassium concentration is linear and stable over the range of concentrations used. Sample concentrations were then simply read from the digital display and the potassium contents calculated using the dilution factors and weight of the original samples.

The calculation of an age requires that the concentration of ^{40}K is known. To convert K% to concentration ^{40}K:

$$^{40}K = \frac{K\% \times 0.0001167 \times 40}{100 \times 39.102} = K\% \times 1.194 \times 10^{-6} \text{ ppm.}$$

where 0.0001167 = The proportion of ^{40}K to total potassium at present day

39.102 = Atomic weight of natural potassium.

The errors in potassium analysis arise mainly in the preparation of the sample prior to flame photometry. The IL543 flame photometer produces a direct digital readout and by re-introducing the same sample solutions several times, readings can be reproduced to within 0.0005%.

The laboratory standard used at BGS for potassium and argon analyses during this study was Mo40 Biotite. It is taken from the Ardgour gneiss within the Moine rocks of NW Scotland. The reproducibility of potassium analyses can be assessed by taking a mean and standard deviation for the standards analysed during the survey. Forty-four analyses were undertaken giving a mean of 7.003% potassium with a standard deviation of 0.062 (0.88%) at the 1σ level. An uncertainty of 1% in potassium content was used in determinations of the final K-Ar ages.

The errors in potassium analysis are introduced mainly during the wet chemistry. Errors in weighing and in the accuracy of the IL543 flame photometer are insignificant.

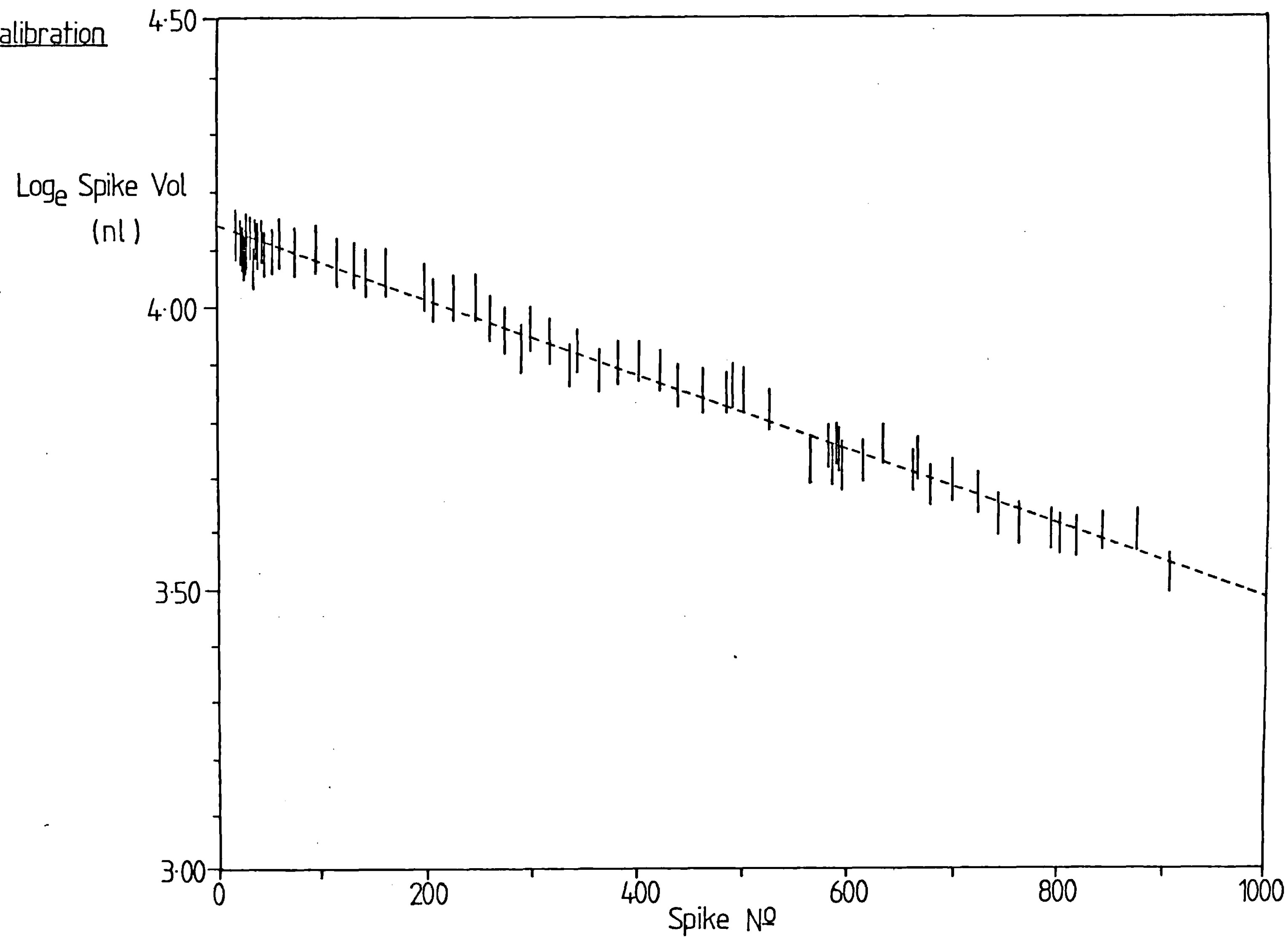
(Potassium results are tabulated in Appendix 1).

GC.1.c <u>Argon Analysis</u>

The argon analyses were undertaken by fusing samples under high vacuum conditions and measuring the quantities of released argon isotopes by the isotope dilution method in an AEI MS10A mass spectrometer.

Approximately 0.5g of sample were weighed accurately into a molybdenum crucible and packed in with glass wool to prevent sample loss during fusion. The crucibles were supported in pyrex furnaces which form part of the high vacuum system attached to the MS10 (fig. 97). Air was removed from the extraction system using a rotary pump and when the pressure had fallen sufficiently, a diffusion pump backed by a rotary pump.

FIGURE 99. Spike calibration graph



The system was baked for at least 8 hours at approximately 150°C to remove gases adsorbed to the sample and glass walls of the system. The pressure resulting was generally below 10^{-7} torr unless the system was leaking. Argon was measured by introducing a known amount of ^{38}Ar spike into the system from a reservoir via a calibrated pipette (fig. 97). The spike was calibrated using air capsules of known volume. Spike calibrations were performed at intervals of about 20 analyses. The amount of spike delivered decays exponentially as the amount of gas in the reservoir is reduced and its partial pressure falls. The volume of spike delivered is predicted by plotting the logarithm of calibration volumes against spike number. This produces, by regression, a straight line of the type Log_e Vol = Mx + c (where x = spike number, c= initial volume). Future spike volumes are predicted by extrapolation (fig. 99).

A known amount of sample was fused at 1200-1300°C using an RF induction coil which was external to the system. The gases evolved mixed with the ^{38}Ar spike which had previously been introduced into the system via the pipette. H_2O and CO_2 were removed by freezing down in a cold finger condenser with liquid nitrogen (fig. 97). Other active gases were removed by adsorption onto a hot titanium sponge. These processes resulted in a mixture of inert gases (of which argon is the most abundant) produced during the fusion and the ^{38}Ar spike. The gas sample was then equilibrated into a smaller volume (pipette) and a secondary 'clean up' was achieved using a heated zirconium/aluminium filament to adsorb the remnants of any active gases (fig. 97). The aliquot was then allowed to equilibrate statically into the Mass spectrometer.

The isotopes ^{40}Ar, ^{38}Ar and ^{36}Ar were measured and the ratios $^{40}Ar/^{38}Ar$ and $^{38}Ar/^{36}Ar$ calculated. Very nearly all ^{36}Ar present is due to atmospheric contamination (^{36}Ar is not a product in any terrestrial radioactive decay series) the contribution of ^{36}Ar in the spike is negligible. The natural $^{40}Ar/^{36}Ar$ ratio of the atmosphere is known to be 295.5 (Steiger & Jäger, 1977) and hence the ^{40}Ar can be corrected for air contamination. The spike contains 99.9% ^{38}Ar and insignificant amounts of ^{40}Ar and ^{36}Ar in comparison to sample size (Section GC.1.d). The amount of ^{40}Ar present due to the decay of ^{40}K can therefore be calculated:

$$\frac{^{40}Ar}{^{38}Ar} - \left(\frac{^{36}Ar}{^{38}Ar}_m - \underbrace{\frac{^{36}Ar}{^{38}Ar}_{Sp}}_{\text{negligible}}\right) \times 295.5 - \underbrace{\frac{^{40}Ar}{^{38}Ar}_{Sp}}_{\text{negligible}} = \frac{\text{Vol } ^{40}Ar \text{ radiogenic}}{\text{Vol } ^{38}Ar \text{ spike}}$$

The volume of ^{40}Ar rad must be converted to gram atoms per gram in order to substitute into the age equation. This is achieved by assuming that 1 mole of gas occupies 22.414l at s.t.p.

The errors inherent in argon analysis stem mainly from the calibration of the ^{38}Ar spike and measurement of atmospheric contamination. The atmospheric contamination measured in the muscovite and biotite samples analysed in this survey amounted to generally less than 10% (see Appendix 1) of the total ^{40}Ar present. The error in measuring the $^{36}Ar/^{38}Ar$ ratio has therefore relatively little effect on the final error on the age due to an error diminishing effect in the calculation (fig.98).

The error in calibrating the ^{38}Ar spike volume is more important to the final error analysis. The main problem occurs in the use of calibrated air capsules. The capsules were made by pulling out thin glass tubing, sealing one end and filling with mercury. They were weighed and the mercury removed as completely as possible. The capsules were then weighed empty and care was taken to assure the air within them was dry before they were finally sealed. Capsules with known volume were placed in the vacuum system and broken after an aliquot of

^{38}Ar spike had been introduced. The ratios of $^{40}Ar/^{38}Ar$ and $^{36}Ar/^{38}Ar$ were measured and the volume of spike calculated from the ratios using the known volume of air in the capsule and $^{40}Ar/^{38}Ar$ ratio of the atmosphere.

The $^{40}Ar/^{36}Ar$ ratio of the air aliquot can be deduced independently and used to correct analyses for discrimination of argon isotopes in the mass spectrometer.

The capsules were made in batches of between 5 and 15. No correlation between batches and fluctuations in the spike calibration line are seen (fig. 99). It is concluded that series errors between batches are smaller than random errors introduced during the making of capsules. Duplicates performed on several aliquots of the same gas sample show that errors in measurement of the isotopic ratios are very much smaller than fluctuations about the calibration line.

Individual determinations deviate from the line of best fit (fig. 99) by an average of 0.5%. These deviations are attributable to variations in the amount of spike introduced by the spike pipette system and more importantly, introduced during the making of the spike calibration capsules.

The accuracy of spike predictions increases as the number of calibrations increases, however there are indications of gentle fluctuations in the volumes of spike, perhaps due to irreproducible pipetting effects. Therefore only the preceding 30 calibrations were used to predict new spike volumes. The error in prediction of new spike volumes was approximately 1% at the 1 σ level when calculated in this manner.

GC.1.d Age Calculation

Calculation of the ages from potassium content and radiogenic argon content was carried out by a microcomputer (HP.9810A) which processed the Mass spectrometer data 'on-line'. However, a calculation is presented here in order to illustrate the method used to arrive at the age and state some of the assumptions which have to be made.

The argon isotope proportions measured are composite, consisting of radiogenic, atmospheric and spike components. It is necessary to know the contribution that spike ^{38}Ar makes to the total ^{38}Ar peak measured and the contribution that radiogenic ^{40}Ar makes to the total ^{40}Ar peak.

Consider the ^{38}Ar peak measured in the mass spectrometer; it can be shown that the contribution to the ^{38}Ar peak by spike ^{38}Ar is:

$$\frac{{}^{36}/_{38}\,\mathrm{At} - {}^{36}/_{38}\,\mathrm{m}}{{}^{36}/_{38}\,\mathrm{At} - {}^{36}/_{38}\,\mathrm{Sp}} \qquad (1)$$

where

At = atmospheric; m = measured; Sp = spike.

The $^{36}/38$ ratio of the spike, measured over a period of time was 0.000028 and regarded as negligible in the calculation. The $^{36}/38$ ratio of atmospheric argon is 5.35. Using the ratios obtained from a coarse biotite, SK14 from Fannich:

$$^{40}/_{38} = 1.7401652$$

$$^{36}/_{38} = 0.0003881.$$

Equation (1) becomes $\dfrac{5.35 - 0.0003881}{5.35} = 0.999928$

The contribution of the spike argon to the measured ^{36}Ar peak is:

$$(1) \times (36/38)_{Sp} = 2.7997 \times 10^{-5} \qquad (2)$$

The total atmospheric contribution to the ^{36}Ar peak is:

$$^{36}/_{38\ m} \ - \ (2) \ = \ 3.601 \times 10^{-4} \qquad (3)$$

The contribution to the ^{40}Ar peak by spike is negligible because the amount of radiogenic ^{40}Ar is large. The atmospheric contribution to the ^{40}Ar peak is:

$$(3) \times {}^{40}/36_{\ At} \ = \ 3.601 \times 295.5 \ = \ 0.1064096 \qquad (4)$$

The radiogenic contribution to the ^{40}Ar peak is:

$$^{40}/39_{\ m} \ - \ (4) \ = \ 1.7401652 - 01/064096$$

$$= \ 1.63376 \qquad (5)$$

The amount of radiogenic ^{40}Ar is now determined using the ratio between the unknown ^{40}Ar and known ^{38}Ar responses in the mass spectrometer.

$(5)/_1$ x Vol of ^{38}Ar spike.

$$\frac{1.63376}{0.999928} \times 40.5019 \times 10^{-9} l \ = \ 66.175 \times 10^{-9} l.$$

22.4141 of any gas weighs 1 gram-atom at s.t.p. (i.e. 40g for ^{40}Ar)

1l ^{40}Ar weighs 1.785g

$66.175 \times 10^{-9} \times 1.785 \ = \ 1.18122 \times 10^{-7} g \ ^{40}Ar$.

The biotite sample weighed 0.4998g

$1.18122 \times 10^{-7} g \ ^{40}Ar$ is 0.4998g biotite = 0.2363 ppm ^{40}Ar.

Normal potassium contains 0.01167% ^{40}K (Steiger & Jäger, 1977).

Therefore 1g Potassium contains $\dfrac{0.0001167 \times \text{At wt } ^{40}K}{\text{At wt normal K}}$

$$= \ 1.194 \times 10^{-6} g \ ^{40}K$$

Thus K% x 1.194 gives ^{40}K ppm directly.

SK14 biotite contains 6.96% potassium.

$6.96 \times 1.194 = 8.3102$ ppm ^{40}K.

Substituting into the K-Ar age equation (Section GC.1.a)

$$t = \frac{1}{5.543 \times 10^{-10}} \ \ln\left[\frac{^{40}Ar\ rad}{^{40}K} \ \frac{5.543 \times 10^{-10}}{0.581 \times 10^{-10}} + 1\right]$$

$\underline{t = 433.02 \text{ Ma.}}$

The standard deviation on this age can be calculated by combining known errors from experimental data using the age calculation to indicate methods of combination.

$$\text{Assume} \left({}^{36}/{}_{38}\right)_m - \left({}^{36}/{}_{38}\right)_{sp} = Z$$

$$\text{Then } A = \text{Error on } Z = \sqrt{\left(E36/38_m\right)^2 + \left(E36/38_{sp}\right)^2}$$

$$B = E \text{ on } Z \times 295.5 = \sqrt{\left(\frac{A}{Z}\right)^2 + \left(\frac{E\left({}^{40}/{}_{36}\right)At}{\left({}^{40}/{}_{36}\right)_{At}}\right)^2}$$

$$C = \text{Error on } \frac{40\ rad}{38\ sp} = \sqrt{\left(B \times 2 \times 295.5\right)^2 + \left(E\left({}^{40}/{}_{38}\right)_{sp}\right)^2 + \left(E\left({}^{40}/{}_{38}\right)_m\right)^2}$$

$$D = \text{Error on Vol of } {}^{40}Ar \text{ rad} = \sqrt{\left(\frac{C}{\left(\frac{40\ rad}{38\ sp}\right)}\right)^2 + \left(\frac{E \text{ on spike Vol}}{\text{spike Vol}}\right)^2}$$

$$F = \text{Error on } \frac{^{40}Ar}{^{40}K} = \sqrt{D^2 + \left(\frac{E \text{ on } k\%}{k\%}\right)^2}$$

Error on the final age =

$$\text{Age} \times F \times \left[\frac{R \times 9.5407}{1 + (R \times 9.5407) \ \ln(1 + R \times 9.5407)}\right]$$

The precision of $\left({}^{40}Ar/{}^{38}Ar\right)_m$ and $\left({}^{36}Ar/{}^{38}Ar\right)_m$ is obtained from an average of 6 scans of the peaks in the mass spectrometer. In this case (SK14 biotite) they were 0.000115 and 0.0000047 respectively. From calculations during the atmospheric ${}^{40}Ar/{}^{36}Ar$ measurement, the error that ratio is about 0.5%. The error on $\left({}^{40}Ar/{}^{38}Ar\right)_{sp}$ is negligible since the radiogenic argon peak is large. The error on $\left({}^{36}Ar/{}^{38}Ar\right)_{sp}$ is 2.8×10^{-5} calculated from many measurements of spike composition on the MS10 mass spectrometer. The error on spike volume (fig. 99) predicted for any particular run is less than 1% (see Section GC.1.c). However, the figure of 1% will be used since this is likely to be a maximum. The error in measurement of potassium content is also less than 1% (see Section GC.1.b) but here again 1% will be used to give a conservative estimate of the errors.

Substituting the above errors into the error equations, the error on the age of SK14 biotite is 5.4Ma. This represents an error of 1.25%.

It is conventional to quote K-Ar ages with errors of two standard deviations (2σ) representing a 95% probability of the real answer lying within the defined zone.

The age of SK14 biotite is therefore 433.0 $\pm$ 10.8 Ma

GC.2 The Rb-Sr Dating Method

GC.2.a Theory

The Rb-Sr dating method has been used to a small extent during this study. Use was initiated in order to detect the presence of excess radiogenic argon (Sections GC.1.a, GC.3.b) in samples which had already been analysed by the K-Ar method. It has also been used to illustrate the relationship between the different dating methods in relation to muscovite and biotite in the Fannich area.

Rb-Sr dating uses the decay of ^{87}Rb to ^{87}Sr in the reaction:

$$^{87}_{37}Rb \rightarrow ^{87}_{38}Sr + \bar{\nu} + \beta^- + Q$$

So the general decay equation; $D = Do + P(e^{\lambda t} - 1)$ becomes:

$$^{87}Sr = ^{87}Sr_o + ^{87}Rb(e^{\lambda t} - 1)$$

FIGURE 100 Mineral Isochron, SK14

Age = 404±4Ma (2-sigma)
Intercept = 0·7297±0·0002

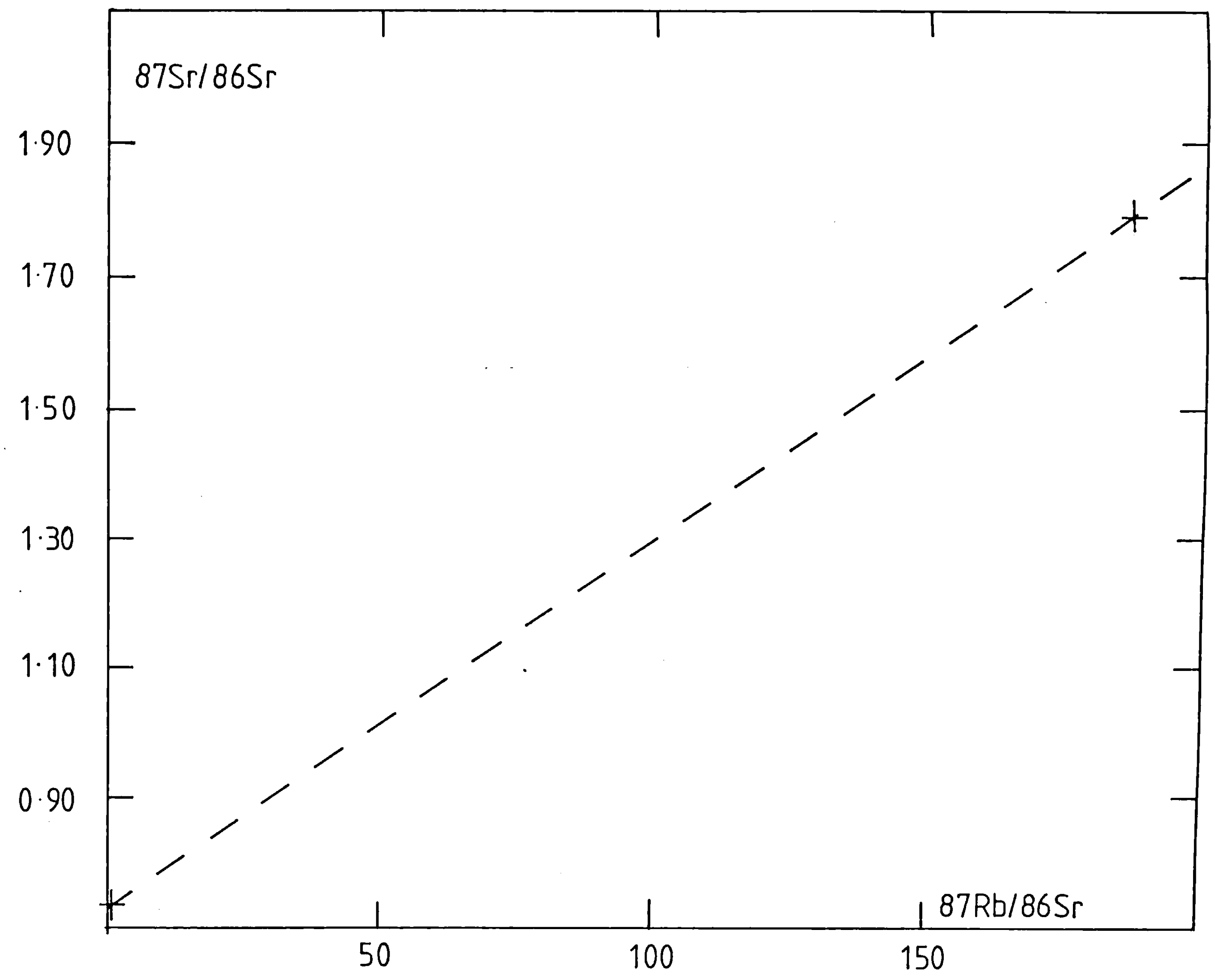
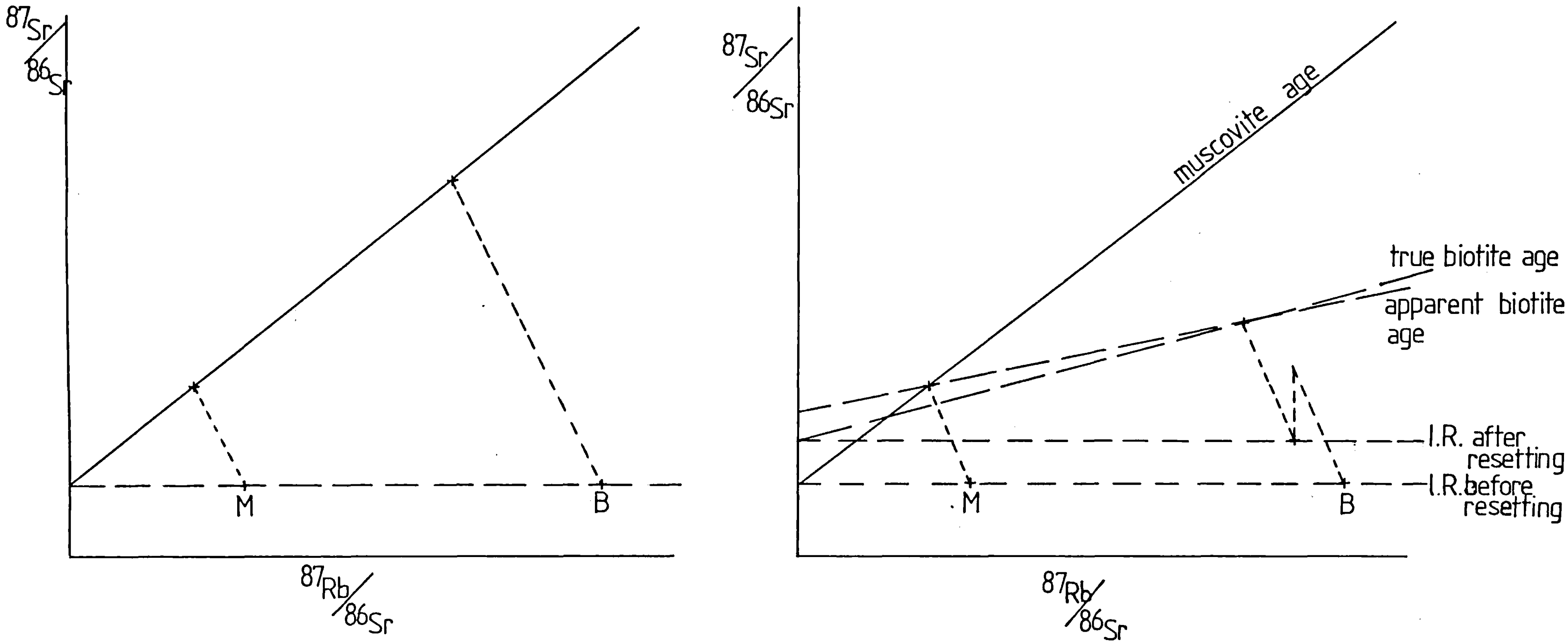


FIGURE 102. (see folder for fig 101)
Effect of an "old" muscovite on a muscovite/biotite Rb-Sr age



Both minerals closed at approximately the same time yeilding a single Rb-Sr age

If the biotite system is reset and the muscovite system remained open, the muscovite/biotite tie line yeilds spurious ages. These may be low (above) or high

($^{87}Sr_o$ is the amount of daughter present in the mineral at the time of closure to strontium diffusion.)

This equation can be modified by dividing the equation by ^{86}Sr, a naturally occurring isotope of strontium, not produced during any terrestrial decay series:

$$\frac{^{87}Sr}{^{86}Sr} = \frac{^{87}Sr}{^{86}Sr_o} + \frac{^{87}Rb}{^{86}Sr}\left(e^{\lambda t} - 1\right) \qquad (A)$$

$\left(\frac{^{87}Sr}{^{86}Sr}\right)_o$ is known as the 'initial ratio'.

Solving for t, the equation becomes:

$$t = \frac{2.303}{\lambda} \log_{10} \left[\frac{\left(\frac{^{87}Sr}{^{86}Sr}\right) - \left(\frac{^{87}Sr}{^{86}Sr}\right)_o}{\left(\frac{^{87}Rb}{^{86}Sr}\right)} + 1 \right]$$

The decay constant (λ) for ^{87}Rb has an agreed value of $1.42 \times 10^{-11} yr.^{-1}$ (Steiger & Jäger, 1977) which is important because many of the ages determined in the 1960s and early 70s in Scotland have to be recalculated before comparison with those of more recent determinations.

The value of t obtained is geologically meaningful only when the system has remained closed. The only change in concentration and ratios of rubidium and strontium are due to the radioactive decay of ^{87}Rb.

The assumptions made are similar to those of the K-Ar method except that the amount of daughter product present at the time of initial cooling is calculated, (not assumed to be negligible as in the K-Ar method). In this instance the Rb-Sr method has been applied to muscovite/biotite mica pairs, solving the isochron equation (A) simultaneously. The isotope dilution method has been applied so both rubidium and strontium were measured using an MM30 solid source mass spectrometer (Section GC.2.b). Strontium contents of the biotite were often as low as 10ppm (see Appendix 2).

In most of the samples used, biotite has large Rb/Sr ratios (50-250 - see Appendix 2) and muscovite has lower Rb/Sr ratios (0.5-10 - see Appendix 2). When plotted on an isochron diagram, co-existing muscovites plot close to the origin whereas biotites plot away from both axes. In this way, the muscovites act as points to tie the initial $\frac{^{87}Sr}{^{86}Sr}$ ratio for the corresponding biotites. The initial ratio for biotites can be estimated and an approximate age obtained. However, the muscovite point reduces the errors on the age although it does not contribute to the age significantly.

The age obtained from such plots is a 'model' age for biotite by the Rb-Sr method.

Several large muscovite books were analysed in the same way, by isotope dilution, but in the case of SKND42, plagioclase was used to tie down the initial ratio. The same initial ratio was applied to SKND35, since both are thought to originate from the Carn Chuinneag granite.

GC.2.b <u>Isotope dilution and Mass spectrometry</u>

Both rubidium and strontium were extracted from the samples and measured using the MM30 solid source mass spectrometer. The isotope dilution technique involves adding a known amount of spike which is highly enriched in one of the stable isotopes of rubidium and strontium (^{85}Rb and ^{84}Sr, respectively) to a known amount of sample at a ratio of approximately 1:1. The isotopic ratios are measured, and thus the

$^{87}Sr/^{86}Sr$ ratio and absolute proportions of rubidium and strontium in the sample can be calculated.

The samples with spikes added, were digested in HNO_3 and HF to dryness. Fluorides were removed by adding 2cm^3 HNO_3, then 10cm^3 6M HCl was added and evaporated to dryness. The residue was taken up in 5cm^3 of 2.5 M HCl and due to their different chemical properties, rubidium moved through the column faster than strontium. In this way, rubidium and strontium were separated from the main interfering ions (ions which have mass numbers close to rubidium and strontium; (see Webster in Smales and Wager, 'Methods in Geochemistry').

The solutions concentrated in rubidium and strontium were collected and evaporated to dryness. They were then taken up in weak HNO_3 and evaporated on to fine tantalum filaments. Rubidium was loaded on triple filaments. In the mass spectrometer, the two side filaments were heated gently to evaporate the rubidium, the central filament was heated to a much higher temperature in order to ionize the vapour. In this way the fractionation of rubidium isotopes was reduced to a minimum. This arrangement is essential since only two isotopes of rubidium occur in significant amounts in nature and therefore no correction can be applied for the instrumental fractionation.

Strontium was loaded in the same way but onto a single filament. Four isotopes of strontium occur naturally and therefore a correction can be applied for fractionation (Boelrijk, 1968).

The precision required for rubidium analysis was 0.25%. The precision for strontium, about 0.005% on the $^{87}Sr/^{86}Sr$ ratio and 0.1% on the strontium concentration. The analysis of strontium therefore required a greater number of measurements to achieve the increased precision. The results are tabulated in Appendix 2.

GC.2.c Age Calculation

The Rb-Sr mica pair ages were calculated from the measurements of rubidium and strontium isotopes. Corrections for fractionation were made at this stage by an HP9845B computer, on line with the mass spectrometer (see Section GC.2.b).

Calculation of the age was carried out using another 9845B computer. In order to obtain a common age for the muscovite/biotite pairs they were plotted together on an isochron diagram (fig.100).

Rearranging equation (A), Section GC.2.a, it can be seen that the equation takes the standard form $y = mx + c$.

$$\left(\frac{^{87}Sr}{^{86}Sr}\right) = (e^{\lambda t} - 1)\left(\frac{^{87}Rb}{^{86}Sr}\right) + \left(\frac{^{87}Sr}{^{86}Sr}\right)_o$$

where = decay constant for $^{87}Rb = 1.42 \times 10^{-11} yr^{-1}$ (Steiger & Jäger,1977)

t = Age of the sample.

$\left(\frac{^{87}Sr}{^{86}Sr}\right)_o$ = ratio at the time of closure to Sr diffusion.

Hence the age of the sample is proportional to the slope of the graph.

Consider a mica pair for sample SKND14:

Muscovite: Rb = 170.781 ppm , Sr = 116.156 ppm
$^{87}Sr/^{86}Sr = 0.73554$

Biotite: Rb = 398.355 ppm , Sr = 10.332 ppm
$^{87}Sr/^{86}Sr = 1.41134$.

The following equation is used to convert the rubidium and strontium contents to an $^{87}Rb/^{86}Sr$ ratio:

$$\frac{^{87}Rb}{^{86}Sr} = \left(\frac{Rb}{Sr}\right)_c \times \frac{Ab\ ^{87}Rb \times At\ wt\ St}{Ab\ ^{86}Sr \times At\ wt\ Rb}$$

where $\left(\frac{Rb}{Sr}\right)_c$ is the ratio of amounts of "common" elements

Ab ^{87}Rb and Ab^{86}Sr are the relative natural abundances of the isotopes to all other isotopes of the element.

At wt is the atomic weight of the natural element.

A line through the data points for muscovite and biotite gives a slope of 5.734×10^{-3} (fig.100).

Slope = $e^{\lambda t} - 1$ $\qquad$ $\underline{\underline{t = 403.77\ Ma}}$

The analytical errors on each data point are plotted on the ^{87}Rb/^{86}Sr v's ^{87}Sr/^{86}Sr graph. Since the line fitted to these two points is unique, errors are restricted to the analytical errors of each point. When more than two points are involved, the line is fitted using the method outlined by York (1966) by weighing each point with its analytical errors.

When normally plotting an isochron diagram, 5 or more points are used. In this case the goodness of fit to a straight line is the mean square weighted deviation (MSWD), this is a measure of the scatter of points about the line. However, with only two data points, both are needed to define the slope and a measurement of MSWD is meaningless. In the case of a two-point isochron, such as those of the present survey, it must be assumed that the line is a valid isochron, since it cannot be checked using the MSWD.

In order to test the reproducibility of analytical errors, the analysis was repeated using two new aliquots of muscovite and biotite (Appendix 2). The analysis was found to be reproducible to within analytical errors. The analytical errors on the ages resulting from this analysis are approximately 1% (2σ). This is more precise than ages produced by the K-Ar method but in this case there must be the assumption that the isochron is valid.

The use of the muscovite/biotite pair effectively produces a biotite Rb-Sr cooling age. As discussed in Section GC.2.a, the age is very sensitive to changes in the biotite point and less sensitive to the muscovite point. It might even be suggested that muscovites could be of a different age without upsetting the Rb-Sr model age of the biotite since changes of up to 3% in the muscovite $^{87}Sr/^{86}Sr$ ratio make little difference to the final age. This is an important point in the present survey and will be discussed further in Section GC.4.

GC.3 Cooling and Uplift from Fannich to the Moine Thrust

GC.3.a Cooling outside the Thrust zone

The initial objective of the K-Ar investigations was to determine the cooling and uplift history from the internal mobile zone to the Moine Thrust belt and then to correlate this with the tectonometamorphic history of the area. To this end an 18km traverse from the thrust belt, just south of Ullapool to the eastern end of Loch Fannich, was studied using K-Ar and Rb-Sr determination on muscovite and biotite from samples, including 13 pelites and semi-pelites, from which muscovite and biotite were extracted and analysed (except localities 64 and 74). The samples had grainsizes of 125-200μ (except samples 64 and 74, from which grainsizes 75-125μ were used because they had undergone more intense mylonitization destroying larger grainsizes) The results obtained from these finer grainsizes indicates that this particular difference between the samples did not affect the argon content or the time of closure (but see Section GC.4).

In 7 localities (19, 31, 34, 36, 47, 49, 51; fig.101) the K-Ar muscovite age is older than the biotite (fig.101) but in 4 localities (14, 72, 198, 207) the reverse is true. This is consistent with previous studies in metamorphic terrains where it has been found that K-Ar muscovite ages are generally slightly older (circa 1Ma - 10Ma) than the biotite (Miller & Brown, 1965; Harper, 1967; Dewey & Pankhurst, 1970; Purdy & Jäger, 1976). This relationship has been interpreted in terms of the uplift and cooling of the rocks through the blocking temperature of, first muscovite and then at a lower temperature, biotite. Purdy and Jäger (1976) from work carried out in the Central Alps estimated that the blocking temperatures for muscovite and biotite are $350^{\circ} \pm 50^{\circ}$C and $300^{\circ} \pm 50^{\circ}$C respectively. More recently, direct measurements using ^{40}Ar - ^{39}Ar step-heating spectra have been produced by Berger and York (1981) but doubt can be thrown on the applicability of their experimental data (Melenevskiy et al.,1978).

In addition Harrison and MacDougall (1981) have suggested the biotite blocking temperature to be $280^{\circ} \pm 40^{\circ}$C combining experimental and borehole data, although they have not yet published the data supporting this estimate.

Purdy and Jäger (op.cit.) found that the difference between co-existing muscovite and biotite ages varied from 1-8 Ma depending upon the uplift rate, whereas Dewey and Pankhurst (1970) determined that the mean differences between muscovite and biotite for the Caledonides are in the range 5-10 Ma. Neglecting samples 31, 74 and 108, which will be discussed later (Section GC.3.b), in this study muscovites give an over-all mean age 3 Ma older than that for biotites. Unfortunately, no reliable cooling rates can be estimated from this data since the mean difference is 3 ± 6 Ma due to an error enhancement of the calculation.

This effect is not seen in Alpine ages since they are much younger and the determinations are therefore more precise. In the Moine examples reported here, the age differences are of the same order as those obtained in the Alps, thus confirming the 'normal' relationships of the Moine micas.

Neglecting localities 31, 74 and 108 (see Section GC.3.b) the plot of data for the traverse from Fannich to the Moine Thrust is relatively flat (fig.101). Both muscovite and biotite exhibit similar ages along the traverse. Muscovite gives the greater spread of ages: 424.5 ± 5.4 Ma (1σ), biotite gives a smaller spread: 421.4 ± 2.5 Ma (1σ).

Six of the micas were analysed by Rb-Sr isotope dilution method (Section GC.2). The results obtained show a flat distribution (fig.101) with ages below that for K-Ar in biotites (fig.101). They have a mean age of 411 Ma and standard deviation of 5 Ma. This age is considerably younger than the mean K-Ar biotite age of 421.4 Ma, in contrast to many other studies which argue that the blocking of Rb-Sr and K-Ar systems are roughly equal in biotite (Purdy & Jäger, 1976; Hanson & Gast, 1967; Roddick <u>et al.</u>, 1980).

Several possible explanations can be advanced for the relationships seen in the Moine traverse:

a) The muscovites and biotites are of different ages, muscovite carrying relict ages older than the biotite. Data from the micas would not therefore lie upon the same isochron. Such analyses could give an apparently young age for biotite.

b) The K-Ar system is giving an old age due to excess argon in the biotite lattice.

c) The biotite lattice, for some reason, closed to Sr diffusion later than it did to movement of argon.

These possibilities can be considered in more detail:

a) Whether or not muscovites had different ages to biotite cannot be detected in this survey (the muscovite $^{87}Sr/^{86}Sr$ ratios are too low and no accurate model ages can be calculated). An isochron diagram drawn between muscovite and biotite might give an anomalous age if the muscovite had been closed significantly earlier (fig.102) than biotite (e.g. biotite reset by a later event). The mica pair age may be anomalously high or low depending upon the relative Rb/Sr ratios of the micas and whole rock and this illustrates the major weakness of the two point mineral isochron; it gives no indication of a partially reset system. However, in the present case, if the problem does exist, it does not seem to be important. The ages obtained from 6 mica pairs were consistent, giving a spread similar to that seen from the K-Ar dating study. This may be due to the fact that any resetting event occurred before the build-up of radiogenic ^{87}Sr had altered the bulk $^{87}Sr/^{86}Sr$ ratio significantly.

b) A biotite sample yielding K-Ar ages older than the Rb-Sr age would normally be interpreted as containing excess argon. However, in the present case, other relationships do not indicate an excess of ^{40}Ar. The biotites have reproducible K-Ar ages, slightly younger than co-existing muscovites, which is the 'normal' relationship in metamorphic terrains. In areas exhibiting excess argon, biotites yield K-Ar ages which are apparently older than co-existing muscovites (Brewer, 1969; Roddick et al., 1980).

Further argument against excess argon is provided by the relationship between different grainsizes in Fannich (Section GC.4) and development of an identifiable excess argon zone adjacent to the Moine Thrust.

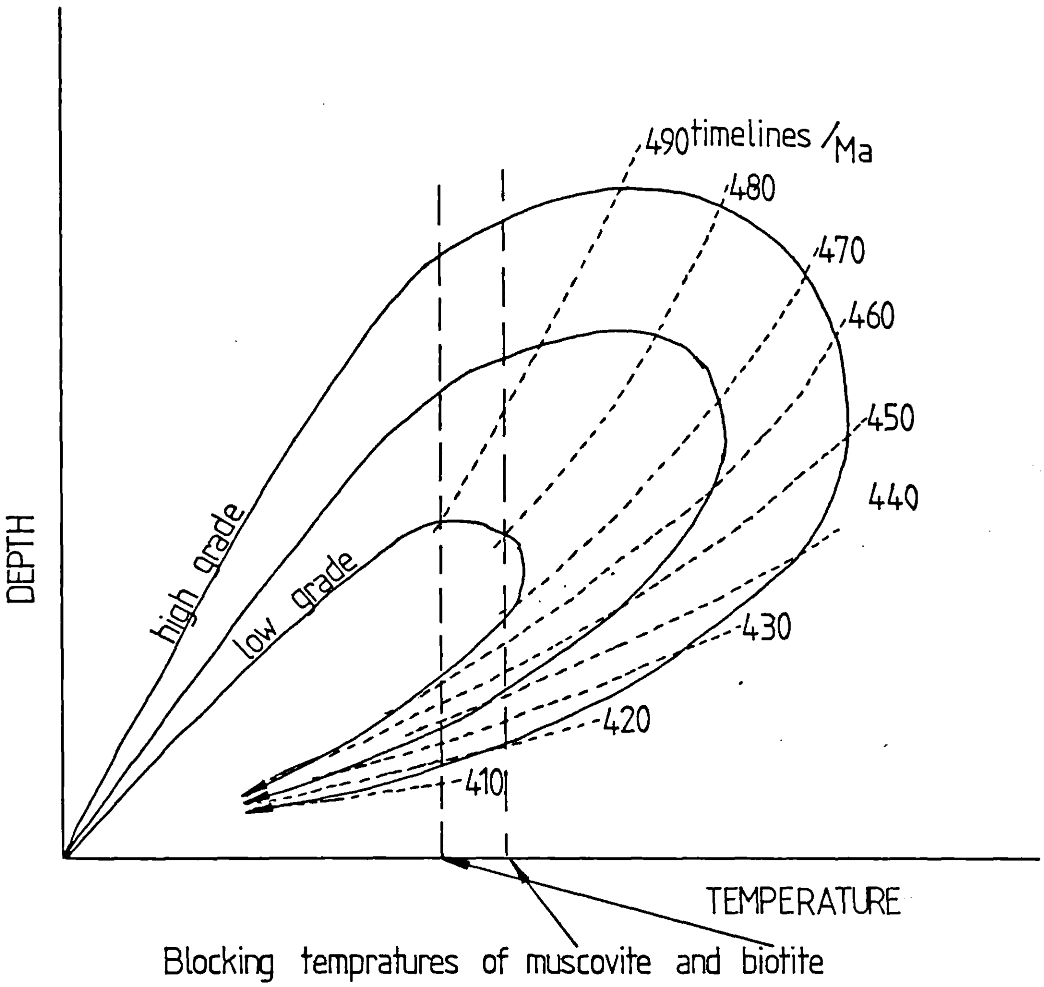
c) It is therefore concluded that the Rb-Sr ages for biotite are genuinely younger than the K-Ar ages. This may have been caused by partial resetting of the K-Ar system and total resetting of the Rb-Sr system or simply closure of the K-Ar system before the Rb-Sr system. The arguments concerning the resetting of the biotites with grainsizes 125-200μ are stated in detail in Section GC.4. It is thought unlikely that they represent partially reset systems because: (i) they are reproducible and therefore homogeneous, if they were partially reset the spread of ages would have been greater; (ii) they exhibit the 'normal' muscovite/biotite relationship; (iii) their relationship to less reproducible analyses from coarser grainsizes indicates that they are not partially reset. (The coarser samples extracted from the same rock give older ages with a wider spreading, indicating an inhomogeneous distribution of argon.)

Thus, although in many cases the Rb-Sr and K-Ar systems give approximately equal ages, in this case the Rb-Sr system seems to have been more easily disturbed and become blocked later than the K-Ar system.

Discussion

The implication from K-Ar and Rb-Sr data on the micas is that the whole area from the Moine Thrust belt toward the centre of the mobile zone was uplifted and cooled virtually simultaneously. The variation of both muscovite and biotite ages along the traverse is within the range expected from purely analytical error. This does not fit the model postulated by many authors for the margin of an orogen (Harper, 1967; Dewey & Pankhurst, 1970; Purdy & Jäger, 1976). They predicted that the outer margins of metamorphic belts should yield the oldest ages since

FIGURE 103 Relationship between cooling age and peak metamorphic grade

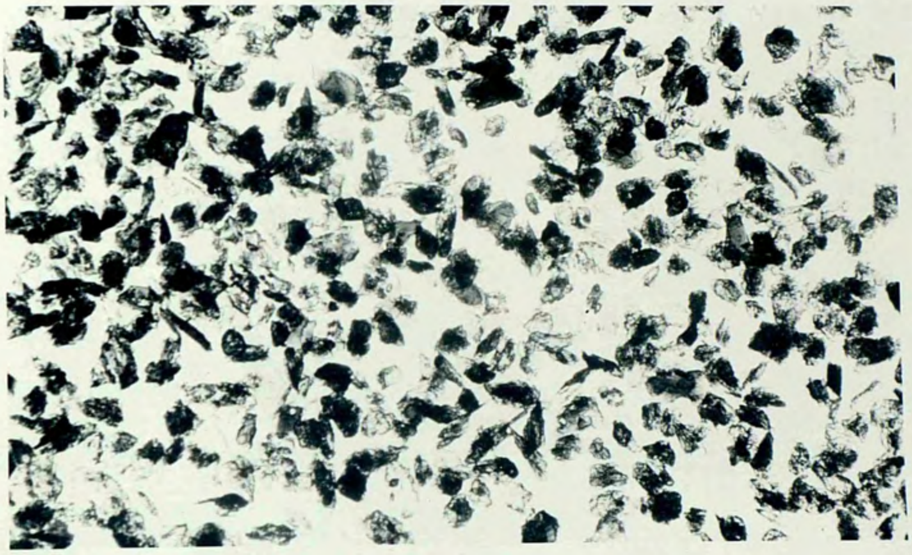


Rock buried to a greater depth and metamorphosed to higher grade yield younger cooling ages than those of lower grade they were coolest and least deeply buried during peak metamorphism. They were therefore uplifted and cooled through the blocking temperatures for argon migration before the internal or central areas (figs. 103,114). In the present survey, although the oldest muscovite ages are found close to the Moine Thrust, the difference between the oldest muscovites against the thrust and those of the internal zone (fig.101) is less than the limits of error imposed by analysis. However, the muscovite ages are consistently less than 430 Ma further than 9km from the thrust and consistently greater than 430 Ma within 9km of the thrust. The variation is equivocal, but if accepted, implies that the initial cooling through the muscovite blocking temperature took place close to the thrust zone at around 435-430 Ma.

It has been shown from structural and metamorphic considerations (see Moine Thrust chapter) that the Moine rocks were mylonitized and thrust over foreland rocks subsequent to the peak metamorphism and movement on the Sgurr Beag Slide. It is thought unlikely that either the K-Ar or Rb-Sr systems for muscovite and biotite could survive mylonitization without being reset at least to some extent. This implies that the systems became closed during or after the mylonitization event because the traverse includes samples which have undergone considerable grainsize reduction and mechanical damage to the micas (Section MT.5.b). The state of breakdown of the micas does not affect the age relative to those not affected by grainsize reduction, suggesting that cooling and thrusting were closely contemporaneous.

FIGURE 104. A slide showing the intergrown nature of the micas in SKRD 74. It proved impossible to separate the muscovite and biotite completely in this sample.

The grains range from 75μ to 125μ.



GC.3.b Excess argon at the Moine Thrust zone

Samples SK31, 74 and 108, lying closest to the Moine Thrust zone contain excessive amounts of radiogenic ^{40}Ar. They lie within a zone less than 3km from the thrust. SK108 is the best controlled sample, the muscovite gives a mean age of 434.23 Ma and the biotite 448.88 Ma. This is the reverse of the normal relationship for these two micas (Brewer, 1969; Moorbath, 1969; Purdy & Jäger, 1976, Roddick et al., 1980). It is taken to indicate an excess of radiogenic argon within biotite. An Rb-Sr mica pair age for this sample yielded 416 $\pm$ 4 Ma which is compatible with the Rb-Sr mica pairs in Fannich. This result implies that the "true" K-Ar age for biotite is similar to those in Fannich and the biotite contains excess ^{40}Ar. It is more difficult to determine whether the muscovite contains excess argon. Noteworthy, however, in this context is the consistency of ages from samples over 8km from the thrust where there is no indication of excess argon in the biotites. The muscovite from SK108 yields a K-Ar age 18 Ma older than the Rb-Sr mica pair age. In other samples further east, the muscovite K-Ar ages are also considerably older than the Rb-Sr mica pair ages, ranging from 5 Ma to 18 Ma. This implies that the separation between muscovite K-Ar and biotite Rb-Sr is within errors, the same as that seen further east and that the SK108 muscovite carries little or no excess ^{40}Ar.

The presence of excess argon in the biotite from sample SK31 is equivocal. The muscovite/biotite relationship is normal, biotite being approximately one million years younger than muscovite and therefore the same age within errors. However, the biotite age is 433.5 $\pm$ 10 Ma, an age which is considerably older than any biotite further than 3km from the thrust where the mean age is 421 $\pm$ 5 Ma. Using the 'critical value' test (Dalrymple & Lanphere, 1967) the difference in age appears to be significant.

Sample SK74 is approximately 1km from the thrust and was completely broken down during mylonitization (see Section MT.4.a). The grainsize of the micas is generally smaller than those in Fannich and therefore a smaller grainsize fraction was used (75-125μ) for analysis. (It proved impossible to separate single grains from the coarser grained fraction.) The same problem applied to SK64 which gives a mean age of 420 $\pm$ 7 Ma, implying that significant amounts of argon were not lost as a result of crushing to this grainsize and that the particular difference of grainsize did not affect the age (see Section GC.4). In sample SK74, great difficulty was experienced in separating muscovite due to the intergrown nature of the two micas (fig.104). Only an impure fraction of biotite was separable. A section was prepared of the separate and the proportion of muscovite was determined as 27%. As a whole the sample gives a mean age of 440 $\pm$ 10 Ma. That is significantly older than any other biotite age from the traverse further than 3km from the thrust. The biotites have a mean age of 421 $\pm$ 5 Ma (2 σ). Again using the 'critical value' test (Dalrymple & Lanphere, 1969) the ages are significantly different.

An estimate of the age which would be given by a pure biotite separate can be made by assuming the K-Ar muscovite age. Samples SK31, 34 and 108 indicate an age of around 435 Ma. In order to make a conservative estimate, an age of 440 Ma was assumed for the muscovite in SK74. The muscovites in this survey show a range between 7% and 9% (fig.105) so an upper and lower estimate of the biotite age was made using the end members of this range. The range of ages for the biotite part of the sample, corresponding to the range of assumed K content, is 439-460 Ma. However, the remaining 73% was also found to contain approximately 15% chlorite which was intergrown with the biotite and therefore the ages quoted are subject to considerable uncertainty since both argon and potassium could be lost in the biotite → chlorite transition.

Possible explanations for the occurrence of oldest ages adjacent to the thrust are:

a) Cooling close to the thrust zone due to thrusting over colder rocks at or before 450 Ma (Maximum biotite ages - Section GC.4).

b) Partial resetting of older systems during thrusting

c) Introduction of excess argon due to a high partial pressure of ^{40}Ar as the biotite became closed to argon diffusion.

Hypothesis (a) is unlikely since the Rb-Sr dates for biotite and K-Ar dates for muscovite are compatible with those further east (fig.101) indicating near simultaneous cooling between around 400°C and 300°C over the whole traverse (Section GC.3.a). Although to a much lesser extent, the K-Ar muscovite dates may indicate earlier cooling adjacent to the thrust. (Note however that the differences are within experimental error - see Section GC.3.a).

Hypothesis (b) seems extremely unlikely since it would imply that fine-grained biotites were more resistant to resetting than muscovites and both micas of larger grainsizes in Fannich (see Section GC.4). The area close to the Moine Thrust has undergone intense grainsize reduction and mechanical damage to the micas during mylonitization at a late stage in the tectonometamorphic history (Section MT.2.b and MT.5.a,b).

Hypothesis (c) is preferred although only one sample (SK108) demonstrates this clearly. Sample SK31 probably contains a smaller amount of excess argon and no muscovite sample exists to confirm the presence of excess argon in SK74. It is preferred over hypothesis (b) because it best explains ages in the fine grainsizes and over (a), because the muscovite ages and Rb-Sr mica pair ages show little or no increase into the thrust zone and demonstrate continuity with those further east. They probably reflect the true uplift dates.

It is thought that excess argon in biotites was introduced during the time that they were "open" to argon diffusion because if argon could diffuse into the biotite lattice after it had "closed" it would not be retained.

From the continuity of Rb-Sr mica pair ages and muscovite K-Ar ages close to the Moine Thrust and further east, it can be implied that the high partial pressure of ^{40}Ar was developed in spatial association with the thrust zone. At the same time, rocks in Fannich cooled through the blocking point of biotite without a high partial pressure of ^{40}Ar. It seems possible that the thrust zone acted as a conduit for ^{40}Ar escaping from deeper levels during reworking of older material (Lewisian gneisses are the best candidate) during mylonitization and thrusting, as has been suggested for some Australian gneisses (Allen & Stubbs, 1982).

The Rb-Sr mica pair and K-Ar muscovite ages show continuity between areas adjacent to the Moine Thrust and those in Fannich. It seems likely that the biotite K-Ar system also cooled virtually simultaneously in both areas. Since excess argon was introduced into the biotites only in the west, adjacent to the Moine Thrust, it is concluded that the thrust zone was active during late cooling.

GC.4 Grainsize Study in Fannich

GC.4.a Initial grainsize (200-250μ)

The initial K-Ar study was undertaken in a 5km traverse along the north shore of Loch Fannich, involving pelites and semi-pelites of the Morar Division and migmatitic pelites of the Glenfinnan Division (fig.94).

Sixteen muscovite/biotite pairs with grainsizes 200-250μ were analysed, the results are presented in Appendix 1b and plotted in Figure 106. Separate argon analyses are plotted using the mean potassium content for each mineral, since all potassium analyses were reproducible to within 1%, whereas argon analyses showed a greater spread. Calculation of analytical errors indicates that 95% (2 σ) of all the analyses should be reproducible to within 3% for a single sample (see Section GC.1.d). As illustrated (fig.106) many of the micas, both muscovite and biotite, exhibit ranges much greater than 3%. Not all analyses were duplicated but of those which were: 36,47 muscovite and 66,49,47,36,14 biotite showed ranges greater than 3%. Other samples did exhibit close reproducibility, even when repeated in triplicate: 14,43,54,207 muscovite and 3,23,25,43,54 biotite. The rest of the samples were not repeated; when the pattern of bad reproducibility began to emerge, a further study of different grainsizes was initiated and the original study terminated.

In cases where co-existing muscovite and biotite have duplicate argon analyses, there is very good correspondence between the degree of reproducibility, i.e. when muscovite is reproducible, so is biotite.

This phenomenon may result from several causes:

1) Over-zealous crushing of some samples causing argon loss due to mechanical damage.

2) Non-uniform series errors introduced during argon analyses such as bad peak measurements or incorrect weighing.

3) Inhomogeneous distribution of argon or potassium in some samples.

These hypotheses can be considered further:

1) It is considered unlikely that over-zealous crushing caused bad reproducibility since it would have been expected that the analyses

giving younger ages and therefore those which had suffered most damage would give unreproducible results, spreading to lower ages. However, the youngest ages tend to be more reproducible and give an age plateau 415-425 Ma for biotite, 415-430 for muscovite. The spread of less common ages is towards the oldest ages.

An experiment performed later on a single muscovite sample illustrates that all the grainsizes being used for this survey are sufficiently large that crushing did not cause significant argon loss (see later in this Section).

2) Series errors produced by bad analytical techniques would be detected because an internal standard (Mo 40 Biotite) was regularly analysed and this was found to be reproducible. Errors in the determination of the radiogenic argon content of the standard were less than 1% (1σ). The errors in determination of potassium content were less than 1% (1σ) for the standard too (Sections GC.1.b,1.c). Potassium contents of all the samples were reproducible to within 1% (1σ). However, the radiogenic argon content of many of the samples did not repeat to within the 1% (1σ) level of the standard. This seems to indicate an inhomogeneous distribution of argon. The amount of atmospheric argon present in the samples varied from 1-12%, it correlated closely with the pressure in the vacuum system, i.e. all the samples on some days gave low atmospheric contents and the same samples another day gave higher atmospheric contents. However the variations in atmospheric contents were too small to cause the differences in radiogenic argon contents, even if they had been totally due to bad measurement (fig.98).

3) It seems therefore that there is an inhomogeneous distribution of argon, attributable to some geological cause, which has not been investigated further. The effect may have been purely due to the grainsize,

in other words unreproducible analyses were obtained because too few grains were analysed each time to remove the variation between grains. Alternatively, it might have been a function of excess argon introduced into the micas or partial loss of argon from the micas.

GC.4.b <u>Coarse and Fine grainsizes</u> (125-200μ and 250-500μ).

In order to test the possible alternatives, a finer grainsize (125-200μ) and a coarser grainsize (250-500μ) were separated from a selection of the already crushed and sieved samples. No further crushing was undertaken at this stage because the intention was to try to obtain several separate grainsizes from the original rock rather than simply crush the coarser mica grains through a finer mesh. The validity of this method might be thought questionable since the coarser-grained micas within the rock are bound to be partially reduced during the crushing process. However, it must be noted that during the crushing and sieving, the coarser fractions always contained a higher proportion of micas. This is taken to indicate resistance of muscovite and biotite to the crushing process. A measure of the "crushability" of micas has been represented in the experimental law "Bond's Law".

$$W = 10\ Wi\left(\frac{1}{\sqrt{P}} + \frac{1}{\sqrt{F}}\right)$$

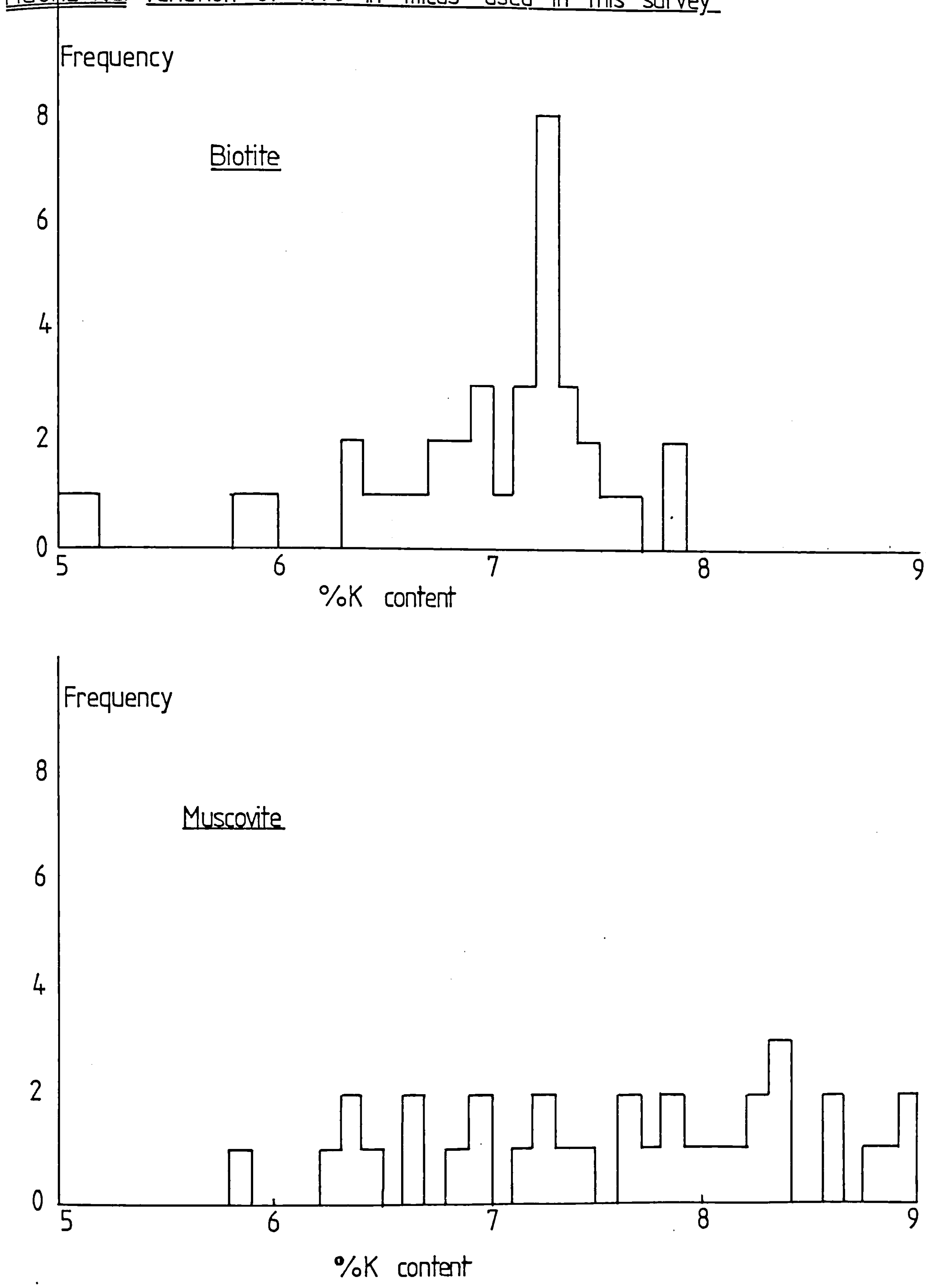
W = work done to crush from feed to product size
Wi = work index
P = Product size P and F values are standardised as 80% passing size
F = Feed size

The work index Wi gives an indication of how difficult it is to reduce the size of a particular material. Quartz, feldspars and garnet all fall into the range Wi = 10 to 15. This is the general range for the great majority of rock forming minerals. However, micas have work indices of around 130, they are therefore of an order of magnitude more difficult to crush. This is because the methods used to crush rocks are generally ineffective on the flaky, flexible mica grains. It is therefore inferred that the coarsest fractions from which micas can be extracted will contain micas which were originally the coarsest micas in the rock. The finest fractions from which micas were extracted will contain micas which were originally small in the rock as well as parts of larger micas reduced during the crushing.

Seven samples were selected for analysis of the fine grainsize (125-200μ) (see Appendix 1A). All 7 muscovite samples analysed were reproducible to within the expected analytical errors except SK54 (which gave a spread of ages from 417 Ma to 428 Ma and will be discussed in detail later in this section). They are plotted on Figure 107 in the same way as previously, using the mean potassium contents to illustrate the variation of ^{40}Ar content. The mean value of the ages from the fine fractions of samples 14,19,36,49,54,72 and 207, is 423.53 Ma. The standard deviation on the mean is 6.4 Ma, compared with ca. 5 Ma expected from analytical errors. This indicates that the 7 muscovite samples are near to having a uniform age, forming a single population which became closed to argon diffusion at 423.5 Ma.

The 7 biotite samples, co-existing with muscovite were reproducible in an even more closely defined population. All samples were reproducible to within the analytical errors in contrast to the coarser grainsizes. The mean of the population is 421.56 Ma with a standard deviation of 3.25 Ma. This indicates that all 7 biotite samples closed to ^{40}Ar diffusion virtually simultaneously but that there may be some geological scatter in the muscovites.

FIGURE 105. Variation of K% in micas used in this survey



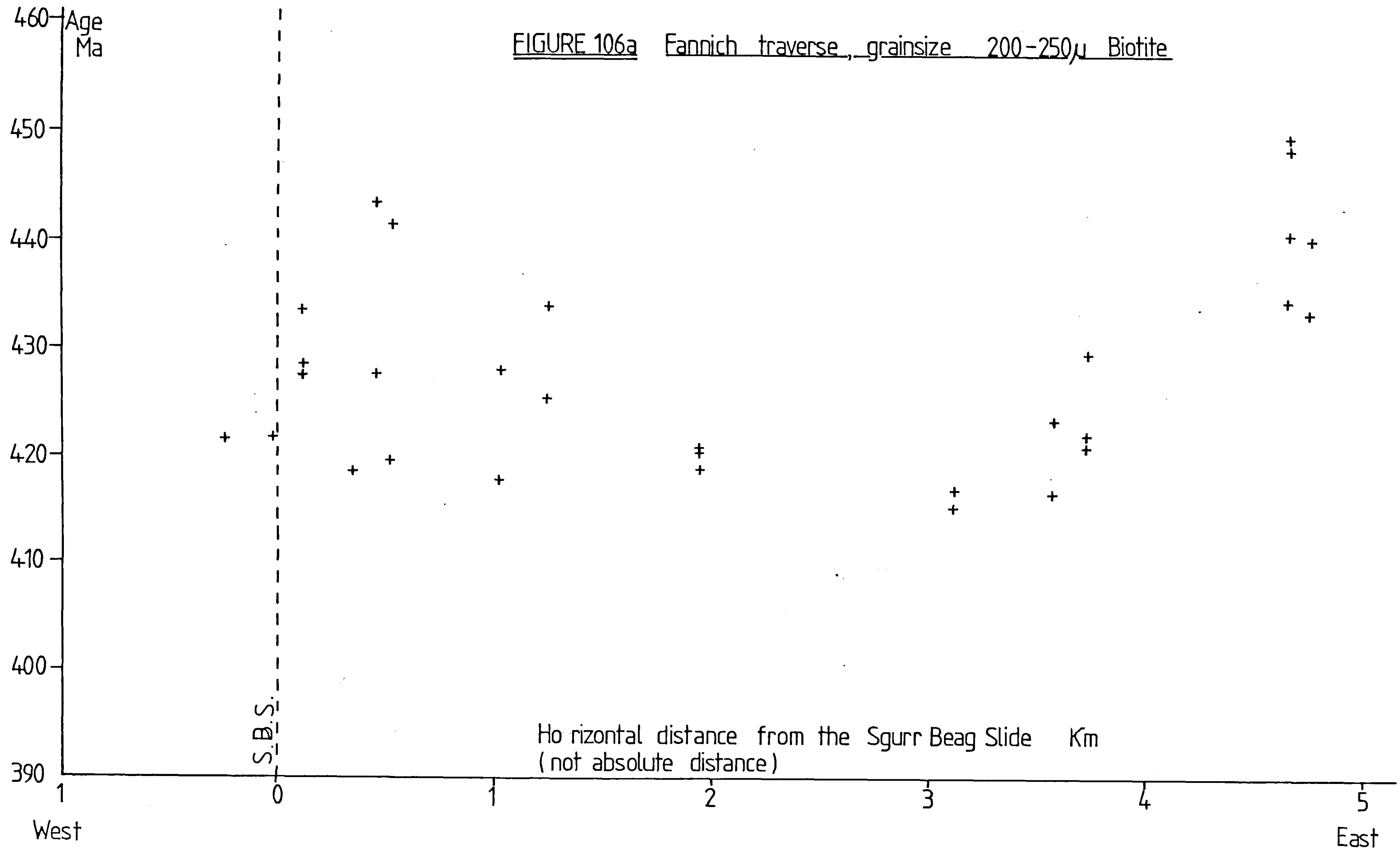


FIGURE 106a Fannich traverse, grainsize 200-250μ Biotite
Age
Ma
460
450
440
430
420
410
400
390
S.B.S.
Ho rizontal distance from the Sgurr Beag Slide Km
(not absolute distance)
1
0
1
2
3
4
5
West
East

FIGURE 106b Fannich traverse, grainsize 200-250μ Muscovite

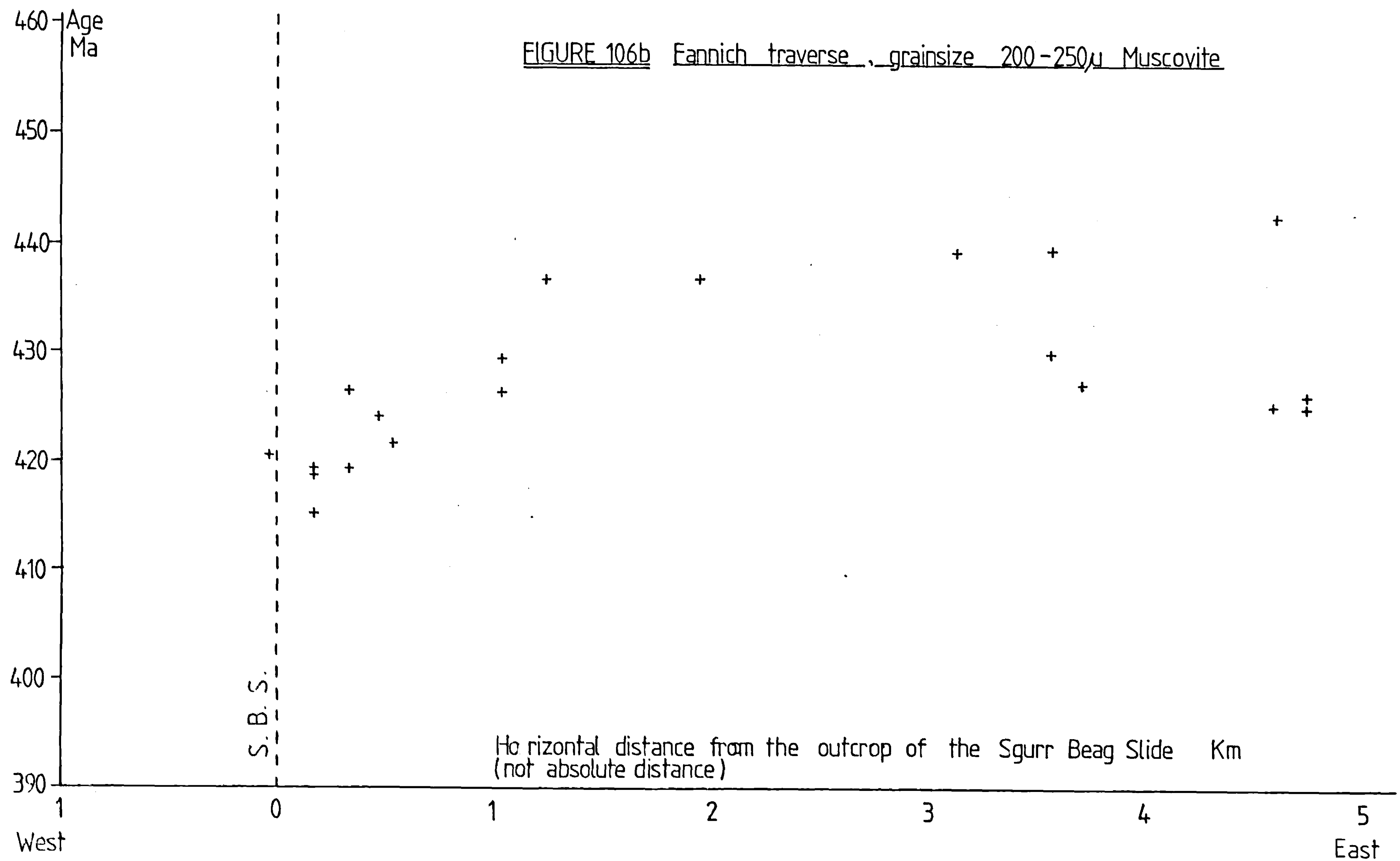
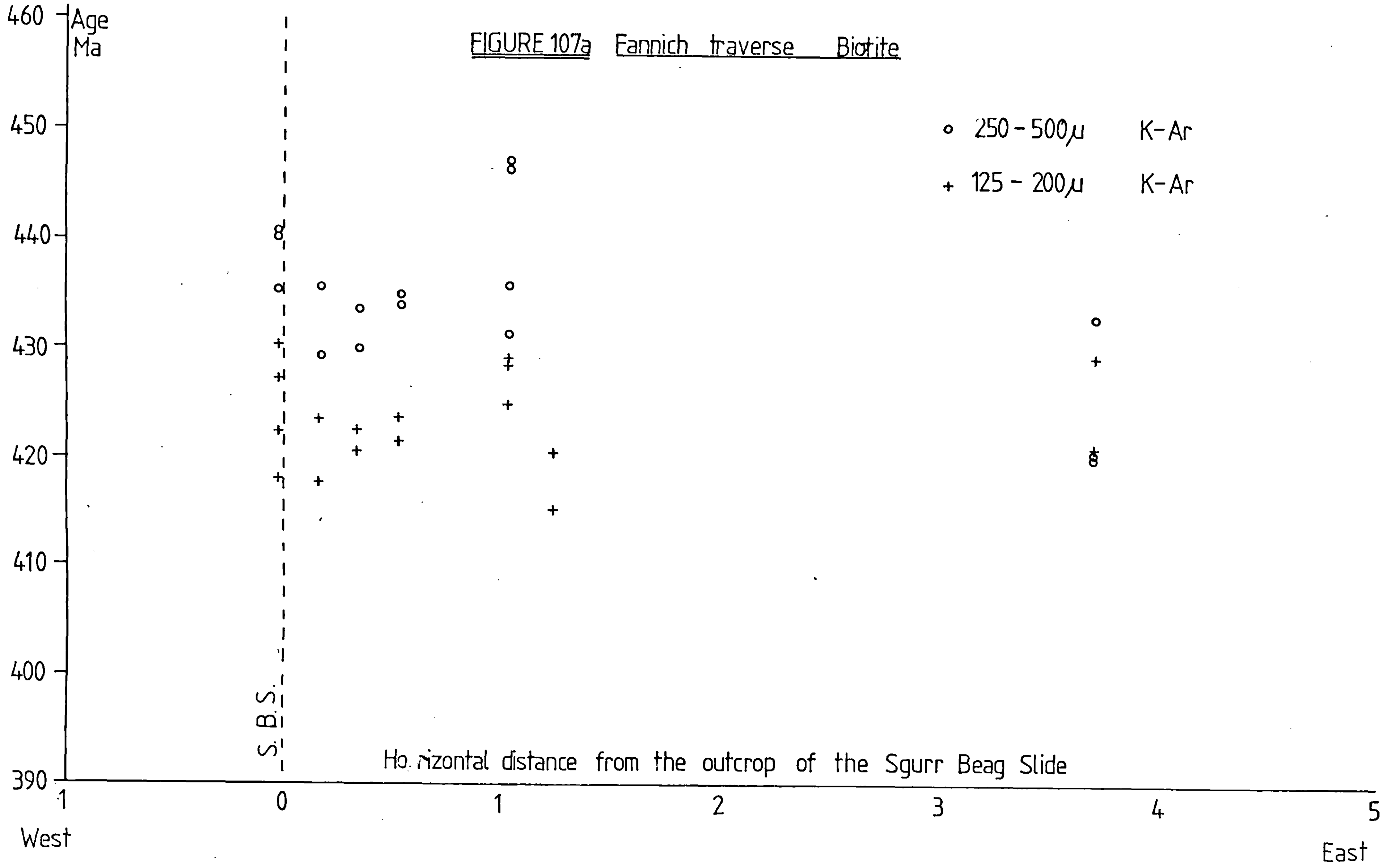


FIGURE 107a Fannich traverse Biotite



The reproducibility of the fine-grained samples confirms the accuracy and precision of the K-Ar method deduced from repetition of the internal standard Mo^{40} biotite.

Six muscovite/biotite pairs were separated from the coarse grainsize fraction (250-500μ) (see Appendix 1c). When analysed: two samples (49,207) were reproducible to within analytical errors; two samples (36,54) contained one mica which was reproducible and one which was unreproducible; in two samples (14,72) neither mica was reproducible (fig.107). Since the spread of grainsizes in this fraction is larger than the previous fractions, the spread caused by sample inhomogeneities might be expected to be larger. However, all the samples were reproducible with respect to potassium content and since the radiogenic ^{40}Ar is derived from ^{40}K, the same should apply to ^{40}Ar. The ages range from 417 Ma to 447 Ma for biotite, 416 Ma to 454 Ma for muscovite. The mean age of biotites is 433.4 ± 7.6 Ma (1σ) and for muscovites is 433.7 ± 8.7 Ma. The analyses exhibit a larger spread than that predicted from the calculation of errors in analysis and larger than the equivalent fine grainsize micas (see earlier in this Section). Thus the scatter in the coarse grainsizes is of geological origin.

It might be suggested that some of the scatter between localities is due to proximity to a major movement zone, the Sgurr Beag Slide. The slide is known to have disrupted contemporary isotherms and was a major influence on the cooling of rocks in its proximity (Powell et al.,1981).

The profile of coarse biotite samples (fig.107) shows no evidence of a relationship between the spatial distance from the slide but muscovites do exhibit a possible correlation. Approaching the slide from the Morar division, the muscovites get older, SK54 exhibiting the oldest mean age. Across the slide in the Glenfinnan Division, the age drops to the lowest value seen in either division. It is difficult to determine whether the differences are significant since they are within the analytical errors (2) of the K-Ar determinations. Hence no absolute conclusion can be drawn on the spatial relationship between the Sgurr Beag Slide and K-Ar ages for the coarse samples.

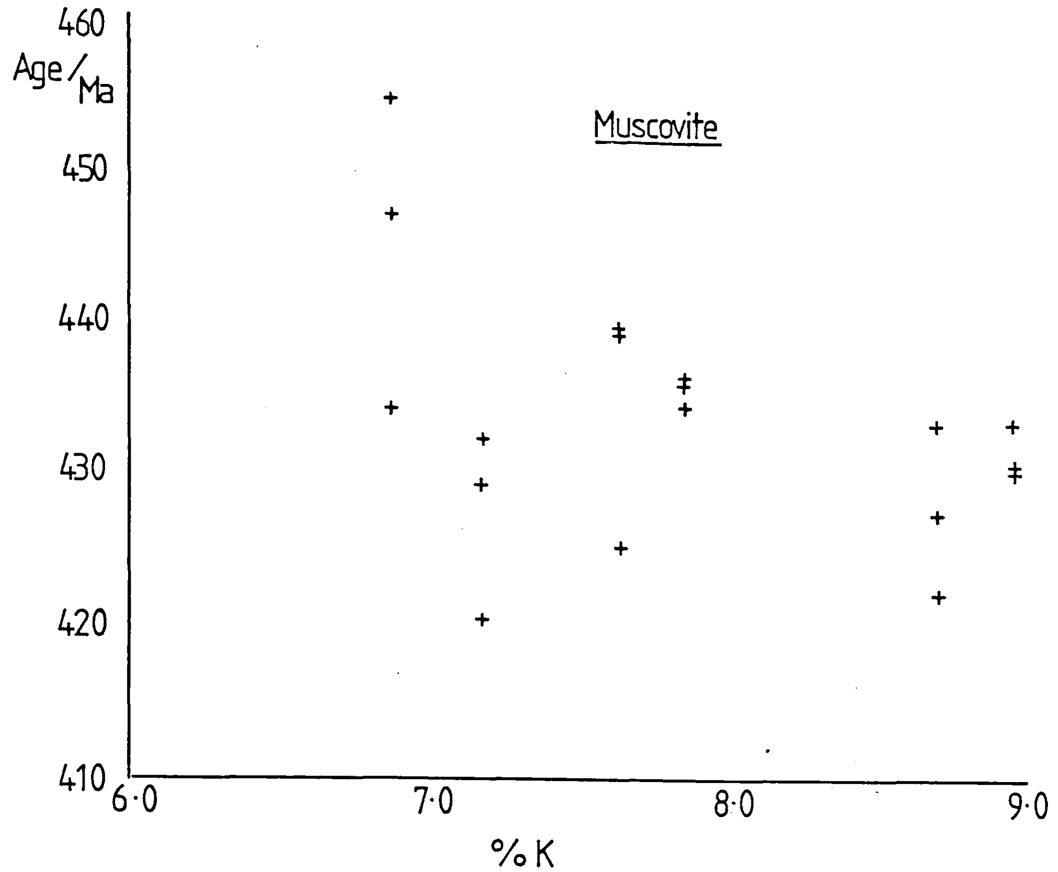
Two Rb-Sr mica pair ages for SK72 and SK54 were obtained using coarse 250-500μ grain sizes. They yielded ages of 410.38 ± 4.23 Ma and 406.66 ± 4.21 Ma. These are within errors the same as Rb-Sr mica pair ages (essentially Rb-Sr biotite ages) are not dependent upon grainsizes as are the K-Ar ages.

The Rb-Sr dates obtained from the coarse samples are over 30 Ma younger than the K-Ar dates for the same samples (see Appendix 1C and Appendix 2). This is not the normal case in metamorphic terrains where K-Ar and Rb-Sr ages for biotite are usually roughly equivalent (Purdy & Jäger,1976; Roddick et al.,1980). If no other data existed, these biotites would probably be interpreted as containing excess radiogenic argon. However, evidence from other grainsizes indicates the possibility that older K-Ar ages seen in the coarse samples are due to greater retention of argon than strontium during a later resetting event. Biotites have been reported retaining their K-Ar ages after subsequent metamorphism to 400°C, well above the blocking temperature generally accepted for either K-Ar or Rb-Sr biotite ages (Verschure et al., 1980; Chopin & Maluski,1980; Del Moro et al., 1983).

GC.4.c Discussion of the relationship between grainsize and age in Fannich micas

The analysis of coarse and fine grainsizes co-existing in the same rocks from Fannich shows a notable variation. The difference between mean ages for the coarse and fine biotite fractions is 11.8 Ma; between

FIGURE 108 Plots of age against % K for the 250-500μ grainsize



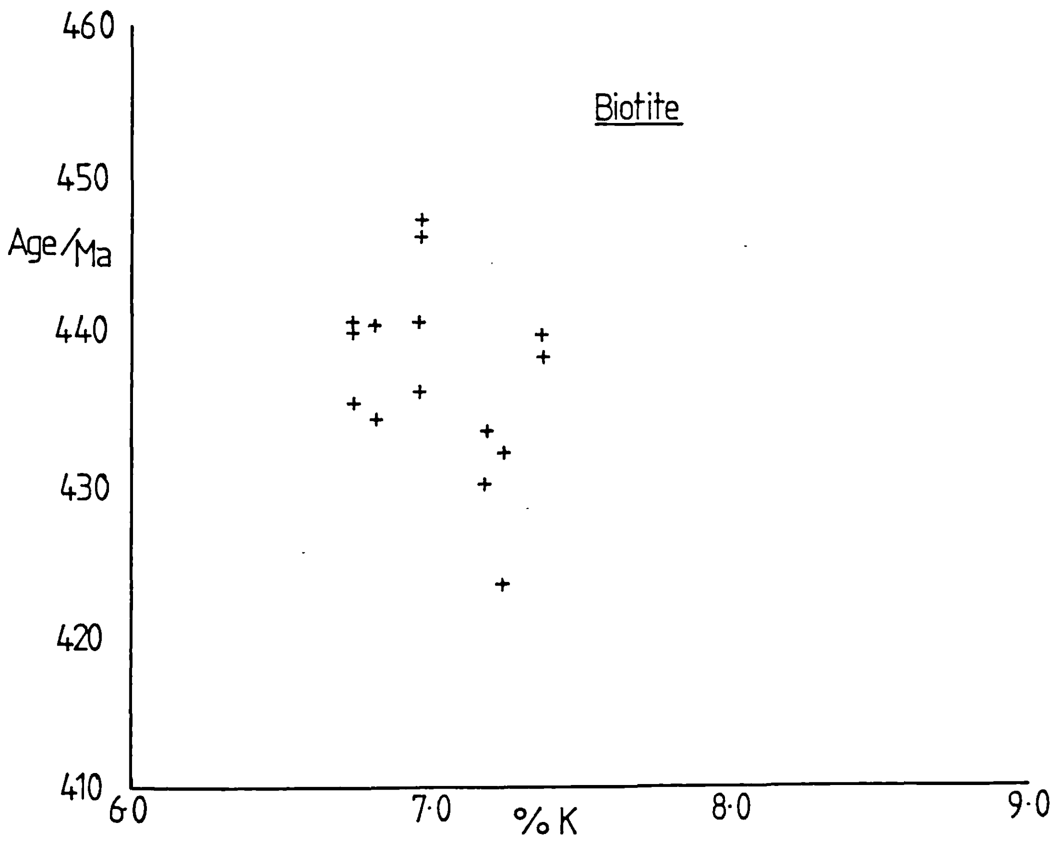
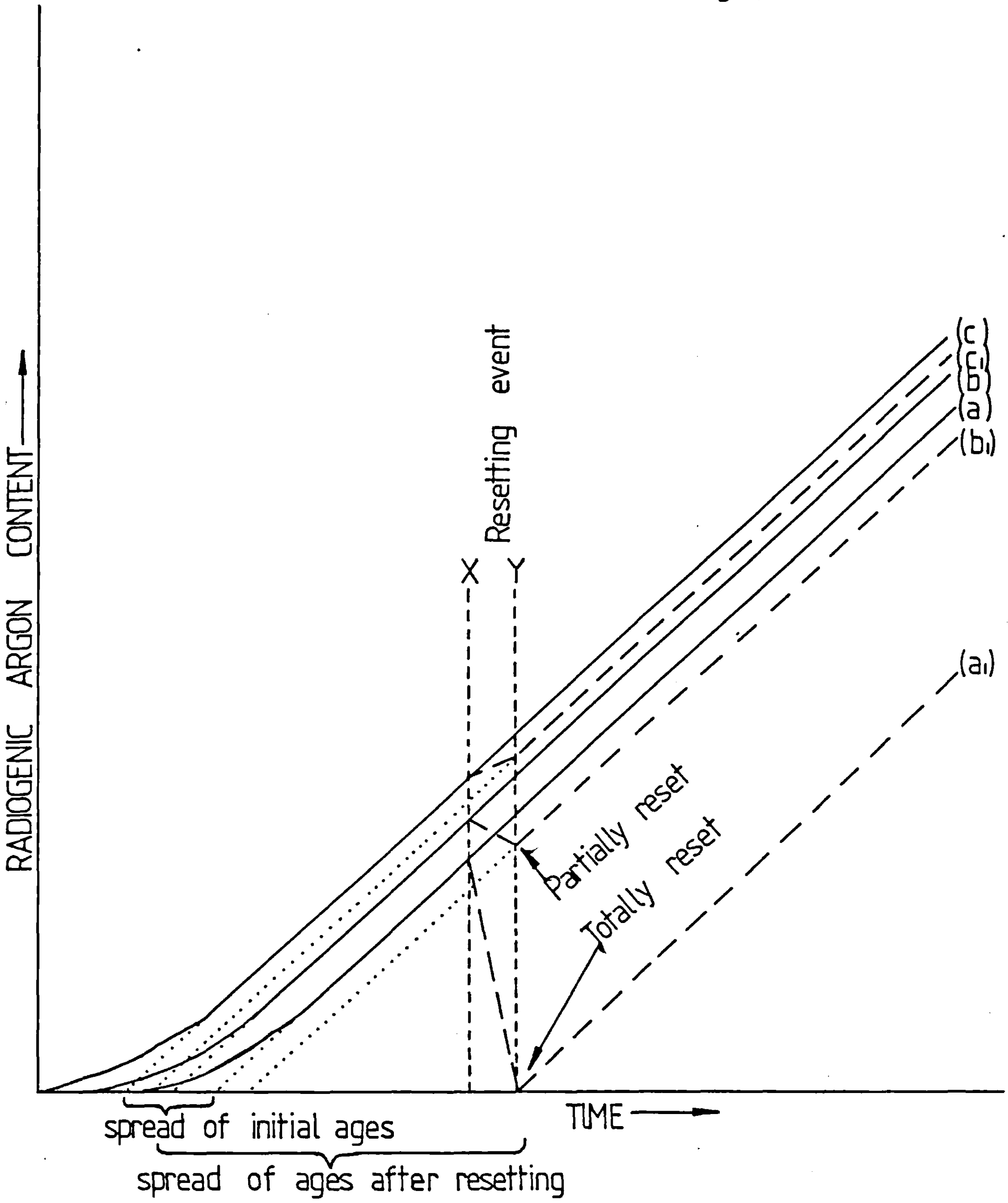


FIGURE 109.

Diagramatic representation of resetting of K-Ar ages



The difference in radiogenic argon content between grainsizes a, b and c is less than the analytical error on the determination of their argon content. During reseting event XY, the differences may be enhanced due to diffusion effects. The spread of ages is greater than analytical errors and can therefore be detected.

the mean ages for coarse and fine muscovite fractions it is 10.2 Ma. Such a difference due to grainsize variation is not recorded in the literature although it is covered in the theory of blocking temperatures (Dodson, 1973,1979).

Two questions with regard to the differences are:

a) Is the difference statistically significant?

b) If the difference is significant what are the possible interpretations?

Dalrymple and Lanphere (1967) defined a 'critical value' test. When the value is exceeded, it indicates a 95% probability that the samples are significantly different.

$$\text{Critical value} = 1.96 \sqrt{\frac{(\sigma_1)^2}{n_1} + \frac{(\sigma_2)^2}{n_2}}$$

σ = standard deviations of the ages obtained.

n = number of repetitions of the analyses.

When applied to the coarse and fine samples of Fannich, half the samples show significant differences between coarse and fine fractions (muscovite 54,207; biotite 14,49,54,207) and half are not significantly different (muscovite 14,36,49,72; biotite 36,72), thus no definite answer was obtained. If the coarse and fine fractions are treated simply as two statistical populations with normal distribution, they can be tested for significant difference using the 't' test. Applying this test to the coarse and fine fractions for both muscovite and biotite, it is found that in both cases, the probability that coarse and fine fractions are distinct is greater than 99.9%.

Several explanations are possible for the separation between the ages of coarse and fine samples:

a) Series errors introduced between batches of analyses (most of the fine fractions were analysed approximately 3 months before the coarse fractions).

b) The fine fractions produced younger ages due to argon loss during crushing.

c) Geological characteristics inherited while the micas were still in the rocks.

Mechanism (a) is considered unlikely because at all stages, the potassium and argon internal standard (Mo40 biotite) was analysed alongside the samples. In both potassium and argon analyses throughout all the different grainsize analyses, the standard was reproducible to within 1%. This indicates that any series errors introduced into the results were at a level below 1% and therefore cannot explain the discrepancy between coarse and fine fractions which are greater than 2.5%.

In order to test hypothesis (b), a sample of SKND49 muscovite 250-500μ grainsize was crushed to a grainsize of 75-250μ. The fine-grained sample and coarse-grained sample were analysed for radiogenic ^{40}Ar content. Analyses were done in triplicate, giving a mean and standard deviation for the coarse-grained sample of $159.7 \pm 2.2(1\sigma) \times 10^{-9} lg^{-1}$ and for the fine-grained sample of $162.3 \pm 2.0(1\sigma) \times 10^{-9} lg^{-1}$. The two ^{40}Ar contents were indistinguishable and it is therefore concluded that the age separation is not a product of the crushing process.

Thus the separation is an original geological feature. It has normally been accepted that in regional metamorphic terrains, the grainsize had very little effect upon the age obtained. Purdy and Jäger (1976) stated that the K-Ar ages from the staurolite and higher grades of metamorphism in Alpine terrains were independent of grainsize, pegmatitic books several cms in diameter giving the same age as fine-grained micas extracted from surrounding metasediments. However, it has been noted that in contact metamorphic zones (Hart,1964; Hanson & Gast,1967) grain-size of the mineral analysed does affect the age obtained, the coarser grains being less susceptible to overprinting during igneous heating. The difference between the two situations has been attributed to the short times and relatively high temperatures involved in contact metamorphism. However, no evidence has been found in Fannich for late igneous activity or any thermal metamorphic effects.

The disparity in ages between grainsizes seems therefore to be related to the regional metamorphism. Two possible causes may be postulated:

a) Excess radiogenic argon is present preferentially in the coarse grainsizes.

b) Extraneous radiogenic argon is present preferentially in the coarse grainsizes, i.e. the finer grainsizes have suffered resetting.

The concept of excess radiogenic argon in the coarser grainsizes (a) is superficially supported by the data from Rb-Sr mica pairs (fig.101). For all grainsizes these yield ages younger than but compatible with K-Ar ages obtained from the finest grainsizes. This implies that all grainsizes closed at the same young age and that the ages of the apparently older, coarse grainsizes reflect the presence of a high partial pressure of ^{40}Ar. In an attempt to test this theory, potassium content was plotted against age for the coarse grainsize (fig.108). There should be a change from high to low values of potassium assuming that all the micas absorbed an approximately equal amount of excess argon. The relative change in age from high to low potassium values can be calculated by assuming an approximately constant volume of excess argon for all potassium contents and a true original age equal to that of the fine fractions. The difference between micas with potassium contents between 6% and 9% would be 12 Ma. Extrapolating the plots for biotite and muscovite, both show increases apparently far greater than that predicted. However in both plots, the variability between individual argon determinations of single samples is greater than the variation between samples and this approach is therefore inconclusive. The plot of potassium against age is useful for young samples, e.g. the Alpine micas, but in samples of Caledonian age, small amounts of excess argon are swamped by radiogenic argon produced during the decay of ^{40}K. It is therefore not proved possible to determine whether there is a significant correlation between apparent age and potassium values of the micas.

The interpretation of excess argon relying purely on the Rb-Sr mica pair ages implied earlier in this Section is also problematical since no study has been found in the literature dealing with a similar problem of grainsize/age correlation.

The only studies relating K-Ar and Rb-Sr studies to grainsize are those in contact metamorphic zones (Hart,1964; Hanson & Gast,1967). These studies indicate that Rb-Sr biotite dates are approximately equal to K-Ar biotite dates on samples of similar grainsize but do not discuss differences due to grainsize variation.

The results of the present study imply that in this area strontium was more mobile than argon in biotite (Section GC.3.a) and therefore the Rb-Sr ages are younger. There is a possibility that even the finest grainsizes contain excess argon but this is unlikely since they are reproducible and exhibit the 'normal muscovite/biotite relationships.

Another factor arguing against the presence of excess argon is that it is difficult to envisage a situation in which excess radiogenic argon caused by a high partial pressure in the rock could be preferentially trapped in the larger grainsizes. Giletti (1974) showed that argon diffuses preferentially parallel to mica cleavages. The argon might be expected to have completely diffused into the smaller grainsizes during an excess argon 'event' before the larger grains had equilibrated. This would cause the finer grainsizes to have apparently older ages but this is not the case in Fannich.

It is possible to produce the effect seen in Fannich if the system, initially under high ^{40}Ar partial pressures, was suddenly released. The excess argon would diffuse out of the finer grains before coarser grains. If the system were then "frozen" at this point, both muscovite and biotite becoming closed, the coarser grainsize would yield older ages due to their greater retentivity. However, since no excess argon atmosphere has been proven in Fannich, this hypothesis remains purely conjecture.

The second hypothesis (b), that different ages are due to extraneous argon in the coarser samples, is also favoured because situations involving excess argon generally yield biotite ages considerably older than muscovite (Brewer,1967; Roddick et al.,1980).

Diffusion theory predicts that finer grainsizes will give ages younger than co-existing coarse grains (Dodson,1973). However, the period during which blocking takes place is very short in comparison to the total age and it is not normally possible to resolve the age difference between grainsizes (fig.109). Two mechanisms may explain the differences seen in Fannich:

a) Slow cooling; i.e. the period during which blocking takes place is not insignificant with respect to the time since blocking.

b) Resetting; i.e. the grains all become closed to argon diffusion and a later event causes greater loss from the fine grainsizes than the coarse grainsizes.

Coincidentally, a controversy long discussed with respect to K-Ar dates from Moine rocks is the problem of distinguishing slow cooling from resetting. The 'slow cooling' hypothesis being advanced by Harper (1967) and Dewey and Pankhurst (1970); the 'overprinting' hypothesis being preferred by Miller and Brown (1965), Brown et al. (1965) and Fitch et al. (1964,1970). This early work will be discussed more fully in Section GC.6.c.

From the present work, it is very difficult to distinguish between hypothesis (a) and hypothesis (b). However, the evidence tends to indicate that (b) is the more probable explanation. If the micas had cooled slowly, it is probable that the fine-grained samples would produce the same variation in ages as coarse-grained samples. This was not the case in Fannich. In an overprinting situation, partially outgassed large micas would exhibit a large range of ages whereas totally outgassed fine-grained micas would have a narrow, younger age range.

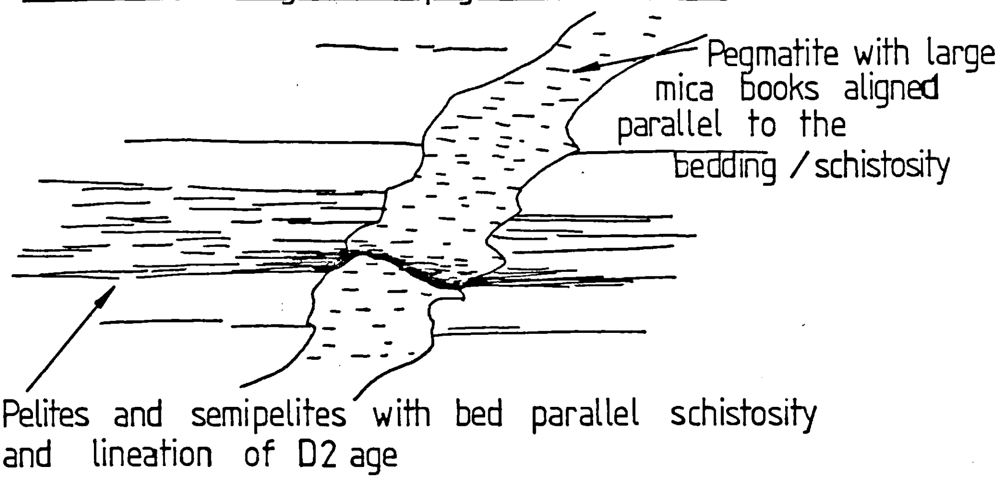
The relationship between muscovite and co-existing biotite ages provides further evidence in favour of resetting. If the pattern of ages in Fannich were simply a product of slow cooling the muscovites would have been expected to cool through their blocking temperature before the biotites and the difference be significant. However, both coarse-grained micas yield ages around 433 Ma and both fine-grained micas have ages around 422 Ma. Thus the grainsize is relatively more important than the mica type in this area. This has not usually been found to be the case in other areas (Purdy & Jäger,1976; Hanson & Gast,1967).

GC.4.d Conclusions

The variation of K-Ar ages, obtained from the Fannich area, with grainsize can be explained in two ways. It has not been possible, on present evidence, to rule out the possibility that the coarse grainsizes may be partially retaining an earlier excess ^{40}Ar atmosphere, which was completely outgassed from the fine grainsizes. The more likely hypothesis is that all grainsizes cooled through their blocking temperatures and the coarse grainsizes are partially reset (overprinted) during a later event which completely outgassed finer grainsizes. Whichever is true, the study establishes the existence of a late tectonometamorphic event in the Fannich area, resetting fine-grained fractions of both muscovite and biotite to give mean ages of 423.5 Ma and 421.5 Ma. They had already, at an early stage, become closed to argon, probably prior to 433 Ma. It seems unlikely from the present data, that the rocks cooled slowly from peak metamorphic temperatures between 460 Ma and 450 Ma to the final closure of muscovite and biotite at 423.5 Ma and 421.5 Ma respectively.

It is significant that Rb-Sr mica pairs (essentially Rb-Sr biotite ages) are consistently younger than the K-Ar ages; their relationship to the K-Ar dates is not wholly understood. Their validity as true ages may be challenged if the biotite has been reset. However, if accepted they indicate that, in the Fannich area, strontium has been more mobile in biotite than argon. This finding is in contrast to previously reported studies, where the mobility of strontium and argon in biotite has been more or less equated.

FIGURE 110 Sketch of transgressive pegmatite SKND 55



GC.5 Geochronology of the Early Pegmatites

Three ages of pegmatite intrusion can be structurally defined as:

Pre-D2

Post-D2/Pre-D3

Post-D3

Pegmatites from the second and third suites were not dated because they do not contain minerals suitable for K-Ar dating. Their mineralogy is restricted to quartz, plagioclase feldspar with only very rare mica flakes. The mica plates were not used since they may have been taken from the walls of the pegmatite during intrusion (see Section MT.2.c)

The early pegmatite suite has a mineralogy of plagioclase and quartz with large mica books up to 2cm in diameter. Although only three such pegmatites were discovered, their tectonometamorphic history is well defined. Two pegmatites lie within the Sgurr Mor pelite and one within the Inverbroom psammites. Samples were extracted from SKND42 and SKND35 from the Sgurr Mor pelite and Inverbroom psammite respectively.

SKND42 (G.R.248664) is a pegmatite approximately 10cm wide extending for at least 5m with generally parallel sided form. It occurs in the nose of an F3 fold and contains a strong planar fabric defined by mica alignment which is folded by F3. The mica alignment is parallel to the bedding/foliation planes of S2 in the adjacent pelites and pegmatite boundaries are sub-parallel to the pelite bedding.

SKND35 (G.R.229667) lies within the nose of an F2 isocline and carries a strong axial planar fabric. It consists of a lense of pegmatitic material 50cm long and 10cm wide at its widest point. It is highly deformed and attenuated. The mineralogy is extremely micaceous at this locality.

SKND55 (G.R.198668), the third pegmatite, which was not dated (fig.110) was intruded transgressing the bedding planes of a semi-pelite but contains a strong mica alignment parallel to the bedding/foliation.

Both SKND42 and SKND55 contain garnet which is thought to be of metamorphic origin although this cannot be proved purely on fabric evidence.

The pegmatites were intruded pre-D2 (Sgurr Beag Slide deformation) and were not deformed until D2 when they also suffered amphibolite facies metamorphism. They lie between 10 and 15km from the present outcrop of the Carn Chuinneag granite (intruded at 550 $\pm$ 10 Ma, Pidgeon & Johnson,1974) which may have extended further west above the present level of erosion (see Section SBS.5). It may be that the pegmatites represent apophyses of the granite. Another possibility is that they are equivalent to the Carn Gorm pegmatites within the Glenfinnan division rocks at Garve which give a mean age of 731 $\pm$ 44 Ma (Van Breemen et al.,1974) although ages as low as 662 Ma have been obtained (Long & Lambert,1963). There is also the possibility that these pegmatites are not connected with any other intrusions so far described in the area.

The results of Rb-Sr and K-Ar studies are presented in Appendix 1D. SKND42 gave an Rb-Sr mineral age of 457 $\pm$ 6 Ma using plagioclase to define the initial ratio of the rock since muscovite has too low an Rb/Sr ratio to be able to assume an initial ratio. The K-Ar date obtained from the same mica flake as used for the Rb-Sr determination was 443 $\pm$ 11 Ma.

It was not possible to extract plagioclase from sample SKND35 to define the initial ratio. Therefore the initial ratio of the other pegmatite from the same suite (SKND42) was used (0.714 $\pm$ 0.005). The age of 450 $\pm$ 3 Ma is, within errors, the same as that for SKND42, the larger errors are caused by the assumed uncertainty in the initial ratio because it was not taken from this pegmatite. The K-Ar age is 483 $\pm$ 13 Ma may indicate the presence of excess argon within the muscovite or be due to relict argon from an earlier closing event. A K-Ar age obtained from a finer-grained fraction (200-250μ) of SKND35 gave an age of 448 $\pm$ 11 Ma which is within errors the same as that for SKND42. The coarse fraction represents only a single determination and the old age may be either excess or relict radiogenic argon within the lattice. The separation of K-Ar ages between the grainsizes of SKND35 again illustrates the importance of grainsize in the area.

The Rb-Sr ages for both pegmatites are younger than ages which might be expected from either Carn Gorm age pegmatites or apophyses of the Carn Chuinneag granite. However, they do not correspond to the "Caledonian pegmatite suite" seen in the SW Moine (Van Breemen et al.,1974) which post-date Sgurr Beag Slide movement. It is more likely that the later suite of pre-D3/post-D2 pegmatites in Fannich and towards the Moine Thrust correspond to those in the SW Moine.

The muscovite books used for this survey were only approximately 1cm in diameter compared with books over 10cm diameter used by Van Breeman et al.(1974) from Carn Gorm. Muscovites used in the mineral isochron produced for the Carn Chuinneag granite (Long & Lambert,1963) were taken from the undeformed part of the intrusion. Those in Fannich are from within the Sgurr Beag Slide zone. It seems probable that the dates obtained from the pegmatites were reset during the Caledonian orogeny. The correspondence of the dates indicates that they were probably totally reset and the date of 457 $\pm$ 6 Ma for SKND42 represents a minimum date for peak Caledonian metamorphism in the area. This had cooled to muscovite K-Ar blocking temperatures by 448 $\pm$ 11 Ma earlier than the dates indicates by the finer-grained muscovites extracted from the surrounding metasediments. It is noteworthy that this relationship contrasts with that found in Alpine rocks of equivalent grades (Purdy & Jäger,1976).

GC.6 Dating within the Moine Thrust Zone

Three previous studies are of importance to the interpretation of the present data in terms of thrust movement:

a) The date of intrusion and structural interpretation of the Ross of Mull granite (Beckinsale & Obradovich, 1973).

b) The date of intrusion and structural interpretation of the Borralan Complex (Van Breemen et al., 1979).

c) The study of cooling patterns for the Caledonides as a whole (Dewey & Pankhurst, 1970).

The Ross of Mull granite gives a minimum estimate of the time of cessation of movement along the Moine Thrust plane. The Borralan Complex and the alkaline intrusive suite of Assynt overlap thrusting and impose important constraints on the age of early movements. The study of Dewey and Pankhurst (1970) used data from the literature to produce a model for cooling and early (Llandeilian) uplift along the Moine Thrust zone.

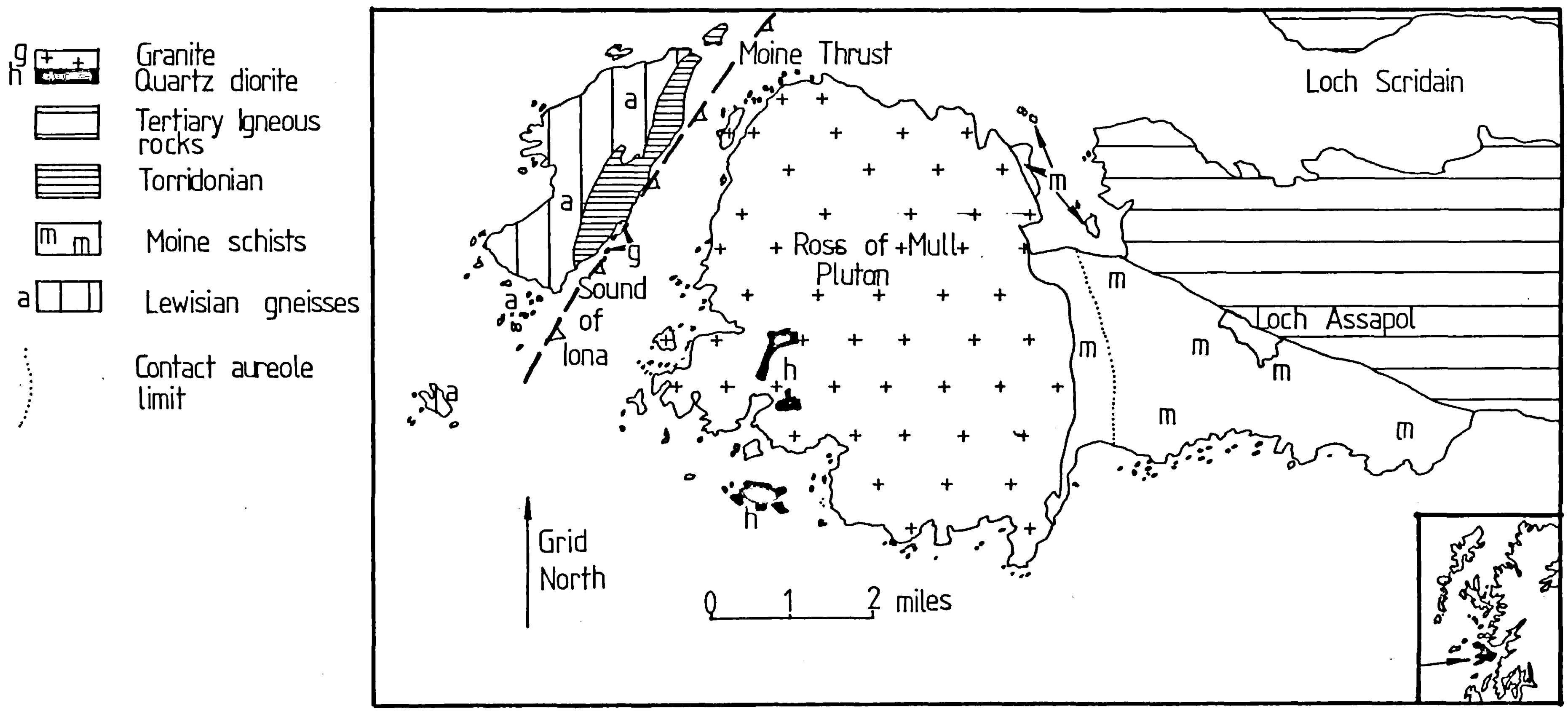
These studies are reported along with discussion of recent literature on the subjects and the implications of the new findings to earlier conclusions.

GC.6.a The Ross of Mull granite

It has long been thought that the Moine Thrust plane runs through the Sound of Iona between Iona and the Ross of Mull (fig.111; Clough, 1911; Cunningham Craig et al., 1911). The Sound separates Lewisian and Torridonian rocks of Iona from Moine schists of both Morar and Glenfinnan Divisions (Harris et al., in prep.) on the Ross of Mull.

The Ross of Mull granite and quartz diorites are intruded into Moine schists of the Morar Division and a thermal aureole approximately

FIGURE 111. The Ross of Mull Granite : after Beckinsale and Obradovich (1973)



500m wide is developed in the schists. At one point, a prominent band of kyanite bearing pelite enters the aureole, and kyanite is replaced by andalusite within the aureole (Mackenzie,1949). Within the granite, andalusite and sillimanite are found in pelitic xenoliths.

Jehu (1922) determined that Torridonian and Lewisian rocks of Iona show effects of thermal metamorphism due to the Ross of Mull granite. The granite also outcrops on several small islands close to the eastern shore of Iona (fig.111).

The Moine Thrust, or a structural equivalent, must run between Iona and Mull. Unpublished gravity data quoted by Beckinsale and Obradovich (1973) also indicates a plane dipping at approximately 30° to the east from the shore of Iona. The characteristics of this interface imply that it is a granite/basement interface and presumed thrust plane.

If the plane is the Moine Thrust plane, the date of intrusion of the Ross of Mull granite is extremely important as it provides a minimum age for final movements of the Moine Thrust. This date would apply to the entire length of the Moine Thrust because diachroneity of movement along the thrust plane is less than that which could be distinguished using the K-Ar method.

Early determinations of the age of the Ross of Mull pluton were obtained from biotites (Miller & Brown,1965). The ages have a mean of 399 $\pm$ 14 Ma (recalculated using the constants recommended by Steiger and Jäger, 1977). Other ages obtained were 410 $\pm$ 20 Ma (Hamilton,1966) and a series of determinations with means of 402 and 407 Ma (Brown et al., 1968).

Beckinsale and Obradovich (1973) obtained ages for 7 biotites and 2 hornblendes from the pluton of 423 $\pm$ 4 Ma and 416 $\pm$ 4 Ma respectively.

They argued that the younger hornblende ages were due to the Fe^{2+} rich nature of the amphibolites. O'Nions (1969) had previously shown that these tended to give ages similar to medium grained biotites. It is also possible that the biotites took excess argon from the country rocks which are known to contain excess argon themselves (Brook, pers. comm.). However, the hornblendes do not seem to exhibit any features of excess argon. Their age is within errors the same as an Rb-Sr mineral isochron (Halliday et al., 1979) yielding a date of 414 ± 4 Ma for the intrusion of the pluton.

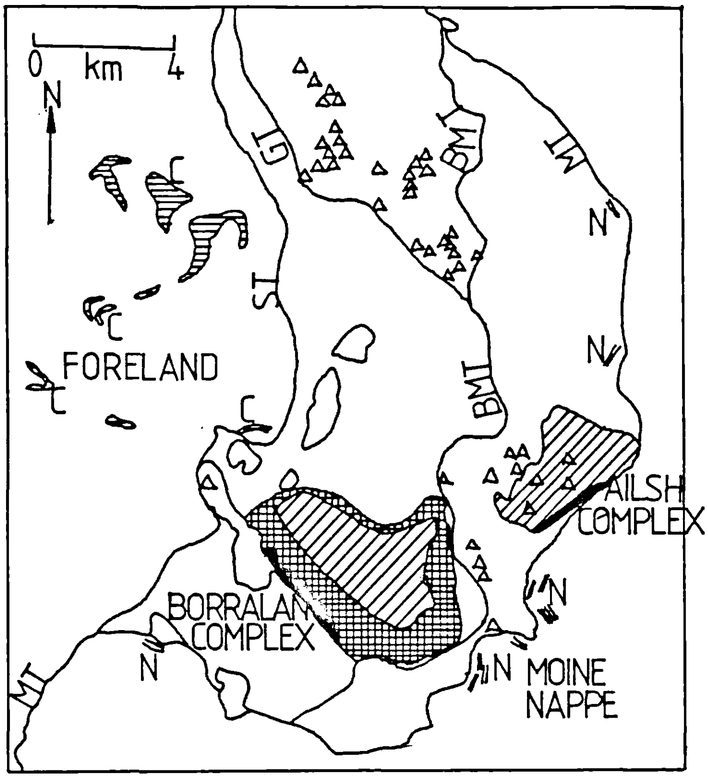
If it is accepted that the Moine Thrust does run through the Sound of Iona and that the pluton post-dates the Moine Thrust, as seen elsewhere at the western boundary of the Moine rocks, all movement had ceased along the whole of the Moine Thrust plane before 414 ± 4 Ma.

GC.6.b The Borralan Complex

The Borralan Complex is a composite pluton forming part of the suite of Caledonian alkaline intrusive rocks in Assynt (fig.112). Major and minor intrusions of the Assynt area overlap the major movements on thrusts within the Moine Thrust belt. However, the intrusions are difficult to date using the Rb-Sr technique since they contain large amounts of "common" strontium, which would mask the ^{87}Sr produced by radioactive decay (R.J. Pankhurst, pers. comm.).

K-Ar mineral data might be problematical; they may only reveal cooling dates from the uplift of the thrust belt, not dates of intrusion. Little K-Ar data exists for this area (Brown et al., 1968). The only reliable intrusion date available is 430 ± 4 Ma for the Borralan Complex using the U-Pb method on zircons (Van Breemen et al., 1979).

FIGURE 112. The Assynt Igneous suite



Saturated s yenites
Nephelene s yenites
Pyroxenites

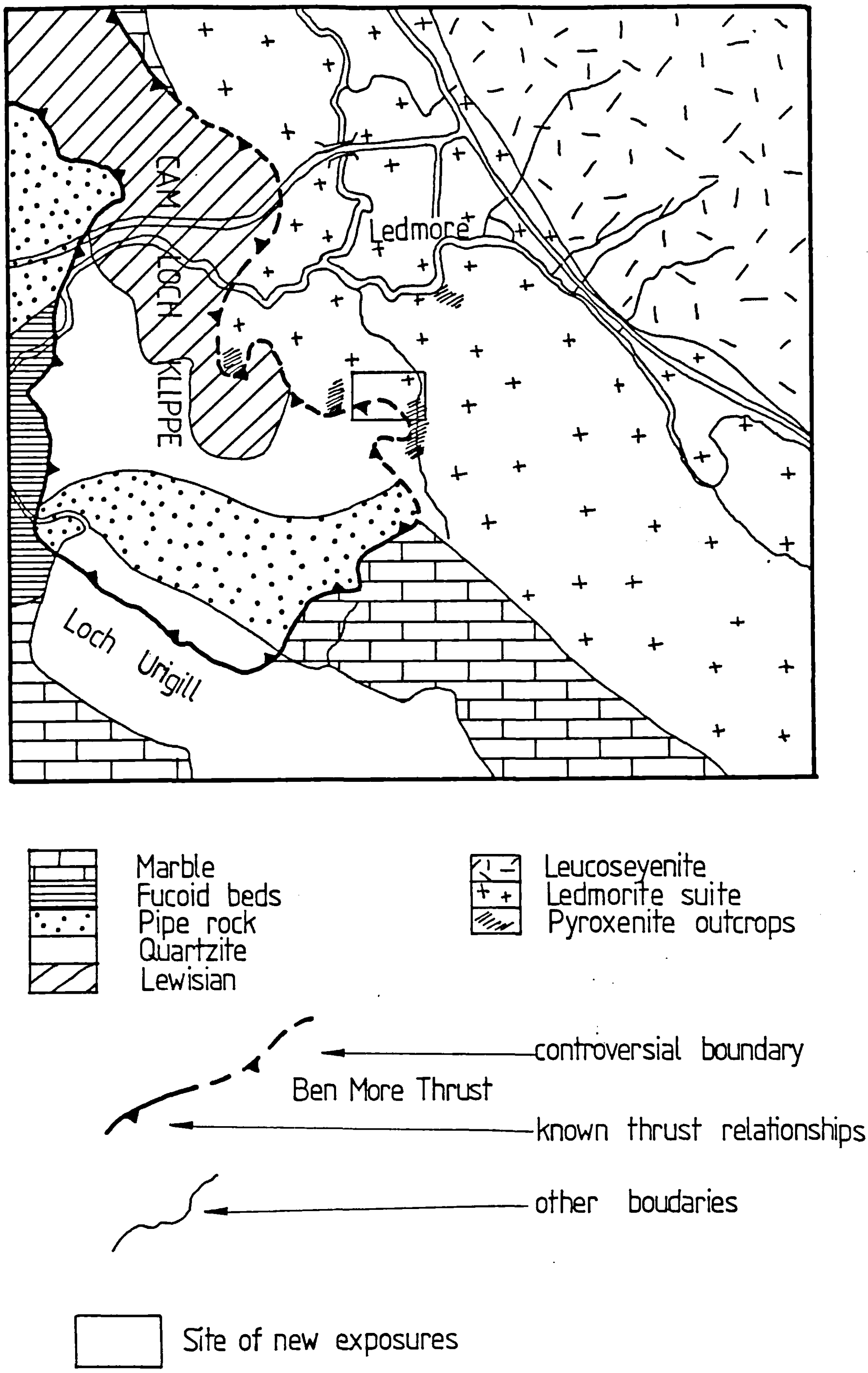
Canisp porphry
Grorudites
Nordmarkites

BMT	Ben More	thrust
GT	Glencoul	"
MT	Moine	"
ST	Sole	"

After Parsons (1979)

FIGURE 113.

Relationship between the Borralan Complex and the Cam Loch Klippe: after Parsons and McKirdy (1983)



The relative dating of thrust movements in the Moine Thrust belt as a whole has been radically revised in recent years. Older ideas of the relative history (Peach & Horne,1907; Bailey,1935; Soper & Wilkinson,1975) involved early movement on the most western thrusts and sequentially later movement on eastern thrusts. The last movement was considered to be on the Moine Thrust itself. However, in recent years this hypothesis has been challenged by an interpretation developed in the Canadian Rocky Mountains. It involves initial movement on the Moine Thrust plane then subsequent movement on more westerly thrusts carrying the early thrusts "piggy back" until finally the sole thrust moved (Barton,1978; Elliott & Johnsón, 1980; Coward & McClay, 1982).

Most of the detailed work on the structural relationships of the igneous rocks of the alkaline suite was undertaken before the new interpretation of relative movements was formulated (Sabine,1953; Wooley,1970; Parsons,1972,1979). Parsons (1979) in a review of the Assynt alkaline suite showed that the spatial distribution of intrusives between the thrust sheets and the relative timing of intrusion and thrust movement, indicates that the Moine Thrust moved late in the thrusting sequence.

However, Elliott and Johnson (1980) used the alkaline suite to demonstrate the relationships between thrusts implying a westward stacking sequence. They did not use Parsons' (1979) inferences that the Moine Thrust moved last in the sequence. Hence, although the structural relationships of the individual intrusions have been well studied and have generally been taken to indicate late movement of the Moine Thrust, they are not conclusive evidence on their own.

The Borralan Complex is the most intensively studied member of the suite (fig.113). This is because for many years, the three phases of intrusion were thought to overlap movement on the Ben More Thrust (Wooley, 1970; Parsons,1972; Van Breemen et al.,1979; Elliott & Johnson,1980). The early pyroxenites and nepheline seyenite sheet were thought to pre-date or, in some interpretations be partially contemporaneous with, movement on the Ben More Thrust.

Wooley (1970) showed that the deformation of under-saturated pseudo-leucite-bearing rocks formed a foliation defined by sub-parallel biotites and pseudoleucites (altered to K-feldspar) elongated by ratios of up to 5:1. He postulated that this deformation was caused by movement on the Ben More Thrust which passes very close to the pluton in the Carn Loch Klippe.

Elliott and Johnson (1980) postulated that the deformation within the nepheline seyenite was protoclastic, i.e. it formed during and as a result of intrusion. Recent excavations (Parsons & McKirdy,1983) have revealed that the early pyroxenites, seyenite suite and therefore the entire Borralan Complex, post-dates all movement on the Ben More Thrust plane.

An early K-Ar date from a biotite of the Borralan Complex gave an age of 394 ± 16 Ma (Miller & Brown,1965). Van Breemen et al. (1979) quote a recalculated age of 394 ± 8 Ma from Brown et al. (1968) but it is not known why the analytical errors halves between 1965 and 1968.) The date was used as a proposed cooling age by Van Breemen et al. (1979) and Elliott and Johnson (1980) used it as a date for late movement on the Ledbeag, Glencoul and Sole thrusts.

The implication of the age and position of the Borralan Complex to the present work are discussed in Section CC.7.

GC.6.c The Model of Dewey and Pankhurst (1970)

Dewey and Pankhurst (1970) produced a model for the cooling history of the Scottish Caledonides based upon the available geochronological, structural and stratigraphical evidence.

Although they did not directly date the thrust zone, their interpretation had important implications for the timing of initial movement along the thrust zone and the length of time for which the north-westward movement continued.

The then available geochronological data, mainly K-Ar dates from micas, showed a range of Caledonian ages from 500 to 380 Ma. This is an extremely large range and two hypotheses had been advanced before 1970 to explain the range:

The 'Overprint' hypothesis, stated that soon after peak metamorphism (perhaps as early as 500 Ma), the rocks cooled through the blocking temperatures and became closed to argon diffusion. They were then re-opened or partially re-opened at a later stage giving the wide spread of ages.

The 'Slow cooling' hypothesis, accounted for the spread of ages by cooling through the blocking temperatures during slow uplift at varying intervals, up to 100 Ma, after the peak of metamorphism.

Evidence cited for the overprint hypothesis by Brown et al. (1965) and Fitch et al. (1970) was the observation of peaks in K-Ar histograms for the Moine series at 420-430 Ma and for the Dalradian series at 430-440 Ma, both of which tailed off to higher ages. They also claimed that slow cooling would have produced a more systematic distribution of dates throughout the Scottish Caledonides than that seen.

FIGURE 114a

K-Ar chrontours for the Scottish Caledonides for muscovite and whole rocks: after Dewey and Pankhurst (1970)

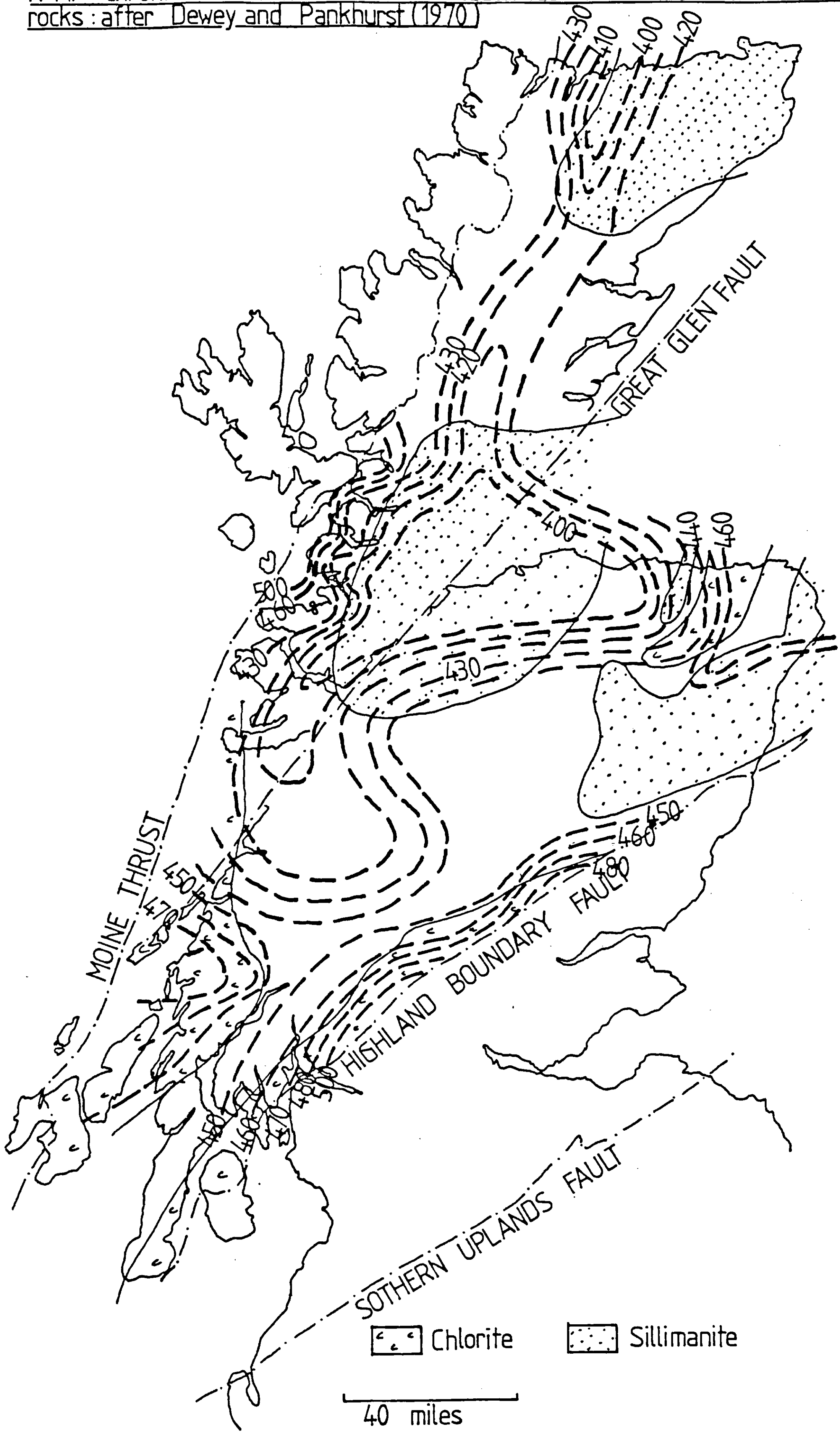
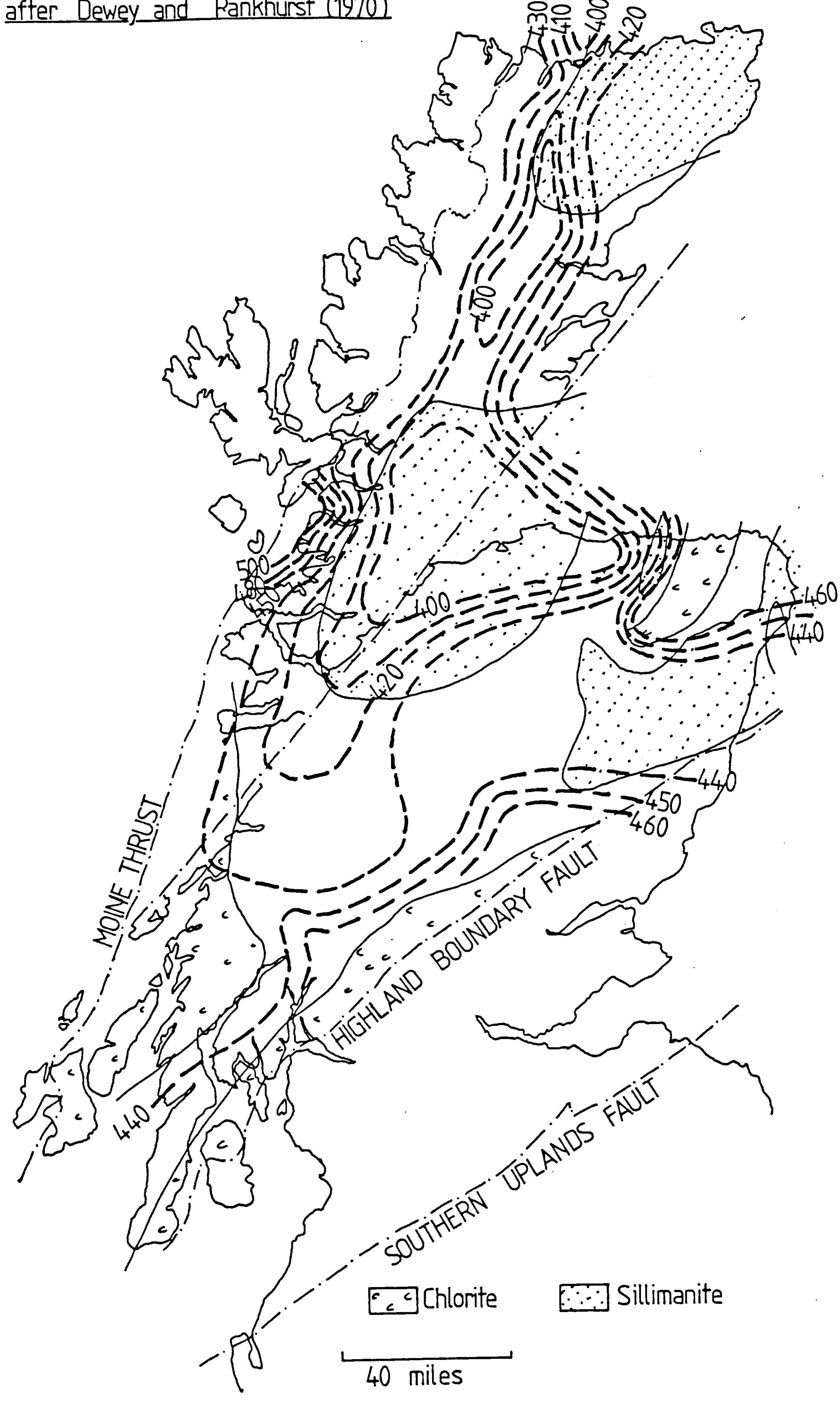


FIGURE 114b

K-Ar chrontours for the Scottish Caledonides for biotites after Dewey and Pankhurst (1970)



Harper (1964,1967a) interpreted whole rock K-Ar ages from the Upper Dalradian slates ranging from 455 Ma to 508 Ma as resulting from uplift very early in the Caledonian orogeny. Similar interpretations have been placed upon mineral ages from western Ireland (Moorbath,1969) and in the Central Alps (Purdy & Jäger, 1976).

Dewey and Pankhurst showed, using data from their own studies and from earlier literature, that a regular pattern of ages from K-Ar determinations of biotites and whole rock or muscovites could be demonstrated in the Scottish Caledonides (fig.114). The oldest ages, at the edge of the orogen in low grade terrain, were interpreted as early uplift and younger ages in the high grade terrain were interpreted as late closure to argon diffusion after cooling from higher peak metamorphic temperatures and greater depth of burial (fig.103).

Within the Moine rocks, Miller and Brown (1965) had already illustrated a westerly increase in K-Ar ages from Morar and Knoydart towards the Moine Thrust. Dewey and Pankhurst extended the chrontours (lines of equal age) over the whole of the Moine outcrop (fig.114). They expressed certain reservations about the validity of the lines but considered that the fit of the data was good. The areas where they experienced most problems fitting data were all within Moine rocks. The Ardnamurchan peninsula contained several erratic dates which they ascribed to Tertiary intrusives. Areas in the Tarskavaig nappe on Skye and around Glenelg produced seemingly young ages (416 Ma and 417 Ma respectively) which were ascribed to late outgassing during the brittle thrust movements.

According to Dewey and Pankhurst, the slow cooling model provided the best explanation of the data and fitted models produced to explain structural, metamorphic and stratigraphical relationships. They termed the lines of equal age "thermochrons", interpreting them as sequential cooling of different parts of the orogen from peak metamorphism at different times after the climatic event (fig.103).

The model works particularly well in the Dalradian where 'thermochrons' mirror the metamorphic isograds. The oldest ages are found in low grade slates of the Aberfoyle anticline and youngest ages are found in the sillimanite grade areas where early intrusions yield K-Ar mineral ages identical to the cooling dates for surrounding metasediments.

Stratigraphical evidence in the form of detrital metamorphic minerals from the Highlands in the Glen App conglomerates (Walton,1965) indicates deep erosion of the Caledonian orogen by Llandeilian or Caradocian times. Therefore there must have been previous uplift and cooling which continued into the Devonian.

In the NW Highlands, the most important implication of the Dewey and Pankhurst model is the early movement on the Moine Thrust and other thrust planes. According to this model, some movement on the Moine Thrust probably occurred before Llandeilian times and repeated movement occurred into the Lower Devonian.

This hypothesis was reinforced by data from structural studies available at the time. It was believed that a four-phase deformation history could be traced from the Moine Thrust into the mobile zone of the orogen and it was known from other studies that D2 in the mobile zone roughly coincided with peak metamorphism. Therefore it was concluded that D2 deformation in the thrust zone coincided with peak metamorphism. Mylonite formation was deduced to have occurred prior to or during the climatic metamorphic episode (Johnson,1960; Christie,1963; Ramsay,1963).

More recent structural studies have illustrated the highly complex nature of folding within the Moine (Tobish et al.,1971; Powell,1974) and recent work on the K-Ar mineral dates in Knoydart (Brook, pers.comm.) has shown that an excess argon problem in many of the micas obscures the true ages. These studies, along with the present work (see Sections MT.2 and GC.3.a,b) indicate that the model of Dewey and Pankhurst requires revision in the Moine rocks although it still seems to be applicable in the Dalradian Supergroup.

GC.7 Conclusions and Delusions

The main conclusions drawn from this survey of K-Ar and Rb-Sr mica pair ages, from Fannich to the Moine Thrust, are split into two sections:

1) Conclusions regarding the timing of uplift and thrust movement.

2) Conclusions regarding the relationships between K-Ar age and grainsize of micas in the Fannich area.

1) K-Ar ages for muscovite and biotite remain very similar from 18km to within 3km of the Moine Thrust (fig.101). This is not the pattern predicted by the model of Dewey and Pankhurst (1970). The mean muscovite age of 424.5 ± 5.4 Ma (1σ) is slightly older than the mean biotite age of 421.4 ± 2.5 Ma (1σ). Both ages show variations similar to or less than analytical errors (see Section GC.1.d). Hence, the whole area from the inner mobile zone to the Moine Thrust belt was uplifted and cooled simultaneously, most probably as a consequence of movement along the Moine Thrust zone.

As the thrust is approached to within 3km, ages from biotites become older than co-existing muscovites. This is interpreted as excess ^{40}Ar within the biotites (Section GC.3.b). The similarity of Rb-Sr biotite and K-Ar muscovite ages to those seen in Fannich indicates that the biotites close to the thrust cooled through their blocking temperature at the same time as those in Fannich. Thus the excess ^{40}Ar was introduced into biotites adjacent to the thrust, while those in Fannich cooled through their blocking point without experiencing a high partial pressure of ^{40}Ar. It follows that during uplift, the thrust zone was an active conduit for argon. This fact strengthens the proposition that cooling from Fannich to the Moine Thrust was consequent upon movement on the Moine Thrust. Further, the muscovite K-Ar ages give a possible indication of initial cooling through their blocking temperature, close to the thrust zone as early as 435 Ma, cooling slightly later in the Fannich area. This difference might indicate uplift along the thrust zone; however, the difference is within analytical errors.

2) Three grainsizes were separated from samples taken in the Fannich area (figs.106,107), 125-200μ, 200-250μ and 250-500μ. The finest grainsize gave reproducible results with means of 423.6 ± 6.4 Ma (1σ) for muscovite and 421.6 ± 3.3 Ma (1σ) for biotite. These were used in the survey from Fannich to the Moine Thrust zone. The two coarser grainsizes gave unreproducible results. The coarsest grainsize yielded mean ages of 433.7 ± 8.7 Ma (1σ) for muscovite and 433.4 ± 7.6 Ma (1σ) for biotite. The standard deviations on these are greater than analytical errors, implying that there is geological scatter in the population.

Using the statistical 't' test, it was found that the difference between coarse and fine mica ages is significant. It has also been deduced that the variation is not due to systematic analytical errors.

The correspondence in ages of muscovite and biotite samples of the same grainsize and the variation between grainsizes indicate a late partial resetting event. The finest grainsizes being totally reset and larger grainsizes being less affected. It is also possible, but considered less likely, that the micas contained excess ^{40}Ar before the resetting event but this cannot be ruled out on the present evidence.

The interpretation of Rb-Sr biotite ages (mica pairs) presents a problem since all grainsizes together give a mean age of 411.5 $\pm$ 6.5 Ma along the traverse; significantly younger than K-Ar biotite ages. This is not the normal case in metamorphic terrains (Purdy & Jäger,1976). K-Ar and Rb-Sr ages for biotites have generally been found to be similar although some variations have been recorded (Verschure et al.,1980).

Normally the younger Rb-Sr biotite ages would be taken to indicate that biotites contain excess ^{40}Ar. However, it seems that in this case strontium was more mobile than argon during the late cooling event in Fannich.

The situation in the Fannich area differs from that found in high grade areas of the Alps, according to Purdy and Jäger (1976) all grainsizes give similar ages and muscovites are systematically older than biotites. In recent work from lower grade Alpine terrains, some indication has been found that the blocking temperature concept may be brought into question (Chopin & Maluski,1980,1982; Desmons et al.,1982). Work in Scandinavia (Verschure et al.,1980) also indicates that the concept of a single temperature at which minerals open and close may not be valid.

In the Fannich area, the greater variation of ages between grainsizes compared to the variation of ages between types of mica, is difficult to explain by conventional, temperature dependent, diffusion theory (Dodson, 1973,1979). The Rb-Sr ages of biotite are anomalously low compared to the results of studies from other areas and this too is difficult to reconcile. It is therefore tentatively suggested that factors other than temperature may be involved in the explanation (cf. Chopin & Maluski, 1980,1982).

Large micas (1-2cm) extracted from pegmatites which pre-date the Sgurr Beag Slide yield K-Ar ages in excess of 445 Ma. A two point

Rb-Sr isochron (muscovite-plagioclase) from one gave an age of 457 $\pm$ 6 Ma which represents a minimum age of peak Caledonian metamorphism. The K-Ar ages from pegmatitic micas are older than the mean muscovite age of the metasedimentary muscovites. If grainsize does have an effect upon the 'resettability' of micas in Fannich, the pegmatitic muscovite ages may represent the original closing age of muscovite in the metasediments.

The age of intrusion of the Ross of Mull granite at 414 $\pm$ 4 Ma is slightly later than the cooling of K-Ar biotite ages in the Fannich and within errors the same as the Rb-Sr biotite ages. If the Ross of Mull granite was intruded within a few million years of the cessation of movement along the Moine Thrust, then the two results are compatible with late movement on the thrust between 414 Ma and 425 Ma.

The Assynt alkaline suite presents several problems of interpretation. as discussed in Section GC.5.b. The most important to the present discussion is the interpretation of the Borralan Complex. Since the recent work by Parsons and McKirdy (1983) it is known that the intrusion entirely post-dates movement on the Ben More Thrust. Under the foreland stacking duplex model for Assynt, the Ben More Thrust post-dates the Moine Thrust. However, late movement on the Moine Thrust is also indicated (Coward,1983; Johnson et al. in prep.). The Borralan intrusion was dated at 430 $\pm$ 4 Ma (U-Pb on zircons) and the present survey indicates that the Moine Thrust was active after 425 Ma. Hence, if the Ben More Thrust pre-dates 430 Ma, the Moine Thrust must be active subsequently. No date yet exists for the Loch Ailsh intrusion which pre-dates the Ben More Thrust but post dates early mylonitization. This intrusion, if dated accurately, would provide an important bracket for the age of movement on the Ben More Thrust.

Little significance is attached to the K-Ar date of 394 $\pm$ 16 Ma on the Borralan complex since the date was obtained in 1965 (Miller & Brown, 1965) and other ages from the same source have proved to be up to 5% out on more recent determinations (Beckinsale & Obradovich,1973).

The slow cooling model of Dewey and Pankhurst (1970) seems to work well in the Dalradian, with ages mirroring the metamorphic grade. In the Fannich/Ullapool area however, rocks which were at, or above, almandine amphibolite grade during the peak of Caledonian metamorphism are juxtaposed against unmetamorphosed rocks of the foreland. The flat age traverse from the inner mobile zone of the thrust belt is in contrast to the "thermal bulge" suggested by Dewey and Pankhurst (fig.114) with old ages adjacent to the thrust and younger ones toward the centre of the orogen. Further, the rocks of the Fannich area have undergone a late resetting event, not a "slow cooling" gradually from peak metamorphism.

Therefore the interpretation due to Dewey and Pankhurst (op.cit.) of early, Llanvirn, uplift along the thrust zone does not fit the data of the present survey. Other work in progress on the Knoydart area (M. Brook, pers. comm.) tends to confirm this, showing that many of the old ages which led Dewey and Pankhurst to postulate early uplift along the thrust belt are due to the presence of excess ^{40}Ar.

CONCLUSIONS

1. A complex pre-Caledonian tectonometamorphic history, including pervasive deformation and migmatization has been recognised within the Glenfinnan Division rocks above the Sgurr Beag Slide in Fannich. This contrasts with rocks of the Morar Division, below the Slide, which exhibit only minor pre-Caledonian deformation and greenschist facies metamorphism.

A further striking difference in complexity of tectonometamorphic histories is observed between the Morar division rocks in Fannich, which are relatively simple, and Morar Division rocks in the SW Moine area which are more complex.

2. The presence of the Sgurr Beag Slide is confirmed in the Fannich area at the boundary between the Meall an t'sithe Pelite and Meall a Chrasgaidh Psammite. From the quartz c-axis fabrics, it is inferred that overthrusting was towards the northwest. By reconstructing the situation prior to F3 folding, it is seen that movement on the slide exceeded 55km and observation of mineral textures and fabrics of calc-silicates and pelites, indicates that the movement occurred under amphibolite conditions.

3. It is proposed that the Sgurr Beag Slide passes over the present outcrop of the Carn Chuinneag granite and it seems most likely that the path of the slide was deflected by the granite.

4. The mylonites above the Moine Thrust cross-cut the bedding parallel to the Sgurr Beag Slide fabrics. These mylonites post-date the Sgurr Beag Slide and a regional pegmatite suite; they formed syn- to post-F3 (a set of major N-S trending a symetrical folds). A significant time interval is indicated between the two movements by the pegmatite suite, regional fold phase and decay of regional metamorphic grade from mid-amphibolite to greenschist facies.

5. Planar, anisotropic fabrics were formed in the pelites and psammites of Fannich as a result of deformation associated with the Sgurr Beag Slide. In the area west of Fannich these fabrics were near co-planar with the Moine Thrust mylonites and this led to the formation of 'shear banding' in pelites and 'foliation boudinage' in psammites. The shear bands formed in pelites at biotite grade and continued to form as the rocks retrogressed to chlorite grade. The mean angle between shear bands and pre-existing fabric decreases with increasing deformation and shear band formation plays an important part in the grainsize reduction process.

6. An 18km traverse of K-Ar ages for muscovite and biotite and Rb-Sr mica pair ages from Fannich to the Moine Thrust indicate that the whole area was uplifted and cooled virtually simultaneously. The area had cooled to c.350°C by 424.5 ± 5.4 (1σ) Ma and c.280°C by 421.4 ± 2.5 (1σ) Ma. No excess ^{40}Ar was detected away from the thrust zone but within 3km of the Moine Thrust, biotite K-Ar ages become anomalously old, indicating that the Moine Thrust zone was active during late cooling, acting as a conduit for ^{40}Ar escaping from areas deeper in the thrust zone. It seems likely that cooling through the muscovite and biotite blocking points was consequent upon movement on the Moine Thrust zone.

7. In the area north of Loch Fannich, an intensive study of several grainsize separates from co-existing micas indicated a correlation between K-Ar age and grainsize. The largest grainsize (250-500μ) yielded a widespread range of ages with a mean of 433 Ma while the finest grainsize (125-200μ)yielded reproducible ages with a mean of 423.5 ± 6 Ma for muscovite and 421.6 ± 3 Ma for biotite. Rb-Sr mica pair ages are consistently lower than both K-Ar ages in all samples analysed. It has been inferred that a partial resetting event occurred in this area. The nature of the event has not been determined but it may be connected with deformation and movement on the Moine Thrust zone.

REFERENCES

Allen, A.R. & Stubb, D. (1982): An ^{40}Ar-^{39}Ar Study of a polymetamorphic complex in the Arunta block, central Australia. Contrib. to Min. & Pet., 79 (3), pp 319-332.

Anderson, D.E & Olimpio, J.C. (1977): Progressive homogenisation of metamorphic garnets, S. Morar, Scotland. Can. Mineral., 15, p. 205, 216.

Bailey, E.B. (1955): Moine Tectonics and Metamorphism in Skye. Trans. Edin. Geol. Soc., 16 Pt II, pp. 93-106.

Baird, A. (1982): The Sgurr Beag Slide within Moine rocks at Loch Eilt, Inverness -shire. J. Geol. Soc.Lond., 139, pp.647-653.

Barber, A.J. (1965): The history of the Moine Thrust zone, Loch Carron and Loch Alsh, Scotland. Proc. Geol. Assoc., 76, pp 215-242.

Barton,C.M.(1978): An Appalachian view of the Moine Thrust. Scot.J.Geol., 14, pp 247-257.

Beckinsale, R.P. & Obradavich, J.D. (1973): Potassium-Argon ages for minerals from the Ross of Mull, Argyllshire, Scotland. Scott.J. Geol., 9, pp 147-156.

Bell, T.H. (1978): Progressive deformation and reorientation of fold axes in a ductile mylonite zone: The Woodroffe Thrust. Tectonophysics, 44, pp 285-320.

Berger, G.W. & York, D. (1981): Geothermometry from ^{40}Ar-^{39}Ar dating experiments. Geochim.et Cosmochim.Acta, 45, pp 795-811.

Berthe, D., Choukroune, P. & Jegouzo, P. (1979): Orthogneiss, mylonite and non-coaxial deformation of granites: the example of the South American shear zone. J. Struc. Geol., 1, pp 31-42.

Biot, A. (1965): Mechanisms of Incremental Deformation. Wiley, New York.

Boelrijk, N.A.I.M. (1968): A general formula for "double" isotope dilution analysis. Chem. Geol., 3, pp 323-325.

Borradaile, G.J. (1972): Variably oriented co-planar primary folds. Geol. Mag., 109, pp 89-98.

Bouchez, J-L & Pecher, A. (1976): Plasticité du quartz et sense de cisaillement dans les quartzites du Grand Chevauchement Central Himalayen. Bull. Soc. geol. Fr., 6, pp 1375-1384.

Brewer, M.S. (1969): Excess radiogenic argon in metamorphic micas from the eastern Alps, Austria. E.P.S.L., 6, pp 321-331.

Brewer, M.S., Brook, M. & Powell, D. (1979): Dating of the tectono-metamorphic history of the south-western Moine, Scotland. The Caledonides of the British Isles. Reviewed 1979, Geol. Soc. of London.

*

Brown, P.E., Miller, J.A., Soper, N.J. & York, D. (1965): Potassium-argon age pattern of the British Caledonides. Proc. Yorks. Geol. Soc., 35, pp 103-138.

*

Bryant, B. & Reed, J. C. (1969): Significance of lineation and minor folds near major thrust faults in the S. Appalachians and the British and Norwegian Caledonides. Geol. Mag., 106, pp 412-429.

Burg, J.P. & Laurant, P.L. (1978): Strain analysis of a shear zone in a Granodiorite. Tectonophysics, 47, pp 15-42.

Burnett, D.S., Lippolt, H.J. & Wasserburg, G.J. (1960): The relative isotopic abundance of ^{40}K in terrestrial and meteoritic samples. J. Geophys. Res., 71, pp 1249-1269.

Carreras, J., Estrada, A. & White, S.H. (1977): The effects of folding on the c-axis fabrics of a quartz mylonite. Tectonophysics, 39, pp 3-24

Charnley, N. (1976): Geochemical Studies of the Moine rocks in western Inverness-shire. (Unpublished Ph.D. thesis).

Chopin, C. & Maluski, H. (1980): ^{40}Ar-^{39}Ar Dating of High pressure metamorphic micas from the Gran Paradiso area (western Alps): Evidence against the blocking temperature concept. Contrib. Mineral Petrol., 74, pp 109-122.

Chopin, C. & Maluski, H. (1982): Unconvincing evidence against the blocking temperature concept? A reply. Contrib. Mineral. Petrol., 80, pp 391-394.

Christie, J.M. (1960): Mylonitic rocks of the Moine Thrust zone in the Assynt region, north-west Scotland. Edin. Geol. Soc. Trans., 18, pp 79-93.

Christie, J.M. (1963): The Moine Thrust zone in Assynt region, north-west Scotland. Univ. California Publ., Geol. Sci., 40, No.6, pp 345-440.

Cobbold, P. R., Cosgrove, J.W. & Summers, J.M. (1971): Development of internal structures in deformed Anisotropic rocks. Tectonophysics, 12, pp 23-53.

Cobbold, P.R. (1977a): Description and origin of banded deformation structures: I. Regional strain, local perturbations and deformation bands. Can. J. Earth Sci., 14, pp 1721-1731.

Cobbold, P.R. (1977b): Description and origin of banded deformation structures: II. Rheology and the growth of banded perturbations. Can. J. Earth Sci., 14, pp 2510-2523.

Cosgrove, J.W. (1976): The formation of crenulation cleavage. J. Geol. Soc. Lond., 132, pp 155-178.

Coward, H.P. (1980): The Caledonian thrusts and shear zones of N.W.Scotland. J. Struc. Geol., 2, pp 11-17.

Coward, H.P. & Kim, J.H. (1982): Strain within thrust sheets. In: Thrust and Nappe Tectonics 1981, spec. publ. Geol. Soc. Lond.

Coward, H.P. (1983): Thrust and shear zones of the Moine Thrust zone and the NW Scottish Caledonides. J. Geol. Soc. Lond., 140, pp 795-813.

Cowie, J.W. (Ed.) (1982): The Precambrian-Cambrian Boundary. Precamb. Res., 17, pp 75-146.

Cunningham Craig, E.H., Wright, W.B. & Bailey, E.B. (1911): The Geology of Colonsay and Oronsay, with part of the Ross of Mull. Mem. Geol. Surv.UK.

Dalrymple, G.B. & Lanphere, M.A. (1967): Potassium-Argon dating. Principles and application to Geochronology. W.H. Freeman, San Francisco.

De L'Apparent, J. (1935): Rythmes du metamorphisme dans les Highlands. Bull. Soc. Geo. Fr., 5, pp 281-317.

Del Moro, A., Puxeddu, M. Radicati Di Brozas, F. & Villa, I.M. (1983): Rb-Sr and K-Ar Ages on minerals at temperatures of 300-400°C from deep wells in the Larderello Geothermal field (Italy). Contrib. Min. Petrol., 81, pp 340-349.

Desmons, J., Hunziker, J.C. & Delaloye, M. (1982): Unconvincing evidence against the blocking temperature concept. Contrib. Mineral. Petrol., 80, pp 386-390.

Dewey, J.F. & Pankhurst, R.J. (1970): The evolution of Scottish Caledonides in relation to their isotopic age pattern. Trans. Roy. Soc. Edinb., 68, pp 361-389.

Dodson, M.H. (1973): Closure temperature in cooling geochronological and petrological systems. Contrib. Mineral. Petrol., 40, pp 259-274.

Dodson, M.H. (1979): Theory of cooling ages. In: Jager & Hunziker (eds.) Lectures in Isotope Geoloey, pp 194-202.

Elliott, D. & Johnson, M.R.W. (1980): Structural evolution in the northern part of the Moine Thrust belt, NW Scotland. Trans. Roy. Soc. Edin. Earth Sci., 71, pp 69-96.

Escher, A. & Watterson, J. (1974): Stretching fabrics, folds and crustal shortening. Tectonophys., 22, pp 223-231.

Etheridge, M.A. & Vernon, R.H. (1981): A deformed polymictic conglomerate - the influence of grain size and composition on the mechanism and rate of deformation. Tectonophysics, 79, pp 237-254.

Etheridge, M.A. & Wilkie, J.C. (1978): Grainsize reduction, grain boundary sliding and the flow strength of mylonites. Tectonophysics, 58, pp 159-178.

* Fergusson, I.W. (1978): Structural age of the Vagastie Bridge Granite. Scot. J. Geol., 14, pp 89-92.

Fitch, F.J., Miller, J. A. & Brown, P.E. (1964): Age of Caledonian Orogeny and Metamorphism in Britain. Nature, Lond, 203, pp 275-278.

Fitch, F.J., Miller, J.A. & Mitchell, J.G. (1970): A new approach to isotopic dating in orogenic belts. In: Time and Place in orogeny, Spec. Publ. Geol. Soc. Lond.

Gapias, D. & White, S. (1982): Ductile Shear bands in a naturally deformed quartzite. Textures and Microstructures, 5, pp 1-17.

Geological Survey of Great Britain (1898): Summary of Progress.

Gunn,W Horne, PeachB & PocockD (1913): Memoir of the Geological Survey for sheet 92 (Fannich Mountains).

Giletti, B.J. (1974): Studies in diffusion - I. Argon in phlogopite mica. In: Geochemical Transport and Kinetics, pp 108-116. A.W. Hoffman, B.J. Giletti, H.S. Yoder Jr & R.A. Yung (Eds) Carnegie Institution of Washington, Publ. 634, p.353.

Hambrey, M.J. & Milnes, A.G. (1975): Boudinage in Glacier Ice - some examples. J.Glaciology, 14, pp 383-393.

Hamilton, E.I. (1960): The isotopic composition of lead in igneous rocks. 1. The origin of some Tertiary granites. E.P.S.L., 1, pp 30-37.

Hanson, G.I. & Gast, P.W. (1967): Kinetic studies on contact metamorphic zones. Geochim. Acta., 31, pp 1119-1153.

Harker, R.I. (1954): Further data on the petrology of the pelitic hornfels of the Carn Chuinneag-Inchbae region, Ross-shire, with special reference to the status of almandine. Geol. Mag., 91, pp 445-462.

Harker, R.I. (1970): Preservation of folding in the Carn Chuinneag-Inchbae hornfels. Scot. J. Geol., 6, pp 226-227.

Harper, C.T. (1964): Potassium-argon ages of slates and their geological significance. Nature (Lond.), 203, pp 468-470.

Harper, C.T. (1967): On the interpretation of K-Ar ages from Precambrian shields and Phanerozoic orogens. Earth Planet. Sci. Letters, 3, pp 123-132.

Harrison, T.M. & McDougall, I. (1980a): Investigations of an intrusive contact, north-west Nelson, New Zealand - I. Thermal, chronological and isotopic constraints. Geochim. Cosmochim. Acta, 44, pp 1985-2003.

Harrison, T.M. & McDougall, I. (1980b): Investigations of an intrusive contact, northwest Nelson, New Zealand - II. Diffusion of radiogenic and excess ^{40}Ar in hornblende revealed by $^{40}Ar/^{39}Ar$ age spectrum analysis. Geochim. Cosmochim. Acta, 44, pp 2005-2020.

Harrison, T.M. & McDougall, I. (1981): Excess ^{40}Ar in metamorphic rocks from Broken Hill, New South Wales: Implications for ^{40}Ar/^{39}Ar age spectra and the thermal history of the region. E.P.S.L., 55, pp 123-149.

Hart, S.R. (1964): The petrology and isotopic mineral age relations of a contact zone in the Front Range, Colorado. J. Geophys. Res., 72, pp 493-525.

Haugh, I. (1964): The Structure and metamorphic history of the S.E. part of the Carn Chuinneag Complex. Ph.D. thesis (Unpublished).

Higgins, A.C. (1967): The age of the Durine member of the Durness Limestone formation at Durness. Scott. J. Geol., 3, pp 382-388.

Holdsworth, R.E. & Roberts, A.M. (1984): Early curvilinear fold structures and strain in the Moine of the Glen Gary region, Inverness-shire. J. Geol. Soc. Lond., 141, pp 327-339.

* Jehu, T.J. (1922): The Archaean and Torridonian formations and the later intrusive igneous rocks of Iona. Trans. Roy. Soc. Edin., 53, pp 165-187.

Johnson, M.R.W. (1957): The Structural Geology of the Moine Thrust Zone in Coul in Forest, Wester Ross. Q.J.G.S., 113, pp 241-270.

Johnson, M.R.W. (1960): The structural history of the Moine Thrust Zone at Loch Carron, Wester Ross. Trans. Roy. Soc. Edin., 64, pp 139-168.

Johnson, M.R.W. (1960): Linear structures in the Moine Thrust belt of NW Scotland and their relation to Caledonian orogenesis. Int. Geol. Cong. Norden, Pt.XIX, pp 89-95.

Johnson, M.R.W. (1961): Polymetamorphism in movement zones in the Caledonian thrust belt of north-west Scotland. J. Geol., 69 No.4, pp 417-432.

Johnson, M.R.W. (1965): The Moine Thrust: a discussion. J. Geol., 73, pp 672-676.

Johnson, M.R.W. (1967): Mylonite zones and mylonite banding. Nature, 213, pp 246-247.

Johnson, M.R.W. & Shepherd, J. (1970): Notes on the age of metamorphism of the Moinian. Scott. J. Geol., 6, pp 228-229.

Johnson, M.R.W., Kelley, S.P., Oliver, G.J.H. & Winter, D.A. (in prep.), The thermal effects of thrusting in the Moine Thrust zone: Their tectonic and regional implications.

Johnstone, G.S., Smith, D. I. & Harris, A.L. (1969): The Ionian assemblage of Scotland. M. Kay (ed) N. Atlantic - geology and continental drift, a symposium. Memo. Amer. Ass. Petrol. Geol., 12, pp 159-180.

Johnstone, G.S. (1978): The Moine Succession. A Correlation of the Precambrian rocks in the British Isles. Special Report No. 6 to the Geol. Soc.

Kendall, B.R.F. (1960): Isotopic composition of potassium. Nature, 186, pp 225-226.

Kennedy, W.Q. (1949): Zones of progressive regional metamorphism in the Moine schists of the Western Highlands of Scotland. Geol. Mag., 86, pp 43-56.

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: A. L. Harris et al. The Caledonides of the British Isles - reviewed Spec. Publ. Geol. Soc. Lond.

Lister, G.S. & Williams, P.F. (1979): Fabric development in shear zones: theoretical controls and observed phenomena. J. Struc. Geol., 1, pp 283-298.

Long, L.E. & Lambert, R.St J. (1963): Rb-Sr isotopic ages from the Moine Series. In: M. R.W. Johnson & P.H. Stewart (eds), The British Caledonides, Edinburgh.

Long, L.E. (1964): Rb-Sr Chronology of the Carn Chuinneag Intrusion, Ross-shire, Scotland. J. Geophys. Res., 69, pp 1589-1697.

McClay, K.R. & Coward, M.P. (1982): The Moine Thrust: an overview. In: Thrust and Nappe tectonics. Spec. publ., Geol. Soc. Lond.

McIntyre, D.B. (1954): The Moine Thrust - Its discovery, age and tectonic significance. Geologists Assoc. Proc., 65, pp 203-219.

Maltman, G (1977): Some microstructures of experimentally deformed argillaceous sediments. Tectonophysics, 39, pp 417-452.

Means, W.D. & Williams, P.F. (1972): Crenulation cleavage and faulting in an artificial salt-mica schist. J. Geol., 80, pp 569-591.

Melenevskiy, V.N., Morozova, I.M. & Yurgina, Y.K. (1978): The migration of radiogenic argon and biotite dehydration. Geochem. Int., 15, No.6, pp 7-16.

MacKenzie, B (1949): Kyanite-gneiss within a thermal aureole. Geol. Mag., 80, pp 215-254.

MacQueen, J.A. & Powell, D. (1977): Relationships between deformation and garnet growth in Moine (Precambrian) rocks of western Scotland. Geol. Soc. of America Bulletin, 88, pp 235-240.

Miller, I.A. & Brown, P.E. (1965): Potassium-argon age studies in Scotland. Geol. Mag., 102, pp 106-134.

Moorbath, S. (1969): Evidence for the age of deposition of the Torridonian sediments of NW Scotland. Scot. J. Geol., 5, pp 154-170.

O'Nions, R.K., Smith, D.G.W., Baadsgaard, H. & Morton, R.D. (1969): Influence of chemical composition on argon retentivity in metamorphic calcic amphiboles from South Norway. E.P.S.L., 5, pp 339-345.

Pankhurst, R.J. (1982): Geochronological tables for British Igneous rocks. In: D. S. Sutherland (ed) Igneous Rocks of the British Isles, Wiley.

Parsons, I. (1972): Comparative petrology of the leucocratic seyenites of the Northwest Highlands of Scotland. Geol. J., 8, pp 71-82.

Parsons, I. (1979): The Assynt Alkaline suite. In: The Caledonides of the British Isles - Reviewed. Spec. Publ. Geol. Soc. Lond.

Parsons, I. & McKirdy, A.P. (1983): Inter-relationship of igneous activity and thrusting in Assynt: excavations at Loch Borralan. Scott. J. Geol., 19, pp 59-66.

Peach,B Gunn,W Clough,C Hinxman,L, Crampton,C & Anderson,E with pet. by Flett.J (1912) The Geology of Carn Chuinneag, Inchbae and the surrounding country - An Explanation of Sheet 93. Mem. Geol. Surv. Scotland.

Peach, B.N. & Horne, J. (1907): The Geological structure of the NW Highlands of Scotland. Mem. Geol. Surv. Scotland.

Pfiffner, O.A. (1981): Fold and Thrust tectonics in the Helvetic Nappes (E. Switzerland). Thrust & Nappe Tectonics (Geol. Soc. Publ).

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 & Planetary Science Letters, 24, pp 105-112.

Pidgeon, R.T. & Aftalion, M. (1978): Cogenetic and inherited zircon U-Pb systems in granites: Palaeozoic granites of Scotland and England. In: Crustal Evolution in NW Britain and adjacent regions. Eds: Bowes & Leake.

Platt, J.P. & Visers, R.L.M. (1980): Extensional structures in anisotropic rocks. J. Struc. Geol., 2 No.4, pp 397-410.

Powell, D. (1964): The stratigraphical succession of the Moine schists around Lochailort (Inverness-shire) and its regional significance. Proc. Geol. Assoc. Lond., 75, pp 223-250.

Powell, D. (1974): Stratigraphy and Structure of the Western Moine and the problem of Moine orogenesis. Quarterly Journal Geol. Soc., 130, pp 575-593.

Powell, D. & MacQueen, J.A. (1976): Relationships between garnet shape, rotational inclusion fabrics and strain in some Moine metamorphic rocks of Skye, Scotland. Tectonophysics, 35, pp 391-402.

Powell, D., 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. Lond., 138, pp 661-673.

Powell, D., Brook, M. & Baird, A.W. (1983): Structural dating of a Pre-cambrian pegmatite in Moine rocks of northern Scotland and its bearing on the status of the 'Morarian Orogeny'. J. Geol. Soc. Lond., 140, pp 813-823.

Purdy, J.W. & Jager, E. (1976): K-Ar ages of rock forming minerals from the Central Alps. Mem. Inst. Geol. Min. Univ. Padova, 30, pp 1-31.

Ramsay, J.G. (1957): Moine-Lewisian relations at Glenelg. Q. J. Geol. Soc. 113, pp 487-523.

Ramsay, J. & Spring, J. (1962): Moine Stratigraphy in the western Highlands of Scotland. Proc. Geol. Assoc., 73, pp 295-326.

Ramsay, J.G. (1967): Folding and Fracturing of Rocks. McGraw-Hill.

Ramsay, J.G. & Graham, R.H. (1970): Strain variation in shear belts. Can. J. Earth Sci., 7, p.786.

Rathbone, P.A. (1982): Basement/cover relationships on the Moine series of Scotland, with particular reference to the Sgurr Beag Slide. Ph.D. thesis (unpublished) Liverpool.

Rathbone, P.A. & Harris, A.L. (1979): Basement-cover relationships at Lewisian Inliers in the Moine rocks. Geol. Soc. Lond - The Caledonides reviewed, pp 101-107.

* Read, H.H. (1934): Age problems of the Moine Series of Scotland. Geol. Mag., 71, pp 302-317.

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. Lond., 140, pp 883-892.

Richey, J.E. & Kennedy, W.Q. (1939): The Moine and sub-Moine series of Morar, Inverness-shire. Bull. Geol. Surv. Gt Br., 2, pp 26-45.

Roberts, J.L. & Sanderson, D.J. (1974): Oblique fold axes in the Dalradian rocks of the southwest Highlands. Scott. J. Geol., 9, pp 281-296.

Roermund, H. Van, Lister, G.S. & Williams, P.F. (1979): Progressive development of quartz fabrics in a shear zone from Monte Mucrone, Sesia-Lanzo zone, Italian Alps. J. Struc. Geol., 1, pp 43-52.

Roddick, J.C., Cliff, R.A. & Rex, D.C. (1980): The evolution of excess argon in Alpine biotites - a ^{40}Ar-^{39}Ar analysis. E.P.S.L., 48, pp 185-208.

Rutledge, H. (1952): The Structure of the Fannich Forest Area. Trans. Edin. Geol. Soc., 115, pp 317-321.

Sabine, P.A. (1953): The petrology and geological significance of post-Cambrian minor intrusions of Assynt and the adjoining districts of NW Scotland. Q.J. Geol. Soc. Lond., 109, pp 137-171.

Sanderson, D.J. (1973): The development of fold axes oblique to the regional trend. Tectonophysics, 16, pp 55-70.

Shepherd, J. (1970): The structure and metamorphism of the Carn Chuinneag area, Ross-Shire. Ph.D. thesis (unpublished) Edinburgh.

Shepherd, J. (1973): The structure and structural dating of the Carn Chuinneag intrusion, Ross-shire. Scott. J. Geol., 9-10, pp 63-88.

Sibson, R.H. (1977): Fault rocks and fault mechanisms. J. Geol. Soc. Lond., 133, pp 191-213.

Smythe, D.K., Dobinson, A., McQuillin, R., Brewer, J.A., Matthews, D.H., Blundell, D.J. & Kelk, B. (1982): Deep structure of the Scottish Caledonides revealed by the MOIST reflection profiles. Nature, 299, pp 338-340.

Soper, N.J. (1971): The earliest Caledonian structures in the Moine Thrust belt. Scott. J. Geol., 7, pp 241-247.

Soper, N.J. & Brown, P.E. (1971): Relationship between metamorphism and migmatization in the northern part of the Moine nappe. Scott. J. Geol., 7, pp 305-325.

Soper, N.J. & Wilkinson, P. (1975): The Moine Thrust and Moine Nappe at Loch Eriboll, Scotland. Scott. J. Geol., 11, pp 339-359.

Soper, N.J. & Barber, A.J. (1982): A model for the deep structure of the Moine Thurst zone. J. Geol. Soc. Lond., 139, pp 127-138.

Steiger, R.H. & Jäger, E. (1977): Sub-commission on geochronology: convention on the use of decay constants in geo- and cosmochronology. E.P.S.L., 36, pp 259-262.

Stewart, A.D. (1982): Late Proterozoic rifting in NW Scotland: the genesis of the 'Torridonian'. J. Geol. Soc. Lond., 139, pp 413-420.

Strachan, R.A. (1982): Tectonic sliding within the Moinian Loch Eil Division near Kinlocheil, W. Inverness-shire. Scott. J. Geol., 18, pp 187-203.

Sutton, J. & Watson, J. (1953): The supposed Lewisian inlier of Scardroy, central Ross-shire, and its relations with the surrounding Moine rocks. Q.J. Geol. Soc., 108, pp 99-126.

Sutton, J. & Watson, J. (1954): The Structure and Stratigraphical succession of the Moines of Fannich Forest and Strath-Bran, Ross-shire. Q. J. Geol. Soc., 60, pp 21-54.

Sutton, J. & Watson, J. (1958): Structures in the Caledonides between Loch Duich and Glenelg, NW Highlands. Q. J. Geol. Soc., 114, pp 231-267.

Sutton, J. & Watson, J.V. (1962): An interpretation of Moine-Lewisian relations in central Ross-Shire. Geol. Mag., 99, pp 527-541.

Tanner P.W.G., Johnstone,S., Smith, D. & Harris, A.L. (1970): Moinian Stratigra and the problem of the central Ross-shire inliers. Geol. Soc. Amer. Bull., 81, pp 299-306.

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., 126, pp 435-463.

Tilley, C.E. (1935): The role of Kyanite in the 'hornfels zone' of the Carn Chuinneag granite (Ross-shire). Min. Mag., 24, pp 92-97.

Tobish, D.T., Fleuty, M.J., Merh, S.S., Mukhopadhyay, D. & Ramsay, J.G. (1971): Deformational and metamorphic history of Moinian and Lewisian rocks between Strathconon and Glen Affric. Scott. J. Geol., 6, pp 243-265.

Treagus, J.E. & Treagus, S.H. (1981): Folds and the strain ellipsoid: a general model. J. Struc. Geol., 3, pp 1-17.

Van Breemen, O., Pigeon, R.T. & Johnson, M.R.W. (1974): Precambrian and Palaeozoic pegmatites in the Moines of northern Scotland. J. Geol. Soc. Lond., 130, pp 493-507.

Van Breemen, O., Aftalion, M. & Johnson, M.R.W. (1979): Age of the Loch Borrolan complex, Assynt, and late movements along the Moine Thrust zone. J. Geol. Soc., 136, pp 489-497.

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 Fault. Scott. J. Geol., 15.

Verschure, R.H. et al. (1980): On the Thermal Stability of Rb-Sr and K-Ar Biotite Systems: Evidence from coexisting Sveconorwegian (ca. 870 Ma) and Caledonian (ca.400 Ma) Biotites in SW Norway. Contrib. Mineral. Petrol., 74, pp 245-252.

Vauchez, A. (1980): Ribbon texture and deformation mechanisms of quartz in a mylonitized granite of great Kabylie (Algeria). Tectonophysics, 67, pp 1-12.

Verbeek, A.A. & Schreiner, G.D.L. (1967): Variations in ^{39}K : ^{41}K ratio and movement of potassium in a granite-amphibolite contact region. Geochim. Cosmochim. Acta., 31, pp 2125-2133.

Walton, E.K. (1956): Lower Palaeozoic rocks - stratigraphy. Trans. Geol. Soc. Glasg., 22.

Watson, J.V. & Dunning, F.W. (1979): Basement-cover relations in the British Caledonides. In: A.L. Harris et al. The Caledonides of the British Isles - reviewed. Spec. Publ. Geol. Soc. Lond.

Watts, M.J. & Williams, G.D. (1979): Fault rocks as indicators of progressive shear deformation in the Guingamp region, Brittany. J. Struc. Geol., 1, pp 323-332.

Webster, R.K. (1960): Mass spectrometric isotope dilution analysis. Chapter in A.A. Smales & L.R. Wager "Methods in Geochemistry", Interscience Publishers Ltd.

White, S. (1973): Syntectonic recrystallization and texture development in quartz. Nature (Lond.), 244, pp 276-278.

White, S. (1976): The effects of strain on the microstructures, fabrics and deformation mechanisms in quartzites. Phil. Trans. Roy. Soc. Lond., 283A, pp 69-86.

White, S.H. (1979): Grain size and sub-grain size variation across a mylonite zone. Contrib. Mineral. Petrol., 70, pp 193-202.

White, S.H., Burrows, S.E. Carreras, J., Shaw, N.D. & Humphreys, F.J. (1980): On mylonites in ductile shear zones. J. Struc. Geol., 2, pp 175-187.

White, S.H., Evans, D.J. & Zhong, D-L. (1982): Fault rocks of the Moine Thrust zone: Microstructures and textures of selected mylonites. Tex & Microstruc., 5, pp 33-61.

Williams, G.D. (1978): Rotation of contemporary folds into the X-direction during overthrust processes in Laksefjord, Finmark. Tectonophys., 48, pp 29-40.

Wilson, D. (1975): Structure and Metamorphism of the Ben Wyvis district. Ph.D. thesis (Unpublished).

Wilson, C.J.L. & Bell, I.A. (1979): Deformation of Biotite and Muscovite: Optical Microstructure. Tectonophysics, 58, pp 179-200.

Wilson, D. & Shepherd, J. (1979): The Carn Chuinneag granite and its aureole. In: The Caledonides of the British Isles - reviewed.

Winchester, A.J. (1970): The geology of Fannich Forest. Ph.D. thesis (Unpublished), Oxford.

Winchester, J.A. (1974): The zonal pattern of regional metamorphism in the Scottish Caledonides. Q. J. Geol. Soc., 130, pp 509-524.

Winchester, J.A. (1973): Extensions to the Lewisian Inlier of Fannich Forest, Ross-shire. Proc. Geol. Assoc., 84, pp 273-282.

Winchester, J.A. (1973): Pattern of regional metamorphism suggests a sinistral displacement of 160km along the Great Glen Fault. Nature (Physical Science), 246, pp 81-84.

Winchester, J.A. (1974): Three dimensionsl pattern of polyphase metamorphism in the Moinian assemblage of northern Ross-shire. J. Geol., 82, pp 637-649.

Winchester, J.A. (1974): The Petrology of Moinian Calc-Silicate Gneisses from Fannich Forest, and their Significance as indicators of Metamorphic Grade. J. Petrol., 13, pp 405-424.

Winchester, J.A. (1974): The control of the whole rock content of CaO, and Al_2O_3 on the occurrence of the aluminosilicate polymorphs in amphibolite facies pelites. Geol. Mag., 111, pp 205-211.

Winchester, J. (1976): Different Moine amphibolite suites in northern Ross-shire. Scott. J. Geol., 12, pp 187-204.

Winchester, J.A. & Whittles, K.H. (1979): The pattern of three-dimensional metamorphism in the Killin area, Inverness-shire; a direct method of determining the thickness of metamorphic zones in the Caledonides. In: The Caledonides of the British Isles - reviewed. (Ed: A.L. Harris et al.).

Winkler, H.G.F. & Von Platen, H. (1961): Geochim. Cosmochim. Acta., 24, pp 48-69, 250-259.

Winkler, H.J.F. (1976): Petrogenesis of Metamorphic Rocks. 4th Edition. Springer-Verlag.

Wooley, A.R. (1970): The structural relationships of the Loch Borralan complex, Scotland. Geol. J., 7, pp 171-182.

York, D. (1966): Least-squares fitting of a straight line. Can.J.Physics, 44, pp 1079-1086. [see also Williamson, J.H., Can.J.Physics, 46, pp 1945-1847, 1968]

*

Brook,M.Powell,D.Brewer,M.S.(1976):Grenville age for rocks in the Moine of North-Western Scotland.
Nature 260 pp515-517.

Brown,P.E.Millar,J.A.Grasty,R.L.(1968):Isotopic ages of late Caledonian granitic intrusions in the British Isles.
Proc. York.Geol Soc 36 pp251-276.

Faure,G.(1977)Principles of Isotope Geology.
Publ.Wiley.

Hutton,D.H.W.(1979):Tectonic slides,a review and reappraisal.
Earth-Sci.Rev.15.pp151-172.

Read,H.H.(1931):The Geology of Sutherland,Northern Scotland.
Mem.Geol.Surv.Scot.

APPENDIX 1A

GRAINSIZES 125-200μ AND 75-125μ

K-Ar DATA

Sample wt	$^{40}Ar/^{38}Ar$	$^{36}Ar/^{38}Ar$	Atm Ar	Spike ^{38}Ar	Radiogenic ^{40}Ar	Age (2σ)
g	x 1	x 10^{-4}	%	x 10^{-9} l	x 10^{-9} l	x 10^{6} years
	SK14 Muscovite (NH202 695)			K = 8.08±0.08%		
0.5254	2.235±0.0031	2.342±0.049	2.97	35.705	147.40	417.25±10.60
0.4020	1.7509±0.0006	2.234±0.018	3.61	35.407	148.64	420.40±10.62
	SK19 Muscovite (NH202 696)			K = 8.10±0.08%		
0.4994	1.9669±0.0002	1.572±0.016	2.21	39.179	150.88	425.24±10.72
0.5109	1.8895±0.0015	3.039±0.030	4.60	42.219	148.95	420.38±10.63
0.4976	1.8225±0.0001	2.342±0.032	3.64	41.701	147.16	415.87±10.52
0.5093	1.8434±0.0006	1.111±0.026	1.62	41.517	147.82	417.55±10.55
	SKRD31 Muscovite (NH109 753)			K = 7.31±0.07%		
0.5011	1.9194±0.0003	1.963±0.034	2.87	36.997	137.64	429.33±10.82
0.5016	2.1406±0.0006	6.037±0.037	8.21	36.063	141.26	429.35±11.05
	SKRD34 Muscovite (NH185 807)			K = 8.30±0.08%		
0.4956	2.2290±0.0013	1.123±0.026	1.36	35.356	156.86	430.77±10.85
0.4976	1.9260±0.0006	1.090±0.030	1.52	41.061	156.50	431.74±10.87
	SK36 Muscovite (NH238 665)			K = 7.56±0.08%		
0.5045	1.8905±0.0002	5.195±0.033	7.98	41.451	142.92	430.89±10.86
0.5114	1.7654±0.0021	2.296±0.017	3.62	42.336	140.76	428.11±10.82
	SK49 Muscovite (NH206 676)			K = 8.25±0.08%		
0.5025	2.0242±0.0004	5.670±0.023	8.15	41.680	154.21	426.57±10.76
0.5078	1.9948±0.0004	3.616±0.045	5.22	41.565	154.75	427.90±10.79

Sample wt	$^{40}Ar/^{38}Ar$	$^{36}Ar/^{38}Ar$	Atm Ar	Spike ^{38}Ar	Radiogenic ^{40}Ar	Age (2σ)
g	x 1	x 10^{-4}	%	x 10^{-9} l	x 10^{-9} l	x 10^{6} years
	SKRD49 Muscovite (NH168 836)			K = 755±0.08%		
0.5039	1.9893±0.0006	1.515±0.034	2.11	36.581	141.37	427.37±10.77
0.5439	2.262±0.0005	2.333±0.035	2.92	35.755	144.36	435.41±10.95
	SK54 Muscovite (NH207 668)			K - 6.65±0.07%		
0.5509	1.6478±0.0004	1.266±0.026	2.09	41.299	120.93	416.23±10.52
0.5066	1.5467±0.0004	1.231±0.046	2.16	41.670	124.46	427.05±10.78
0.4991	1.4868±0.0004	1.092±0.024	1.98	41.396	120.87	416.05±10.52
0.5002	1.5490±0.0002	1.029±0.030	1.78	41.149	125.16	429.17±10.81
	SK72 Muscovite (NH119 665)			K = 7.44±0.07%		
0.5789	2.3135±0.0011	2.252±0.034	2.75	35.780	139.05	426.30±10.75
0.5014	1.9870±0.0008	2.383±0.014	3.40	35.432	135.63	416.94±10.54
	SK108 Muscovite (NH104 747)			K = 6.34±0.06%		
0.6012	1.9997±0.0006	2.034±0.023	1.11	37.128	119.95	431.49±10.86
0.5024	1.7430±0.0003	1.333±0.019	2.09	36.319	123.36	442.35±11.10
0.5124	1.8225±0.0008	3.505±0.035	5.53	35.457	119.13	428.85±10.82
	SK207 Muscovite (NH203 691)			K = 7.20±0.007%		
0.5124	1.6103±0.0003	2.473±0.039	4.36	44.360	133.32	423.00±10.68
0.5003	1.6395±0.0004	1.986±0.030	3.41	41.639	131.79	418.67±10.58

Sample wt	$^{40}Ar/^{38}Ar$	$^{36}Ar/^{38}Ar$	Atm Ar	Spike ^{38}Ar	Radiogenic ^{40}Ar	Age (2σ)
g	x 1	$\times 10^{-4}$	%	$\times 10^{-9}$l	$\times 10^{-9}$l	$\times 10^{6}$ years
	SK14 Biotite (NH202 695)			K = 7.22±0.07%		
0.5446	1.9849±0.0014	0.795±0.034	1.04	37.049	133.62	422.50±10.83
0.5151	1.9261±0.0002	1.827±0.012	2.65	36.659	133.43	421.96±10.80
0.4836	1.7706±0.0001	1.06±0.031	1.61	36.633	131.96	417.81±10.71
0.6134	2.353±0.0006	1.732±0.036	2.05	35.556	133.59	422.42±10.82
	SK19 Biotite (NH202 696)			K - 7.45±0.07%		
0.5628	1.8257±0.0002	0.984±0.023	1.41	41.527	135.09	415.15±10.49
0.5006	1.6919±0.0005	1.8629±0.016	3.08	41.833	137.02	420.45±10.62
	SKRD31 Biotite (NH109 753)			K = 6.31±0.06%		
0.4402	1.5341±0.0005	1.505±0.029	2.71	35.581	120.64	435.20±10.95
0.5053	1.7312±0.0006	1.111±0.032	1.73	35.531	119.42	431.95±10.88
	SKRD34 Biotite (NH185 807)			K = 7.43±0.07%		
0.5616	2.1513±0.0025	0.562±0.018	0.64	36.427	138.65	425.94±10.77
0.5273	1.9812±0.0002	1.029±0.037	1.39	36.344	134.66	414.97±10.49
0.5423	2.0878±0.0002	0.913±0.032	1.15	35.308	134.36	414.14±10.47
	SK36 Biotite (NH238 665)			K = 7.80±0.08%		
0.5433	1.9922±0.0004	2.3407±0.018	3.33	41.489	147.06	429.85±10.82
0.5146	1.8206±0.0015	2.077±0.036	3.21	42.297	144.82	424.02±10.71
0.5052	1.7682±0.0003	0.624±0.024	0.88	41.456	143.82	421.39±10.63
	SK49 Biotite (NH206 676)			K = 7.32±0.07%		
0.4993	1.8119±0.0004	6.549±0.029	10.54	41.795	135.68	423.38±10.70
0.5019	1.6862±0.001	2.216±0.021	3.71	41.718	134.95	421.34±10.63

Sample wt g	$^{40}Ar/^{38}Ar$ x 1	$^{36}Ar/^{38}Ar$ x 10^{-4}	Atm Ar %	Spike ^{38}Ar x 10^{-9}l	Radiogenic ^{40}Ar x 10^{-9}l	Age (2σ) x 10^{6} years
	SKRD49 Biotite (NH168 846)			K = 7.52±0.08%		
0.4917	1.9228±0.00009	0.946±0.015	1.35	35.730	137.84	419.33±10.59
0.5413	2.1316±0.0004	0.903±0.012	1.11	36.401	141.75	429.89±10.82
	SK54 Biotite (NH207 668)			K = 6.54±0.07%		
0.5032	1.5013±0.0003	1.141±0.036	2.05	41.413	121.01	423.30±10.69
0.5292	1.5391±0.0003	0.841±0.025	1.43	41.609	119.28	417.33±10.55
	SK64 Biotite (NH154 874)			K = 6.34±0.06%		
0.5061	1.6260±0.0004	0.945±0.042	1.54	36.762	116.28	419.53±10.61
0.4953	1.6076±0.0009	0.928±0.019	1.52	35.830	114.52	413.84±10.47
0.4992	1.7153±0.0032	2.449±0.046	4.05	35.631	117.46	420.36±10.74
	SKRD74 Biotite (NH156 895)			K = 5.31±0.05%		
0.1148	1.2683±0.01	4.850±0.05	11.30	1.048	102.71	439.98±11.07
	SK72 Biotite (NH199 665)			K = 6.70±0.07%		
0.4995	1.7015±0.0022	0.722±0.043	1.08	37.102	125.01	425.30±10.90
0.4983	1.7144±0.0008	0.893±0.044	1.37	36.088	122.46	417.55±10.68
0.4864	1.743±0.0007	1.866±0.024	3.00	35.655	123.86	421.81±10.68
	SKRD108 Biotite (NH203 691)			K = 5.83±0.06%		
0.5001	1.6176±0.0012	0.553±0.027	0.83	36.737	117.83	457.18±11.45
0.5556	1.7798±0.0004	0.612±0.036	0.85	36.685	116.51	452.62±11.33
0.4894	1.4958±0.0005	1.171±0.040	2.12	35.680	106.74	418.75±10.59
		(some sample lost)				
0.2078	0.7239±0.0002	0.225±0.0020	8.79	35.234	111.94	436.86±11.01
	SK207 Biotite (NH203 691)			K = 7.24±0.07%		
0.5158	1.7101±0.0003	1.795±0.016	2.93	41.375	133.14	420.41±10.61
0.5238	1.7052±0.0005	0.734±0.038	1.10	41.548	133.76	422.15±10.66

APPENDIX 1B

GRAINSIZES 200-250μ

K-Ar DATA

Sample wt g	$^{40}Ar/^{38}Ar$ x 1	$^{36}Ar/^{38}Ar$ x 10^{-4}	Atm Ar %	Spike ^{38}Ar x 10^{-9}1	Radiogenic ^{40}Ar x 10^{-9}1	Age (2σ) x 10^{6} years
	SK3 Muscovite (NH220 667)			K = 8.26±0.08%		
0.5058	1.9131±0.0014	6.446±0.03	9.82	46.55	158.79	437.53±11.03
	SK14 Muscovite (NH202 695)			K = 6.20±0.06%		
0.5490	1.5100±0.0010	4.655±0.032	8.93	46.253	115.84	426.41±10.79
0.5025	1.4504±0.0003	1.0932±0.023	2.03	41.295	116.77	429.45±10.81
	SK19 Muscovite (NH202 696)			K = 5.88±0.06%		
0.5272	1.4026±0.0004	4.021±0.055	8.27	36.198	112.73	436.29±11.02
	SK23 Muscovite (NH237 667)			K = 8.32±0.08%		
0.5222	1.9370±0.0006	5.683±0.058	8.53	46.24	156.95	430.07±10.86
0.5077	1.8821±0.0006	3.25±0.032	4.95	45.662	160.88	440.09±11.06
	SK25 Muscovite (NH229 668)			K = 8.80±0.09%		
0.5163	1.9212±0.0014	6.365±0.07	6.95	46.348	155.80	406.39±10.36
		(some sample lost)				
0.5092	1.9994±0.0009	3.634±0.07	5.23	45.758	170.26	439.85±11.08
	SK36 Muscovite (NH238 665)			K = 7.18±0.07%		
0.4993	1.8418±0.0004	7.372±0.029	11.69	45.923	149.59	469.74±11.73
0.5930	1.9602±0.0011	1.306±0.016	1.82	41.426	134.44	427.21±10.77
0.5341	1.7512±0.0004	0.786±0.015	1.16	41.335	133.95	425.82±10.73
	SK43 Muscovite (NH249 662)			K = 6.42±0.06%		
0.5024	1.4135±0.0004	3.635±0.034	7.40	45.868	119.48	425.52±10.71
0.4951	1.4010±0.0003	3.837±0.032	7.90	46.032	119.96	426.72±10.78

Sample wt g	$^{40}Ar/^{38}Ar$ x 1	$^{36}Ar/^{38}Ar$ x 10^{-4}	Atm Ar %	Spike ^{38}Ar x 10^{-9}l	Radiogenic ^{40}Ar x 10^{-9}l	Age (2σ) x 10^{6} years
	SK47 Muscovite (NH249 661)			K = 7.97±0.08%		
0.5006	1.7395±0.0006	4.506±0.053	7.50	46.208	148.52	425.39±10.76
0.4958	1.6989±0.0009	1.356±0.032	2.11	45.554	155.47	443.04±11.12
	SK49 Muscovite (NH206 676)			K = 8.78±0.09%		
0.5180	2.1103±0.0004	2.693±0.030	3.64	41.032	161.07	419.50±10.60
0.5106	1.9188±0.0071	6.648±0.039	10.10	46.585	162.16	421.86±10.67
	SK51 Muscovite (impure) (NH206 675)			K - 2.01±0.03%		
0.4945	0.4937±0.0005	3.457±0.025	20.13	45.950	36.63	418.61±13.66
	SK54 Muscovite (NH207 668)			K = 6.34±0.06%		
0.4992	1.3506±0.0007	3.793±0.052	8.09	46.225	114.93	415.05±10.55
0.5035	1.4360±0.0008	5.561±0.051	11.26	46.005	116.43	419.87±10.67
	SK65 Muscovite (NH203 668)			K = 7.90±0.08%		
0.5116	1.6572±0.0004	4.219±0.051	7.36	46.320	138.99	404.36±10.28
	SK66 Muscovite (NH203 668)			K = 6.66±0.07%		
0.5003	1.4446±0.0005	3.848±0.022	7.68	46.142	122.98	421.98±10.75
	SK72 Muscovite (NH199 665)			K = 6.90±0.07%		
0.5061	1.4906±0.0024	4.448±0.039	8.64	46.724	127.01	420.77±10.65
	SK207 Muscovite (NH203 691)			K = 7.27±0.07%		
0.5037	1.5809±0.0006	4.704±0.064	8.62	46.403	133.08	418.68±10.63
0.5088	1.7087±0.0004	0.970±0.025	1.51	41.119	136.01	426.88±10.76
	SK209 Muscovite (NH221 683)			K = 6.94±0.07%		
0.4968	4.7656±0.0038	111.277±0.113	69.06	45.690	135.59	443.66±16.90

Sample wt g	$^{40}Ar/^{38}Ar$ x 1	$^{36}Ar/^{38}Ar$ x 10^{-4}	Atm Ar %	Spike ^{38}Ar x 10^{-9}l	Radiogenic ^{40}Ar x 10^{-9}l	Age (2σ) x 10^{6} years
	SK3 Biotite (NH220 667)			K = 7.25±0.07%		
0.4868	1.5223±0.0023	5.533±0.029	10.57	47.717	133.44	420.88±10.72
0.5025	1.5018±0.0007	3.492±0.027	6.69	46.236	128.93	408.52±10.36
		(some sample lost)				
0.5031	1.5691±0.0002	4.666±0.022	8.61	46.758	133.26	420.38±10.62
0.5010	1.6430±0.0022	0.937±0.029	1.51	41.090	132.71	418.83±10.58
	SK14 Biotite (NH202 695)			K = 7.63±0.08%		
0.4949	1.5616±0.0004	2.576±0.016	4.69	47.629	143.22	428.36±10.79
0.5017	1.6239±0.00003	4.522±0.045	8.06	47.787	139.22	417.68±10.58
	SK19 Biotite (NH202 696)			K = 7.29±0.07%		
0.5180	1.5260±0.0005	1.689±0.03	3.08	47.599	135.89	425.54±10.74
0.5007	1.6448±0.0005	4.805±0.086	8.47	46.170	138.81	433.65±11.00
	SK23 Biotite (NH237 667)			K = 7.21±0.07%		
0.5015	1.3955±0.0004	0.597±0.017	1.05	47.687	141.29	416.77±10.52
0.5106	1.5701±0.00009	3.966±0.016	7.29	46.873	144.62	423.36±10.69
	SK25 Biotite (NH228 668)			K = 7.86±0.08%		
0.5020	1.5374±0.0003	1.36±0.033	2.43	47.658	142.41	415.09±10.50
0.4993	1.6762±0.0001	4.728±0.069	8.17	46.375	142.95	416.75±10.57
	SK36 Biotite (NH235 665)			K = 7.25±0.07%		
0.5112	1.6196±0.0007	4.691±0.113	8.39	46.087	133.75	421.61±10.79
0.4975	1.5603±0.0002	3.750±0.038	6.93	47.454	138.51	434.94±10.96
0.5064	1.6631±0.0004	0.890±0.031	1.41	41.244	133.54	421.01±10.63

Sample wt g	$^{40}Ar/^{38}Ar$ x 1	$^{36}Ar/^{38}Ar$ x 10^{-4}	Atm Ar %	Spike ^{38}Ar x 10^{-9}l	Radiogenic ^{40}Ar x 10^{-9}l	Age (2σ) x 10^6 years
	SK43 Biotite (NH249 662)			K = 6.91±0.07%		
0.4998	1.5309±0.0002	4.099±0.033	7.73	47.367	133.86	440.31±11.07
0.4998	1.6246±0.0007	1.142±0.043	1.90	41.2137	131.42	433.18±10.92
	SK47 Biotite (NH249 661)			K = 6.75±0.07%		
0.5012	1.4959±0.00009	3.463±0.033	6.66	47.338	131.88	440.77±11.08
0.5016	1.5103±0.0009	1.6910±0.005	3.12	45.786	133.55	448.70±11.29
0.5010	1.5951±0.0008	1.190±0.012	2.02	41.266	128.72	434.22±10.92
	SK49 Biotite (NH206 676)			K = 7.21±0.07%		
0.4985	1.47±0.0002	9.761±0.048	1.32	45.895	134.12	419.57±10.62
0.4999	1.6131±0.0002	4.100±0.027	7.26	47.425	141.91	441.16±11.09
	SK41 Biotite (NH206 675)			K = 6.95±0.07%		
0.5049	1.4937±0.0025	2.864±0.022	5.48	47.309	132.28	433.48±10.91
	SK54 Biotite (NH207 668)			K = 6.70±0.07%		
0.4911	1.4635±0.0001	5.043±0.017	10.00	46.902	125.79	428.24±10.80
0.5058	1.4749±0.0004	3.448±0.052	6.72	45.978	125.05	423.84±10.73
0.5020	1.4884±0.00003	1.406±0.026	2.54	41.184	120.23	409.12±10.36
	SK65 Biotite (NH203 558)			K = 7.12±0.07%		
0.5010	1.5915±0.0002	4.193±0.017	7.61	47.396	139.09	443.62±11.14
0.5181	1.5840±0.0002	3.884±0.032	7.07	46.930	133.32	427.24±10.77
	SK66 Biotite (NH203 668)			K = 3.81±0.04%		
0.5101	0.8204±0.0004	3.349±0.028	11.72	47.483	67.41	406.61±10.36

Sample wt g	$^{40}Ar/^{38}Ar$ x 1	$^{36}Ar/^{38}Ar$ x 10^{-4}	Atm Ar %	Spike ^{38}Ar x 10^{-9}l	Radiogenic ^{40}Ar x 10^{-9}l	Age (2σ) x 10^{6} years
	SK72 Biotite (NH199 665)			K = 5.90±0.06%		
0.4991	1.2929±0.0002	5.068±0.017	11.37	47.542	109.14	421.78±10.67
	SK80 Biotite (NH194 674)			K = 7.50±0.08%		
0.5011	1.5642±0.0005	3.629±0.045	6.68	47.513	138.40	421.89±10.67
	SK207 Biotite (NH203 691)			K = 7.24±0.07%		
0.5030	1.6249±0.0005	0.667±0.027	1.03	41.325	132.06	417.36±10.55
	SK209 Biotite (NH221 683)			K = 7.15±0.07%		
0.4906	1.4909±0.0005	1.031±0.076	1.85	45.636	130.11	433.54±10.94

APPENDIX 1C

GRAINSIZES 250-500μ

K-Ar DATA

Sample wt g	$^{40}Ar/^{38}Ar$ x 1	$^{36}Ar/^{38}Ar$ x 10^{-4}	Atm Ar %	Spike ^{38}Ar x 10^{-9}l	Radiogenic ^{40}Ar x 10^{-9}l	Age (2σ) x 10^{6} years
	SK14 Muscovite	(NH202 695)		K = 8.69±0.09%		
0.5099	2.1253±0.0026	3.606±0.068	4.88	40.5030	160.67	422.44±10.73
0.4984	2.1469±0.0007	3.418±0.016	4.58	39.638	162.92	427.70±10.78
0.5020	2.2717±0.0015	5.383±0.021	6.88	39.260	165.42	433.53±10.92
	SK36 Muscovite	(NH238 665)		K = 7.63±0.08%		
0.6653	2.3986±0.0031	2.428±0.080	2.87	40.586	142.11	425.20±10.79
0.5320	2.0223±0.0004	1.465±0.023	2.00	39.611	147.60	439.76±11.04
0.4988	1.9356±0.0006	2.093±0.027	3.05	39.233	147.59	439.76±11.05
	SK49 Muscovite	(NH206 676)		K = 8.95±0.09%		
0.4988	2.1481±0.0010	2.994±0.025	3.99	40.922	169.20	430.88±10.85
0.5134	2.340±0.0004	4.509±0.038	5.58	39.584	170.33	433.44±10.91
0.5021	2.2302±0.0020	2.180±0.031	2.76	39.124	168.97	430.36±10.86
	SK54 Muscovite	(NH207 668)		K = 6.85±0.07%		
0.5015	1.7366±0.0005	3.689±0.017	6.12	40.175	130.60	434.14±10.93
0.5080	1.7665±0.0001	1.216±0.016	1.87	39.557	134.97	447.00±11.20
0.5030	1.7770±0.0017	1.011±0.017	1.50	39.502	137.46	454.28±11.39
0.5065	1.9673±0.0003	2.297±0.017	3.31	35.283	132.50	437.21±10.99
	SK72 Muscovite	(NH199 665)		K = 7.16±0.07%		
0.5033	1.7594±0.0004	4.724±0.058	7.78	40.308	129.93	415.44±10.54
0.4988	1.7476±0.00005	1.121±0.018	1.73	39.475	135.91	432.44±10.88
0.4997	1.7899±0.0017	2.423±0.019	3.84	39.094	134.64	428.85±10.83

Sample wt	$^{40}Ar/^{38}Ar$	$^{36}Ar/^{38}Ar$	Atm Ar	Spike ^{38}Ar	Radiogenic ^{40}Ar	Age (2σ)
g	x 1	x 10^{-4}	%	x 10^{-9}l	x 10^{-9}l	x 10^{6} years
	SK207 Muscovite (NH203 691)			K = 7.84±0.08%		
0.4987	1.9587±0.0002	2.250±0.038	3.25	39.530	150.20	436.02±10.97
0.5031	1.9534±0.0007	1.282±0.031	1.79	39.448	150.41	436.54±10.98
0.5057	1.9780±0.0003	1.45±0.033	2.02	39.067	149.71	434.75±10.94

Sample wt g	$^{40}Ar/^{38}Ar$ x 1	$^{36}Ar/^{38}Ar$ x 10^{-4}	Atm Ar %	Spike ^{38}Ar x 10^{-9} l	Radiogenic ^{40}Ar x 10^{-9} l	Age (2σ) x 10^{6} years
	SK14 Biotite (NH202 695)			K = 6.96±0.07%		
0.4998	1.7373±0.0003	3.887±0.047	6.45	40.502	131.69	431.21±10.88
0.4930	1.7738±0.004	3.431±0.049	5.56	40.419	137.33	447.57±11.24
0.5088	1.8349±0.0003	5.334±0.049	8.44	40.363	133.26	435.78±10.98
0.5000	1.8304±0.0002	3.653±0.024	5.71	39.692	136.99	446.60±11.20
	SK36 Biotite (NH238 665)			K = 7.23±0.07%		
0.4951	1.7601±0.0004	4.271±0.053	7.01	40.558	134.06	423.52±10.71
0.5030	1.7265±0.0005	2.353±0.062	3.86	40.316	134.01	423.37±10.71
0.5198	1.8987±0.0005	3.635±0.021	5.51	39.732	137.13	432.12±10.88
	SK49 Biotite (NH206 676)			K = 7.36±0.07%		
0.5231	1.8201±0.0007	0.823±0.047	1.18	40.894	140.60	434.90±10.95
0.5198	1.9159±0.0004	2.868±0.015	4.28	39.787	140.37	433.74±10.91
	SK54 Biotite (NH207 668)			K = 6.80±0.07%		
0.5356	1.8397±0.0002	3.166±0.027	4.93	39.870	130.19	435.75±10.96
0.4827	1.6102±0.0004	2.002±0.025	3.50	39.760	127.99	429.19±10.81
	SK72 Biotite (NH199 665)			K = 6.73±0.07%		
0.5004	1.6506±0.0002	4.334±0.050	7.59	40.335	122.94	417.91±10.59
0.4846	1.6245±0.0005	1.393±0.022	2.36	39.843	130.41	44-.43±11.06
0.5862	2.0241±0.0009	4.228±0.031	6.04	39.705	128.82	435.65±10.97
0.5117	1.7288±0.0006	1.395±0.025	2.56	39.665	130.57	440.91±11.07
	SK207Biotite (NH203 691)			K = 7.17±0.07%		
0.5030	1.7266±0.0004	1.196±0.033	1.88	40.147	135.21	429.93±10.83
0.5518	1.9247±0.0013	0.965±0.042	1.33	39.677	136.55	433.71±10.92

APPENDIX 1D

Rb-Sr AND K-Ar DATA FOR THE 'OLD' PEGMATITES (SECTION GC.5)

K-Ar ages measured on the same crystal as corresponding Rb-Sr age

Sample wt g	$^{40}Ar/^{38}Ar$ x 1	$^{36}Ar/^{38}Ar$ x 10^{-4}	Atm Ar %	Spike ^{38}Ar x 10^{-9} l	Radiogenic ^{40}Ar x 10^{-9} l	Age (2σ) x 10^{6} years
	SK35 Muscovite					
0.0140	1.2024±0.0001	1.05±0.02	23.39	35.209	231.42	582.8±15.71
	SK42 Muscovite					
0.084	0.3966±0.00005	1.796±0.02	12.66	35.184	145.07	443.47±11.25
	K-Ar age from 200-250μ separate.					
	SK35 Muscovite		K = 8.61±0.08%			
0.5110	1.9892±0.00045	4.284±0.06	6.23	45.608	166.48	439.60±11.07

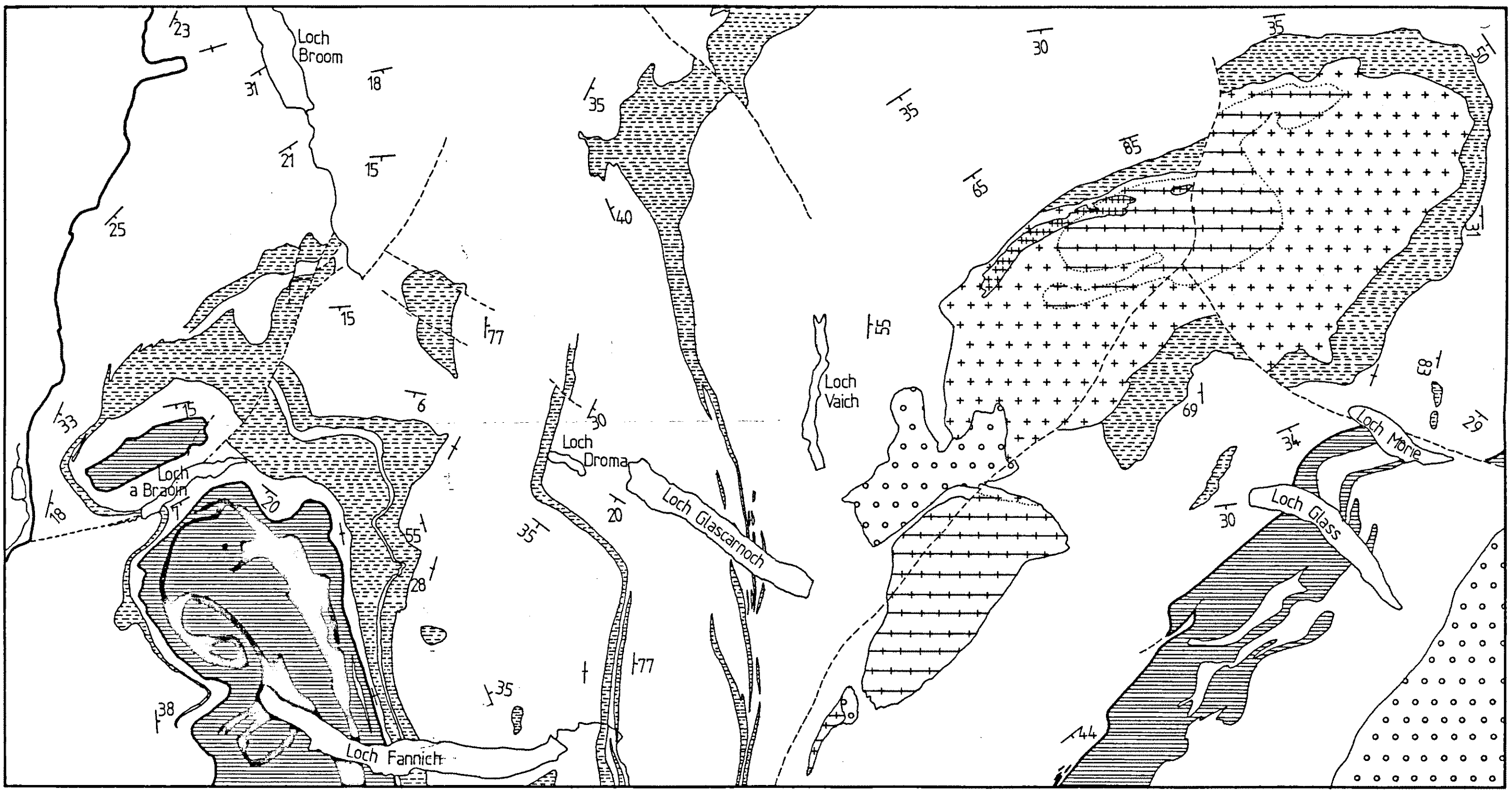
Sample	Rb (ppm)	Sr (ppm)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$	Age (2σ)	Initial $^{87}Sr/^{86}Sr$
SK42 Muscovite	247.63	167.27	4.298 ± 0.022	0.74165 ± 0.00007	457 ± 6 Ma	0.7137 ± 0.0001
Plagioclase	6.04	766.96	0.0228 ± 0.0001	0.71382 ± 0.00007		
SK35 Muscovite	228.9	47.21	14.156 ± 0.071	0.80402 ± 0.00004		
Assumed initial ratio	-	-	0.00	0.00	448 ± 99 Ma	0.7137 ± 0.02

APPENDIX 2

Rb-Sr DATA

Sample	Rb (ppm)	Sr (ppm)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$	Age (2σ)	Initial $^{87}Sr/^{86}Sr$
SK 14 Muscovite	170.8	116.2	1.028 ± 0.05	0.73554 ± 0.00004		
Biotite	412.5	8.73	186.5 ± 0.9	1.79211 ± 0.0001	400 ± 4 Ma	0.727 ± 0.0001
SK 25 Muscovite	289.7	32.12	26.555 ± 0.13	0.89177 ± 0.00004		
Biotite	579.9	7.694	247.99 ± 1.24	2.19564 ± 0.00011	413 ± 5 Ma	0.7354 ± 0.0025
SK 54 Muscovite	163.7	220.3	2.491 ± 0.012	0.73556 ± 0.00004		
Biotite	415.2	13.18	96.21 ± 0.48	1.27833 ± 0.0004	407 ± 4 Ma	0.7211 ± 0.0002
SK72 Muscovite (125-200μ)	230.4	232.2	2.880 ± 0.014	0.73895 ± 0.00004		
Biotite	472.7	16.22	92.40 ± 0.46	1.26218 ± 0.00006	410 ± 4 Ma	0.72215 ± 0.0003
SK 72 Muscovite (200-250μ)	219.1	113.9	5.585 ± 0.028	0.74518 ± 0.00004		
Biotite	491.5	8.209	184.52 ± 0.92	1.80569 ± 0.00016	416 ± 4 Ma	0.7121 ± 0.0005
SK108 Muscovite	230.4	282.9	2.235 ± 0.011	0.74091 ± 0.00003		
Biotite	639.5	13.88	144.79 ± 0.72	1.58515 ± 0.0012	416 ± 4 Ma	0.7277 ± 0.0002

FIGURE 9 Map illustrating the area from the Moine Thrust to Carn Chuinneag



Lewisian gneisses	Morar psammites	Carn Chuinneag granite (and augen gneiss)
Glenfinnan migmatites	" pelites	Old Red sandstone

FIGURE 64.

Extension lineations being rotated from the Sgurr Beag slide to the Moine Thrust

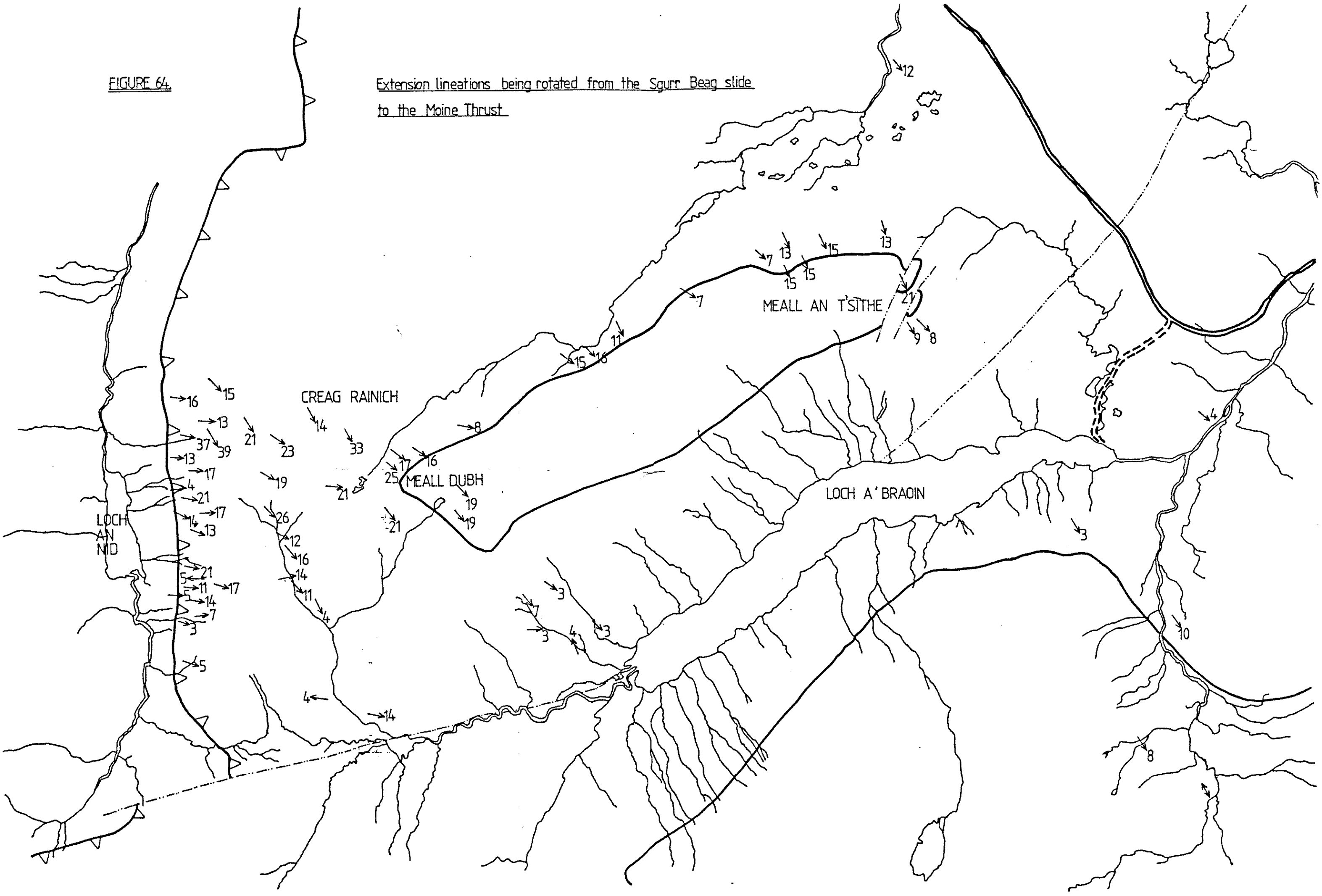
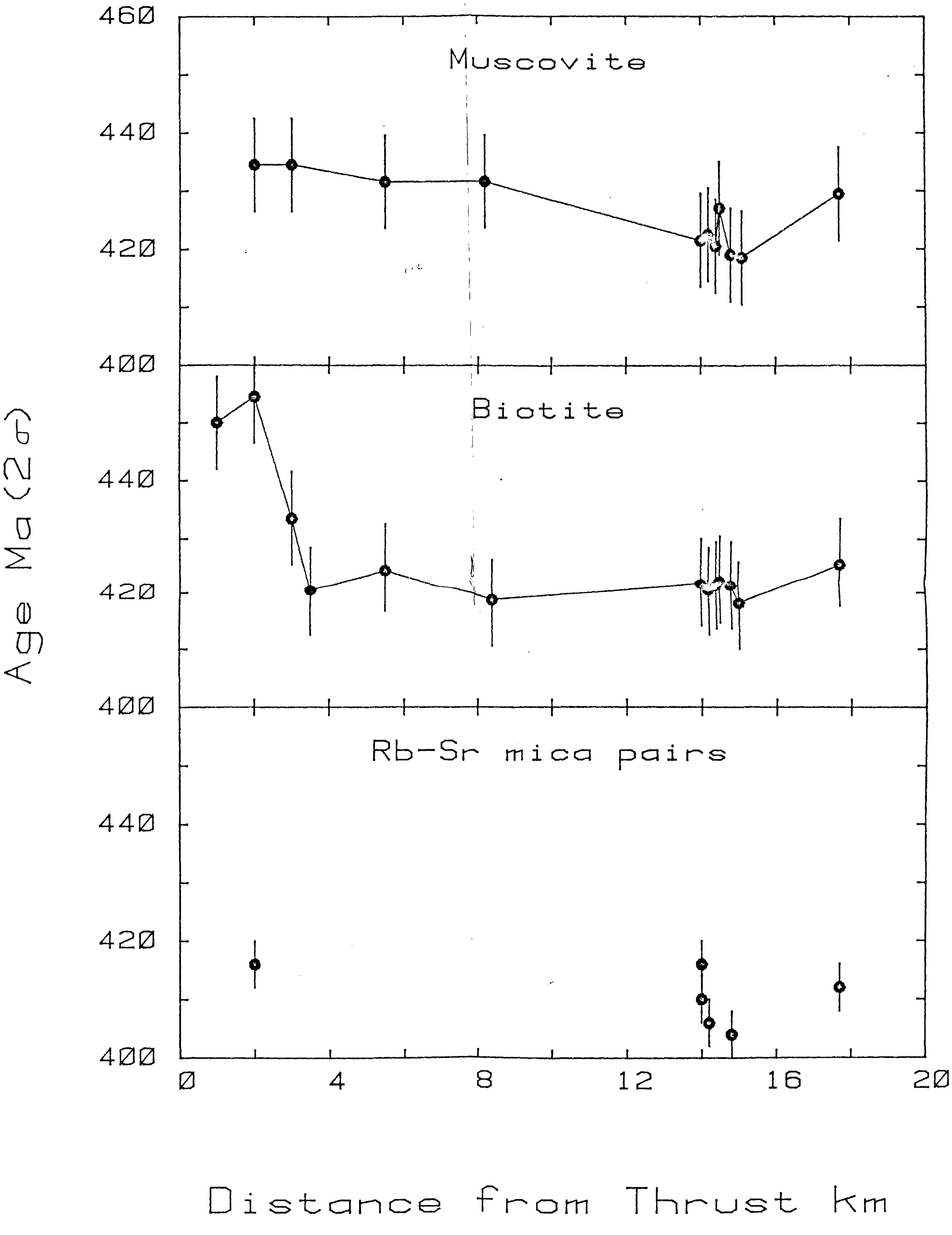


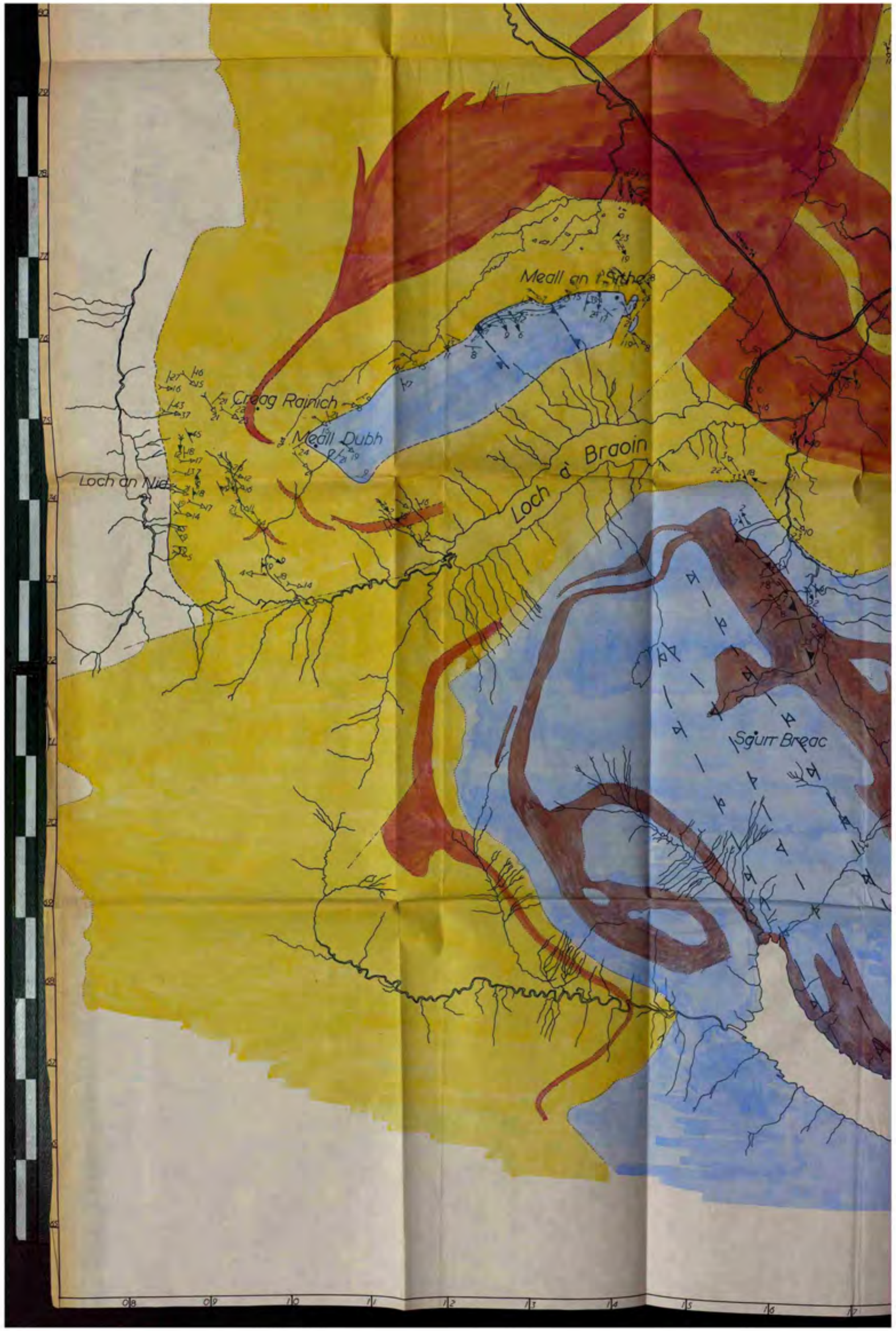
TABLE 1 Stratigraphy of the area between Fannich and the Moine Thrust

Peach, Horne, Gunn and Pocock 1913	Sutton and Watson 1954	Winchester 1970	Kelley 1984
(cut out against Moine Thrust)	(Youngest)	Upper Meall an t'sithe pelite	Meall an t'sithe pelite
No name given(semi-pelitic)	Fannich gneiss	Lewisian gneiss	Lewisian gneiss
		Basal conglomerate	
Sgurr Mor biotite schist (pelitic)	Meall an t'sithe pelite	Lower Meall an t'sithe pelite and Sgurr nan Each psammites	Meall an t'sithe pelite
			Sgurr Beag Slide
Meall a'Chrasgaidh rock (siliceous)	Meall a'Chrasgaidh siliceous group	Meall a'Chrasgaidh psammite	Meall a'Chrasgaidh psammite
Meall an t'sithe rock (pelitic)	Sgurr Mor pelitic group	Sgur Mor pelite and Carn na Criche psammite	Sgurr Mor pelite
Acid gneiss with Lewisian types	Inverbroom semi-pelitic	Inverbroom psammites	Inverbroom psammite (cut out against the Moine Thrust)

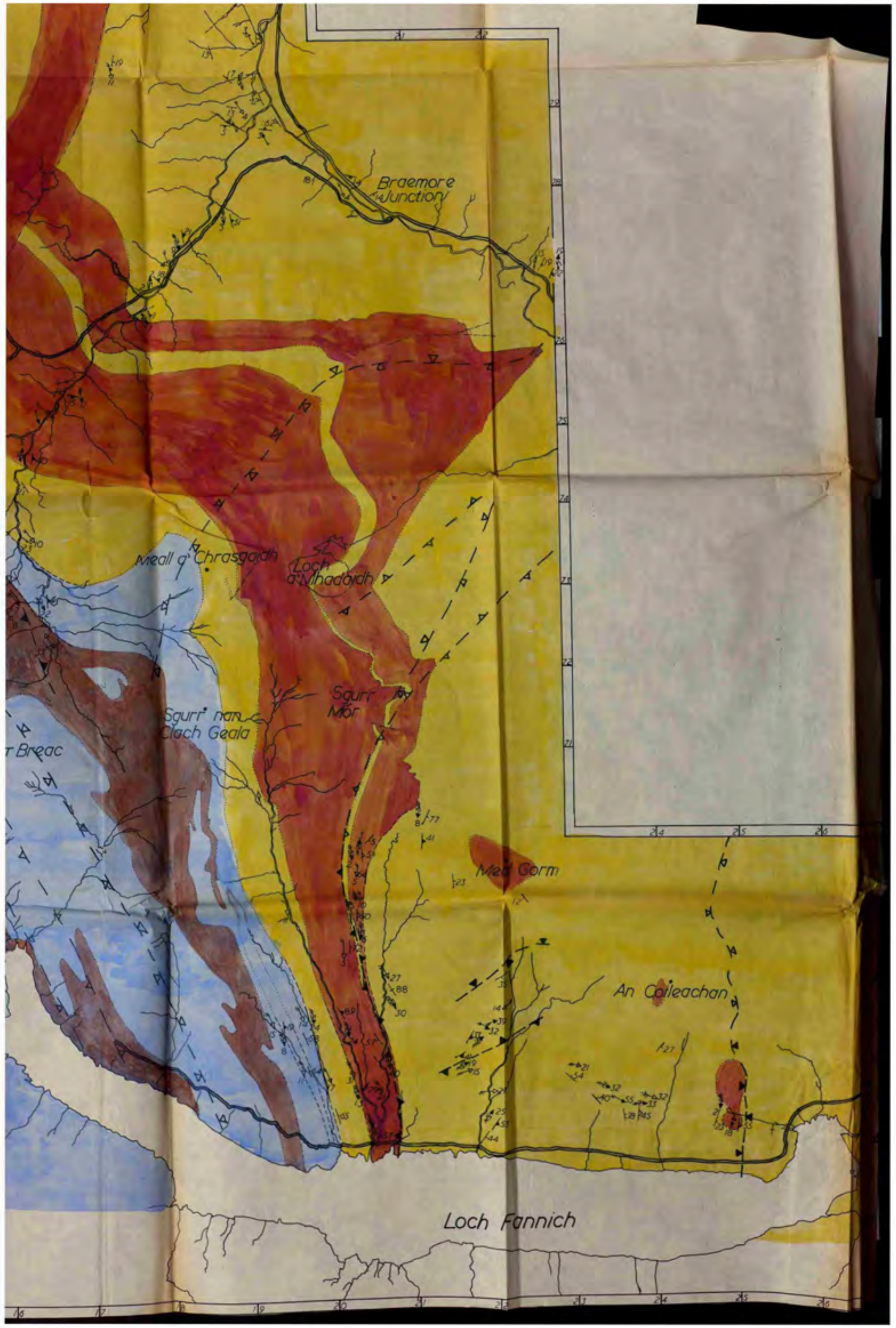
FIGURE 101.

Fannich-Ullapool traverse

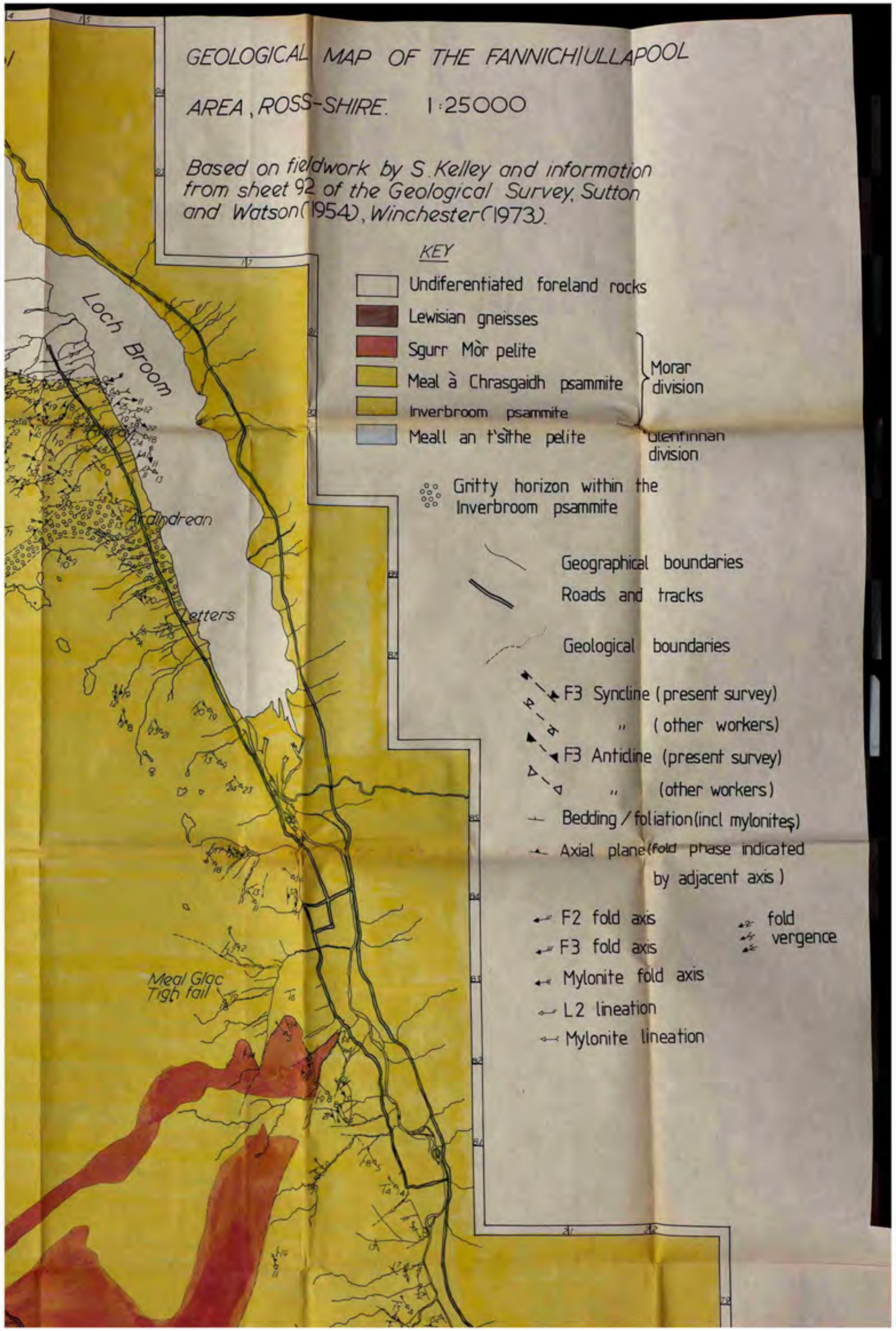




Creag Rainich
Meall Dubh
Loch a' Braoin
Sgurr Breac


Braemore Junction
Meall a' Chrasgaidh
Loch a' Mhadaidh
Sgurr Mòr
Sgurr nan Clach Geala
Breac
Meall Gorm
An Coileachan
Loch Fannich


GEOLOGICA
AREA , ROS
Based on t
from sheet
and Watson
Ullapool
Loch Broom
Little
Loch
Broom
Dundonnell
Ardindrean
Letters
Meal Glac
Tigh fail


GEOLOGICAL MAP OF THE FANNICH/ULLAPOOL AREA, ROSS-SHIRE. 1:25000
Based on fieldwork by S. Kelley and information from sheet 92 of the Geological Survey, Sutton and Watson (1954), Winchester (1973).
KEY
Undiferentiated foreland rocks
Lewisian gneisses
Sgurr Mòr pelite
Meal à Chrasgaidh psammite
Inverbroom psammite
Morar division
Meall an t'sìthe pelite
Glenfinnan division
Gritty horizon within the Inverbroom psammite
Geographical boundaries
Roads and tracks
Geological boundaries
F3 Syncline (present survey)
" (other workers)
F3 Anticline (present survey)
" (other workers)
Bedding / foliation (incl mylonites)
Axial plane (fold phase indicated by adjacent axis)
F2 fold axis
F3 fold axis
Mylonite fold axis
L2 lineation
Mylonite lineation
fold vergence
Loch Broom
Letters
Meal Glac Tigh fail