THE GEOLOGY AND GEOCHEMISTRY OF THE

METASEDIMENTARY ROCKS OF THE LOCH LAGGAN -

UPPER STRATHSPEY AREA, INVERNESS-SHIRE

VOL I

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ABSTRACT

An area of 120Km^2 extending from Loch Laggan to the upper reaches of the River Spey has been mapped in detail on a scale of 1:10,000. The metasediments cropping out in this area have been assigned to the Grampian Division and consist of varied assemblages of psammites, semi-psammites and semi-pelites which can be subdivided into two lithostratigraphic successions separated by a tectonic discontinuity, the Gairbeinn Slide.

Two lithostratigraphic formations were recognised in the lower Glenshirra Succession, while four formations as well as a transitional lithostratigraphic unit were recognised in the overlying Corrieyairack Succession.

Sedimentological analysis suggests that the sediments accumulated in environments varying from alluvial to marine and were mostly transported from the south to the north. The scattered calc-silicate pods and bands were produced by localised precipitation of carbonates during diagenesis.

Geochemical studies suggest that the original sandstones of the Glenshirra Succession were dominantly immature lithic arenites whereas those of the Corrieyairack Succession were greywackes. The semi-pelitic rocks of the Garva Formation appear to be the most chemically "immature". Provenance analysis suggest that their source rocks were upper and middle level crustal complexes composed of gneisses and granites.

The rocks in this area were subjected to three phases of ductile deformation followed by one of brittle movements including faults. Dl resulted in minor, isoclinal, recumbent folds and movement on the Gairbeinn Slide. D2 produced a major tight Laggan Antiform with generally, NE-SW trending axis, while D3 is represented by major crossfolds of the earlier structures on NW-SE axes.

Calc-silicate and pelitic phase assemblages indicate that middle amphibolite facies metamorphic conditions lasted from early Dl to post-D2 times, while retrograde conditions followed D3 folding. Localised contact metamorphism accompanied the intrusion of the Corrieyairack Granite Complex.

Other igneous rocks in the area include the early two-mica granites and pegmatites which were intruded syn-to late -D3. The remainder of the igneous rocks were emplaced later, and are post-orogenic and post-metamorphic.

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CHAPTER 1 : INTRODUCTION

1.1   Location and extent of the area

The area studied lies 40 km north east of Fort William and 20km
southeast of Fort Augustus (Fig 1.1). It is bounded in the south by Loch
Laggan and in the north by the River Spey. The eastern limit is defined by
the broad glen containing Loch Crunachdan and extending southwards to the
Loch Laggan Hotel (NN531897) while the western margin is a line extending
from Loch Spey (NN420938) to Moy Lodge (NN440834). The total area is
approximately 120 square kilometres.

1.2   Physiography

The topography of the area is dominated by the long N.E. - S.W. trending
ridge extending N.E. from Creag Meagaidh (NN419875). Rising over 1000m above
mean sea-level, it extends to Cam Dubh (NN516926) in the east by way of
Carn Liath (NN472904) and Puist Coire Ardair (NN432874) (Fig 1.2). The
elevation falls steeply from the top of this ridge to about 250m on the shores
of Loch Laggan in the South, and more gently to the valley of the River Spey at
a similar elevation in the north.

The terrain has been deeply dissected by the glaciers which formed
the valleys now occupied by River Spey, the Allt Coire Ardair and their
tributaries. These streams commonly rise from lochs situated at the head of
steep corries cut by the glaciers, and follow broad glaciated "U" shaped
alluvial valleys. Glacial debris and moraines form small hills up to 100m
above the surrounding ground as at NN534936, southeast of Garva Bridge.

Much of the ground is covered by peat and bog which limits the
exposure of the underlying rocks.
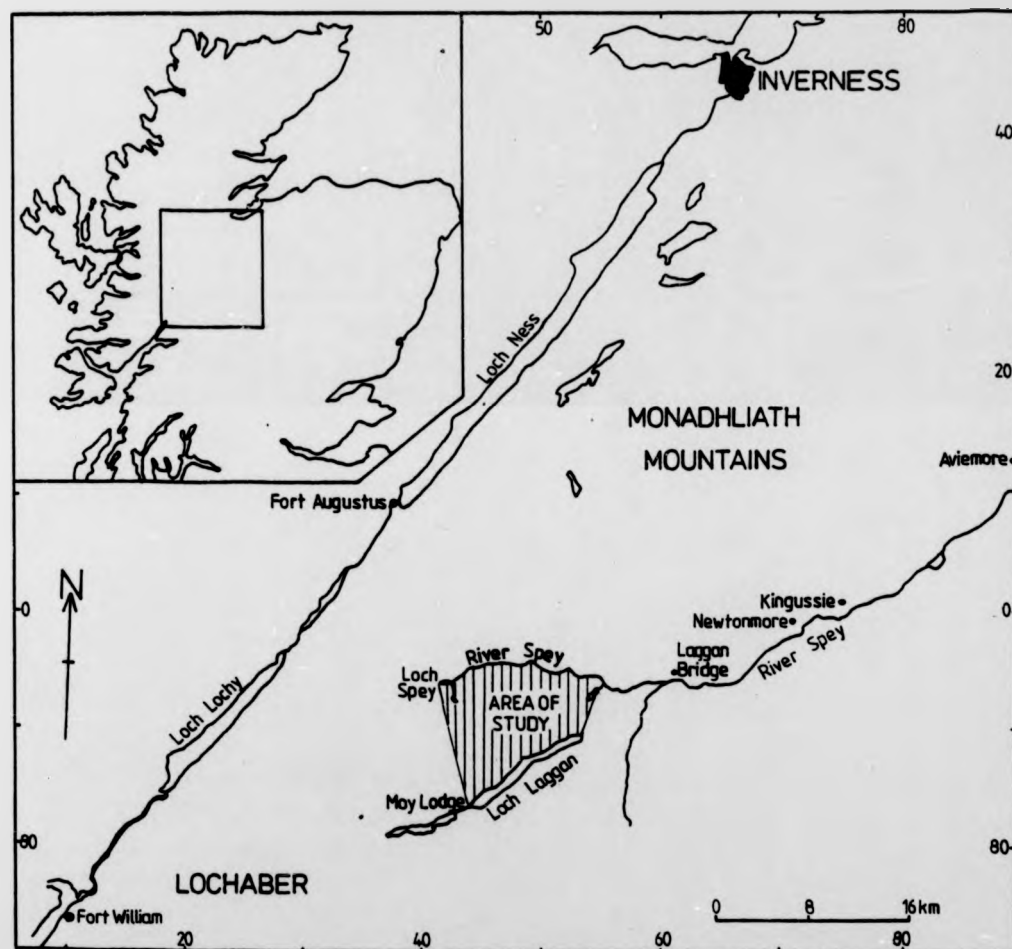
Fig.1.1: The location of the area of study
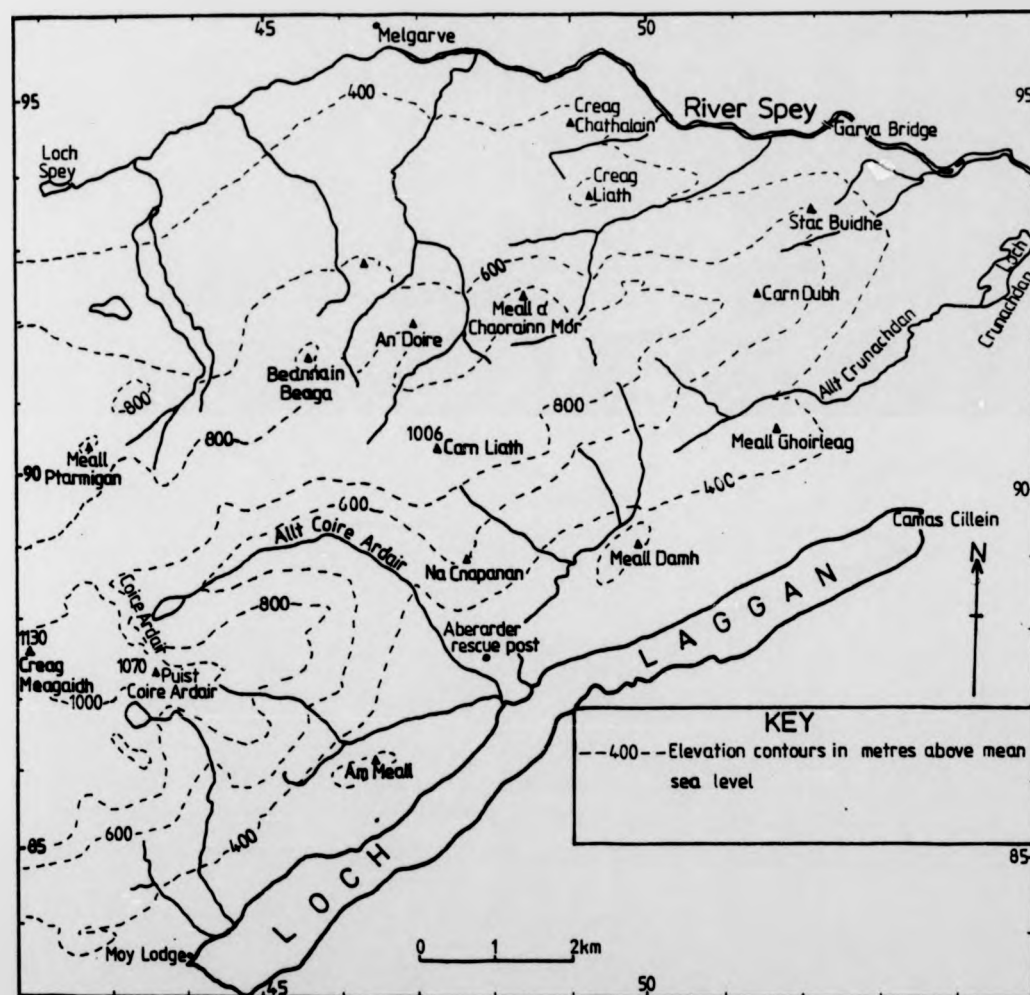
Fig. 1.2: Simplified topography of the study area

Accessibility is very poor in this area, and apart from the Kinlochlaggan-Fort William Road bounding it in the south and General Wade's Military Road bounding it in the north, there are only a few foot-paths. West of Melgarve (NN464960) General Wade's Military Road is only passable to four-wheel drive vehicles.

Human population in this area is very sparse, its main economic activity being sheep-farming and red deer stalking in the hills and some grouse shooting. There are also some forestry plantations.

1.3 Aims and Scope of Research

Detailed mapping on the scale of six inches to the mile was carried out during the spring and summer seasons of 1982 and 1983, with the aim of establishing a local stratigraphic succession and the structural history of the metasedimentary rocks in this area.

The various lithostratigraphical units and their associated calc-silicate bands were extensively sampled for petrographical and geochemical studies. It was intended to use their petrographical and geochemical characteristics to suggest the conditions of their formation and petrological evolution.

The lithostratigraphic units recognised in this study area can be broadly correlated with those of the Corrieyairack Pass and Killin Areas studied by Haselock (1982) and Whittles (1981), respectively.

1.4 Previous Research and Geological Background

Murchison and Geikie (1861) were the first to recognise the similarities between the rocks forming the Central Highlands, of which this area is part, and those of the N.W. Highlands. They contended that folding repeated the psammitic rocks of Loch Eil on the eastern side of the Great Glen, around

Glen Spean.

Hinxman (1913) assigned these "granulites" and "mica schists" of the Central Highlands to the Moine Series. Hinxman and Anderson (1915) introduced the term: Central Highland Granulites, to describe these "undifferentiated siliceous and quartzo-feldspathic schists" with belts of pelitic and semi-pelitic schists and gneisses, and also containing lenses of "garnetiferous zoisite-granulites".

Peach and Horne (1930) recognised a belt of pelitic gneiss affected by repeated folding which extends from the Findhorn to the Monadhliath Mountains. This was later named the "monadhliath Schists" by Anderson (1956).

The first comprehensive account of the geological history of the rocks of this area was that of Anderson (1956). He established a stratigraphic succession for the metasedimentary rocks in the area (Table 1.1) and described the various granitic and porphyritic rocks intruding them.

He noted that the psammitic "Eilde Flags" were overlain by a series of pelites and discontinuous quartzites of the "Transition Group" which form the base of the lower Dalradian Appin Group, and the Monadhliath Schists. These were in turn overlain by the Dalradian limestones, the Kinlochlaggan and the Ballachulish limestones. Anderson (op. cit.) contended that the coarse-grained garnetiferous pelites of the Monadhliath Schist, traceable from Loch Killin to the Corrieyairack Pass area were the same as the fine-grained, silvery-grey schists of Glen Roy and the Leven Schists, and indeed all the schists of the Lochaber Subgroup. He put the base of the Dalradian supergroup at the bottom of the lowest limestone of the Lochaber subgroup (Table 1.1).

However, subsequent workers including Bailey (1934) and Johnstone (1966) place the base of the Dalradian below the Eilde Quartzite and this is now the
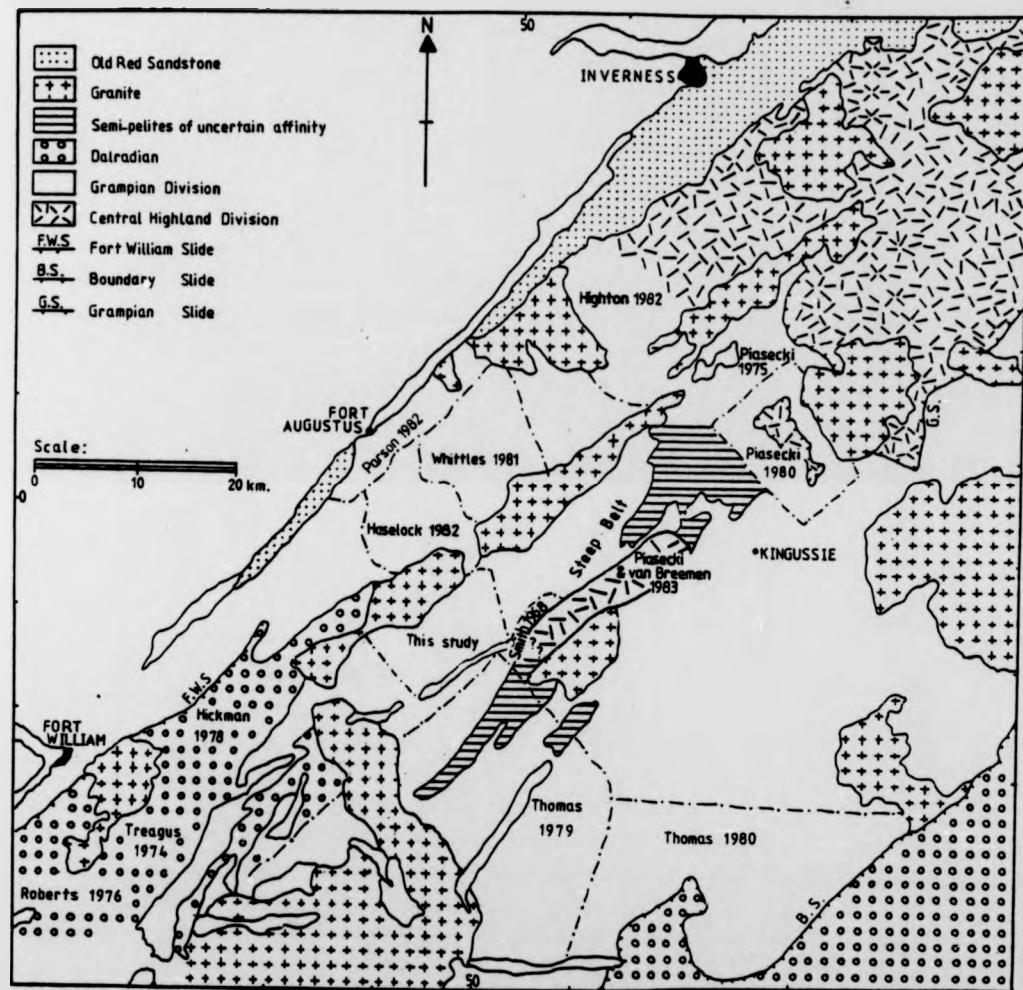
Fig.1.3: General geology and recent research in the Central Highlands of Scotland

Table 1.1 : Stratigraphical Succession of Moinian and Dalradian Strata, Glen Roy to Monadhliaths (Anderson 1956).

generally accepted boundary.

Anderson (op. cit.) recognised three major NE-SW trending folds: the Kinlochlaggan Syncline in the east with the Kinlochlaggan Limestone in its core, the Loch Laggan Anticline to its west, and further west, the Corrieyairack Syncline with the Monadhliath Schist in its core.

Recent work in this region (Whittles, 1981 and Haselock, 1982) involving more detailed mapping and geochemical studies has established two lithostratigraphical successions: the Glenshirra and the Corrieyairack Successions, separated by a tectonic break, the Gairbeinn Slide, within the "Grampian Division" (Piasecki, 1980; Piasecki and van Breemen, 1979a, 1979b, 1983, Piasecki et al. 1981).

Three episodes of deformation were recognised (Whittles, 1981, Haselock, 1982; Haselock et al, 1982) in these rocks. The first episode, D1, involved movements on the Garbeinn Slide, and the second, D2, gave rise to three major NE-SW trending folds: the Creag Mhor Antiform in the east, the Corrieyairack Syncline to its west and the Tarff Anticline further west. The third episode produced a weak cross-folding of the earlier structures.

They also recognised three episodes of metamorphism, the climax of which attained middle amphibolite facies conditions.

Their work and that of Piasecki (op. cit.) and Piasecki and van Breemen (op. cit.) shows that not all the semi-pelitic units in the Monadhliath Mountains assigned by Anderson (1956) to the same sequence are equivalent. In Glen Roy, the Monadhliath Schist is separated from the Leven Schist by a psammitic unit 500m thick termed by Haselock and Winchester (1981) the Carn Leac Psammite, and in Speyside some of the semi-pelitic units are considered to belong to the Central Highland Division (Piasecki, op. cit;

Piasecki and van Breemen, op. cit.).

Johnstone et al (1969, 1975) correlated the "Central Highland Granulites" of the Grampian Highlands with the "younger" Moinian assemblage of the NW Highlands which it was claimed rested unconformably on older Moines affected by the 1000Ma "Grenville" Precambrian orogeny as well as a later Palaeozoic Caledonian event.

However, Harris et al (1978) contended that the Grampian Division rocks (which they termed the Grampian Group) were part of the Dalradian Supergroup. They argued that the rocks of the Loch Eil Division in the NW Highlands which are lithologically similar to those of the "Central Highland Granulites" had undergone a Precambrian orogenic event as indicated by their structural conformity with the Glenfinnan Division intruded by the 1050 Ma old Ardgour Granitic Gneiss (Brook et al, 1976) and the Loch Quoich Granitic Gneiss (Piasecki and van Breemen, 1979) in contrast to the Grampian Division rocks which may have only been affected by Caledonian events.

Recent work in Speyside (Piasecki, 1980; Piasecki and van Breemen, 1979a, 1979b, 1983) has shown that the "Central Highland Granulites" comprise a "basement" : the Central Highland Division which has yielded Rb/Sr ages of between 950 and 1300 Ma, separated from a "cover", the Grampian Division, by a zone of repeated sliding deforming pegmatites which have yielded Rb/Sr ages ranging from 784+8 Ma to 570 $\pm$ 17Ma.

The oldest dates obtained from the lower levels of the Grampian Division are comparable to those obtained from "Morarian" pegmatites, 730 $\pm$ 20 Ma (van Breemen et al. 1978; Lambert, 1969) in the Morar Division of the NW Highlands.

These dates provide evidence that the lower part of the Grampian

Division has been affected by this late Proterozoic event which also involved the early movements on the Grampian Slide (Piasecki, op. cit; Piasecki et al. op. cit.). As yet no evidence for an epeirogenic stage associated with this event has been found; and it has been suggested that it was a localised event which affected only the rocks at depth while the deposition of the upper part of the Grampian Division was still going on (Piasecki and van Breemen, op.cit.)

South-east of Loch Laggan and east of the Ossian Steep Belt, Thomas (1979, 1980) has recognised major nappe structures in supposedly Grampian Division rocks of the Ben Alder and north Schichallion areas. He interpreted the Ossian Steep Belt as the root zone of nappes facing in opposite directions, to the NW and the S.E. , the "formation" of which he attributed to the Grampian Orogeny (Lambert and McKerrow, 1976).

However, recent work (Temperley,pers, comm. 1985) has questioned this structural interpretation and considers the steep belt to be a synclincal belt affected by later, tight folding, the folds being characterised by steep axial planes. As he was able to map individual formations through this zone from NW to SE, he does not consider it to be a root zone for nappes.

CHAPTER 2 : STRATIGRAPHY

2.1   Introduction and previous work

Few attempts were made by early workers to subdivide the metasedimentary rocks of the Grampian Highlands collectively described as the "Eilde Flags" by Bailey (1934). The earliest lithostratigraphic subdivisions were made by Anderson (1956) who, correlating the main lithological types with the stratigraphic succession established by Bailey (1934) for the Lochaber district, described all the pelitic rocks in the region as Monadhliath Schists which he correlated with the Leven Schist of Glen Roy. He equated the associated psammitic rocks with the Eilde Flags.

More recently, others (Piasecki and van Breemen, 1979; Piasecki, 1980) have recognised an older migmatitic Moine basement within the Grampian Highlands which they called the Central Highland Division. This was found to be separated from its younger cover, which they called the Grampian Division by a tectonic break which Piasecki (op cit) has termed the Grampian Slide.

Detailed work in the Killin area (Whittles 1981) and the Corrieyairack Pass areas (Haselock, 1982), (Haselock et al, 1982) has led to the recognition of two lithostratigraphic successions within the Grampian Division rocks separated by a tectonic discontinuity termed by Haselock (op. cit.) the Gairbeinn Slide.

In the Corrieyairack Pass area, northwest of the present study area, Haselock (1982) has subdivided the upper lithostratigraphic succession, which she termed the Corrieyairack Succession, into four major formations which can be recognised in a modified form in the Loch Laggan-Upper-Strathspey area, (Fig 2.2.).

The lithostratigraphic succession underlying the Gairbeinn Slide, termed

by Haselock (op. cit.) the Glenshirra Succession, was also subdivided into
four formations but their correlation with rocks in the present study area
has proved less simple, although a broad similarity may be recognised. (Fig 2.2)

Correlation with the Geal Charn and Ben Alder areas to the southeast
(Thomas 1979) has proved difficult because the intervening ground is yet to be
mapped in detail. However, on a purely lithological basis, most of the rock
types he described are similar to those found in the Loch Laggan-Upper
Strathspey area although rocks similar to the gneisses and quartzites found
in both the Geal Charn and the Ben Alder successions are absent in the present
study area. Lithological similarities also exist between some of the
rocks east of the Glen Markie and some of those of the present study area.
However, until more detailed lithological and sedimentological analyses of
these adjoining areas are available, no definite correlations can be made.

2.2 Lithostratigraphy

In this study area, two lithostratigraphic successions separated by a
zone of high strain have been recognised. These successions can be broadly
correlated with the Glenshirra and the Corrieyairack Successions of Haselock
(1982) and Haselock et al. (1982) in the adjacent Killin and Corrieyairack
Pass areas (Fig 2.2).

Subdivision of the rocks in this area into the different lithostratigraphic
formations was based on overall lithological characteristics including the
type of the dominant lithology, facies variations and associations.
Classification of the metasedimentary rocks into the main lithotypes, viz
psammite, semi-psammite, semi-pelite and pelite was based mainly on the
relative proportions of mica to quartz and feldspar present in the rock (see
Chapter 3).

As far as possible, the lithostratigraphic units recognised in this area
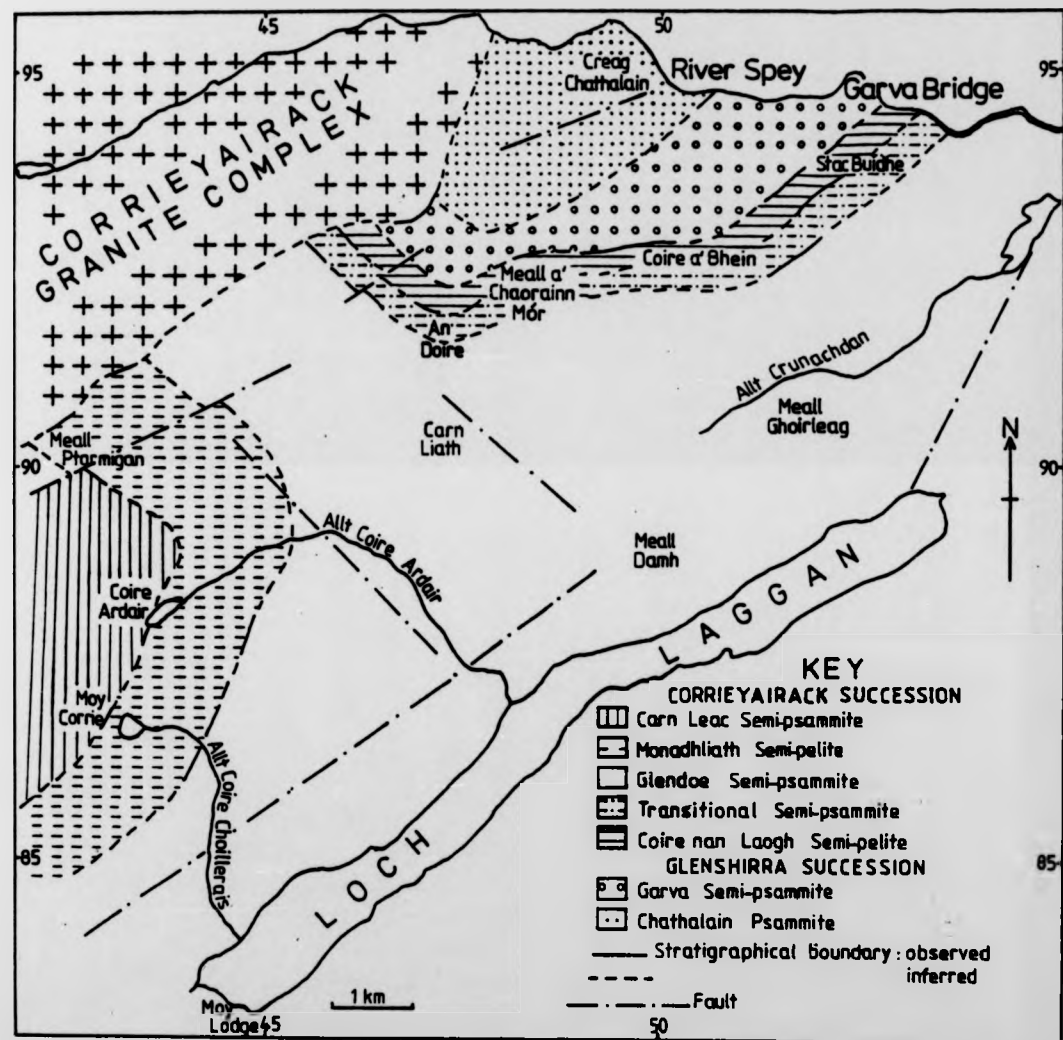
Fig.2.1: The distribution of the lithostratigraphic formations in the

Loch Laggan-Upper Strathspey area

Table 2.1 : Stratigraphical succession of the rocks of the Loch Laggan-Upper Strathspey area.

SUCCESSION		FORMATION	THICKNESS (Metres)
GRAMPIAN DIVISION	CORRIEYAIRACK SUCCESSION	Carn Leac Formation	> 600
		Monadhliath Formation	800 - 1200
		Glendoe Formation	1700 to over 3200
		Transitional Semi-psammite	300 - 400
		Coire nan Laogh Formation	400 - 500
		GAIRBEINN SLIDE	
	GLENSHIRRA SUCCESSION	Garva Formation	650 - 700
		Chathalain Formation	> 700

will be correlated with their equivalents in the adjoining Killin and the Corrieyairack Pass areas. They will therefore be referred to by the formational names of Haselock et al. (1982) where definite correlations are possible.

Observed sedimentary structures indicate that in most places the rocks are uninverted. The lithostratigraphic units are described from the base of the mapped succession upwards.

The distribution of the various lithostratigraphic units mapped in this study area is given in Fig. 2.1.

2.2.1   The Glenshirra Succession

This lithostratigraphic succession crops out in the north of the study area, covering an area of about 8 km^2 and occupying the ground between Garva Bridge (NN522948) and the eastern end of the Corrieyairack Granite Complex (Fig 2.1).

Only two lithostratigraphic formations have been distinguished within this succession. They are:

a. <u>Chathalain Formation</u>

This is the lowest formation, best exposed in this area on Creag Chathalain (NN 490947). It consists dominantly of grey psammites with subordinate semi-psammite and minor semi-pelitic bands. These psammites occur as 20-60 cm thick units commonly showing internal laminations which pass upwards into 2-5 cm thick sequences of laminated semi-psammite and a thin (10-20 cm) semi-pelitic band at the top (Plate 2.2).

Large-scale trough cross bedding and cross lamination are commonly developed in this formation (Plate 2.1). Erosional channels (Plate 4.1a) and water-escape "flame structures" occasionally observed in the sequence

Plate 2.1 : Cross-bedded psammite in the Chathalain Formation

(NN 429946).

(The width of the picture represents 41cm)

Plate 2.2 : Thinly-bedded semi-psammite and thin semi-pelitic band

in the Chathalain Formation (NN 491948).

Hammer head is 10cm long.

Plate 2.1 : Cross-bedded psammite in the Chathalain Formation

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(The width of the picture represents 41cm)

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(NN 429946).
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Plate 2.2 : Thinly-bedded semi-psammite and thin semi-pelitic band
in the Chathalain Formation (NN 491948).
Hammer head is 10cm long.

indicate some channel deposition and rapid sedimentation for the sediments of the Chathalain Formation. Rare, concordant, thin (3-5cm thick) calc-silicate bands are present in the sequence.

The base of the formation was not seen, but its thickness is estimated to exceed 700m.

b.   Garva Formation

Overlying the Chathalain Formation is a sequence of striped, thinly-bedded, medium-grained pinkish semi-psammites with grains of microcline and quartz 0.5 - 3 mm in diameter in some layers, and minor semi-pelitic intercalations.

Near Garva Bridge (NN 522947) the formation shows well-developed crossbedding and ripple cross lamination which are commonly overturned (Plate 2.3), to the north.

Here the bed thicknesses are 20-30 cm and show normal grading with a thin semi-pelitic band, 5-10 cm thick at the top.

The upper 400m of the formation consists of thin (0.5 - 10 cm. thick) flaggy psammites (Plate 2.4) occasionally intercalated with thin (5-10 cm thick) semi-pelitic bands. No sedimentary structures have been observed in this upper section which contains numerous concordant and subconcordant quartzofeldspathic bands 5-8 cm thick which are variously folded and boudinaged. It is possible, therefore, that any original sedimentary structures have been obliterated by intense deformation, and that the flagginess of the units may be in part due to the deformation associated with the Gairbeinn Slide.

Typical whitish calc-silicate bands are absent but thin 0.5 - 2 cm epidotic bands occur at some horizons (Plate 2.3). The thickness of this formation is estimated at 650-700m, but given the presence of tight, isoclinal major D1 folds observed in it this figure must be a maximum although there may

Plate 2.3 : Garva Formation with overturned cross lamination, C, in
semi-psammite.   (NN 522947)
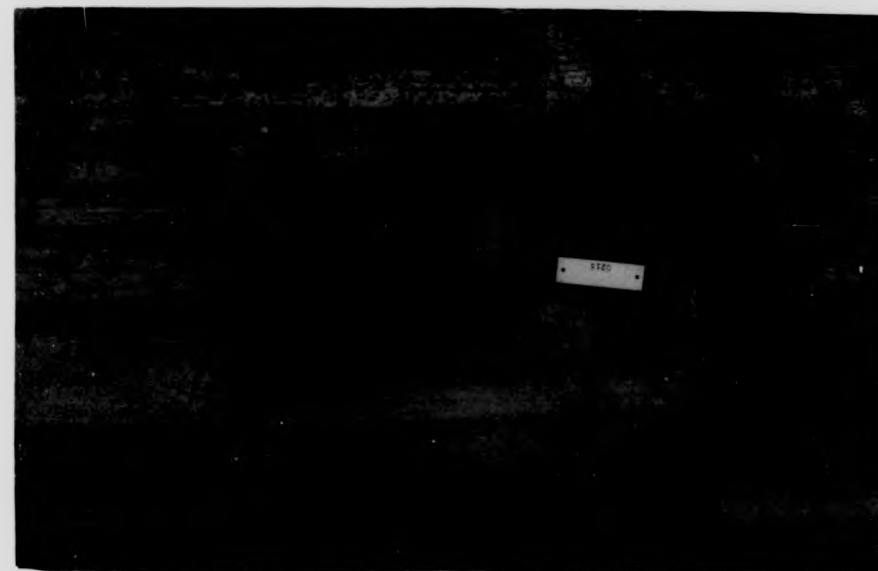
Plate 2.4 : Thinly-bedded and laminated semi-psammite in the
Garva Formation (NN 522943).
Black hammer handle is 17.5 cm long.

Plate 2.3 : Garva Formation with overturned cross lamination, C, in
semi-psammite. (NN 522947)

Plate 2.4 : Thinly-bedded and laminated semi-psammite in the
Garva Formation (NN 522943).
Black hammer handle is 17.5 cm long.

Plate 2.3 : Garva Formation with overturned cross lamination,C,in

semi-psammite.   (NN 522947)

Plate 2.4 : Thinly-bedded and laminated semi-psammite in the

Garva Formation (NN 522943).

Black hammer handle is 17.5 cm long.

also have been some thinning associated with extension and flattening.

Haselock (1982) recognised four lithostratigraphic formations in her Glenshirra Succession. Her lower units : the Creag Mhor Psammite and the Carn Dearg Psammite are lithologically similar to the Chathalain Formation, and hence have been correlated with it (Fig 2.2).

The Allt Luaidhe Semi-psammite (Haselock 1982) is also similar lithologically to the Garva Formation, but no equivalent can be found for the Gairbeinn Pebbly Semi-psammite in the present study area. The absence of this unit may be explained in terms of lateral facies changes or tectonic excision during movements on the Gairbeinn Slide. This will be discussed in greater detail in Chapter 5.

2.2.2. The Corrieyairack Succession

This structurally overlies the Glenshirra Succession from which it is separated by the Gairbeinn Slide. In this study area, the Corrieyairack succession crops out over an area of about 85 km^2, occupying the ground between the Glenshirra Succession, the Corrieyairack Granite Complex and Loch Laggan.

Five lithostratigraphic formations have been recognised in this succession and they correspond broadly to the units described from the Killin and Corrieyairack Pass areas by Whittles (1981), Haselock (1982), and Haselock et al. (1982).

a   Coire nan Laogh Formation

This is the lowest formation of the Corrieyairack Succession and it is separated from the underlying Garva Formation by a zone of high strain containing the Garbeinn Slide.

The formation is exposed in the area extending from Stac Buidhe (NN521939)

Plate 2.5 : Base of the Coire nan Laogh Formation
(NN 520940).
Hammer is 33cm long.

Plate 2.6 : Semi-pelites of the Coire nan Laogh Formation with
thin psammite bands and concordant quartzo-feldspathic
veins. (NN 521936).
Hammer is 33.5cm long.

Plate 2.5 : Base of the Coire nan Laogh Formation

(NN 520940).

Hammer is 33cm long.

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thin psammite bands and concordant quartzo-feldspathic

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Plate 2.5 : Base of the Coire nan Laogh Formation
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Plate 2.6 : Semi-pelites of the Coire nan Laogh Formation with
thin psammite bands and concordant quartzo-feldspathic
veins. (NN 521936).
Hammer is 33.5cm long.

in the east through Coire a Bhein (NN 504926), Meall a Chaorain Mor (NN484926) and then swings north-west as a result of folding. Its contact in the west with the Corrieyairack Granite Complex is inferred beneath poorly exposed ground (Fig. 2.1).

The Coire nan Laogh Semi-pelite is characterised by a coarse, migmatitic fabric with coarse biotite-muscovite (>2mm)- rich bands containing concordant quartzo-feldspathic segregations, 2-10 cm thick, which have been isoclinally folded and boudinaged (Plate 2.6).

The formation also contains thin, 5-10 cm thick, semi-psammite bands which may occasionally contain thin (2-5 cm thick) calc-silicate bands. No sedimentary structures have been observed in this formation; this is probably as a result of their destruction during the intense deformation suffered by these incompetent rocks.

The estimated thickness of this formation is about 450m, but given the occurrence of commonly-observed isoclinal folding this figure must be considered a maximum. However, there may also have been considerable thinning associated with extensional fabrics (Chapter 5) observed in these rocks.

b.   The Transitional semi-psammite

Overlying the migmatitic semi-pelites of the Coire nan Laogh Formation is a sequence of thinly-bedded semi-psammites (10-20 cm) thick which show good normal grading from a psammitic base to a semi-pelitic top. These units also occasionally contain crosslamination which indicates that this formation is uninverted. The top of this formation comprises thinly striped semi-psammite bands 2-5 cm thick, separated by pelitic laminae (Plate 2.7).

The estimated thickness of this formation is 300 - 400m but the occurrence of isoclinal folding suggests that this may be a maximum figure. However, some thinning associated with stretching may have also taken place.

c.   The Glendoe Semi-psammite Formation

The Transitional semi-psammite passes up into grey and more thickly-

Plate 2.7 : Thinly-bedded semi-psammite in the Transitional
Semi-psammite. (NN 528935).

Plate 2.8 : Massive semi-psammites in the Glendoe Formation
(NN 536922)
Hammer is 33.5 cm long.

Plate 2.7 : Thinly-bedded semi-psammite in the Transitional
Semi-psammite. (NN 528935).

Plate 2.8 : Massive semi-psammites in the Glendoe Formation
(NN 536922)
Hammer is 33.5 cm long.

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          Semi-psammite. (NN 528935).

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          Hammer is 33.5 cm long.
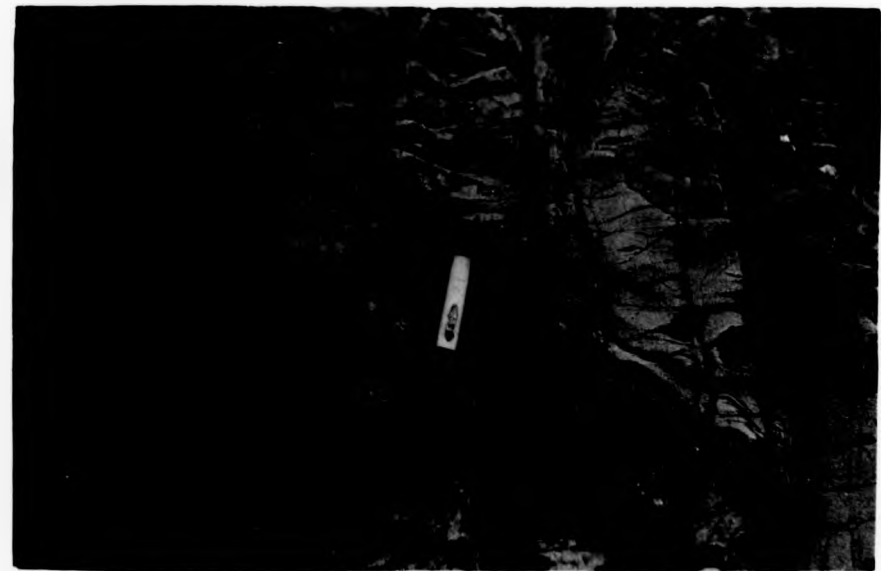
bedded semi-psammitic units (Plate 2.8) which are medium-grained. Bed thicknesses vary from 10 to 100 cm and commonly show normally graded (Bouma) sequences with thin, 2-5 cm thick semi-pelitic tops (Plates 4.8 and 4.12) as well as climbing ripple lamination (Plate 4.9) and slump structures (Plate 4.10).

Thin, 2-8 cm thick whitish calc-silicate bands and lenses are abundantly developed (Plate 4.8). These and the other sedimentological features will be discussed in detail in Chapter 4.

In the Killin area, Whittles (1981) distinguished whitish calc-silicate bearing semi-psammites of the Glendoe Formation from the greenish calc-silicate bearing semi-psammites of the Knockchoilum Formation. In view of this calc-silicate facies variation, and the similarity in the assemblage of sedimentary structures, the semi-psammites mapped at the equivalent stratigraphic level in the present study area have been correlated with the Glendoe Formation of Whittles (op. cit.)

In Fig 2.2 a correlation has been made between the Glendoe Semi-psammite and the Knockchoilum Semi-psammite of the Corrieyairack Pass area (Haselock, 1982). She had not recognised the Glendoe Formation in that area. The bases of this correlation are:

a. The similarity in the lithological type constituting the two formations.

b. The similarity in the assemblage of sedimentary structures, i.e. graded, convolute bedding and cross-lamination.

c. Similarities in their bulk-rock chemistries, discussed in Chapter 7.

According to Whittles (1981) the only differences between the Glendoe and the Knockchoilum Semi-psammites related solely to the colours and mineral composition of their calc-silicate bands. He found no significant difference in the bulk-rock chemistries of these calc-silicates and given this similarity the observed differences in their mineral compositions is possibly due to metamorphic parameters as yet undetermined.

An examination of the maps of the Killin area (Whittles, 1981) shows the proximity of the green calc-silicate-bearing semi-psammites to the Foyers and
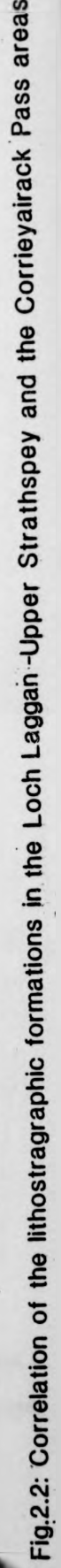
Fig.2.2: Correlation of the lithostragraphic formations in the Loch Laggan-Upper Strathspey and the Corrieyairack Pass areas

and the Allt Crom Granites, in contrast to the white calc-silicate-bearing semi-psammites. The same observation can be made with regard to the adjoining part of the Corrieyairack Pass area (Haselock, 1982).

These observations cannot be mere coincidence. They suggest a cogenetic relationship between the green, pyroxene-bearing calc-silicates and the granites. This inference is supported by the report by both Whittles (op. cit.) and Haselock (op. cit.pers. comm. 1984)., that the areas where these rocks outcrop are extensively veined by these granites.

It may therefore be concluded that both the Glendoe and the Knockchoilum Formations form part of the same thick lithostratigraphic sequence.

The thickness of the Glendoe Formation varies from 1700m in the NW where it appears to have thinned considerably from over 3,200m in the east towards Loch Laggan.

d.   The Monadhliath Formation

The semi-psammites of the Glendoe Formation pass through a sequence of thinly interbanded semi-psammites and semi-pelites 5-10 cm thick, into the normally graded sequences of the Monadhliath Formation. These thinly banded units may be equivalent to the "striped" unit described at an equivalent stratigraphic level by Haselock (1982).

This formation is extensively exposed south-east of the Corrieyairack Granite Complex. Its contact with the complex is marked by a 400m wide aureole in which the rocks have been hornfelsed.

The basal section of the Monadhliath Formation is marked by cyclic sequences of 40-80 cm thick grey psammite beds (Plate 4.13) which are succeeded upwards by a series of 20-120 cm thick normally graded (Bouma) sequences (Plate 4.14) with occasional 20-40cm thick semi-pelitic interbands. Towards the top of the formation, the graded units become thinner, and range from 10-40 cm in thickness.

Some of the thick psammite beds near the base of this formation contain concordant calc-silicate bands up to 20cm thick, whereas thinner, 2-5 cm thick bands occur within the semi-psammitic bands of the upper section.

The estimated formation thickness is 800 - 1200m.

e.   Carn Leac Semi-psammite Formation

This consists dominantly of grey to dark grey semi-psammitic rocks exposed on the high ground west and north-west of Coire Ardair (NN430885) and also west and south-west of Moy Corrie (NN 430870). This formation has a gradational contact with the underlying Monadhliath Formation.

The basal section of the Carn Leac Formation consists of fine-grained thinly-bedded "flaggy" semi-psammite units, 5 -10 cm thick with subordinate psammitic and semi-pelitic bands, 0.5-2cm thick. Higher in the stratigraphic succesion, the thickness of the beds increases and may reach 20-25cm.

The Carn Leac Formation occasionally contains rippled and lenticular semi-psammite bands (Plate 4.7a). Some of the semi-psammitic units also show poor normal grading (Plate 4.7b). Rare, whitish, thin, 2-5cm thick, calc-silicate bands are also present.

The thickness of the Carn Leac Formation exposed in the present study area is about 600m, but as its top was not seen it may be thicker than this.

CHAPTER 3 : PETROGRAPHY OF THE METASEDIMENTS

3.1 Introduction

The non-calcareous metasedimentary rocks of the Loch Laggan-Upper
Strathspey area have been subdivided into four main lithological types on
the basis of their relative modal proportions of mica to quartz and feldspar
present. These are:

Pelite   :   70% mica,   30% quartz and feldspar

Semi-pelite   :   40-70% mica, 30-60% quartz and feldspar

Semi-psammite :   20-40% mica, 60-80% quartz and feldspar

Psammite   :   80% quartz and feldspar

The modal analyses of the various rock types obtained by counting 1000
points per thin section are given in Tables 3.1, 3.2 and 3.3.

Plagioclase compositions have been optically determined by the Michel-Levy
method using maximum extinction angles on 010.

Semi-pelites and Pelites

Semi-pelites are the dominant lithological type in the Coire nan Laogh
and the Monadhliath Formations, but they also occur as thin bands and
intercalations in the Chathalain, Garva, Transitional, Glendoe and Carn Leac
Formations. Pelites are very rare and have only been found as thin
discontinuous bands in some parts of the Coire nan Laogh and Monadhliath
Formations.

The principal minerals comprising the semi-pelites and pelites are
biotite, quartz and plagioclase with varying amounts of muscovite, garnet,
epidote, microcline, ilmenite and sphene.

In the semi-pelites, close to the Corrieyairack Granite complex,

Plate 3.1 : Garnet in semi-pelite in varying stages of replacement
by biotite, quartz and magnetite. Garnet at right has
been completely replaced.
(Coire nan Laogh Formation, NN 467926)
Plane polarised light. Scale bar represents 0.1mm

Plate 3.2 : Folded semi-pelitic band defining So/S1 overgrown
by late porphyroblastic muscovite which shows
kink bands.
(Coire nan Laogh Formation, NN 497927).
Crossed micols.   Scale bar represents 0.2mm.

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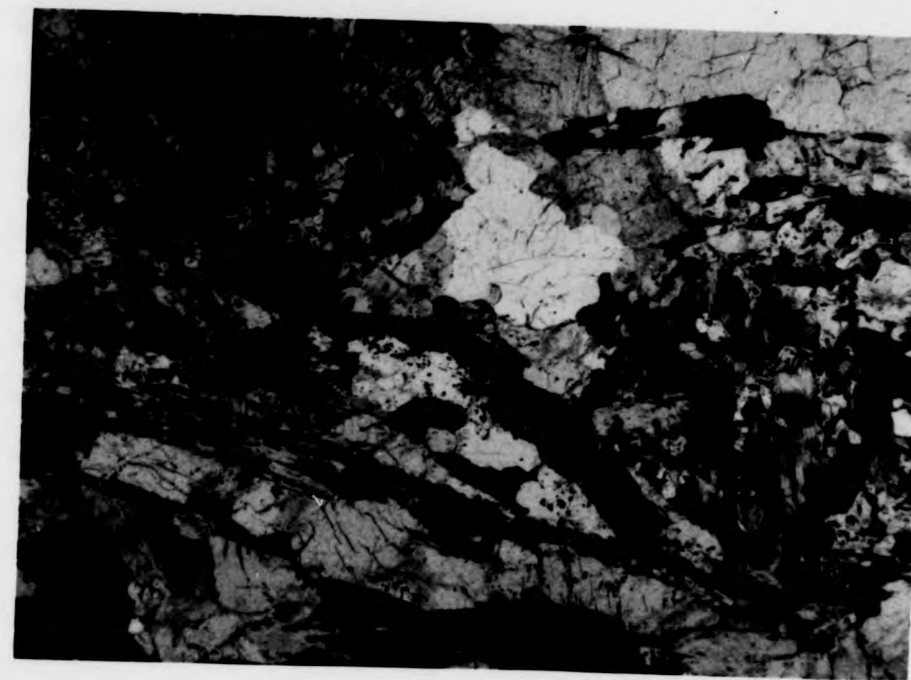Plane polarised light. Scale bar represents 0.1mm

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Crossed micols. Scale bar represents 0.2mm.

andalusite, sillimanite and cordierite have been developed within the
contact aureole. These minerals believed to result from a late thermal
overprint will be considered in detail in Chapter 6.

<u>Biotite</u> is the ubiquitous mica in the semi-pelites and pelites. It occurs
as subidioblastic laths with ragged terminations, usually defining the
dominant Sl fabric. Less commonly it occurs as late post-tectonic grains
cross-cutting this fabric. The laths vary in length from 0.25- 2mm and are
commonly intergrown with muscovite in the muscovite-rich semi-pelites. In
the dominantly semi-psammitic Garva Formation, biotite is pleochroic with
X = yellow, Y = yellowish brown and Z= brownish black, and is intergrown with
epidote, ilmenite, and sphene. In the other formations biotite is usually
pleochroic in shades of brown to reddish-brown.

Occasionally aggregates of small unoriented biotite laths are
associated with relict garnet grains suggesting pseudomorphous replacement
of garnet by biotite. (Plate 3.1.)
Some biotite laths show strained (undulatory) extinction inferred to be
the result of late deformation.

<u>Muscovite</u> forms a major proportion of the mica in most semi-pelites of the
Coire nan Laogh and Monadhliath Formations but occurs in minor proportions in
the semi-pelitic bands of the dominantly semi-psammitic Garva, Glendoe and Carn
Leac Formations. In these formations, muscovite occurs chiefly as late,
disoriented porphyroblasts up to 2.5mm long which contain small inclusions of
quartz and biotite (Plate 3.2.)
Occasionally some cross-cutting grains and some of the late muscovite
porphyroblasts are deformed during the late (D 4) episode and show strained
extinction.

<u>Plagioclase</u> is an essential constituent of all the semi-pelites and occurs

as xenoblastic grains 0.5 - 1mm in diameter which commonly show twinning on the albite law and occasionally on the combined Carlsbad-albite laws. The composition of the plagioclase grains varies from oligoclase (An15) to andesine (An38).

Plagioclase grains commonly contain tiny, rounded quartz inclusions. This feature could be the result of grain boundary diffusion and the need to reduce interfacial free energy (Toriumi, 1979) during growth. Plagioclase grains in the Garva semi-pelites occasionally share myrmekitic boundaries with microcline (Plate 3.3.).

Plagioclase grains in semi-pelites from the Coire nan Laogh and Garva Formations show pericline twinning and undulose extinction as a result of strain.

K-feldspar occurs mainly in the semi-pelites of the Garva Formation as xenoblastic microcline grains (Plate 3.3.) ranging in size from o.5-2mm in diameter. They contain tiny inclusions of rounded quartz and plagioclase grains. These microcline grains are relatively less common in the Glendoe, Monadhliath and Carn Leac Formations and have not been found in the Coire nan Laogh and Chathalain Formation semi-pelites.

Microcline grains occasionally share myrmekitic boundaries with plagioclase grains (Plate 3.3.) Several explanations have been given for the origin of myrmekites (Phillips, 1974,1980). These vary from cotectic crystallisation (Spencer, 1938) to the replacement of K-feldspar by plagioclase and quartz (Becke, 1908 In Phillips 1980). The Becke replacement model has been favoured for myrmekite in metamorphic rocks (Phillips, 1980).

The large grain size of microcline found in the associated psammites suggest that most may be relict detrital grains. The implications of these observations for the provenance of the metasediments will be discussed later in Chapter 4.

Quartz occurs as xenoblastic grains 0.25-1mm in diameter. It is commonly

Plate 3.3 : Myrmekite developed at the contact between

microcline and plagioclase in the semi-pelites

of the Garva Formation (NN 520941).

Crossed micols.   Scale bar represents 0.1mm.

Plate 3.4 : Garnet in semi-pelite with straight Si, of tiny

quartz and opaques, which is oblique to Se defined by

biotite grains.

(Coire nan Laogh Formation, NN 482925).

Plane polarised light.   Scale bar represents 0.2mm.

Plate 3.3 : Myrmekite developed at the contact between
          microcline and plagioclase in the semi-pelites
          of the Garva Formation (NN 520941).
          Crossed nicols.   Scale bar represents 0.1mm.

Plate 3.4 : Garnet in semi-pelite with straight Si, of tiny
          quartz and opaques, which is oblique to Se defined by
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          (Coire nan Laogh Formation, NN 482925).
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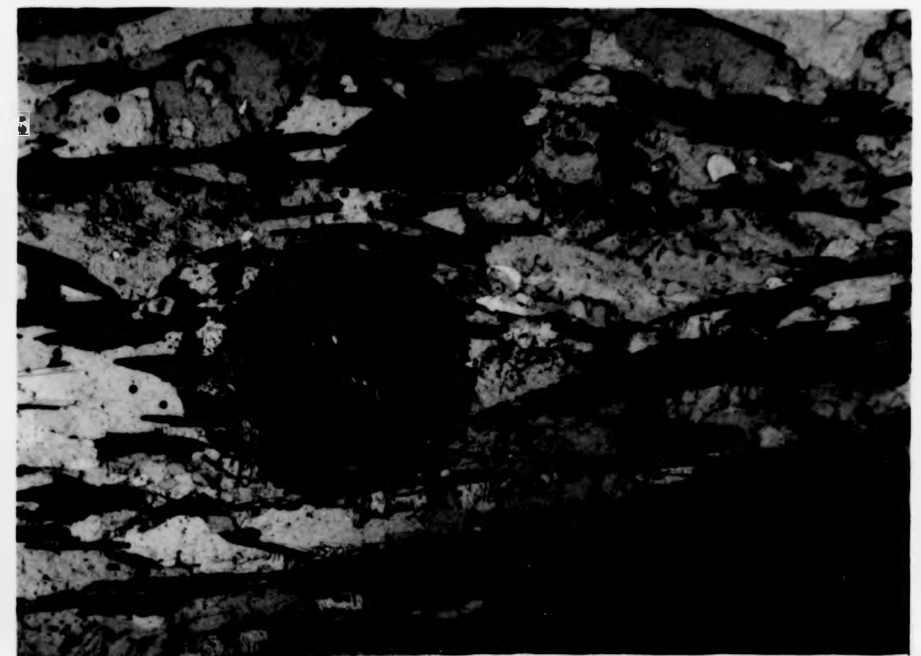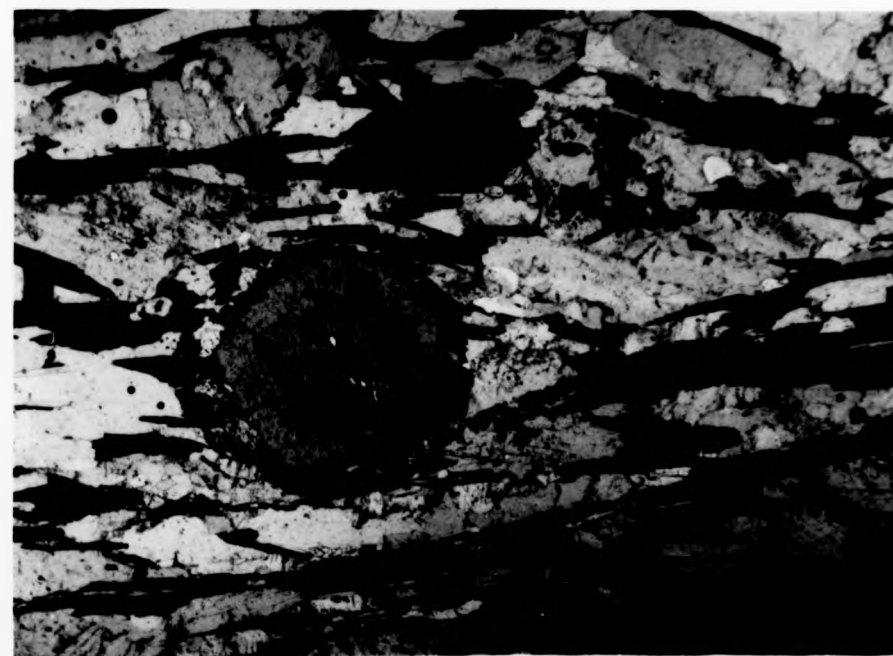Plane polarised light.   Scale bar represents 0.2mm.

associated with plagioclase with which it shares straight to curved
boundaries. Quartz grain boundaries are commonly concave to plagioclase
suggesting quartz adjustment to plagioclase. Quartz-mica boundaries are
usually straight and within mica-rich domains quartz and feldspar grains
are often elongated parallel to mica foliation (S1). This feature suggests
constricted grain boundary migration within mica bands for the tectosilicates
(Vernon, 1976).

Quartz grains generally show undulatory extinction as a result of strain.

Garnet occurs in varying amounts in the semi-pelites. It is a common
constituent of the Coire nan Laogh and Monadhliath semi-pelites often making
up to 5% of the rock, but is rare elsewhere. In the Garva Formation it has
only been found in a few samples. Haselock (1982) reported only a single
occurrence of relict garnet in the Allt Luaidhe Formation and none in the rest
of the Glenshirra Succession of the Corrieyairack Pass area. However several
occurrences of garnet have been observed south of the River Spey, in the
Chathalain Formation. Whole rock chemistry (high iron and low MgO, CaO and MnO
contents), suggests that the garnets in the semi-pelites are probably
almandine.

Garnet occurs both as poikiloblastic porphyroblasts up to 4mm in diameter
and as partially resorbed, relict grains. Garnet grains, especially in the
Coire nan Laogh semi-pelites, have poikiloblastic cores containing tiny
aligned grains of quartz and occasionally opaques and plagioclase probably
defining an early So or S1 fabric (Plate 3.4). These internal inclusions are
oriented at varying angles to the external composite, S1/S2 fabric thus
suggesting variable rotation of the garnets during subsequent deformations.
The garnet rims are usually massive and free of inclusions but some of them
show late post-tectonic alteration of the outer rims to biotite, quartz and
occasionally, chlorite.

At least two stages of garnet growth are thus indicated: an early rapid
growth marked by the inclusion-rich cores and a later, slower growth marked by

Plate 3.5 : Garnet in semi-pelite showing at least two stages

of growth, a poikiloblastic core and a later massive

growth. Outer poikiloblastic rim may either represent

a late partial replacement or another poikiloblastic

stage.

(Coire nan Laogh Formation, NN 482925).

Plane polarised light.   Scale bar represents 0.1mm.

Plate 3.6 : Ilmenite (black) rimmed by sphene in the heavy

mineral-rich bands in the Garva Formation.

(NN 485928).

Plane polarised light.   Scale bar represents 0.1mm.

Plate 3.5 : Garnet in semi-pelite showing at least two stages

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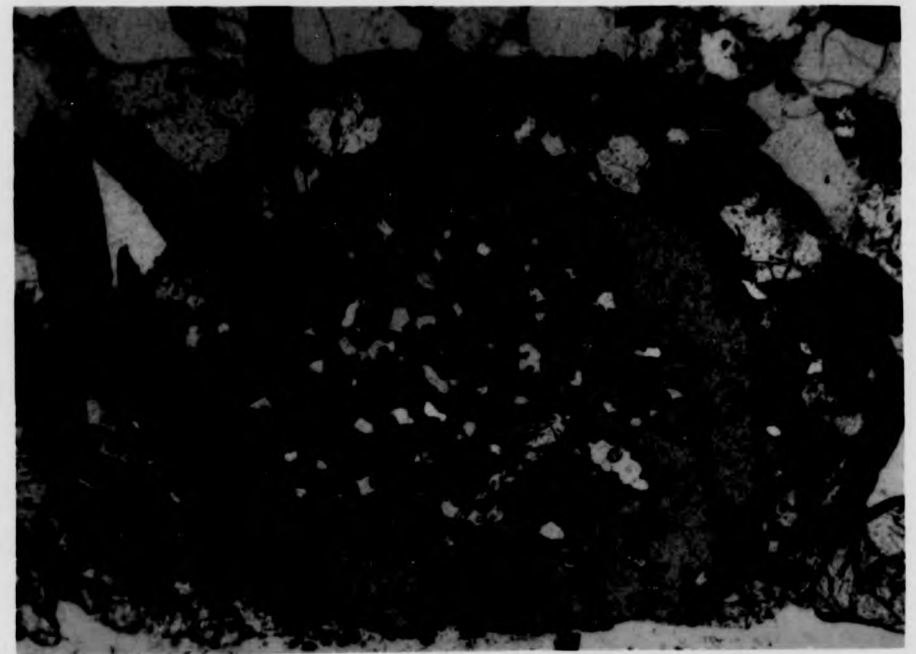Plane polarised light.   Scale bar represents 0.1mm.

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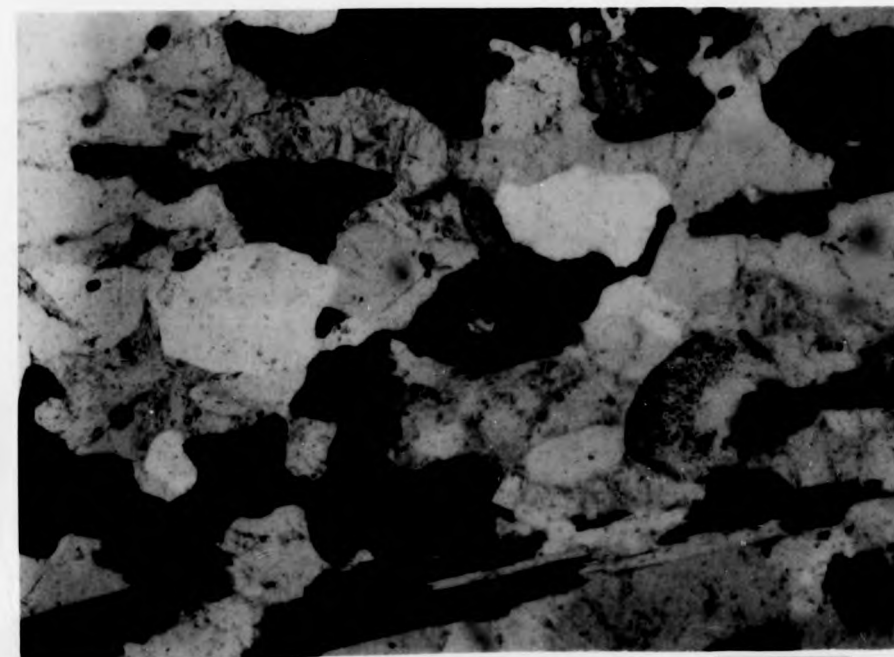Plane polarised light. Scale bar represents 0.1mm.

the more massive rims (Sturt and Harris, 1961) (Plates 3.4 and 3.5).

Garnets from the lower portion of the Coire nan Laogh Formation have been micro-boudinaged and pulled apart, the fractures having been infilled with annealed quartz grains (Plate 5.9). These features are discussed later with regard to movements on the Gairbeinn Slide, in Chapter 5.

Throughout the study area, garnets often show a varying degree of alteration to biotite, quartz and plagioclase.

Epidote occurs most commonly in the semi-pelites of the Garva Formation where it may make up to 4% of the rock. It is also present in smaller amounts in the semi-pelites of the Glendoe and Carn Leac Formations whereas it is relatively rare in the Chathalain, Coire nan Laogh and Monadhliath Formations apart from in some calc-silicate bands discussed later.

It occurs as xenoblastic to idioblastic grains 0.25-1mm in diameter commonly intergrown with olive-brown biotite, ilmenite, sphene and plagioclase (Plate 3.6). Occasionally epidote grains rim allanite and clinozoisite, showing some colour zoning with more iron-rich rims. Two generations of epidote growth are therefore indicated. The subrounded allanitic cores may represent relict detrital grains.

Magnetite commonly occurs as tiny grains in accessory amounts (Table 3.1) in most of the semi-pelites in which it is commonly intergrown with biotite and muscovite. It is also commonly associated with relict garnet and this may suggest that much of the magnetite in these rocks is the product of garnet breakdown (Plate 3.4). Much coarser grains occur in the Garva Formation where magnetite may form up to 2% of the rock (Table 3.1).

Ilmenite occurs most abundantly in the semi-pelites of the Garva Formation where it may form up to 3% of the rock. It is invariably associated with biotite, epidote and sphene. Ilmenite grains are 0.25 - 1.5mm long and

occasionally contain inclusions of quartz, epidote, biotite and plagioclase. They are also occasionally rimmed by sphene (Plate 3.6).

These minerals typically occur within thin bands and may represent original minor heavy mineral concentrations.

Sphene is a common phase in the Garva Formation semi-pelites where it may form up to 3.5% of the rock. It occurs as brown anhedral grains 0.2 - 1mm in diameter intergrown with biotite and ilmenite grains or occasionally rimming ilmenite grains (Plate 3.6). It is rarely seen in the other formations.

Chlorite occurs as greenish or occasionally brown grains. It is a rare retrograde replacement of biotite and typically forms within biotite cleavages It has been found in semi-pelites from all formations. Optical properties such as its very low birefringence and parallel extinction suggest that it could be penninite.

Apatite is a common accessory mineral in all semi-pelites especially those within the Corrieyairack Succession. It occurs sporadically as tiny colourless grains and as inclusions in biotite.

Zircon is also a common accessory phase in all the semi-pelites where it occurs as tiny inclusions with pleochroic haloes in biotite grains and also in the groundmass of the rocks.

Tourmaline is a rare accessory mineral which has been found only in the Coire nan Laogh and Monadhliath Formations. It occurs as subidioblastic to idioblastic laths 0.2 - 0.5 mm long overgrowing biotite (Plate 3.7). The grains occasionally show weak colour zoning with lighter rims (iron-poorer). The grains are pleochroic from light brown to olive green.

The presence of tourmaline in these two formations' semi-pelites suggests that some boron was adsorbed by the detrital argillaceous sediments and may

Plate 3.7 : Zoned tourmaline grains in semi-pelite from

the Monadhliath Formation (NN 448899).

Plane polarised light. Scale bar represents 0.1mm.

Plate 3.8 : Plagioclase (colourless) intergrown with quartz in

the quartzo-feldspathic segregations of the Coire nan

Laogh Formation (NN 522939). Crossed nicols.

Scale bar represents 0.2mm.

Plate 3.7 : Zoned tourmaline grains in semi-pelite from

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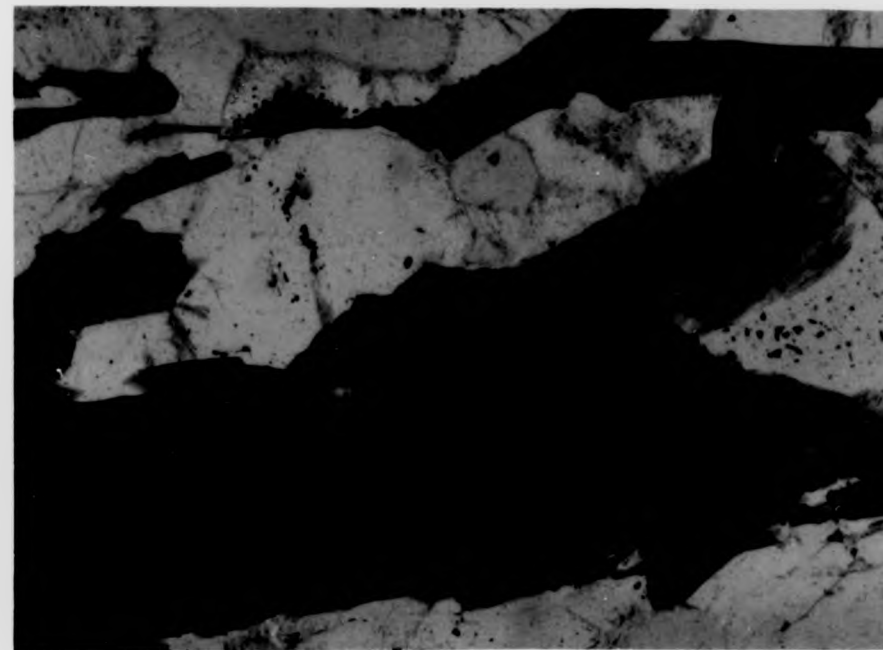Plane polarised light.   Scale bar represents 0.1mm.

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	716	213	487	621	1	311	720	133	318	8401	36	206	158	430	442	8343
Quartz	25.8	26.5	35.5	26.8	21.2	25.2	24.5	30.4	20.5	12.8	17.5	19.5	17.4	24.1	25.5	27.6
Plagioclase	23.2	23.8	31.6	20.5	25.8	34.8	23.2	23.5	18.2	7.8	13.6	42.8	38.1	32.9	8.5	30.2
Microcline	-	-	-	-	8.5	0.6	-	-	-	-	-	0.2	0.4	-	-	-
Biotite	26.5	26.8	29.6	27.5	30.2	19.8	25.5	27.6	23.4	69.2	34.0	36.8	34.7	27.8	24.9	24.8
Muscovite	24.8	22.5	3.2	24.8	0.5	8.5	26.2	17.4	35.5	10.0	32.0	0.5	9.2	14.6	41.1	14.8
Garnet	-	-	trace	-	-	-	-	0.4	2.2	trace	2.5	trace	-	-	-	-
Epidote	-	trace	-	trace	3.5	0.9	0.5	-	-	trace	-	trace	-	-	-	1.6
Magnetite	trace	trace	trace	trace	2.2	4.6	trace	trace	trace	trace	trace	trace	-	0.1	trace	0.1
Ilmenite	-	-	-	-	2.5	3.2	-	-	-	-	-	-	-	-	-	-
Sphene	-	-	-	-	3.8	1.6	trace	-	-	trace	-	trace	-	-	-	trace
Chlorite	trace	trace	-	trace	trace	0.2	trace	trace	0.1	trace	trace	-	trace	trace	trace	trace
Apatite	trace	trace	trace	trace	trace	0.2	trace	trace	trace	trace	trace	trace	trace	trace	trace	trace
Zircon	trace	trace	trace	trace	trace	trace	trace	trace	trace	trace	trace	trace	trace	0.3	trace	trace
Tourmaline	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Total	99.3	99.6	99.9	99.6	99.5	99.8	99.9	99.3	99.9	99.8	99.6	99.8	99.8	99.8	100	99.1

716, 213, 487, 621 : Chathalain Formation semi-pelites

1, 311, 720 : Garva Formation semi-pelites

133, 318, 8401 36 : Coire nan Laogh Formation semi-pelites

206, 158 : Glendoe Formation semi-pelites

430, 442 : Monadhliath Formation semi-pelites

8343 : Carn Leac Formation semi-pelites

Table 3.1 : Modal analyses of semi-pelites

suggest that these semi-pelites were derived from marine sediments (Keith and Degens, 1959).

Migmatisation of the semi-pelites

The semi-pelites of the Coire nan Laogh Formation together with those at the top of the Chathalain Formation and parts of the Monadhliath Formation have been subjected to varying degrees of migmatisation, characterised by the development of swarms of S1 concordant and subconcordant quartz and quartzo-feldspathic veins and stromatic segregations. Individual leucosomes vary in thickness from 1 to 5 cm and may be traced for up to 1 metre. They are usually strongly deformed (Plates 5.5 and 5.6). Structural relationships show that they have been affected by the D1 and subsequent deformation episodes.

The melanosomes are often poorly developed and are indistinguishable from the paleosomes. Petrographically and mineralogically, the melanosomes are similar to the semi-pelites already described.

Quartz is the dominant phase in the leucosome and occurs as xenoblastic porphyroblasts up to 6mm in diameter often containing inclusions of plagioclase grains. (Plate 3.8). Quartz grains show undulose extinction and occasionally deformation bands. The margins of some of the grains are recrystallised into strain-free grains, indicating some dynamic recovery or post-tectonic annealing.

Plagioclase is oligoclase (An_{12} - An_{26}). This range of composition is similar to that observed in the melanosome. Plagioclase occurs mainly as xenoblastic grains 0.5 - 2 mm in diameter, which occasionally contain tiny, rounded quartz grains.

Isolated biotite and muscovite grains, typically smaller than those found in the melanosome, aligned parallel to the S1 fabric may be found in the leucosome. These may represent early mica grains caught

within growing porphyroblasts of quartz and plagioclase during leucosome development.

The features described above, especially the similarity of leucosome and melanosome plagioclase compositions, suggest that the migmatic segregations were largely the product of metamorphic segregation (Yardley, 1978; Robin, 1979) arising from stress induced mass transfer and diffusion.

Some of these segregations may however represent early quartz-rich veins produced during the dewatering of the original sedimentary pile (Fyfe et al, 1978) since they have also been observed in more psammitic sediments which are not as highly strained as the semi-pelites.

3.4.   Semi-psammites and Psammites

These lithological types contain essentially the same minerals as the semi-pelites, but proportionately less mica and more quartz and plagioclase. Representative modal analyses of these rocks obtained from 1000 points per thin section are given in Table 3.2.

The semi-psammites and psammites occur in all formations but constitute the dominant lithologies in the Chathalain, Garva, Glendoe and Carn Leac Formations.   The rocks generally possess a granoblastic-polygonal texture except in highly deformed horizons where quartz grains become elongate and feldspar grains are granulated.   Such microstructures will be described in detail in Chapter 5.

Biotite   forms 0.5 - 2.6 modal percent of these rocks, being more abundant in the semi-psammites.   It forms subidioblastic laths 0.25 - 1mm long which define the dominant So/S1 fabric.   Few discordant post-tectonic biotite grains cross-cut this fabric indicating at least two stages of biotite growth.

Biotite in the Garva Formation semi-psammites is pleochroic from yellow to olive-brown, whereas that from the Coire nan Laogh and Monadhliath Formations is commonly pleochroic from light brown to reddish-brown.   Both colour schemes are obtained in biotite from the Chathalain, Glendoe and Carn Leac Formations.   Generally, the colour of the biotite from the Garva Formation is not as green as that described by Haselock (1982) from biotites in her Glenshirra Succession.

Muscovite   occasionally occurs as a minor constituent in the semi-psammites of the Chathalain, Coire nan Laogh and Glendoe Formations.   Early muscovite is rare in the Garva, Monadhliath and Carn Leac semi-psammites which largely contain late disoriented porphyroblasts up to 2mm long.   Muscovite is relatively subordinate to biotite in psammites of all formations.   Early muscovite is commonly intergrown with biotite to define the S1 fabric.

Plate 3.9 : Myrmekite developed at the contacts between microcline
and plagioclase in a psammite from the Garva Formation
(NN 511946).
Crossed nicols. Scale bar represents 0.05 mm.

Plate 3.10 :Microcline porphyroclast in a psammite from the
Garva Formation (NN 511946).
Crossed nicols. Scale bar represents 0.2mm.

Plate 3.9 : Myrmekite developed at the contacts between microcline
and plagioclase in a psammite from the Garva Formation
(NN 511946).
Crossed nicols. Scale bar represents 0.05 mm.

Plate 3.10 :Microcline porphyroclast in a psammite from the
Garva Formation (NN 511946).
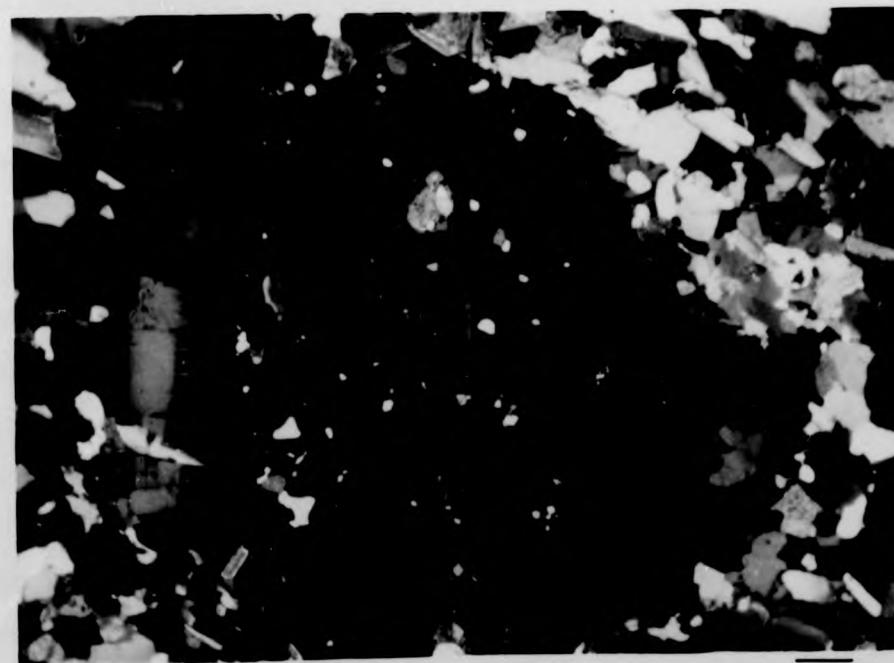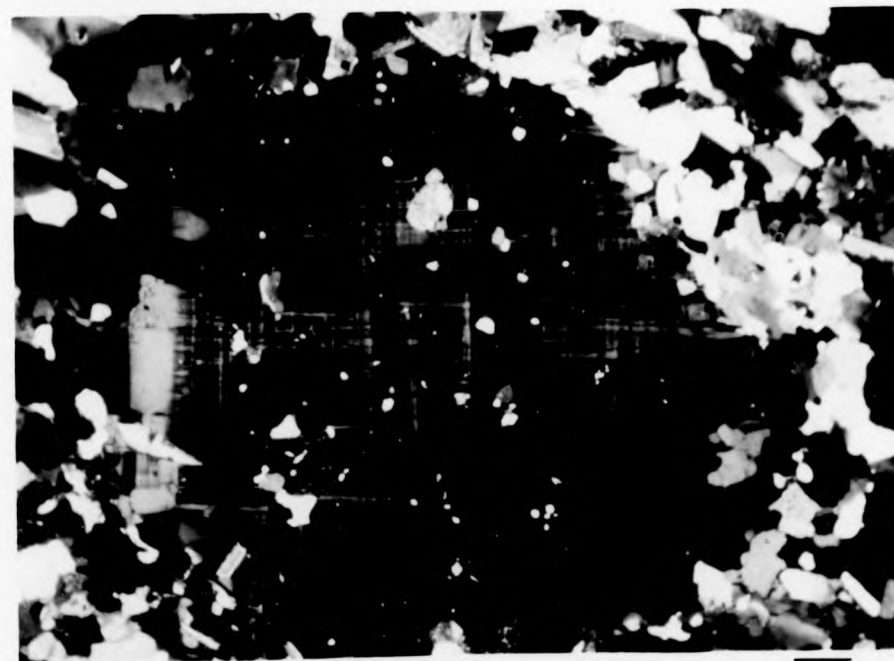Crossed nicols. Scale bar represents 0.2mm.

Plate 3.9 : Myrmekite developed at the contacts between microcline
and plagioclase in a psammite from the Garva Formation
(NN 511946).
Crossed nicols. Scale bar represents 0.05 mm.

Plate 3.10 :Microcline porphyroclast in a psammite from the
Garva Formation (NN 511946).
Crossed nicols. Scale bar represents 0.2mm.

Plagioclase (oligoclase $_{An20}$ - andesine $_{An40}$) occurs in all the semi-
psammites and psammites, as xenoblastic grains 0.25 - 1.5 mm in diameter,
in which it is intergrown with quartz. Plagioclase grains occasionally
contain tiny rounded quartz inclusions representing equilibrium lowest free
energy configuration (Toriumi, op. cit.)

In the Garva Formation plagioclase grains occasionally share
myrmekitic boundaries with microcline grains (Plate 3.9). Plagioclse
grains are also occasionally normally zoned with slightly less calcic rims
Rare sericitic alteration has affected some of the plagioclase grains.

K-feldspar occurs commonly in the Garva Formation as xenoblastic, microcline
grains 0.5 - 4.5 mm in diameter which show the characteristic "grid-iron"
twinning (Plate 3.10). They occasionally share myrmekitic boundaries with
grains of plagioclase (Plate 3.9), and also contain tiny, rounded quartz and
plagioclase inclusions.

Microcline may constitute up to 33 modal percent of some of the psammites
of the Garva Formation (Table 3.3). The size and the rounded habit of some
of these microcline grains suggest that they may have been original detrital
grains derived from a highly evolved granitic complex. This will be discussed
in detail in Chapter 7.

Quartz occurs as xenoblastic grains ranging in diameter from 0.25 - 5 mm, which
commonly show undulose extinction as a result of strain. Some large grains show
evidence of exaggerated grain growth (Vernon, 1976), and contain inclusions of
groundmass minerals (Plate 3.11) probably due to the coalescence of suitably
oriented grains

Garnet occurs in minor amounts as small relict grains in all semi-psammites
except those of the Carn Leac Formation. It is more common in the semi-psammites
of the Chathalain, Coire nan Laogh and Monadhliath Formations. In the Garva
semi-psammite rare relict garnet is found intergrown with plagioclase, chlorite

Table 3.2 : Modal analyses of semi-psammites

	103	718	320	40b	62	645	8310
Quartz	38.2	27.4	47.5	24.4	24.5	33.2	39.5
Plagioclase	36.2	44.5	34.1	48.2	51.8	39.8	43.4
Microcline	0.5	18.6	-	-	1.1	-	-
Biotite	19.4	8.8	15.4	26.2	14.4	26.5	16.5
Muscovite	-	0.4	2.6	-	7.1	-	0.2
Garnet	-	-	trace	trace	-	trace	-
Epidote	2.1	0.2	-	-	trace	-	trace
Magnetite	1.1	trace	trace	trace	trace	trace	0.1
Sphene	1.8	trace	-	trace	trace	-	trace
Chlorite	trace	trace	trace	trace	trace	trace	trace
Apatite	trace	trace	trace	trace	trace	0.1	trace
Zircon	trace	trace	trace	trace	trace	trace	trace
Sericite	trace	trace	trace	0.6	trace	trace	-
Total	99.4	99.9	99.6	99.4	98.9	99.6	99.7

103, 718 : Garva Formation Semi-psammites
320 : Coire nan Laogh semi-psammites
40b : Transitional semi-psammite

62 : Glendoe Formation Semi-psammite
645 : Monadhliath Formation semi-psammite
8310 : Carn Leac Formation semi-psammite.

Plate 3.11 : Quartz porphyroblast showing deformation bands

probably resulting from exaggerated grain growth.

Transitional Semi-psammite (NN 515931).

Crossed nicols. Scale bar represents 0.2mm.

Plate 3.11 : Quartz porphyroblast showing deformation bands
probably resulting from exaggerated grain growth.
Transitional Semi-psammite (NN 515931).
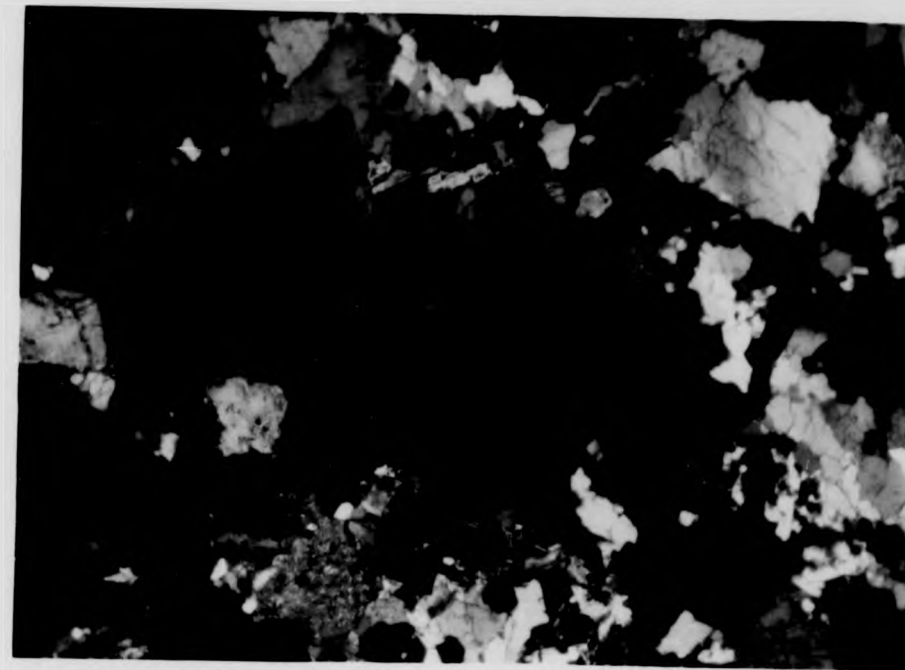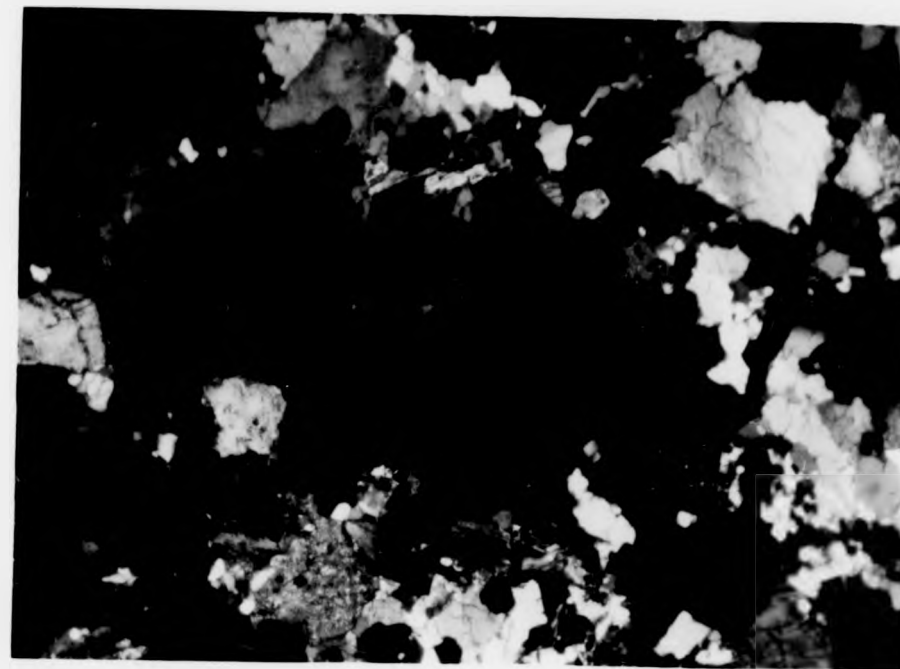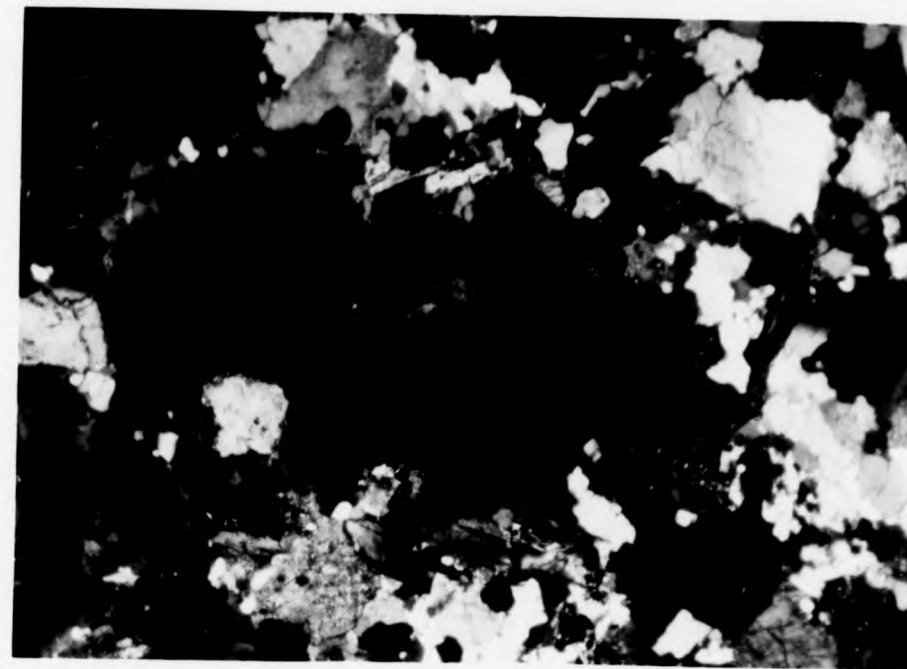Crossed nicols. Scale bar represents 0.2mm.

Plate 3.11 : Quartz porphyroblast showing deformation bands
probably resulting from exaggerated grain growth.
Transitional Semi-psammite (NN 515931).
Crossed nicols. Scale bar represents 0.2mm.

and epidote. It is commonly found in different stages of alteration to biotite, quartz, plagioclase and magnetite.

Epidote occurs as small grains 0.25-0.5 mm in diameter commonly associated with magnetite, biotite and sphene particularly in the Garva Formation.

Magnetite occurs as an accessory phase in many semi-psammites where it is commonly intergrown with biotite and muscovite. Its association with relict garnet in some specimens suggests that it is a product of garnet breakdown. It is also associated with epidote, biotite and sphene in the Garva Formation.

Sphene occurs as brown subdioclastic grains 0.25 - 1 mm in diameter commonly intergrown with epidote, biotite and some magnetite in the Garva Formation semi-psammites.

Apatite is a widely-distributed accessory mineral in the semi-psammites and psammites of all formations. It commonly occurs as tiny colourless grains 0.25 - 0.5 mm in diameter in the groundmass of the rock or is occasionally enclosed in biotite.

Zircon is also a widely-distributed accessory mineral in all semi-psammites and psammites. It is most readily seen as tiny inclusions surrounded by pleochroic haloes in biotite.

Chlorite (green) is a rare retrograde replacement product of biotite, occurring within biotite cleavage planes. Optical properties including its very low birefringence (Berlin blue) and parallel extinction suggest that it could be penninite.

Table 3.3 : Modal analyses of psammites

	74	214	617	3	76	313	8364b	8315	35
Quartz	46.8	56.5	44.9	60.5	45.2	72.5	33.4	37.4	53.4
Plagioclase	41.2	36.3	44.8	19.2	39.8	20.8	19.6	47.8	40.2
Microcline	2.8	-	-	1.5	3.4	4.2	33.5	-	-
Biotite	5.6	6.6	10.1	3.0	2.6	0.5	7.8	11.4	5.4
Muscovite	3.6	0.2	-	14.2	7.6	1.2	4.8	2.8	0.8
Garnet	-	-	trace	1.0	-	-	-	-	trace
Epidote	-	0.3	trace	0.5	0.2	0.2	0.2	0.2	-
Magnetite	trace	trace	trace	trace	0.6	-	trace	trace	trace
Sphene	trace	trace	-	trace	0.2	trace	trace	trace	-
Apatite	trace	trace	trace	trace	trace	trace	trace	trace	-
Zircon	trace	trace	trace	trace	0.2	trace	trace	trace	trace
Chlorite	trace	trace	-	trace	0.2	trace	0.5	-	trace
Sericite	trace	trace	-	trace	trace	trace	0.2	-	trace
Total	100	99.9	99.8	99.9	100	99.4	100	99.6	99.8

74, 214, 617   : Chathalain Formation psammites

3, 76, 313, 8364b   : Garva Formation psammites

8315   : Carn Leac Formation psammites

35   : Transitional psammite

3.5   Deductions from petrography

The metasediments of the Loch Laggan-Upper Strathspey area are all plagioclase-rich rocks ; no quartzites have been found. The plagioclase/quartz ratio is lowest in the Coire nan Laogh Formation, this may indicate a greater maturity for this formation as will be discussed in detail later in Chapter 7.

Petrographically, the Garva Formation is distinctive in its higher amounts of epidote, sphene, ilmenite, and olive brown biotite which are concentrated in thin bands. These probably represent minor heavy mineral concentrations, although they are not as important as those described from some parts of the Moine Assemblage. The Garva Formation is also distinctive in its important modal content of microcline which is a less common mineral in the other formations. This petrographic characteristic suggests that the Garva Formation psammites and semi-psammites were partly arkosic arenites whereas those of the other formations were greywackes.

On the whole, petrographic characteristics suggest that the Garva Formation is the least mature of all the lithostratigraphic units in the present study area, a subject which will be considered in detail in Chapter 7.

CHAPTER 4 : SEDIMENTOLOGY

4.1 Introduction

Little sedimentological analysis of the Grampian Division metasediments
east of the Great Glen Fault has been attempted. Previous workers (Anderson,
1947, 1956; Hickman, 1975; Thomas 1979; and Whittles 1981) observed the
common occurrence of preserved sedimentary structures and suggested shallow
marine environments for their deposition.

Haselock (1982) also suggested a shallow marine environment of deposition
for the Corrieyairack Succession and the lower Creag Mhor and Carn Dearg
Formations of the underlying Glenshirra Succession, but an alluvial origin
for the Gairbeinn Pebbly Semi-psammite and the Allt Luaidhe Semi-psammite
Formations which form the upper part of the Glenshirra Succession.

Detailed sedimentological analysis of the metasedimentary rocks of this
study area has been hampered by lack of sufficient sedimentological information
as a result of poor exposure and the obliteration of much of the sedimentary
textures and structures by post-sedimentary deformation and metamorphism.

However, sedimentary structures, such as cross-bedding, grading and slump
structures can still be recognised in several parts of the study area enabling
some general sedimentological inferences to be made.

4.2 Facies description and interpretation

Reading (1978) defined sedimentary facies on the basis of colour,
bedding, composition, texture, fossils and sedimentary structures. In the
Loch Laggan-Upper Strathspey area, however, the metasediments have undergone
several episodes of metamorphism and deformation so that virtually all primary
sedimentary textures, colour and any microfossils have been obliterated.

Plate 4.1 : Scour based psammites in crossbedded psammites

        in the Chathalain Formation.

        NN 492946.

Plate 4.2 : Trough cross-bedded psammites in the Chathalain

        Formation.

        NN 49009455   Hammer is 33.5   long.

Plate 4.1 : Scour based psammites in crossbedded psammites
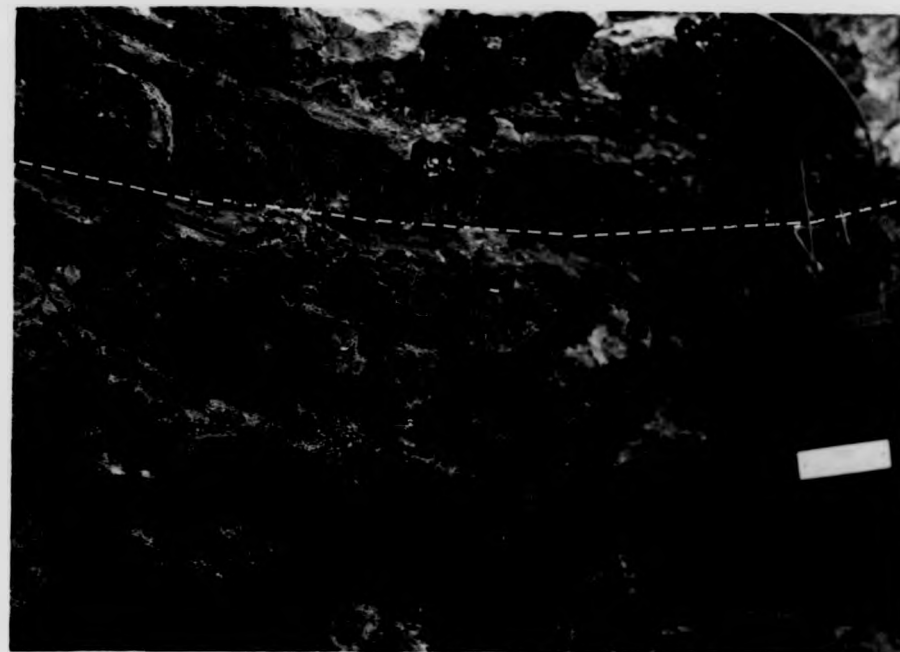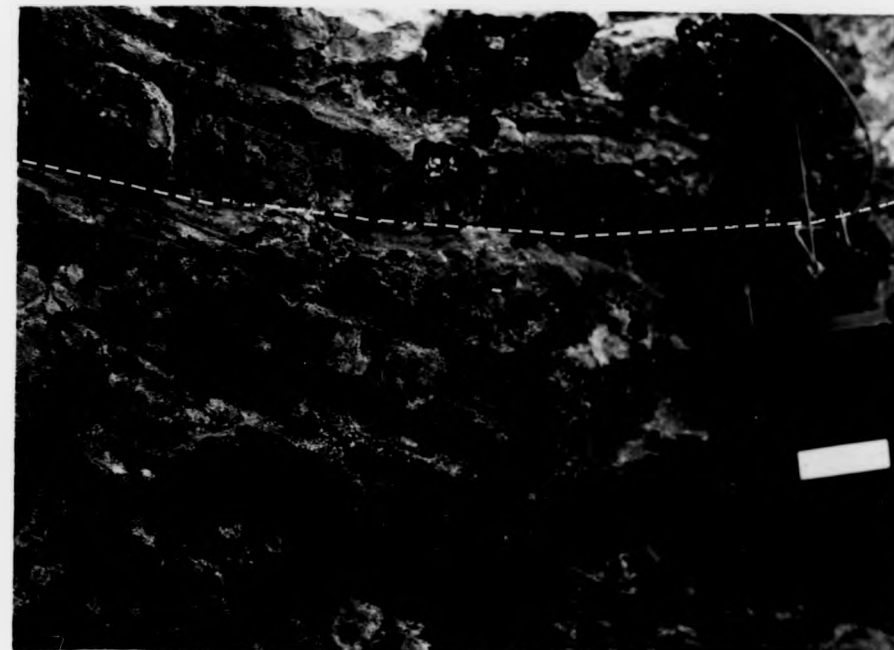in the Chathalain Formation.
NN 492946.

Plate 4.2 : Trough cross-bedded psammites in the Chathalain
Formation.
NN 49009455   Hammer is 33.5   long.

Plate 4.1 : Scour based psammites in crossbedded psammites
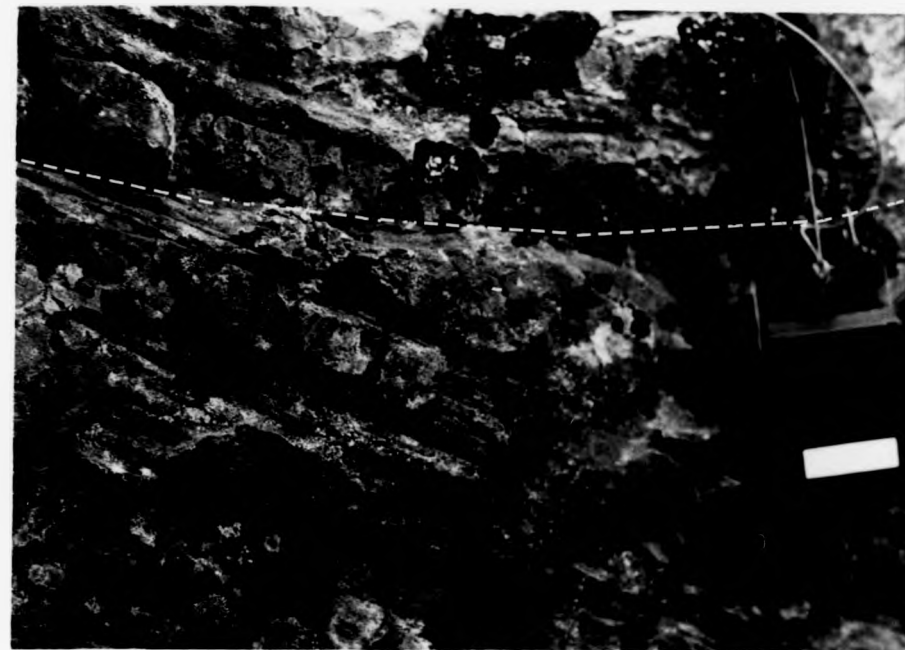
in the Chathalain Formation.

NN 492946.

Plate 4.2 : Trough cross-bedded psammites in the Chathalain

Formation.

NN 49009455   Hammer is 33.5   long.

Consequently, these rocks have been subdivided into facies only on the basis of their modified lithologies and preserved sedimentary structures.

4.2.1 The Chathalain Formation

This formation comprises sequences of the following sedimentary facies:

a. Trough Cross-bedded Psammites

This facies forms over 70% of the Chathalain Formation. It comprises sets of trough cross-bedded psammites (Plates 2.1, 4.1, 4.2). Individual set thicknesses vary from 10-60 cm and foreset lamina thickness ranges from 1mm to 1cm. The foresets show unidirectional dips to the north.

Dunes are the migrating bed-forms that deposit trough cross stratification (Allen, 1963; Harms,et al., 1975). Such bedforms occur in the upper part of the lower flow regime (Simons et al, 1965). The possibility that some of the trough cross beds formed by scour and fill mechanism is indicated by scour and fill structures (Plate 4.1) observed in this facies. Polished sections (Plate 4.3) show water escape (flame) structures which indicate sediment liquefaction and contemporaneous deformation (Lowe, 1975).

b. Thinly-bedded Semi-psammites

Parallel-bedded semi-psammites 3-5cm thick (Plate 4.4) generally overlie the cross-bedded psammites. Such beds are traceable laterally for several metres within the bounds of an outcrop.

Flat beds or horizontal stratification of coarse or micaceous sandstones suggest lower flow regime conditions, whereas an upper flow regime is indicated by the deposition of non-micaceous flat beds of sands with a grain size less than 0.6mm (Harms et al, 1975; Collinson and Thompson 1982). The micas now present in these semi-psammites could have been derived in part from original detrital grains or entirely from the metamorphism of clay minerals present in the original sediment. As the information so far available is insufficient for the purpose of accurate identification of the original source

Plate 4.3:   Water escape "flame" structure in semi-psammite.

Chathalain Formation.   NN 492940.

Plate 4.4 : Thinly-bedded semi-psammite in the Chathalain

Formation.   NN 492946.

Plate 4.3:   Water escape "flame" structure in semi-psammite.
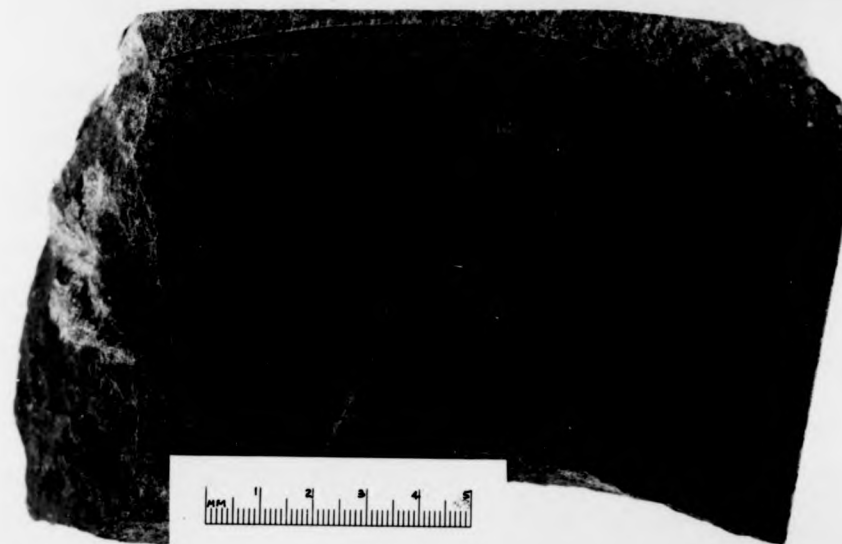   Chathalain Formation.   NN 492940.

Plate 4.4 : Thinly-bedded semi-psammite in the Chathalain
            Formation.   NN 492946.

Plate 4.3: Water escape "flame" structure in semi-psammite.
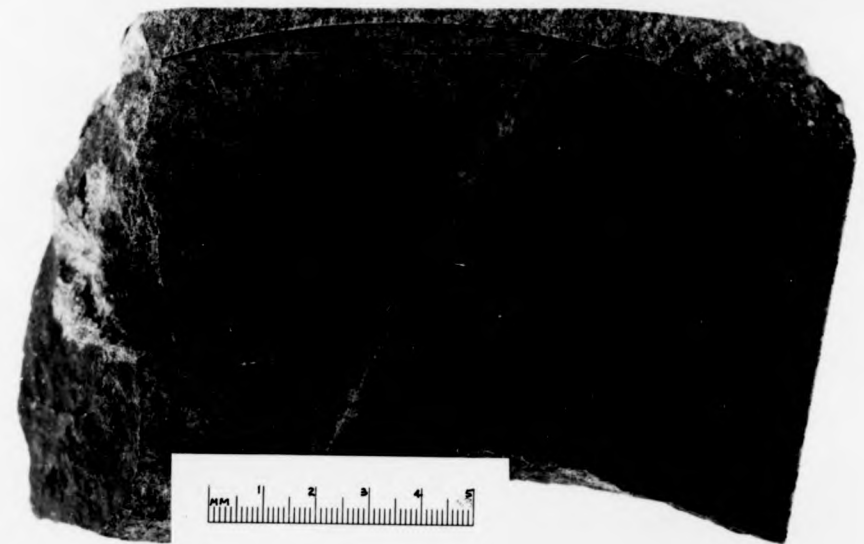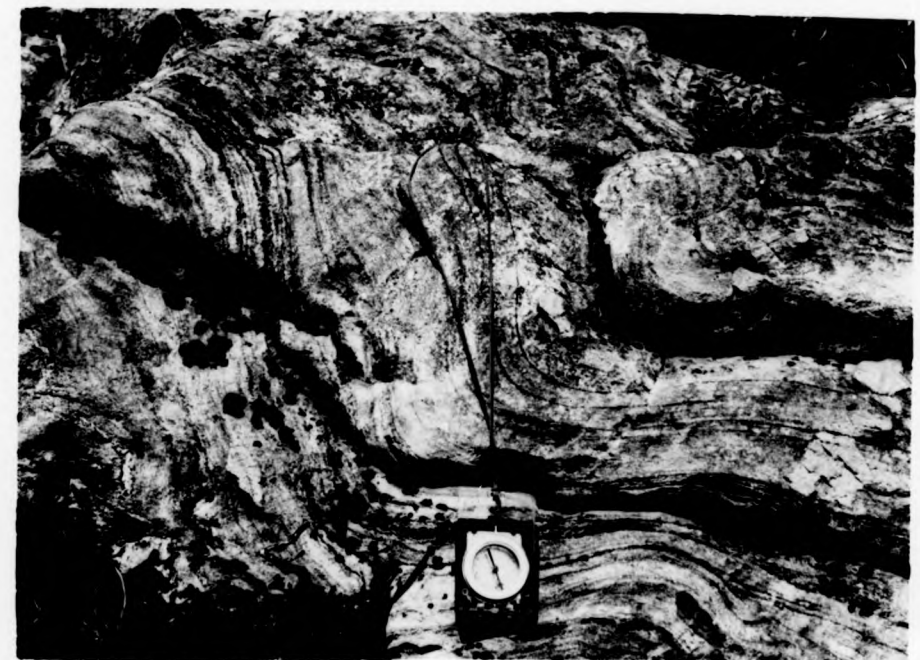
Chathalain Formation.   NN 492940.

Plate 4.4 : Thinly-bedded semi-psammite in the Chathalain

Formation.   NN 492946.

of these micas, this facies cannot be assigned with confidence to either the
low flow or the upper flow regime.

c.   Semi-pelite

A thin semi-pelite band (Plate 2.2) generally overlies the thinly-bedded
semi-psammite. No sedimentary structures were found in this unit; any that
may have existed must have been obliterated by metamorphism, recrystallisation
and deformation

These semi-pelites probably represent metamorphosed siltstones or
silty mudstones and are products of deposition from suspension under low
energy conditions.

d.   Discussion and environmental Interpretation

The fining-up sequences exhibited by the rocks of the Chathalain
Formation indicate deposition in channels (see Plate 4.1) and flanking banks.
The cross-bedded and the parallel-bedded psammites may be interpreted as
channel deposits with the semi-pelites as overbank deposits.

Further evidence is necessary to identify these channels definitely as
either alluvial or marine. Unfortunately, there are no microfossils or any
other environmentally specific indicators.

However as submarine channels are typically associated with turbidites
and other facies of the submarine fan environment (Walker, 1979), and as
no turbidites were seen in this formation it seems that these may not be
submarine channels.

Many workers on ancient fluvial deposits have tried to set up criteria for
differentiating between meandering (high sinuosity) and braided (low sinuosity)
stream deposits. Moody-Stuart (1966) produced the following criteria: sand-
body geometry, type of cross bedding, thickness of the fine-gradined sediment,
and paleocurrent direction variability.

The Chathalain Formation is characterized by abundant large-scale cross
bedding, meagre overbank deposits represented by the thin semi-pelite band,
and very low paleocurrent direction variability. These characteristics therefore

suggest a braided or low sinuosity channel environment for the deposition of the sediments. The question of whether the environment was alluvial or marine remains unresolved.

4.2.2 The Garva Formation

This formation comprises crosslaminated semi-psammites, commonly with overturned foresets, normally graded units and parallel-laminated semi-psammites occasionally with discontinuous psammite laminae and thin semi-pelitic bands.

No sedimentary structures have been observed in the uppermost 200m of the formation adjacent to the Gairbeinn Slide. It is likely that any original sedimentary structures in that section have probably been destroyed by the intense deformation associated with the sliding.

a. Crosslaminated semi-psammite

Foresets of the cross laminae have unidirectional dips to the NNW and set thicknesses vary from 3cm to 4cm (Plate 4.5). Some of the foreset laminae have been variably deformed and overturned (Plate 2.3). That these structures record syn-sedimentary folding of cross lamination is supported by the fact that they are bounded above and below by undisturbed parallel bedding and crosslaminations. The direction of overturning the foresets is the same as that of the foreset dips of undeformed cross lamination found in the same location.

Current scours (Plate 4.1) occur within this facies and indicate high current, energy and channelling (Reineck and Singh, 1980; Miall, 1964).

Overturned foreset laminae are believed to result either from current drag acting on the upper surface of a crosslaminated bed which had been weakened by liquefaction either by a seismic shock, or spontaneously (Allen and Banks, 1972; Collinson and Thompson, 1982), or from frictional drag following the passage of a mass of saturated sand over the surface of a cross-bedded sand without prior liquefaction (Hendry and Stauffer, 1975).

Water escape (flame) structures (Plate 4.6) which are also present in this formation support the case for the liquefaction of the sediments before

Fig. 4.1. A section of the Garva Formation at NN 522947

Laminated semi-psammite
Discontinuous psammite laminae

Normally-graded semi-psammitic units
with thin semi-pelitic tops

Thin epidote-bearing bands

Cross laminated semi-psammite with
current scours and

overturned cross laminae

2m]

Plate 4.5 : Isolated crossbedding in the semi-psammite in
the Garva Formation. Current scours occur at the
top of the thin psammite band.   NN 522947.
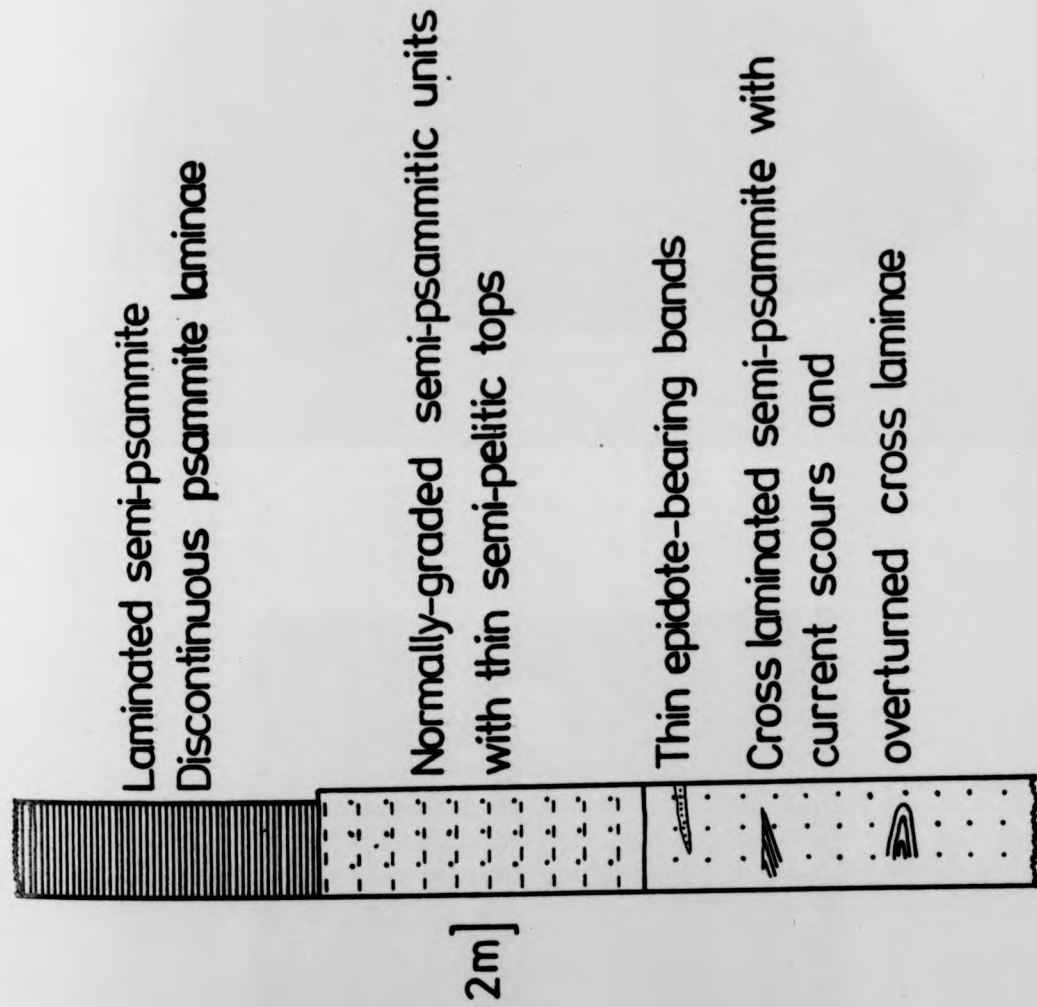
Plate 4.6 : Water-escape "flame" structure in the semi-psammite
of the Garva Formation.   NN 503947.

Plate 4.5 : Isolated crossbedding in the semi-psammite in
the Garva Formation. Current scours occur at the
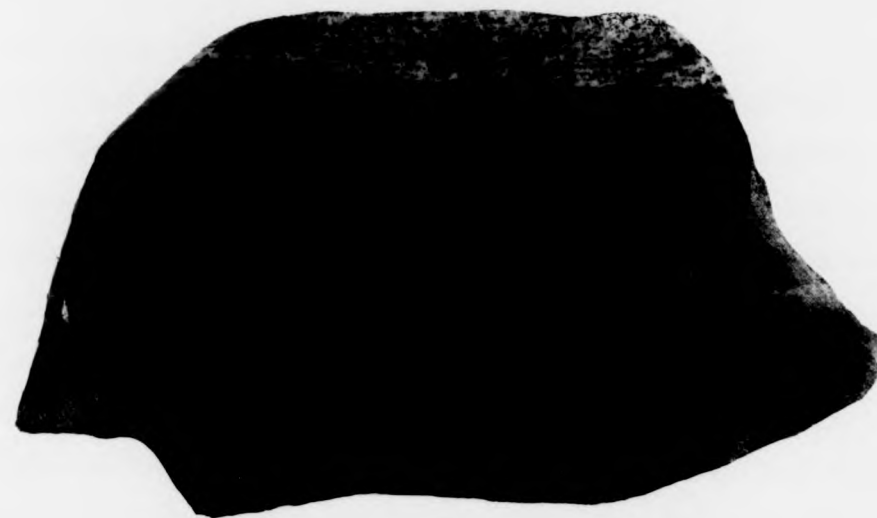top of the thin psammite band. NN 522947.

Plate 4.6 : Water-escape "flame" structure in the semi-psammite
of the Garva Formation. NN 503947.

Plate 4.5 : Isolated crossbedding in the semi-psammite in
the Garva Formation. Current scours occur at the
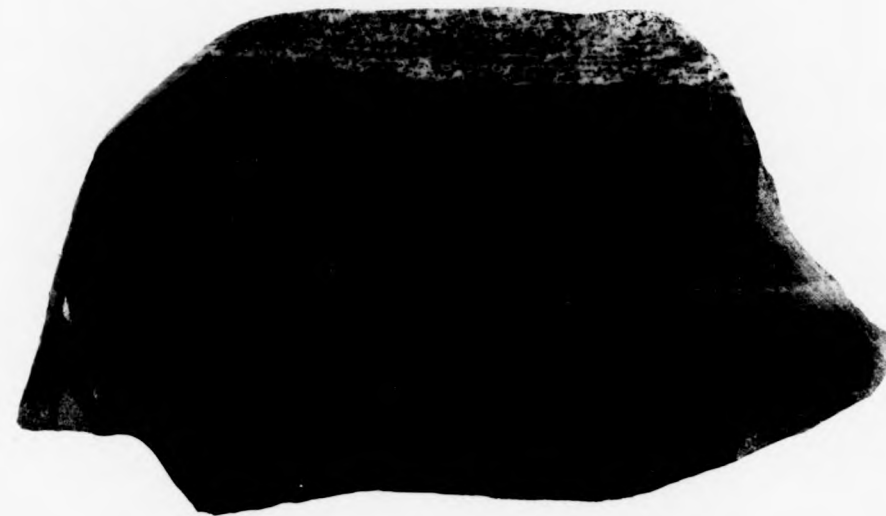top of the thin psammite band.   NN 522947.

Plate 4.6 : Water-escape "flame" structure in the semi-psammite
of the Garva Formation.   NN 503947.

subsequent dewatering (Miall, 1984; Lowe, 1975).

b.   Normally graded semi-psammite

This facies overlies the cross laminated facies. It comprises a cyclic sequence of graded beds 20-30 cm thick passing up from semi-psammite at the base to a thin , 2-10 cm thick, semi-pelitic top (Fig 4.1).

These beds are products of deposition from waning, sediment-laden currents (Reineck and Singh, op. cit.)

c.   Laminated Semi-psammite

This facies commonly overlies the graded and cross-laminated facies (Fig. 4.1). It is characterised by alterations of parallel laminated semi-psammite (up to 5 mm thick) occasionally containing discontinuous psammite laminae which resemble starved ripples (Pettijohn, 1975; Reineck and Singh, 1980

This facies was probably produced in the plane-bed phase òf the upper flow regime (Simons et al, 1965). The discontinuous psammite laminae are the result of insufficient coarse sand material supply, and are also typical of upper flow regime conditions (Simons et al, op cit.).

d.   Discussion and environmental interpretation

Overturned cross-bedding is common in fluvial and deltaic environments (Allen and Banks, 1972; Reineck and Singh, 1980; Miall, 1984). In modern environments, it has been reported from the alluvial bars of the Brahmaputra River (Coleman, 1969).

Thin graded units occasionally occur in shallow-water sediments (Reineck and Singh, op. cit.). Sporadic occurrences of graded beds may be the result of deposition from the last phases of a heavy storm or by periodic silting of deltaic distributaries.

Starved ripples or lenticular bedding may be found in several environments including intertidal flats (Reineck and Singh, op. cit.) levees on alluvial floodplains (Miall 1984), and channel-mouth facies of submarine deepsea fans

(Mutti, 1977).

The unidirectional nature of the ripple cross-laminae found in this formation (see Figs 4.1 and 4.7) rules out tidal environments. The relatively highly oxidised nature of the rocks (see Chapter 7) which was probably inherited from the original sediments (Chinner, 1960) suggest that the depositional environment of the Garva Formation may have been very shallow water, occasionally subaerial. These observations tend to suggest that the alluvial environment was the most probable. They suggest that the Garva Formation sediments were probably laid down in the topographically higher parts of the alluvial system, perhaps on the floodplains.

4.2.3 Coire nan Laogh Formation

This formation consists dominantly of semi-pelitic rocks. The top of it becomes more semi-psammitic with increasing abundance of intercalated thin semi-psammitic units 5-10 cm thick. With the dominance of the semi-psammitic units it grades into the Transitional Semi-psammite.

a.   Semi-pelite

This comprises over 70% of the formation above the tectonic contact with the Garva Formation (Plates 2.4a and 2.4b). Intense deformation and repeated metamorphic recrystallisation have completely obliterated any sedimentary structures that may have been originally present in this facies.

The semi-pelites comprising this facies were probably fine-grained sediments deposited in a low energy environment. The thickness of the semi-pelites (>250m) suggests that the parent sediments were deposited in a steadily subsiding basin from a constant terrigeneous supply. Pettijohn (1975) suggests that thick argillaceous deposits were typical of geosynclinal piles while Potter et al (1980) have suggested that thick argillaceous deposits are more likely to be formed in a pro-deltaic environment.

b.   Thinly-bedded semi-psammite.

Semi-psammite units about 2-10 cm thick occur sporadically within the semi-pelite. These semi-psammite units are in places highly deformed and boudinaged (see Chapter 5) so that primary sedimentary relationships have been

obliterated.   They may occasionally form the loci for calc-silicate formation.

These units probably represent thin sand layers deposited at intervals of higher current activity separated by much longer periods of fine sediment deposition from suspension.

c.   Discussion and environmental interpretation

The absence of sedimentary structures in this formation makes sedimentological interpretation difficult. However, as observed previously the great thickness of the semi-pelites (>250m) suggests that their parent fine-grained sediments were deposited in a steadily subsiding marine basin (Pettijohn 1975, Potter et al, 1980).

Also, petrographic evidence, in the form of sporadic tourmaline occurrences in these semi-pelites suggest that the parent argillaceous sediments had adsorbed some boron, which is generally regarded as preferentially concentrated in marine environments (Degens et al, 1958).   Hence a marine environment of deposition is inferred for this formation.

4.2.4   The Transitional Semi-psammite

This forms the transition from the dominantly semi-pelitic Coire nan Laogh Formation to the largely semi-psammitic Glendoe Formation.

a.   Thinly-bedded semi-psammite with thin semi-pelite

This formation consists dominantly of thin, normally graded semi-psammite units 5-15 cm thick with semi-pelitic tops, occasionally associated with thin semi-pelitic units 5-10 cm thick.

b.   Discussion and environmental interpretation

This transitional formation probably reflects an increasing supply of coarser, sandy material implying a long-term change in sediment supply. This facies may have originated in a variety of environments; for example in shelf environments where mud deposition is interupted by sand deposition from waning currents or in mixed intertidal flats (Reineck and Singh, 1980) where sands deposited by current and wave activity, alternate with mud or silts deposited during slack water conditions.   However the absence of wave or tide-generated structures such as wave-ripples or herringbone cross-

Plate 4.7 : "Massive" channel semi-psammites in the Glendoe
          Formation containing lenticular pods of
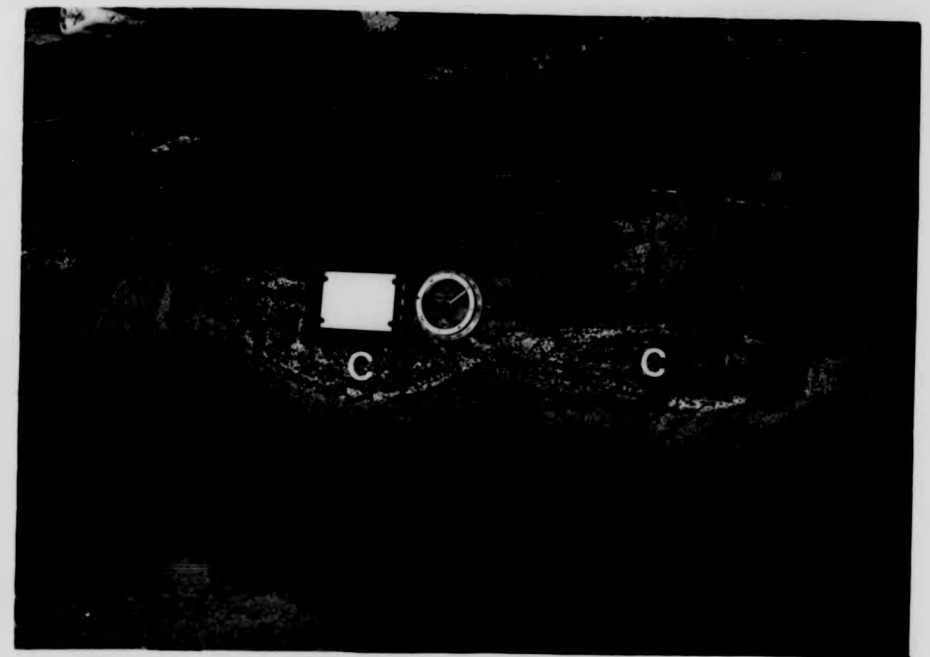   calc-silicates,C.

Plate 4.8 : Turbidite bed in the Glendoe Formation showing delayed
          grading from semi-psammite at the base to a thin
          semi-pelitic top. Bouma A, C, D, E divisions
          Note also the thin lenticular calc-silicate bands,C.
          NN 462850.
          Hammer is 60cm long.

Plate 4.7 : "Massive" channel semi-psammites in the Glendoe
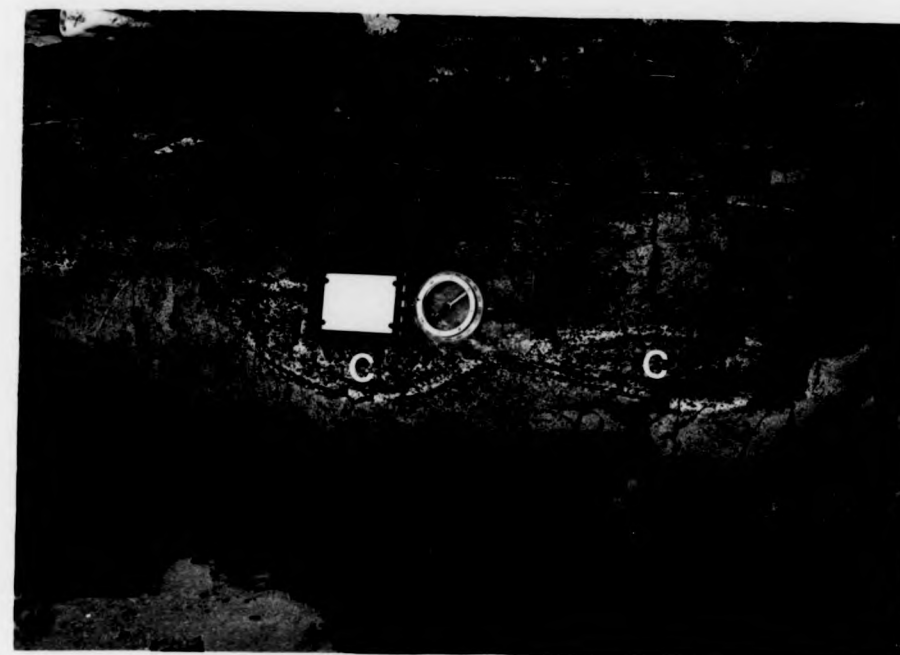Formation containing lenticular pods of
calc-silicates,C.

Plate 4.8 : Turbidite bed in the Glendoe Formation showing delayed
grading from semi-psammite at the base to a thin
semi-pelitic top. Bouma A, C, D, E divisions
Note also the thin lenticular calc-silicate bands,C.
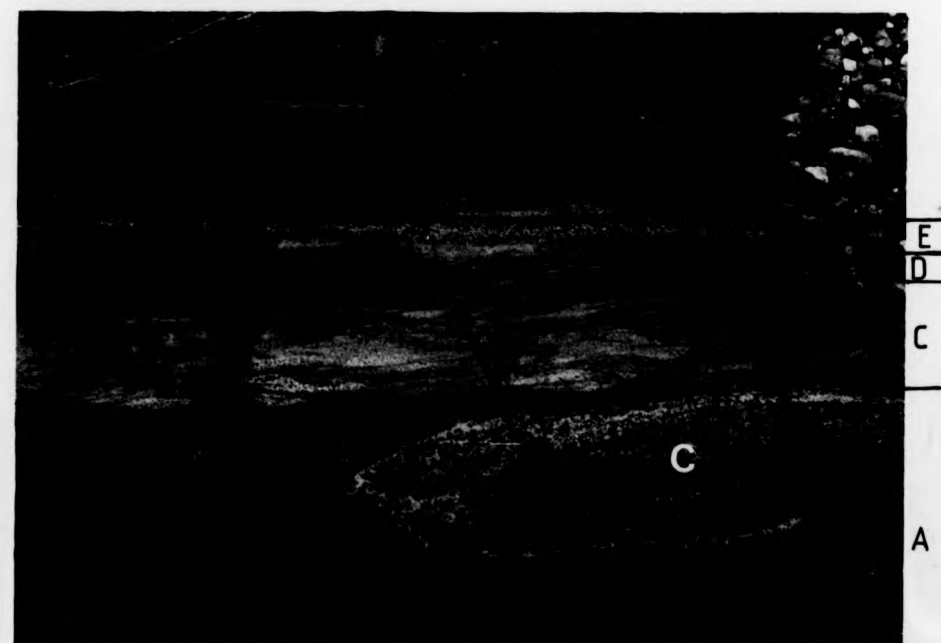NN 462850.
Hammer is 60cm long.

Plate 4.7 : "Massive" channel semi-psammites in the Glendoe
           Formation containing lenticular pods of
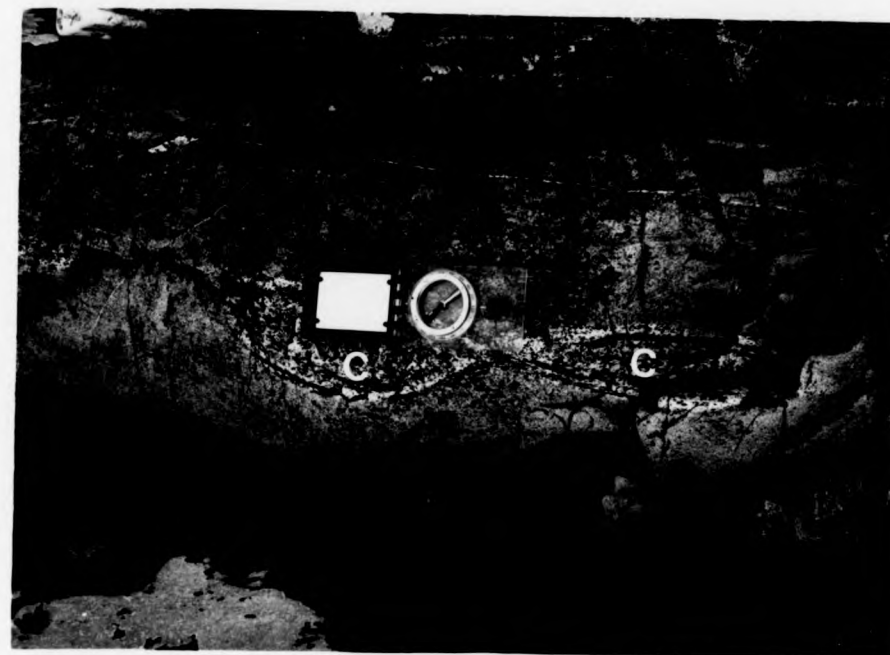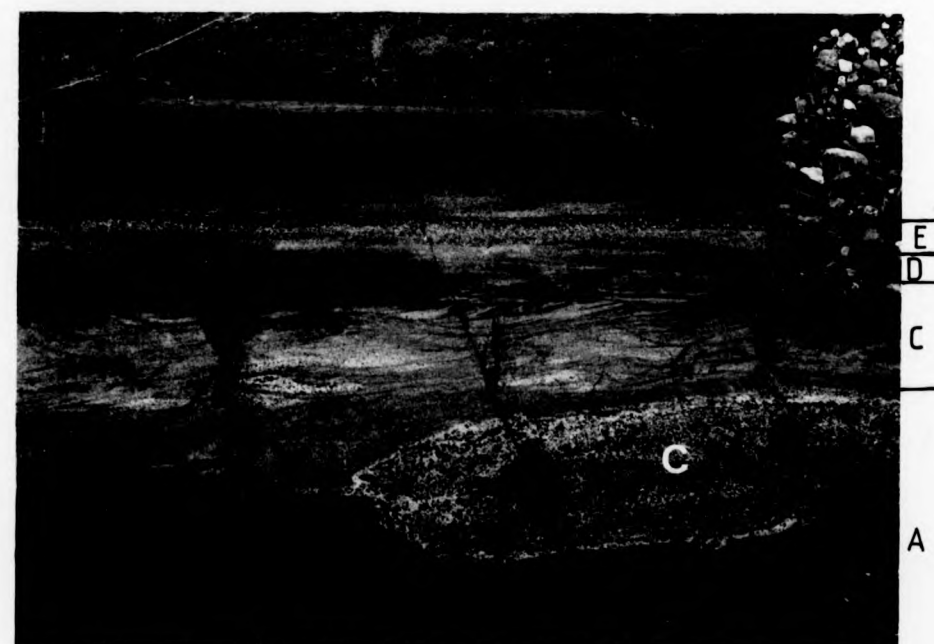   calc-silicates,C.

Plate 4.8 : Turbidite bed in the Glendoe Formation showing delayed
           grading from semi-psammite at the base to a thin
           semi-pelitic top. Bouma A, C, D, E divisions
           Note also the thin lenticular calc-silicate bands,C.
           NN 462850.
           Hammer is 60cm long.

lamination rules out tidal environments, and hence a subtidal marine environment is indicated.

4.2.5 The Glendoe Formation

This very thick formation consists mainly of semi-psammites, occurring as sequences of graded beds with cross laminated and semi-pelitic tops (Plates 4.8, 4.12) convolute laminated beds (Plate 4.10) massive beds (Plates 4.7, 2.8) and a wave-rippled unit (Fig. 4.4).

a. Graded Semi-psammites

These occur as laterally persistent beds ranging from 10-100 cm in thickness. The beds commonly show normal grading from psammitic bases to semi-pelitic tops (Plates 4.8 and 4.12), illustrating a compositional change believed to reflect original sedimentary grading. These graded units may pass up into current rippled and finally semi-pelitic tops (see also Plates 4.8 and 4.12). These features are very well-displayed on the shore of Loch Laggan at NN 489910.

Graded bedding may be produced by several processes, which include sedimentation of suspended material, deposition in the last phases of a heavy flood, periodic silting of deltaic distributaries, and generally, deposition by waning currents (Reineck and Singh, 1980).

b. Cross-laminated semi-psammite

Cross-lamination is commonly developed in the semi-psammites of the Glendoe Formation. Most of the current-ripple lamination is found in the upper part of graded beds (Plate 4.8). Ripple-drift cross-lamination (Plate 4.9) is also occasionally developed in the semi-psammites. Cross-laminae set thicknesses vary from 2 to 6 cm. They possess unidirectional foreset dips to the NE (Fig 4.7).

The current-rippled units occasionally show convolute laminations resulting from syn-sedimentary deformation (Plate 4.10) These convolute folds, with average wavelengths of 15cm and amplitudes of 7 cm, generally overlie parallel laminations of the semi-psammite.

Plate 4.9 : Climbing-ripples in semi-psammite of the Glendoe
Formation.   Semi-psammite has been intruded by late
pegmatite.of the Loch Laggan Complex.
NN 462850.   Hammer is 60 cm long.

Plate 4.10 ; Convolute-bedded semi-psammite, Glendoe Formation
cut by pegmatites of the Loch Laggan complex.
NN 462850.   Hammer is 60cm long

Plate 4.9 : Climbing-ripples in semi-psammite of the Glendoe
Formation. Semi-psammite has been intruded by late
pegmatite.of the Loch Laggan Complex.
NN 462850. Hammer is 60 cm long.

Plate 4.10 ; Convolute-bedded semi-psammite, Glendoe Formation
cut by pegmatites of the Loch Laggan complex.
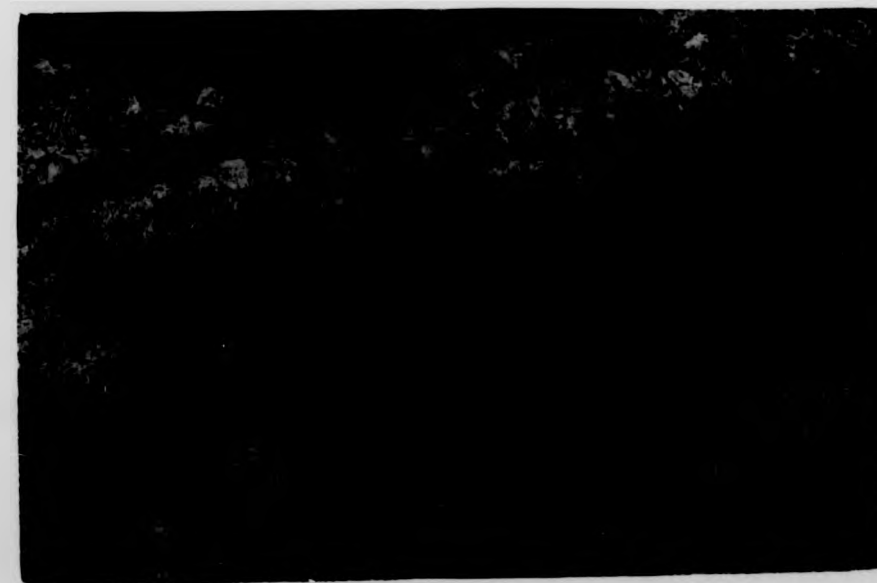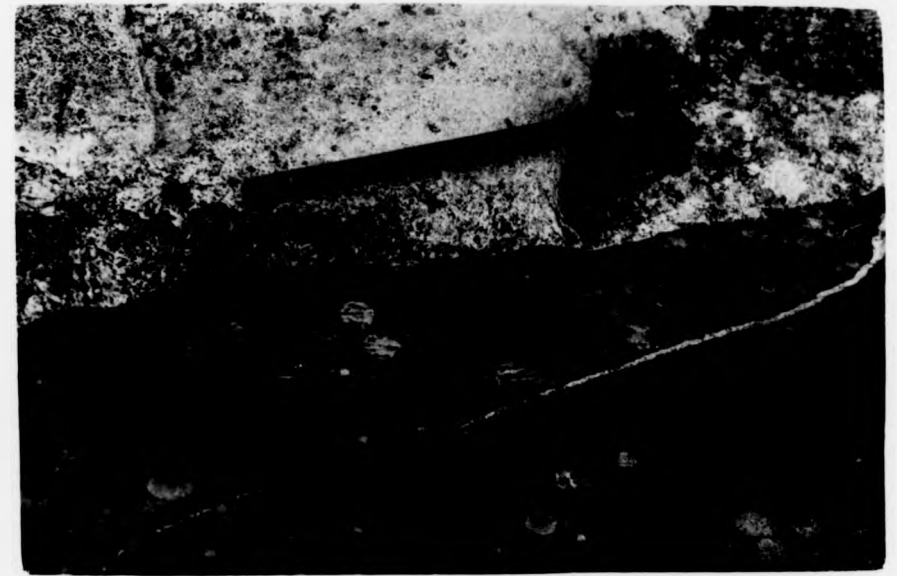NN 462850. Hammer is 60cm long
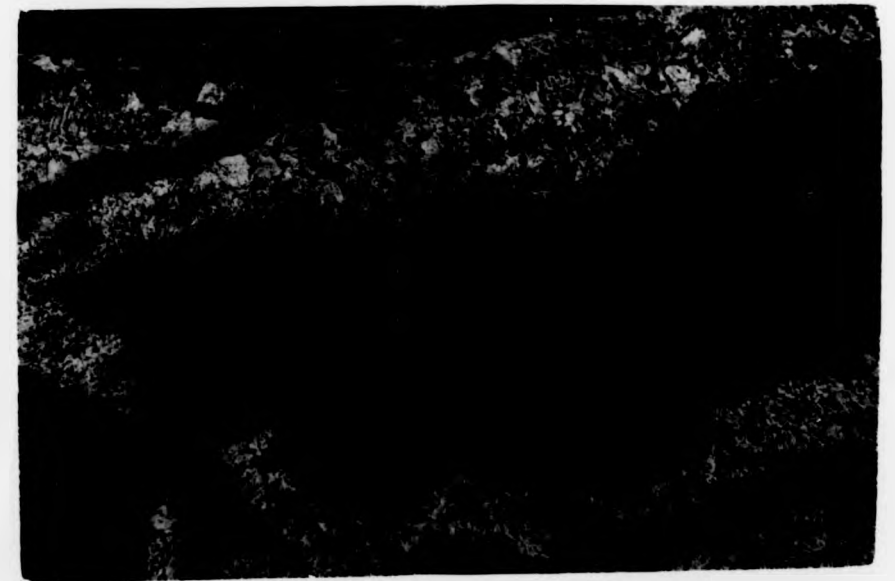
Plate 4.9 : Climbing-ripples in semi-psammite of the Glendoe
          Formation.   Semi-psammite has been intruded by late
          pegmatite of the Loch Laggan Complex.
          NN 462850.   Hammer is 60 cm long.
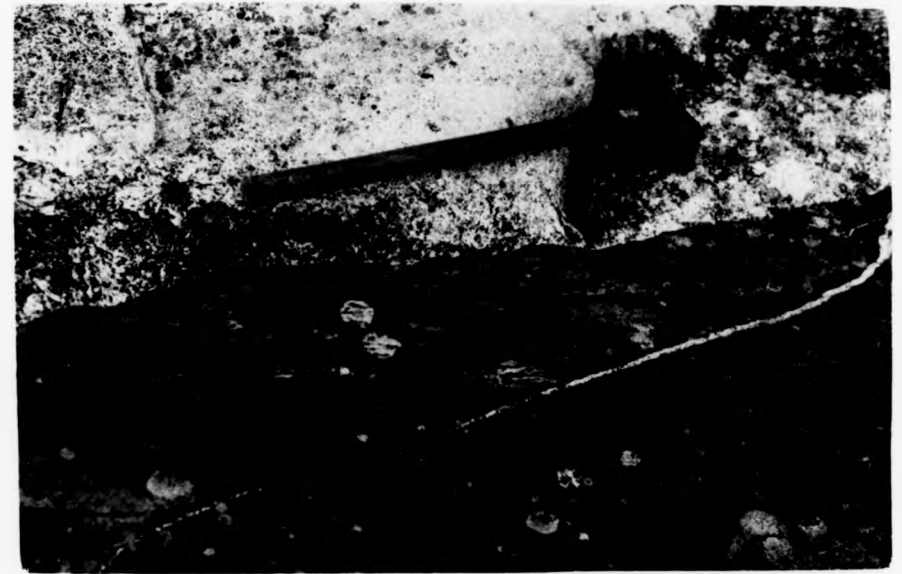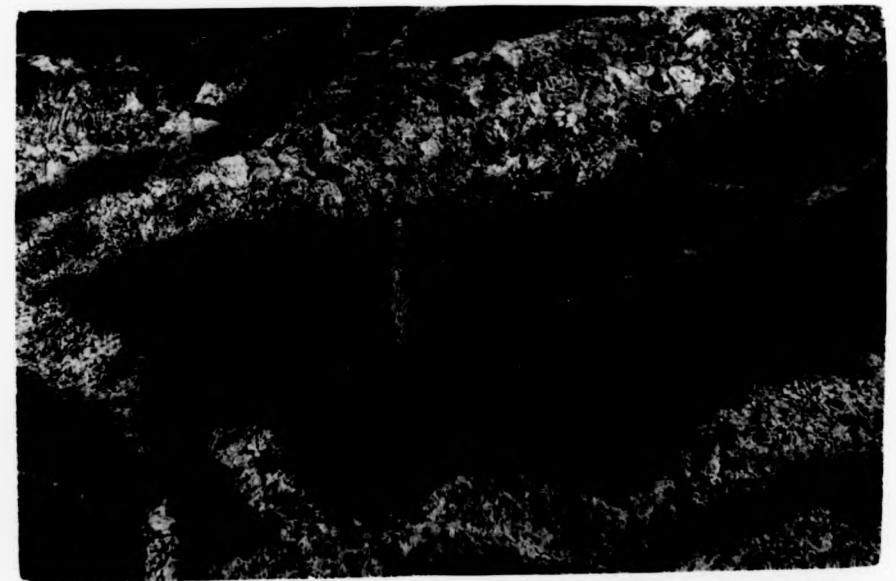

Plate 4.10 ; Convolute-bedded semi-psammite, Glendoe Formation
           cut by pegmatites of the Loch Laggan complex.
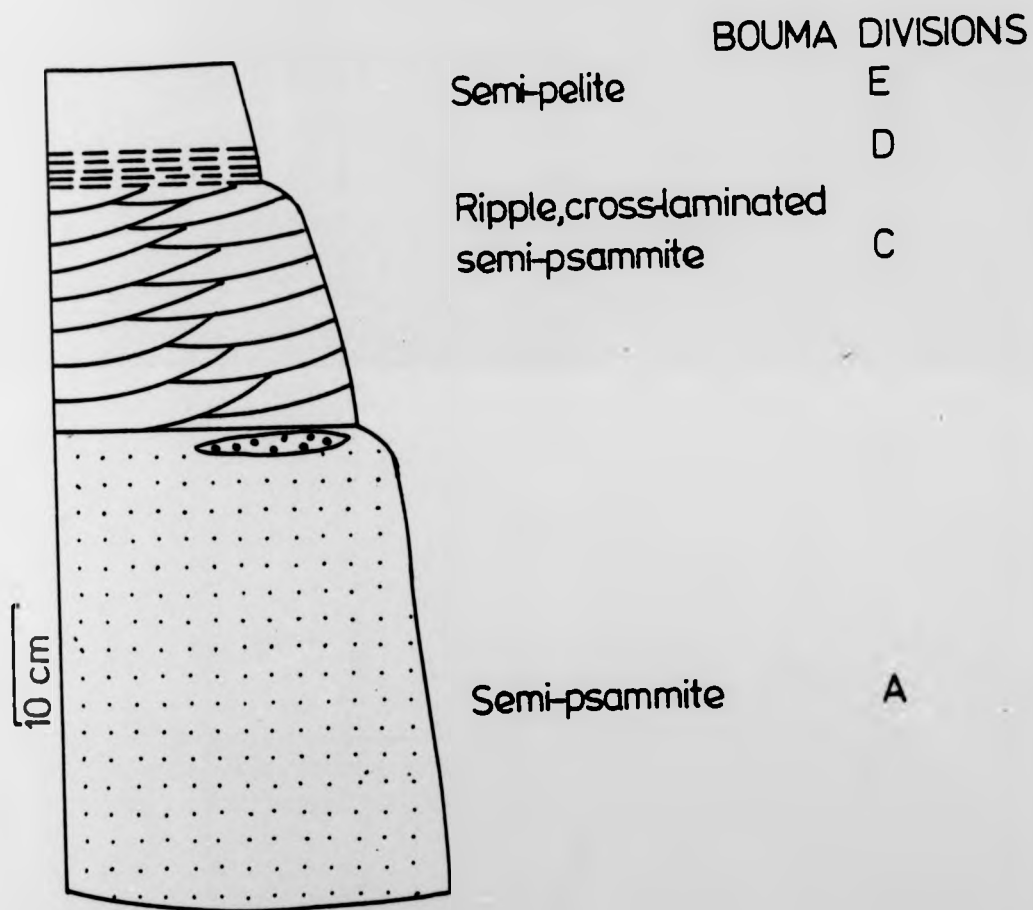   NN 462850.   Hammer is 60cm long

Fig. 4.2: A turbidite sequence in the Glendoe Formation
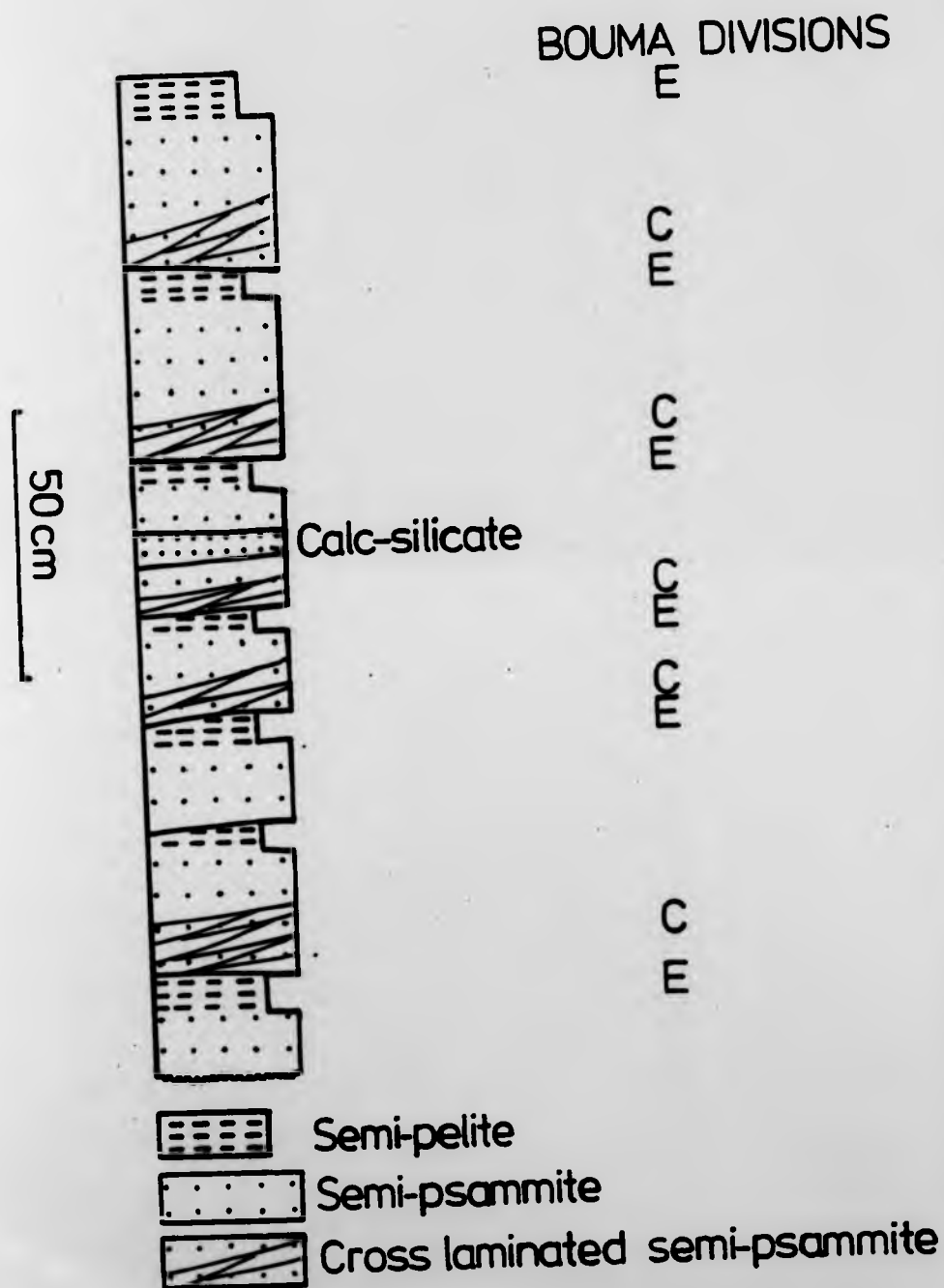
(NN 462850)

Fig. 4.3: Turbidite sequences in the Glendoe Formation
(NN 462850)

The occurrence of ripple cross-lamination indicates low current velocities and is typical of sediments in the size range of fine-to medium-grained sands and coarse silts. Climbing ripple laminations (type A of Joplin and Walker, 1968) may imply moderate suspended load to bed-load ratios. Convolute lamination provides evidence of rapid deposition (Collinson and Thompson 1982) and is generally well-developed in fine-grained, non-cohesive sediments such as fine sands or silty, fine sands (Reineck and Singh, 1980). Liquefaction of a sediment may be a critical factor in its formation (Johnson, 1977); and Lowe (1975) suggested that some of the convolute bedding in turbidites were water-escape structures produced during de-watering and consolidation. Kuenen (1953) however, believed that convolute laminations were caused by the deformation of ripples as a current drag caused suction over the crests and a downward pressure in the troughs.

c. Massive semi-psammite

This consists of units of medium - to thickly-bedded semi-psammites, 30-50 cm thick, with sharp bases (Plates 2.8 and 4.7). No internal structures have been observed in these beds, but they occasionally show a thinning-upwards of the massive beds. They are commonly associated with graded beds.

Such massive beds have been interpreted as deposits of grain-flow processes (Middleton and Hampton, 1976; Reineck and Singh, 1980) under high energy currents.

d. Wave-rippled semi-psammite

This rare sedimentary facies has only been found in semi-psammites near the Allt Crunachdan at NN 536922. Tabular laminae are less than 1mm thick with azimuthal discordances of about 120°. Overall fore-set laminae dips are however, to the NW, similar to some current ripple directions obtained in the other parts of the Glendoe Formation. This facies is spatially associated with massive semi-psammite beds which show a general thinning- and fining-upwards sequence.

The wave-rippled facies was probably deposited by an oscillatory flow or current.

Thinly-bedded and striped semi-psammite

2m]

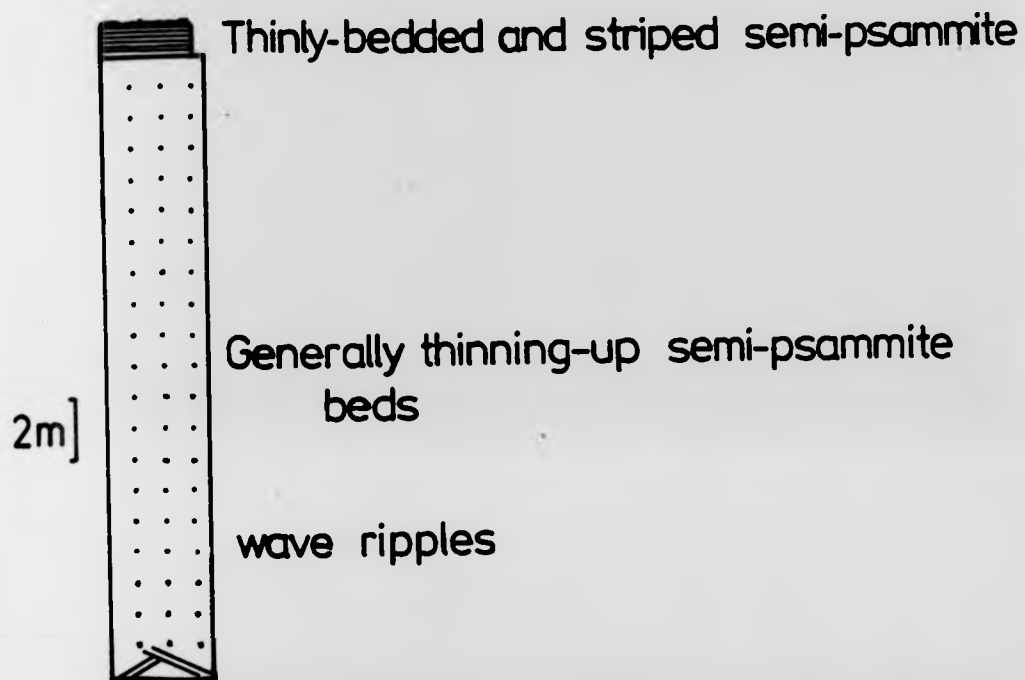Generally thinning-up semi-psammite
beds

wave ripples

Fig.4.4: Wave-rippled and thinning-upwards facies in the Glendoe Formation

Plate 4.11 : Thin calc-silicate band in semi-psammite.

Monadhliath Formation, NN 435877.

(C = calc-silicate)

Plate 4.12 : Cross-laminations deformed around lenticular

calc-silicate pods in the Glendoe Formation.

NN 462850.

Lens cap is 5.5cm in diameter.

(C = calc-silicate)

Plate 4.11 : Thin calc-silicate band in semi-psammite.

Monadhliath Formation, NN 435877.

(C = calc-silicate)

Plate 4.12 : Cross-laminations deformed around lenticular

calc-silicate pods in the Glendoe Formation.

NN 462850.

Lens cap is 5.5cm in diameter.

(C = calc-silicate)

Plate 4.11 : Thin calc-silicate band in semi-psammite.

Monadhliath Formation, NN 435877.
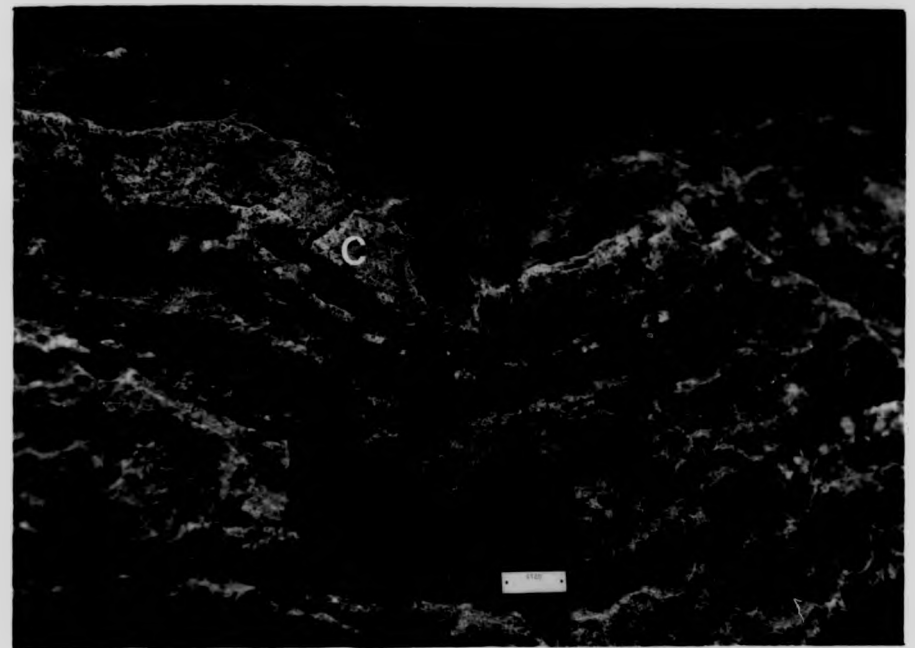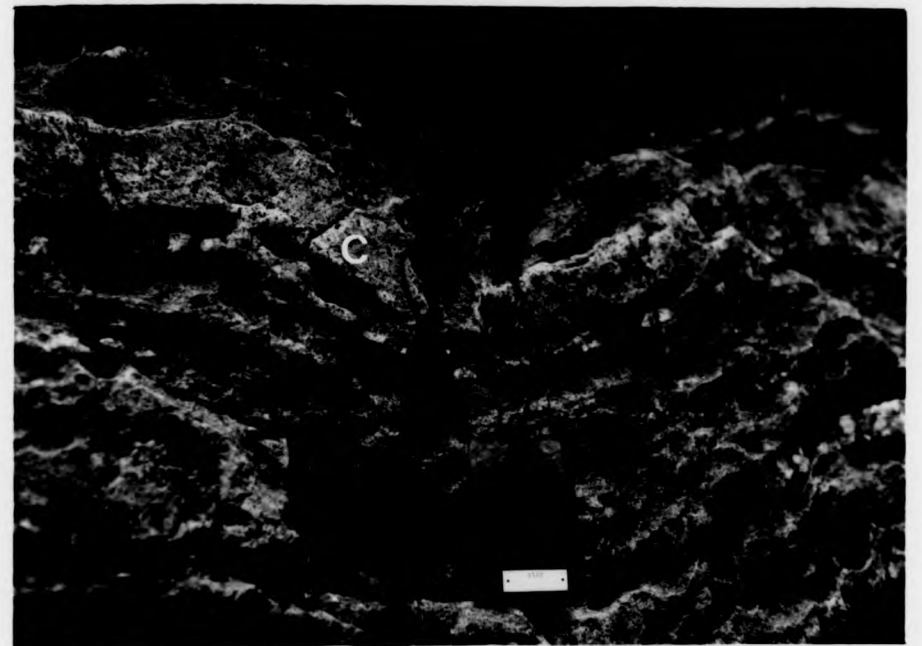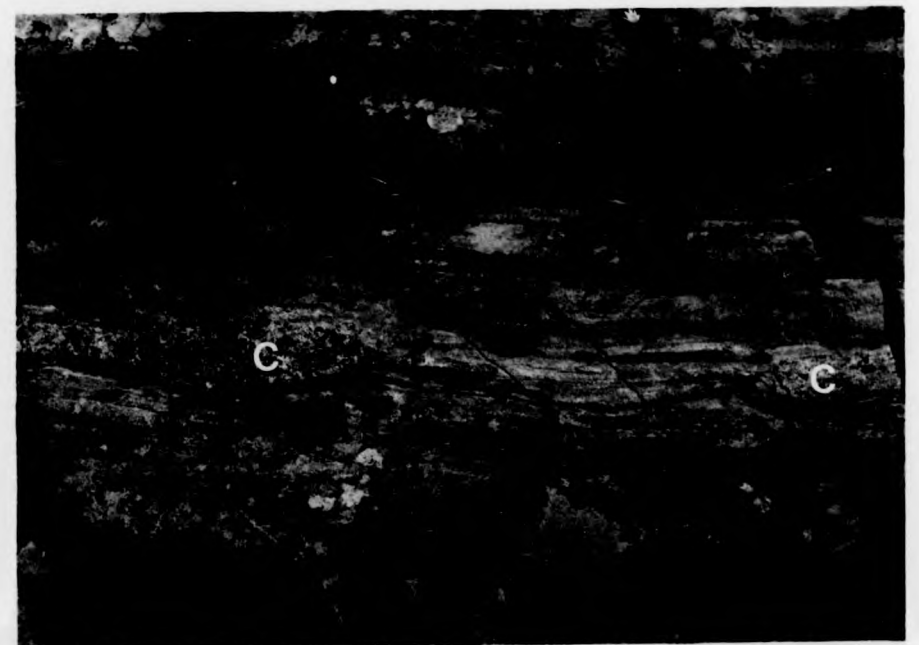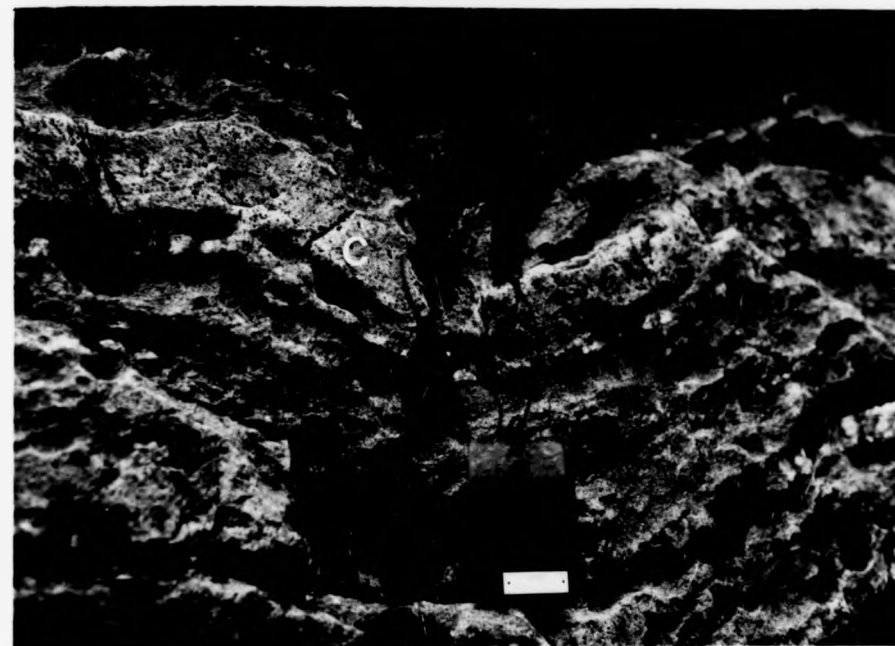
(C = calc-silicate)

Plate 4.12 : Cross-laminations deformed around lenticular

calc-silicate pods in the Glendoe Formation.

NN 462850.

Lens cap is 5.5cm in diameter.

(C = calc-silicate)

e.   Discussion and environmental interpretation

The various facies recognised in the Glendoe Formation may be correlated with specific environments of the submarine fan model of Walker (1979) (See Fig. 4.5).

The massive semi-psammites are inferred to have been massive sandstones which occur as channel fills. The thinning and fining-upwards sequences associated with this facies are typical of channel-fill sequences (Walker, op. cit).   The tabular, wave-rippled semi-psammite may represent rare diverging currents within the channel (Miall, 1984). Mutti et al. (1978) have also suggested that these thinning- and fining-upwards sequences may result from progressive channel abandonment with the deposition of thinner and finer beds from decreasing flows in the channels.

Graded beds are generally inferred as products of turbidity currents (Kuenen and Migliorini, 1959), and thick ones are typical of deep-water sandstones (Pettijohn, 1975).   The complete Bouma turbidite sequence (Bouma, 1962) has not been observed in any unit of the Glendoe Formation but most beds show the A (graded or massive), C(ripple-laminated with occasional convolute laminations), D and E (fine-grained, semi-pelitic) divisions (Figs 4.3 and 4.4 and Plates 4.9, 4.10 and 4.12), while the B (laminated) division is also shown by some beds (Plate 4.8, Fig 4.3) which may lack the C division.

These thick, graded beds may be equated with the proximal turbidites of the suprafan lobes (Walker, op. cit.), while the thinner units, especially those observed in the transition to the Monadhliath Formation may be lower fan distal turbidites.

4.2.6   The Monadhliath Formation

The base of this formation is marked by a thick sequence of massive psammitic beds (Plate 4.13) which passes up into graded semi-psammitic beds (Plate 4.4) with interbedded semi-pelitic bands.
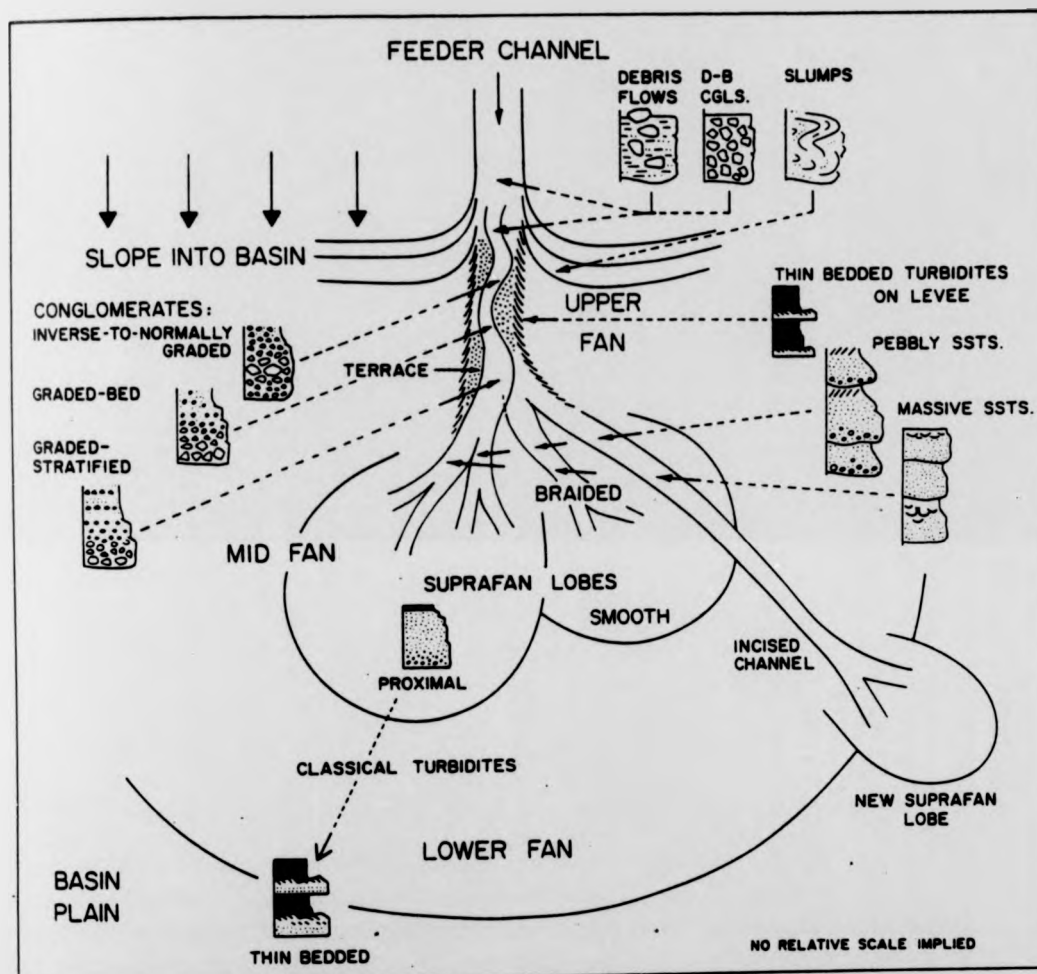
Fig. 4.5: The submarine fan model of Walker (1979)

Plate 4.13 : Amalgamated psammite beds probably representing
channel deposits. Monadhliath Formation.
NN 437867. Hammer is 33 cm long.

Plate 4.14 : A turbidite sequence showing the Bouma A,C,D and E
divisions. Monadhliath Formation. NN 450902.
Hammer is 60 cm long.

Plate 4.13 : Amalgamated psammite beds probably representing
channel deposits. Monadhliath Formation.
NN 437867. Hammer is 33 cm long.

Plate 4.14 : A turbidite sequence showing the Bouma A,C,D and E
divisions. Monadhliath Formation. NN 450902.
Hammer is 60 cm long.

Plate 4.13 : Amalgamated psammite beds probably representing
channel deposits. Monadhliath Formation.
NN 437867.   Hammer is 33 cm long.

Plate 4.14 : A turbidite sequence showing the Bouma A,C,D and E
divisions. Monadhliath Formation. NN 450902.
Hammer is 60 cm long.

MONADHLIATH FORMATION

Interpretation

Semi-pelite — Turbidite

Semi-psammite

Semi-pelite

Turbidite

Semi-psammite

Calc-silicate

Massive
Psammite — Submarine
channel

Semi-pelite

Semi-psammite — Turbidite

Massive
Psammite — Submarine
channel

40 cm

Fig. 4.6: A section of the Monadhliath Formation (NN 450902)

a.   Graded units

These comprise graded units varying in thickness from 10-100 cm, inter-
bedded with semi-pelite, semi-psammites and psammites (Plates 4.13 and 4.14,
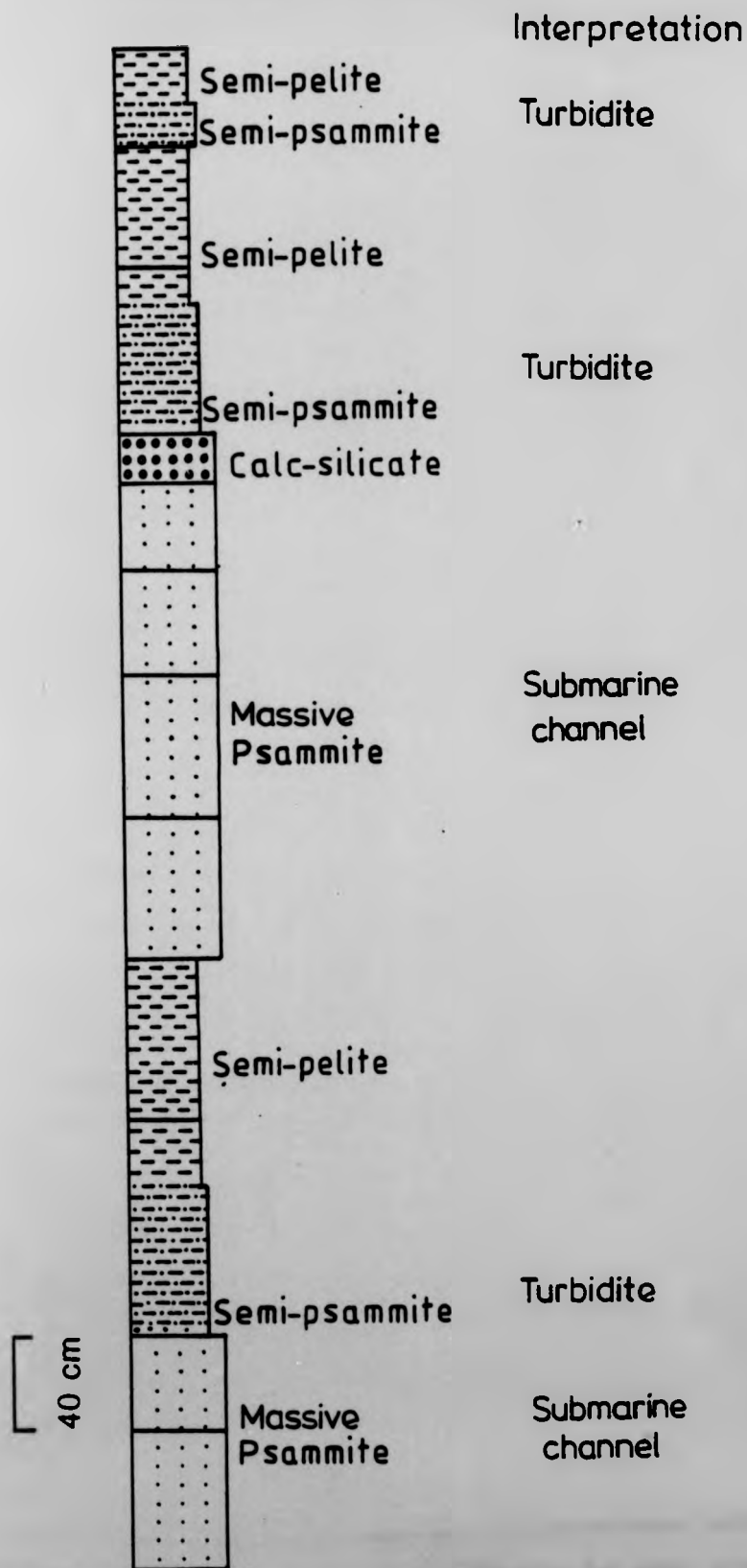Fig 4.6). The thicker graded beds occur in the basal section of the formation
with sharp psammitic or semi-psammitic bases passing into semi-pelite and/or
pelite at the top. This variation in thickness may indicate the proximality
of the units at the base of the formation compared to the stratigraphically
higher ones. The complete Bouma turbidite sequence (Walker, 1979) is
often not developed as most beds show only the A and E (the graded sandstones
and the pelitic) divisions respectively. However, many units show the A,C
(rippled), D (parallel laminated) and E divisions (Plate 4.6b).

b.   Massive psammites

Massive psammitic units which often possess sharp bases, but which
occasionally show some beds merging into one another, occur intermittently
in the Monadhliath Formation associated with the graded units. These units
(Plate 4.13) vary in thickness from 20-50 cm, the thicker units occurring in
the basal section of the formation.

These units probably represent massive sandstones deposited by grain—
flow processes (Middleton and Hampton, 1976; Reineck and Singh, 1980) similar
to those found in the Glendoe Formation.

c.   Semi-pelite and pelite

This lithofacies is intermittently found as 25-40 cm thick beds
interbedded with the turbidites (top of Plate 4.14) No tractional sedimentary
structures have been observed in this lithofacies.

This lithofacies represents metamorphosed silty mudstone. If the absence
of tractional sedimentary structures is an original feature of the sediments,
then deposition from suspension in quiet waters from a steady supply of fine
sediments is indicated.

d.   Discussion and environmental interpretation

The various facies described above from the Monadhliath Formation may be
equated with specific environments of the submarine fan model of Walker (1979),
(Fig 4.5).

The massive units with irregular merging contacts are interpreted as channel fills (Walker,op. cit). They may represent amalgamated beds which resulted from channel shifting and the building-up of coalesced channels.

The thicker graded units in the lower section of the formation represent proximal turbidites of the suprafan lobes, while the thinner units near the top of the formation are equated with distal turbidites of the fan fringe or lower fan (Walker, op.cit).

The semi-pelites and pelites probably represent mudstones deposited on channel levees or hemi-pelagic deposits laid down from suspension between periods of classical turbidite deposition.

4.2.7 The Carn Leac Formation

The basal section of this formation comprises thinly-bedded psammites and semi-psammites, 5-30 cm thick and semi-pelitic units, 5-10 cm thick. The psammitic and semi-psammitic units occasionally show poor grading and cross-lamination (Plate 4.16).

a. Rippled semi-psammite and cross-bedded psammites

This unit comprises thin, discontinuous, lenticular semi-psammite bands 40-60 cm long and 0.5-2cm thick which form symmetrical ripples in a more semi-pelitic band (Plate 4.15). Associated with these units (Plate 4.15) are low-angle cross-bedded and cross laminated psammites occurring as thin, lenticular bands 60-200 cm long and 4-15 cm thick separated by thin semi-pelitic partings.

The presence of ripples indicate lower flow regime (Simons et al 1965) conditions during the deposition of inferred silty units with the thin, discontinuous coarser units suggesting depletion of sand supplies probably associated with fluctuating current conditions (Reineck and Singh, 1980).

By contrast, the cross-bedded, lenticular psammites indicate higher flow regime conditions as well as increased supply of sand under tractional transport.

Plate 4.15 : Discontinuous, lenticular bedded psammites and
semi-psammites. Carn Leac Formation NN 413822.
Hammer is 60 cm long.

Plate 4.16 : Rippled semi-psammites in the Carn Leac Formation.
NN 433883.   The bigger key is 7.5 cm long.

Plate 4.15 : Discontinuous, lenticular bedded psammites and
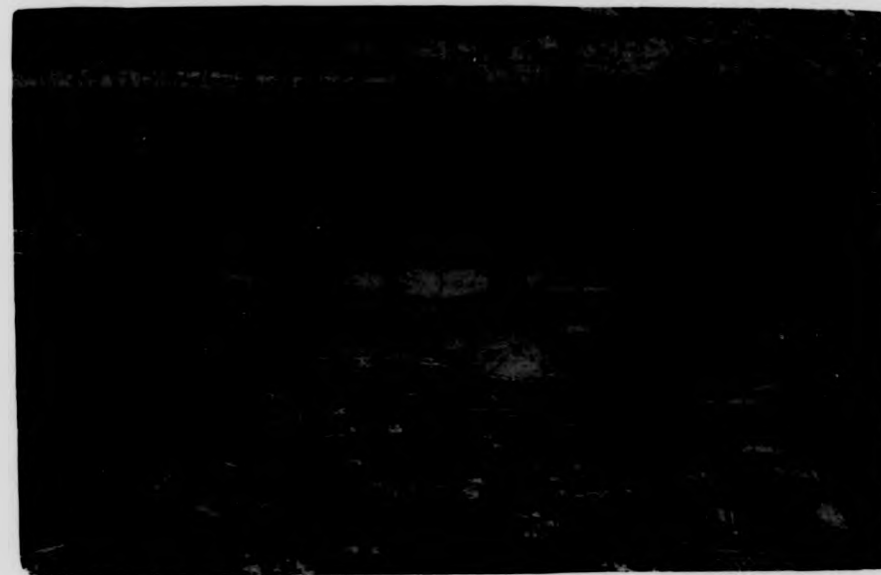semi-psammites. Carn Leac Formation NN 413822.
Hammer is 60 cm long.

Plate 4.16 : Rippled semi-psammites in the Carn Leac Formation.
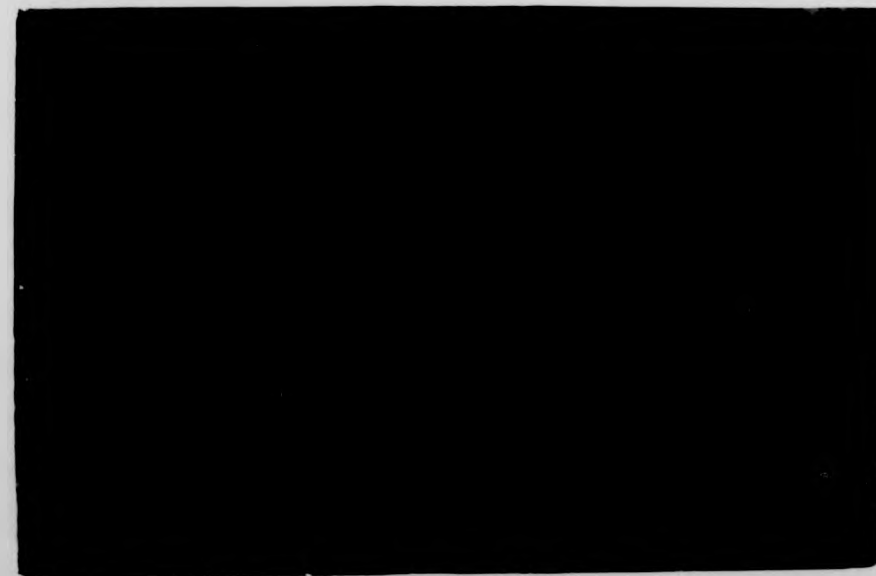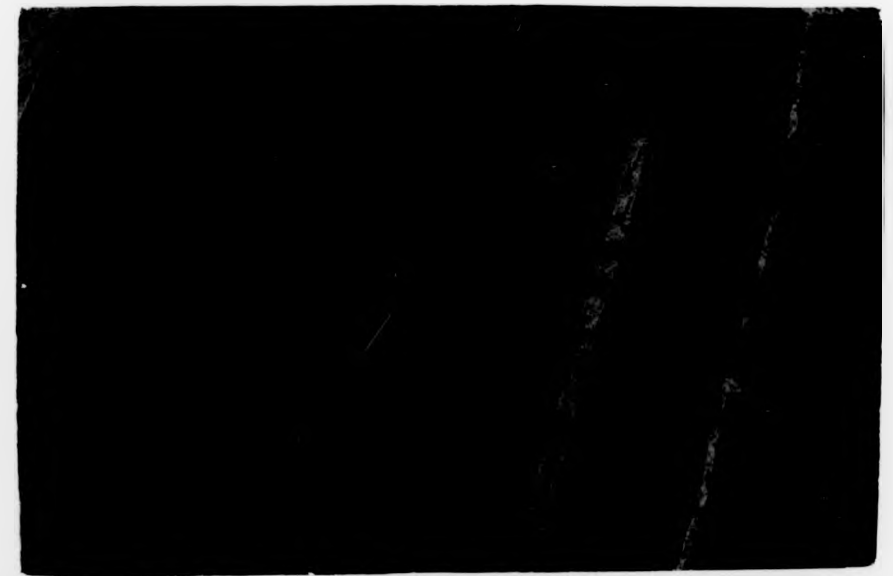NN 433883.   The bigger key is 7.5 cm long.

Plate 4.15 : Discontinuous, lenticular bedded psammites and
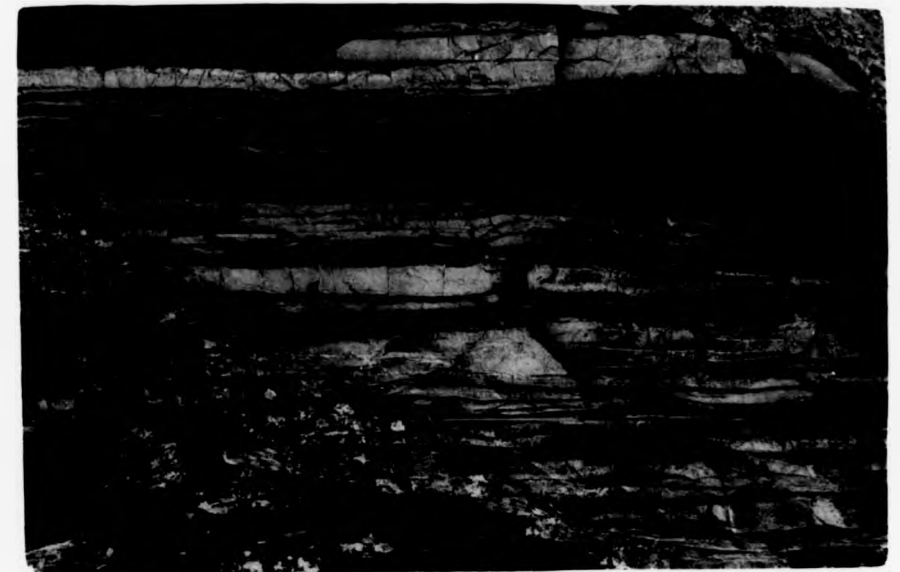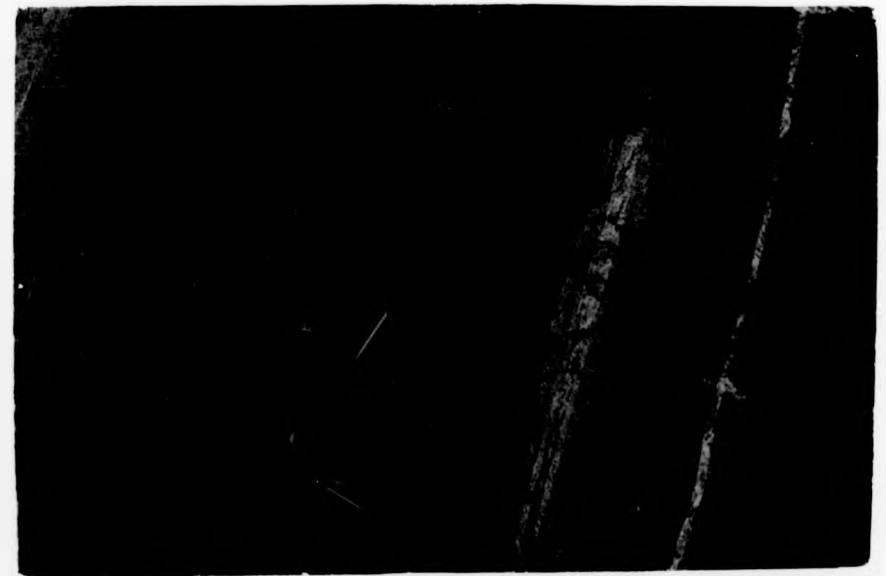semi-psammites. Carn Leac Formation NN 413822.
Hammer is 60 cm long.

Plate 4.16 : Rippled semi-psammites in the Carn Leac Formation.
NN 433883.   The bigger key is 7.5 cm long.

b. Thin, parallel-bedded psammites and semi-psammites

This is made up of psammites and semi-psammites occasionally interbedded with rippled units (Plate 4.16). They are commonly thinly-bedded, with thicknesses ranging from 5-30 cm.

This facies was probably deposited by strong traction currents in the lower part of the upper flow regime (Simons et al, op. cit.)

c. Semi-pelite

This facies which occurs as units 5-10 cm thick, occasionally intercalated with the semi-psammites represents siltstone or mudstone units deposited under slack water conditions, from suspension.

d. Discussion and environmental interpretation

Discontinuous, lenticular bedding is formed in intertidal flats as a result of the migration of tidal channels (Reineck and Singh, 1980). Allen (1965) has also observed that it occurs in river floodplain environments. Thus it is typical of shallow-water environments. The exact environmental context of this facies cannot, however, be determined because of the paucity of available evidence, although foregoing observations indicate a form of shallow water depositional environment.

The vertical lithostratigraphic sequence from the Monadhliath Formation to the Carn Leac Formation may therefore be interpreted as a regressive sequence associated with a shallowing of the depositional basin.

4.3 Paleocurrents

The measurement of current direction parameters from cross-stratified units were hampered by difficulties arising from lack of suitable three-dimensional exposures.

In the Chathalain and Garva Formations a unimodal paleocurrent distribution with some dispersion was obtained (Fig.4.7) after correction for tectonic tilt using the stereographic methods described by Ragan (1973). The results indicate paleoflow towards the NNW.

In the Corrieyairack Succession, paleocurrent data are only available
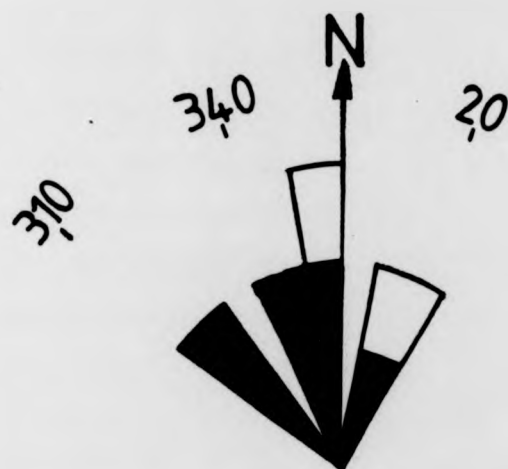
Fig. 4.7: Paleocurrents from crossbedding and ripples
Black roses: Glenshirra Succession
White roses: Glendoe Formation

for the Glendoe Formation. They show a paleoflow to the NNE (Fig 4.7).
The convolute folds observed in this formation are also overturned to this
direction.

Overall, therefore, in both successions, available data show that
the paleocurrent direction is to the northerly quadrants. In view of the
small number of measurements and possible errors due to the paucity of
suitable three-dimensional exposures, these results merely indicate the
general trend of paleocurrents.

For the Glendoe Formation and perhaps also the other formations in the
Corrieyairack Succession as well as the Garva and Chathalain Formations,
these observations suggest a northerly paleoslope and sediment transport from
the southern quadrants.

4.4 Conclusions

In view of the absence of environmenally specific criteria, only
tentative conclusions are possible.

The common pattern of paleocurrent direction towards the northern
quadrants indicates a common direction of sediment transfer for the rocks of
all the formations in this study area, and is further evidence in support of a
common provenance for these rocks (see Chapter 7).

In the Glenshirra Succession, the lithostratigraphic sequence comprising
the Chathalain and Garva Formations probably represents a shallow-water
sequence, possibly associated with alluvial fan progradation.

As observed from both petrography and geochemistry (Chapters 3 and 7,
respectively), the Coire nan Laogh Formation is the most mineralogically
and chemically mature of all the formations studied.

This may imply that it is a distal "facies" of the lithostratigraphic
units mapped in this study area, its sediments have undergone more extended
transport and greater weathering. On the other hand they could have been
derived in part from a more mature source rock.

Unfortunately because the rocks of the Coire nan Laogh Formation have

been subjected to very intense and repeated deformation and metamorphism very little sedimentological evidence is available to enable a more definite interpretation of the relationship between this formation and the rest. However, since it has a gradational boundary with the overlying Transitional and Glendoe Formations, its higher relative maturity may be attributed to a more extended transport and a higher degree of weathering. This is consistent with its predominantly semi-pelitic nature.

The lithostratigraphic sequence from the Coire nan Laogh Formation through the Transitional Semi-psammite to the Glendoe Formation constitutes a major thickening- and coarsening-upwards sequence which may be interpreted in terms of a submarine fan progradation (Walker, 1979).

Lithologically, the Monadhliath Formation is more semi-pelitic than the Glendoe Formation, although both have been inferred to be submarine fans. This more semi-pelitic character of the Monadhliath Formation taken together with its higher maturity (see Chapter 7) and scarcity of tractional sedimentary structures suggest that its sediments were laid down relatively farther away from their source rocks, possibly in deeper waters.

The transition from the Monadhliath Formation to the overlying Carn Leac Formation represents a shallowing of the depositional basin associated with increasing current activities.

4.5   Regional Correlations

North-flowing paleocurrents similar to those observed in the present study area have also been found in the Grampian rocks of the Corrieyairack Pass Area (Haselock, 1982) and in the Eilde Flags of the Lochaber district (Hickman, 1975). A common sediment transfer direction is therefore suggested for the rocks of the three areas.

Haselock (1982) interpreted the Corrieyairack Succession as deposits of a shallow-marine environment with the deposition of the Coire nan Laogh semi-pelite in the deeper distal part of the shallow shelf sea. She suggested an alluvial fan or braided stream environment for the Gairbeinn Pebbly semi-psammite and assigned the rest of the Glenshirra Succession to the shallow

marine environment. The Gairbeinn Pebbly Semi-psammite is absent in the
present study area and the possible reasons for its absence have been discussed
in Chapter 5. If it is due to a primary sedimentary wedging-out as in alluvial
fans, its reported thickening and increase in clast size towards the NW (Haseloch
and Winchester, pers comm, 1985) may suggest a provenance therefrom. The
underlying finer-grained Allt Luaidhe Formation and its equivalent in the
present study area the Garva Formation may therefore be interpreted as more
distal facies of the alluvial fan sequence.

As pointed out earlier (Chapter 2) there are some similarities in the
sedimentary structures observed in the Glendoe, Fechlin and Knockchollum
Formations (Whittles 1981; Haselock, 1982) and in the Glendoe Formation of the
present study area, thus suggesting possible sedimentological correlations.

Hickman (op. cit.) contended that the Eilde Flags of the Lochaber
district were deposited partly on a coastal plain and partly under estuarine
and tidal flat conditions. Since these "flags" probably form the top of
the Grampian Division and may be regarded as equivalent to the upper part of the
Carn Leac Formation, this interpretation is in agreement with the shallowing
of the depositional basin inferred to be associated with the deposition of the
Carn Leac Formation sediments.

4.6   SEDIMENTOLOGICAL ORIGIN OF THE CALC-SILICATE BODIES

Calc-silicate pods and bands are common in the metasedimentary rocks
of the Loch Laggan-Upper Strathspey area. They are generally restricted to
the psammitic lithologies and the thin semi-psammitic bands within the
semi-pelitic formations. Calc-silicate bodies are relatively rare and less
well-developed in the Garva and Chatalain Formations. Possible reasons for the
relative scarcity of well-developed calc-silicate in the rocks of the
Glenshirra Succession are discussed later.

a.   Morphology

The shapes of calc-silicate bodies vary from elliptical pods to more
elongate bands of thicknesses ranging from a few mm to 20 cm.

In the dominantly semi-psammitic Glendoe and Chatalain Formations, they occur mainly as lensoid pods 10-30 cm long and 5-10cm thick, with their long axes generally parallel to the bedding (Plates 4.7 and 4.8). In the thin semi-psammitic bands of the dominantly semi-pelitic Coire nan Laogh and Monadhliath Formations, they more commonly occur as thin elongated bands, 20-80 cm long and 3-5 cm thick (Plate 4.11).

Differences in the permeability anistropy were therefore apparently important in determining the shapes of the calc-silicate bodies.

b. Mode of Formation

The restriction of the calc-silicate bodies to the psammitic lithologies suggests an important lithological control. Their absence in the pelitic lithologies indicates that the permeability of the original sedimentary rock was a more important factor than the rock porosity in the localisation of the calc-silicates.

Deegan (1971), Raiswell (1971, 1976), Sass and Kolodny (1972) have shown that carbonate concretions are products of authigenic carbonate concentration during diagenesis. Raiswell (1971) observed that the calcareous concretions in the upper Cambrian sediments of Wales were developed in the more permeable siltstone units rather than in the surrounding mudstones. He suggested that the formation of the concretions relieved the supersaturation of the mobile carbonate-bearing fluids. Deegan (1971) suggested that the psammitic lithologies provided nuclei or areas conducive to carbonate accumulation, thereby setting up a chemical potential down which carbonate ions migrated from other parts of the host rock to the points of precipitation through the medium of pore waters.

Favourable sites for concretion nucleation are provided by micro environments of localised carbonate supersaturation where carbonate precipitation occurred with a rise in the pH beyond 7.8. This rise in alkalinity has been attributed to the decomposition of organic matter with concomitant

release of CO_2 and ammonia (Weeks, 1957; Sass and Kolodny, 1972; Berner, 1971) or to other microbiological processes (Raiswell 1976). It is possible that such processes associated with organic matter contributed to the localised precipitation of carbonate in the rocks of the present study area.

Ramberg (1952), however, suggested that the detrital carbonate particles already within the sediments collected into spheroidal segregations, these being the lowest energy configuration.

Although it is possible that the calc-silicates contained some original detrital calcite component, it is doubtful whether Ramberg's hypothesis was of any significant importance in the formation of the calc-silicate bodies observed in this study area.

c. Period of Formation

Pantin (1958) suggested that there were three periods during which carbonate concretions were formed in the host rocks:

a. Syngenetic: at the time of the deposition of the enclosing sediments.

b. Diagenetic: when the host sediments were still unconsolidated

c. Epigenetic: after the consolidation of the enclosing sediments.

Those calc-silicate bodies observed to overgrow the cross-bedding in the host rocks could not have been syngenetic. Pantin (op. cit.) argued that syngenetic concretions would show morphological asymmetry since their tops and bottoms would have grown under different conditions. The fact that most calc-silicate bodies, especially the ellipsoidal ones, are generally symmetrical indicates their non-syngenetic origin (Plates 4.7 and 4.8).

In some localities, outer laminations surrounding the calc-silicate pods are observed to be deformed around them (Plate 4.12) while the inner laminations cannot be traced across them. Raiswell (1971) showed that while the inner laminations would pass through the concretions relatively undeformed, the outer ones were deformed around the relatively rigid concretion as it became cemented and compacted. This situation is exemplified by Plate 4.12 and is evidence in support of a diagenetic, pre-compaction origin for the calc-silicate bodies of this study area.

Criteria for the recognition of epigenetic calc-silicates are not

definitive, and it is possible that some of the more elongate ones which occur as bands might be epigenetic. However, without specific contrary evidence a diagenetic origin is still favoured. Hence all available evidence indicates that the calc-silicate bodies were formed by authigenic carbonate concentration at suitable sites during the diagenesis of the sediments.

CHAPTER 5 : STRUCTURAL GEOLOGY

5.1   Previous work

The earliest structural work in this area was undertaken by Anderson (1947, 1956). He recognised north-easterly trending anticlines and synclines which are generally tight and locally overturned. These are the Loch Laggan Anticline and the complementary Kinlochlaggan and Corrieyairack Synclines.

Smith (1968) working in the area to the south east of the present study area, has recognised a set of early major isoclinal, recumbent folds with north-east trending axial planes which were refolded by late asymmetric folds with east trending axial planes.

Thomas (1979, 1980) put forward a model involving major nappe structures which diverge from the NE - SW trending Ossian-Geal Charn steep belt.

In the Corrieyairack Pass and Killin areas, Haselock et al. (1982), Whittles (1981) and Haselock (1982) have recognised three phases of deformation: 1. The D1 Gairbeinn Slide separates two lithostratigraphic successions which share a common structural and metamorphic history. 2. The NE - SW trending D2 Corrieyairack NE - SW trending Corrieyairack Syncline (Anderson 1956) and mapped corresponding Creag Mhor and Tarff anticlines and 3. open, D3, cross folding of earlier structures on a N-S trend.

Southwest of Loch Laggan between Glen Roy and Lismore, Hickman (1978) recognised five phases of deformation of which the first three were the most important folding events.

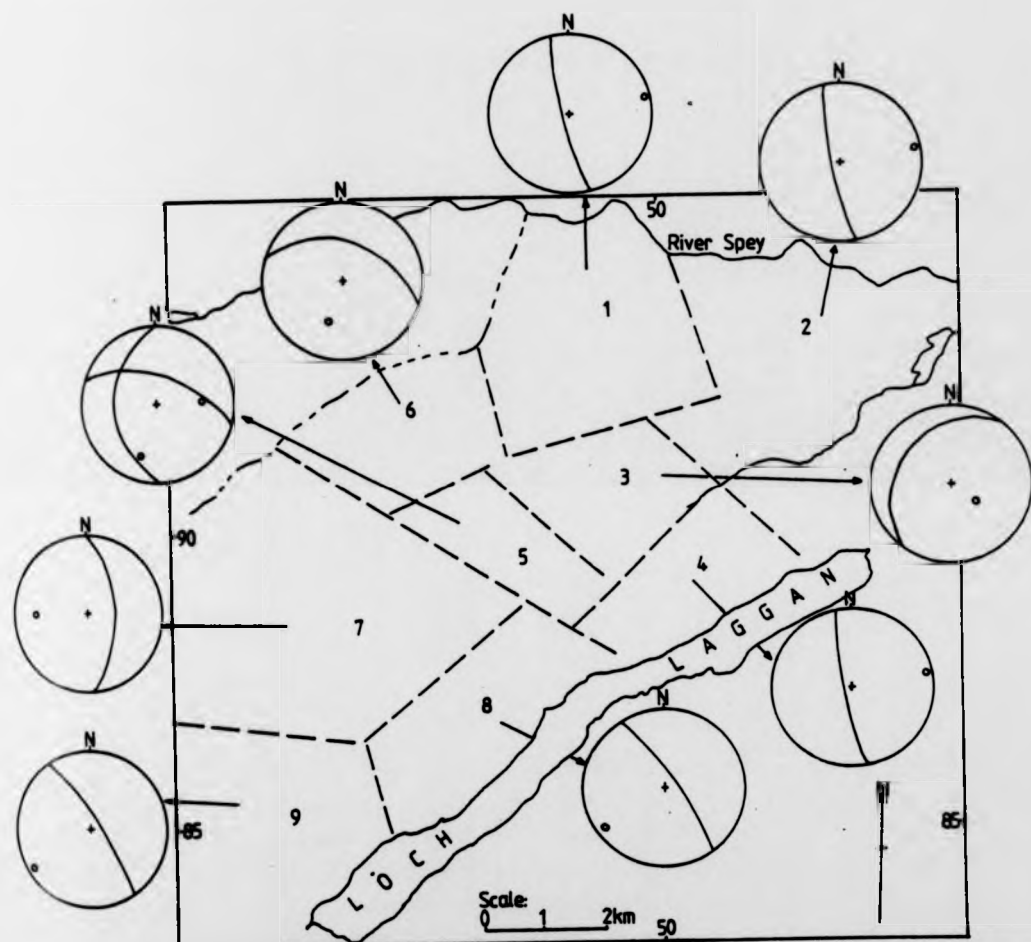An attempt will be made later to find what correlations exist
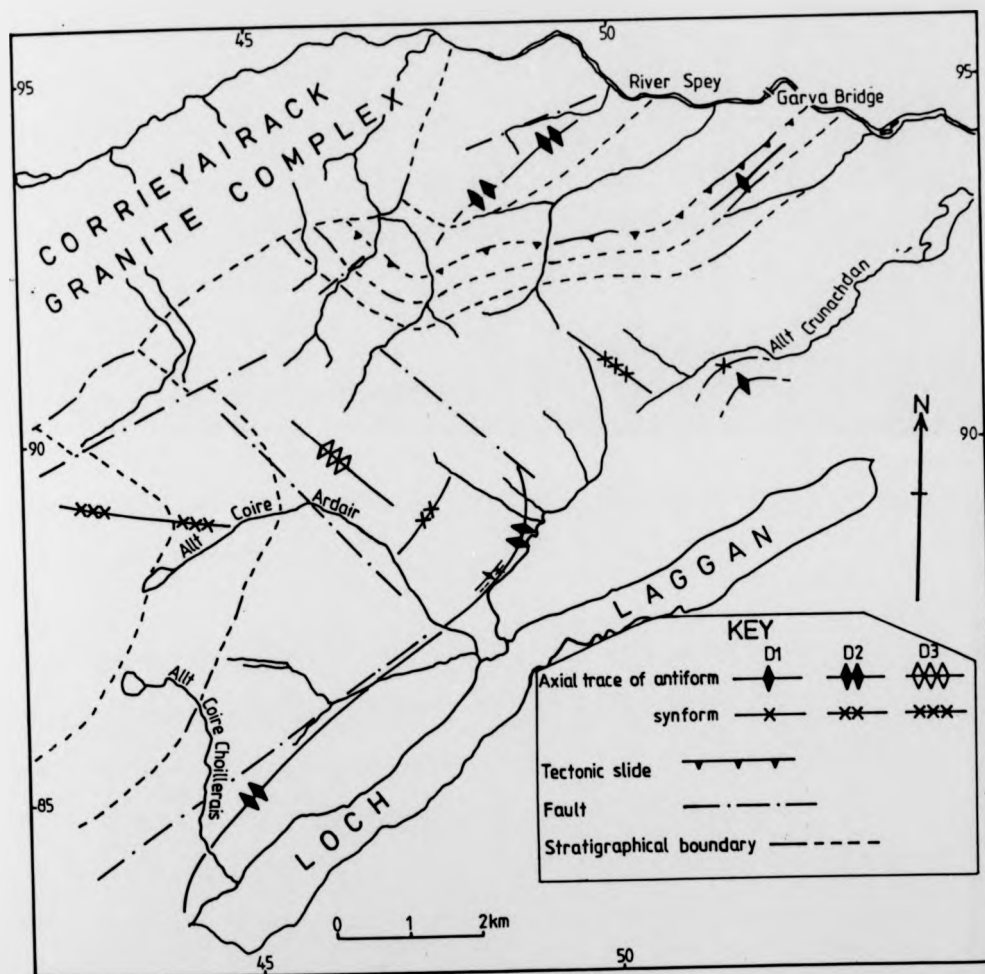
Fig. 5.1: Synopsis of structural subareas

Fig.5.2: Major structural features of the Loch Laggan-Upper Strathspey area

between the structures established by these workers in their various areas and those recognised in the present study area.

5.2   Introduction

Work in the Loch Laggan-Upper Strathspey area has shown that the metasediments have suffered polyphase deformation. Analyses of the geometry of minor and major structures, including their microfabrics, has led to the recognition of four phases of deformation. The structural sequence has been established in places by the recognition of refolding and other overprinting criteria.

The nomenclature of deformational phases, in sequence, is given by Dn with associated folds Fn, a planar fabric Sn and a linear fabric Ln. The various phases and their associated fabrics are given subscripts 1,2,3 etc. So refers to the bedding plane in the metasediments.

The entire map area was subdivided into smaller structural sub-areas (Fig 5.1), each containing statistically homogeneous structures. Measured structural data; lineations and planar surfaces, from each sub-area were then plotted on Lambert equal area nets, and the plots of So/S1 surfaces were contoured by using the Kalsbeek counting net so as to determine the orientations of the folds and other structural elements in each subarea.

A summary of the major structural features is given in Fig. 5.2.

5.3   First Deformation, D1

5.3.1 Minor Structures

The first phase of deformation, in this area, produced the widespread penetrative foliation S1 which is generally parallel or subparallel to the bedding planes (So) as defined by lithological variations.

Plate 5.1 :   D1 minor fold with shearing on the upper limb

marked by elongate quartzofeldspathic veins.
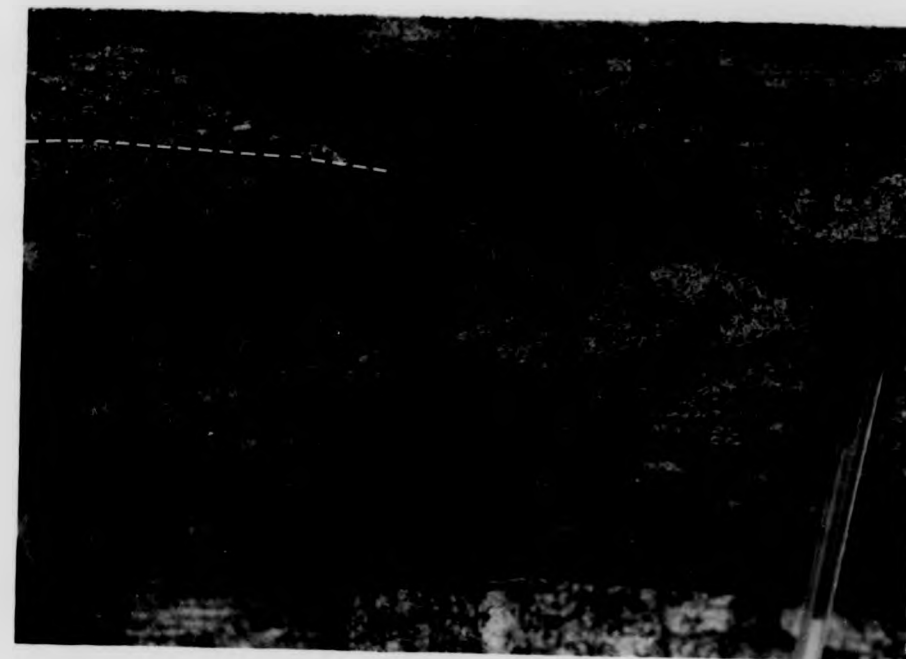
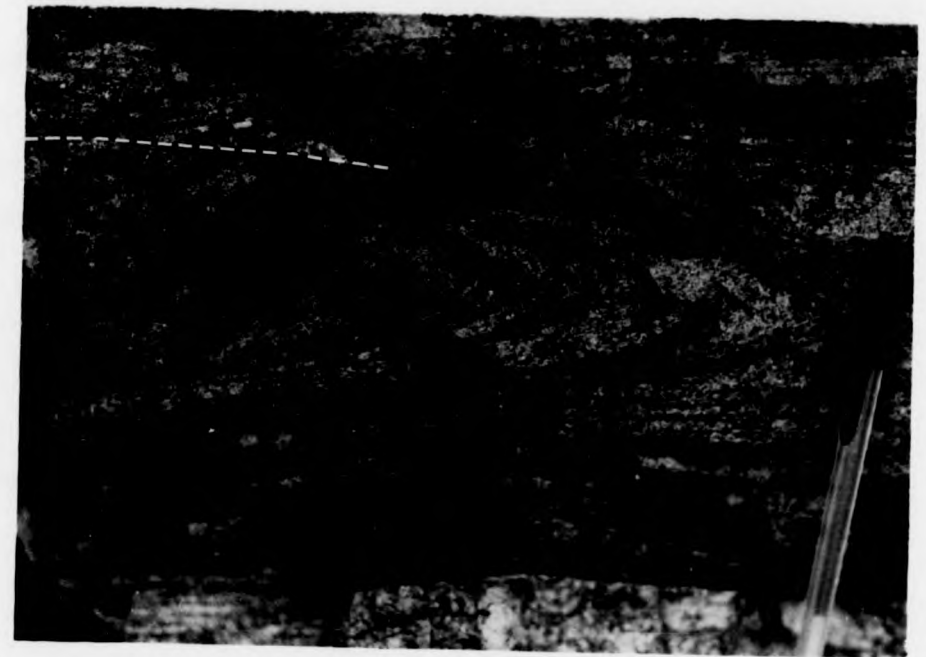(Chathalain Formation, NN 493941).

Plate 5.2 :   D1 minor fold with boudinaged quartz band parallel

to its axial plane.

(Garva Formation, NN 466928).

Hammer is 75 cm long.

Plate 5.1 : D1 minor fold with shearing on the upper limb

marked by elongate quartzofeldspathic veins.

(Chathalain Formation, NN 493941).

Plate 5.2 : D1 minor fold with boudinaged quartz band parallel

to its axial plane.

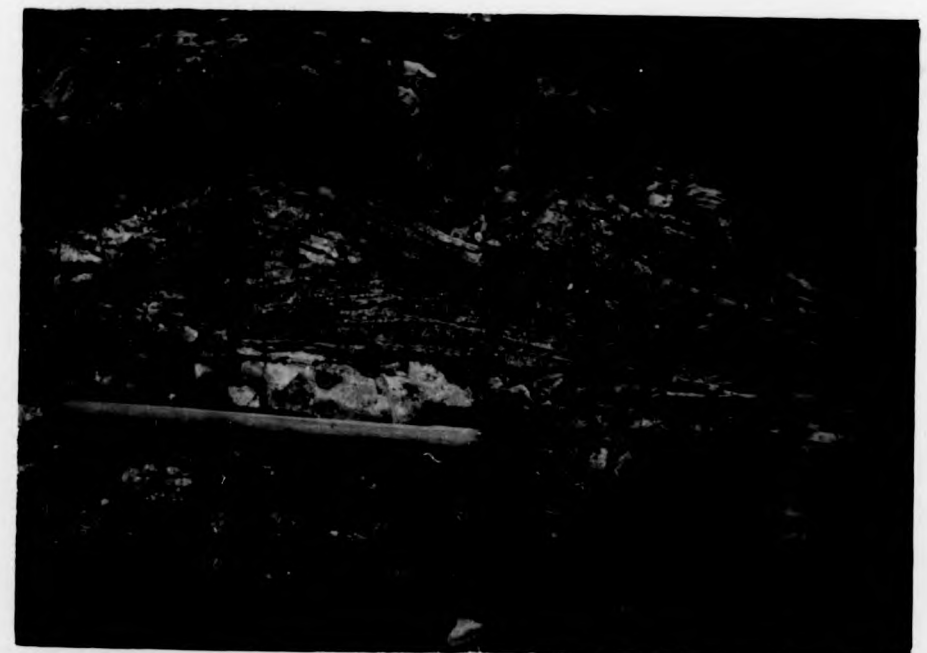(Garva Formation, NN 466928).

Hammer is 75 cm long.

Plate 5.1 : D1 minor fold with shearing on the upper limb

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(Chathalain Formation, NN 493941).

Plate 5.2 : D1 minor fold with boudinaged quartz band parallel

to its axial plane.

(Garva Formation, NN 466928).

Hammer is 75 cm long.

S1 is defined by a weak preferred orientation of mica, mostly biotite, and in strongly deformed rocks also by the dimensional preferred orientation of quartz grains. In more micaceous rocks, it is developed as a domainal schistosity (Hobbs et al, 1976) with anastomosing mica-rich and quartz-feldspar rich domains.

D1 minor folds (F_1) have been mapped in the Chathalain, Garva and Coire nan Laogh Formations, and in the overlying Transitional Semi-psammite. They are relatively rare in the remaining lithostratigraphic units. They deform the bedding, So, and pre-D1 quartz veins and process a penetrative axial planar schistosity, S1. F1 minor folds are tight to isoclinal with thickened hinges and attenuated limbs; they are typically recumbent folds with subhorizontal axial planes (Plates 5.1 and 5.2). Intense shearing on some of the fold limbs have, in some cases, produced dislocations now occupied by quartz veins (Plate 5.1).

The folds commonly possess a very high amplitude to wavelength ratio so that in most instances it is not possible to measure fold amplitudes on the outcrop. Where they have not been affected by later deformation, they show cylindrical hinges, and an intersection lineation, L1, of variable plunge, parallel to the fold axis.

5.3.2 F1 minor fold layer analyses

The geometry of F1 minor fold layers has been analysed using the methods of Ramsay (1967) and Huddleston (1973). Dip isogons were constructed at 10° intervals on tracings of representative folds from field photographs. Orthogonal thicknesses, $t\alpha$, of the layers were measured between the tangents to the layer surfaces at the various angles of dip, α. The angle, $\emptyset$, between an isogon and the normal to the parallel tangents of the folded layer surfaces were also measured. These aforementioned parameters are defined in Fig 5.3.
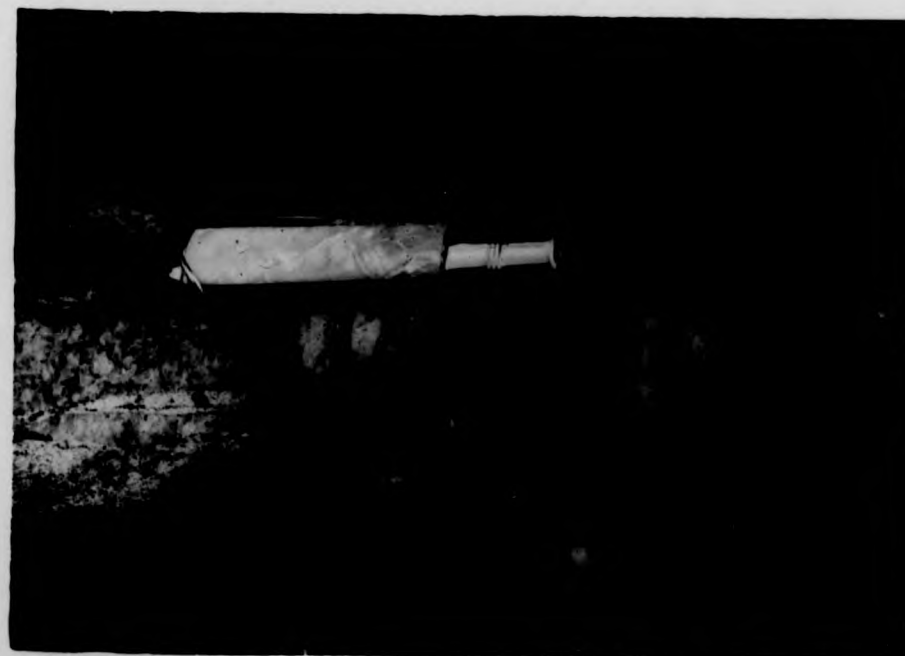
Plate 5.3 : Minor D1 fold of quartz vein in the Garva Formation
just below the Gairbeinn Slide. (NN 578939)
Lens cap is 5.5cm in diameter.

Plate 5.4 : D 1 folded quartzofeldspathic vein with sheared out
lower limb inside platy semi-psammites in the high
strain zone of the Gairbeinn Slide. (Garva Formation
NN 518938).
Pocket knife is 10cm long.

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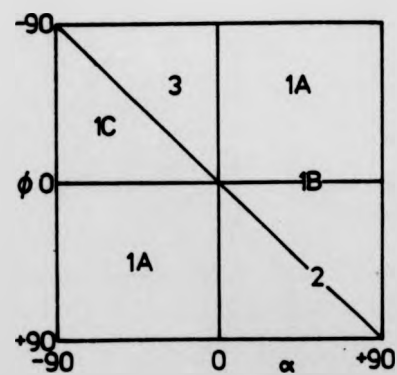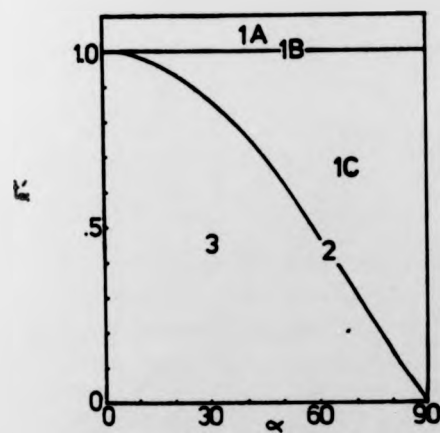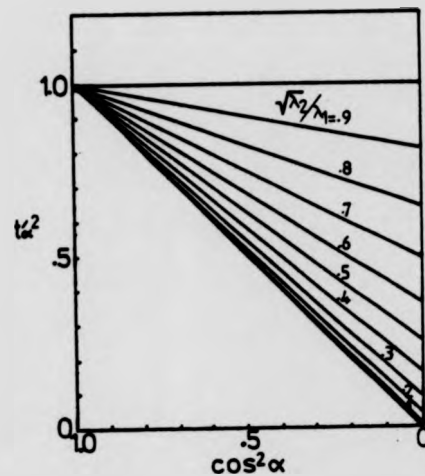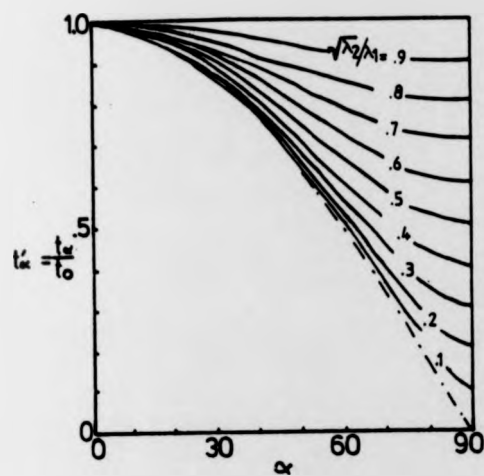   Pocket knife is 10cm long.

Fig.5.3: Positions of fold classes of Ramsay(1967) with curves for determining the values of flattening in terms of the quadratic elongations, λ_1 and λ_2.

Table 5.1 : D1 minor fold profile analyses

| Fold No	Location and Stratigraphic position	Lithology	Fold class from t'α vs α	Fold class from φ vs α	Flattening strain ratios : $\sqrt{\lambda_3/\lambda_1}$	
					from t'α vs cos²α	from φ vs α
1	NN 497940 Chathalain Formation	a)thinly b)bedded c)semi- psammite	2 2-3 2	2 2-3 2	- - -	- - -
2	NN 496939 Chathalain Formation	a)thinly-)bedded b)semi- psammite	1C-3 1C-2-3	2-3 1C-2-3	- -	- -
3	NN 493941 Chathalain Formation	a.thinly- bedded semi- psammite	2	2	-	-
4	NN497941 Chathalain Formation	a)thinly-)bedded b)semi- c)psammite	1C-2-3 1C 3	1C-2-3 1C-2 2-3	0.4 0.2 -	0.2 0.4 -
5	NN 497940 Chathalain Formation	a.thinly- bedded semi- psammite	1C-2-3	1C-2-3	0.1	-
6	NN 496939 Chathalain Formation	a.thinly- bedded semi- psammite	2	1C-2-3	-	0.1
7	NN 500925	a.calc- silicate band in semi- psammite	2	2-3	-	-
8	NN 501925 Transitional Formation	a)thinly-)bedded b)semi- c)psammite	1C 1C 1C-2	1C 1C 1C	0.5 0.45 0.4	0.45 0.4 0.4
9	NN 530936 Transitional Semi- psammite	a.thinly- bedded semi- psammite	2	2	-	-
10	NN 529936 Transitional Semi- psammite	a. Quartz vein in semi- psammite	2	2	-	-
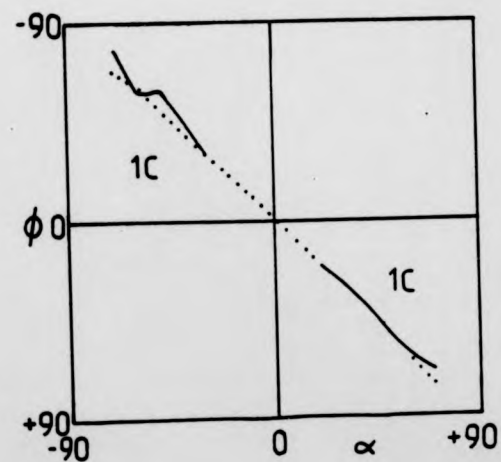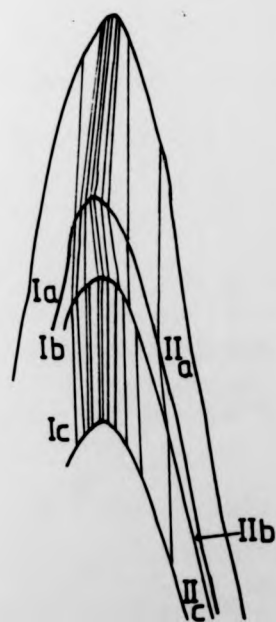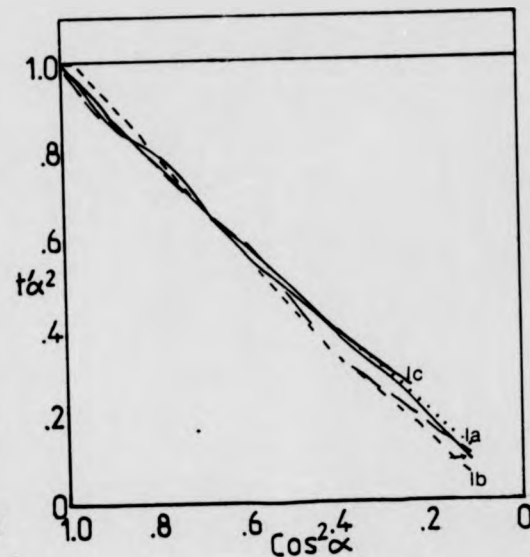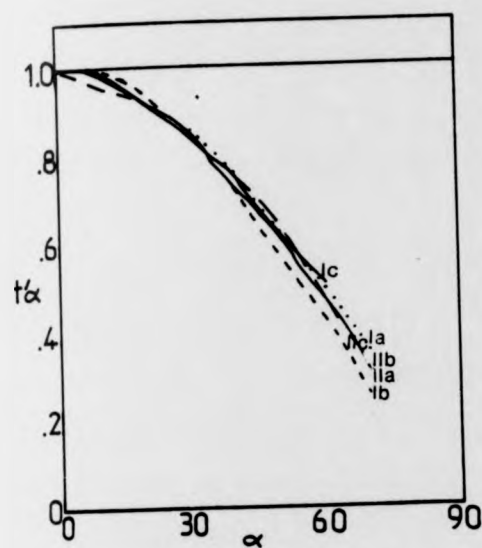
Fig. 5.4: Analysis of the geometry of a D1 minor fold

Another parameter, $t'\alpha$, defined as the orthogonal thickness of the layer, $t\alpha$, divided by the thickness of the layer at the fold hinge, t_o, ($t'\alpha = \dfrac{t\alpha}{t_o}$), is plotted against the various angles of dip, α (Fig 5.4). The curves obtained from this plot as well as from the plots of $\emptyset$ against α are used to classify each fold on the basis of Ramsay's fold classes: 1A, 1B 1C, 2 or 3. (Ramsay, op. cit.)

The results obtained from the analyses of 10 folds comprising 17 layers indicate that the F1 minor folds fall into three main classes: 1C, 2 and 3 (Table 5.1). In some fold layers various segments fall into two or more of these classes. There is, generally, a good agreement between the resulting classifications obtained for each fold from dip isogon, $t'\alpha : \alpha$, and $\emptyset : \alpha$ plots (Table 5.1).

Most of the folds analysed deform thinly-bedded, flaggy semi-psammites or quartz veins, together with a few thin calc-silicate bands. These lithologies produce predominantly class 2 geometries. Where thicker semi-psammite layers are folded with semi-pelitic bands Class 1C geometries are obtained.

5.3.3 D1 major structures

a. Tectonic sliding

A zone of high strain marks the contact between the Garva and the Coire nan Laogh Formations. By correlation with the Corrieyairack Pass area to the NW, this zone of high strain is considered to be associated with the Garbeinn Slide described by Haselock (1982).

In the topmost 40m of the Garva Formation and much of the Coire nan Laogh Semi-pelite, especially the basal 40m, high strain is indicated by a very marked intensification of the foliation with the production of platy and

Plate 5.5 : D1 folded quartzo-feldspathic segregations with
sheared-out limb, and low strain augen of
semi-psammite in the high strain zone. (Coire
nan Laogh Formation, NN 523938).
Lens cap is 5.5cm in diameter.

Plate 5.6 : D1 folded quartzo-feldspathic segregations with
sheared out limbs.
(Coire nan Laogh Formation, NN 522940)
Lens cap is 5.5cm in diameter.

Plate 5.5 :   D1 folded quartzo-feldspathic segregations with
              sheared-out limb, and low strain augen of
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Plate 5.6 :   D1 folded quartzo-feldspathic segregations with
              sheared out limbs.
              (Coire nan Laogh Formation, NN 522940)
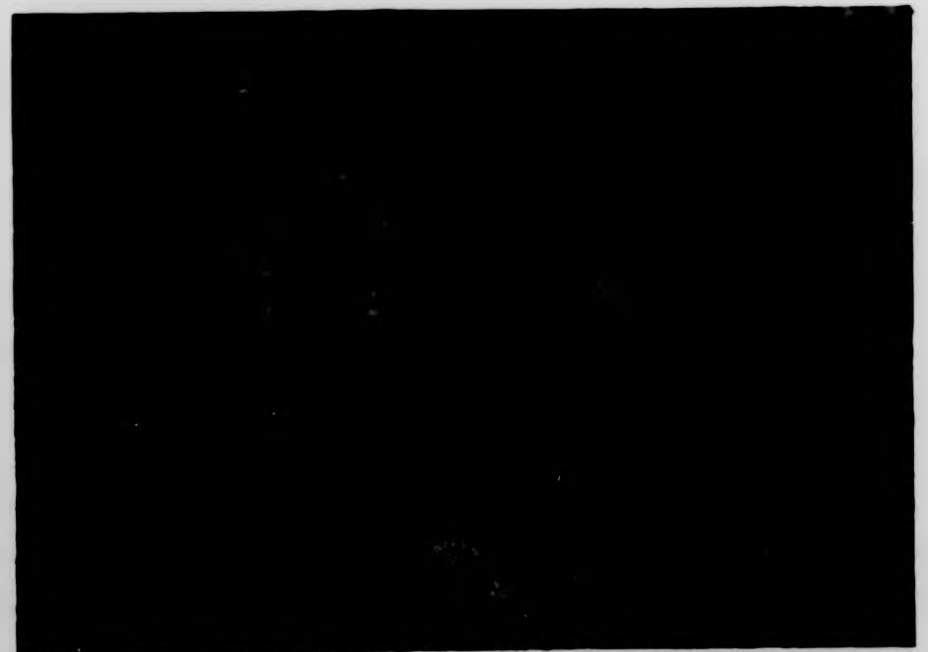   Lens cap is 5.5cm in diameter.
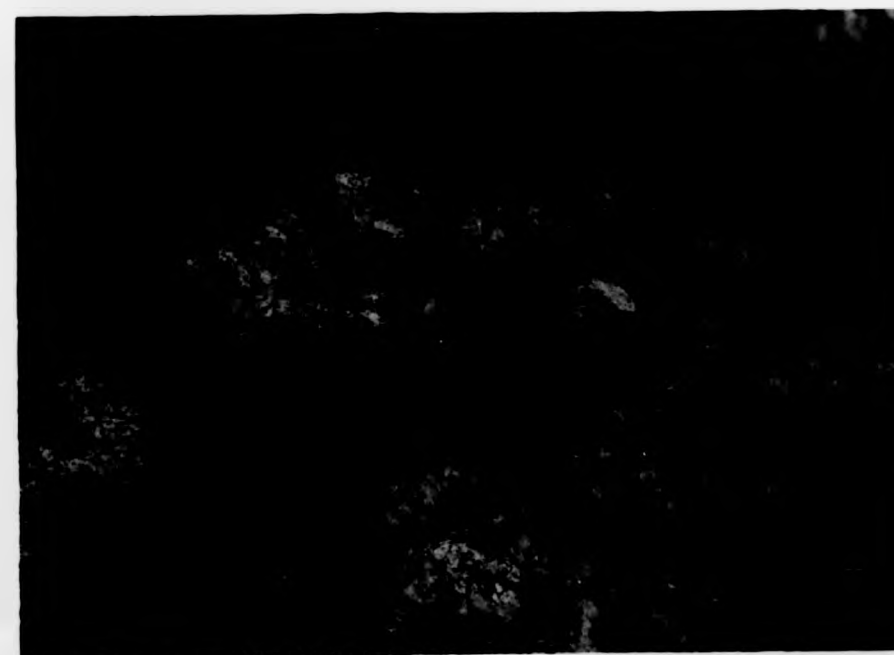
Plate 5.5 : Dl folded quartzo-feldspathic segregations with
            sheared-out limb, and low strain augen of
            semi-psammite in the high strain zone. (Coire
            nan Laogh Formation, NN 523938).
            Lens cap is 5.5cm in diameter.

Plate 5.6 : Dl folded quartzo-feldspathic segregations with
            sheared out limbs.
            (Coire nan Laogh Formation, NN 522940)
            Lens cap is 5.5cm in diameter.

flaggy fabrics. Swarms of concordant and subconcordant quartz and quartzo-feldspathic segregations similar to those described by Piasecki (1980), Piasecki and van Breemen (1983), Haselock (1982) and Haselock et al (1982) in the slide zones of the Central Highlands are also present.

i   Minor structures in the high strain zone

Two types of folds were recognised in the high strain zone. First, there are tightly appressed folds deforming bedding (Plate 5.2) which are recumbent intrafolial isoclines occurring the topmost part of the Garva Formation and in the Coire nan Laogh semi-pelite. Boudinaged and streaked out quartz and quartzo-feldspathic bands as well as the S1 foliation are axial planar to these folds (Plate 5.2).

The second group of folds deform boudinaged quartz and quartzo-feldspathic veins (Plates 5.3, 5.4, 5.5 and 5.6). These are commonly tight, asymmetrical folds with axial planes which are parallel to the intensified shear foliation, S1.

The vergence of these folds changes at NN 522938 suggesting a medium scale   fold closure at this location (Fig 5.2).

The axes of these minor folds and other D1 mineral lineations defined by the alignment biotite grains show a dispersion in trend from NNE to SSW in subarea 2 (Fig 5.8). This dispersion is inferred to be largely due to variable rotation towards the stretching direction during the sliding and indicates the curvilinearity of the D1 fold axes in this area.

The D1 minor structures probably record a continuous sequence of deformation consisting of progressive simple shear; they are consistent with models of progressive simple shear deformation outlined by various workers (Ghosh, 1966, Sanderson, 1979; Skjernaa, 1980; Ramsay 1967, 1980, 1983; Rattey and Sanderson 1979; Evans and White,1984). Under the progressive simple shear regime, the early intrafolial fold axes were progressively rotated into parallelism with the extension lineation (Sanderson 1973, 1979; Escher and

flaggy fabrics. Swarms of concordant and subconcordant quartz and quartzo-feldspathic segregations similar to those described by Piasecki (1980), Piasecki and van Breemen (1983), Haselock (1982) and Haselock et al (1982) in the slide zones of the Central Highlands are also present.

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The axes of these minor folds and other Dl mineral lineations defined by the alignment biotite grains show a dispersion in trend from NNE to SSW in subarea 2 (Fig 5.8). This dispersion is inferred to be largely due to variable rotation towards the stretching direction during the sliding and indicates the curvilinearity of the Dl fold axes in this area.

The Dl minor structures probably record a continuous sequence of deformation consisting of progressive simple shear; they are consistent with models of progressive simple shear deformation outlined by various workers (Ghosh, 1966, Sanderson, 1979; Skjernaa, 1980; Ramsay 1967, 1980, 1983; Rattey and Sanderson 1979; Evans and White,1984). Under the progressive simple shear regime, the early intrafolial fold axes were progressively rotated into parallelism with the extension lineation (Sanderson 1973, 1979; Escher and

Plate 5.7 : Garnet enveloped in S1 defined by biotite and
muscovite in a semi-pelitic band interbanded with
quartzo-feldspathic segregations.
(Base of the Coire nan Laogh Formation, NN 578938).
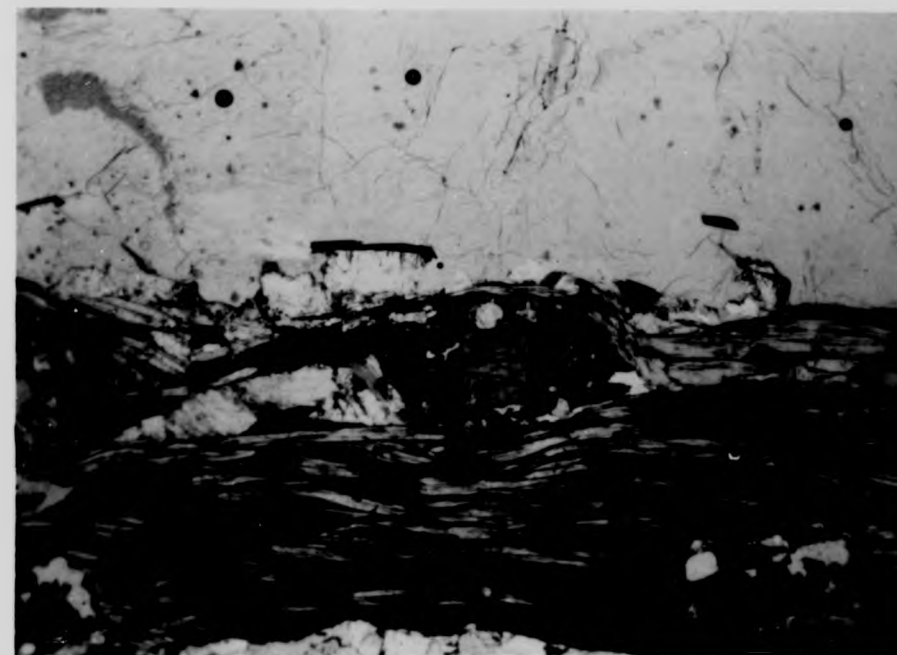Plane polarised light. Scale bar represents 0.2mm.

Plate 5.7 : Garnet enveloped in S1 defined by biotite and

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Plane polarised light. Scale bar represents 0.2mm.

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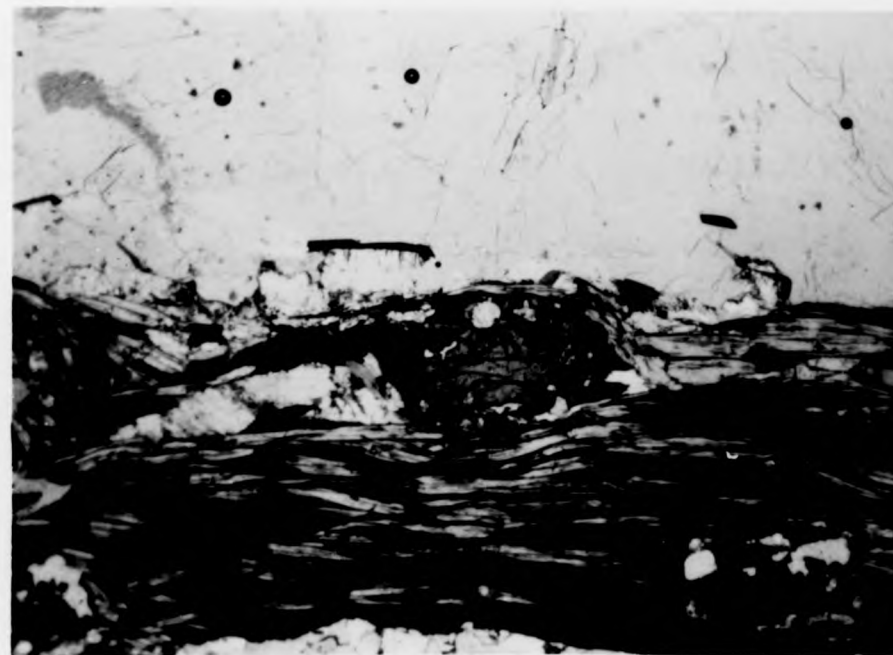Plane polarised light. Scale bar represents 0.2mm.

Watterson 1974).

Directly adjacent to the tectonic contact between the Garva and the Coire nan Laogh Formations (the Gairbeinn Slide) the folds become attenuated through extreme flattening and shearing out of the fold limbs (Plate 5.4).

ii   Microstructures

The basal section of the Coire nan Laogh semipelite consists of a highly tectonised migmatitic schist, containing coarse biotite laths (1-2mm long) intergrown with muscovite. Finely interbanded with these are coarse (0.5-2mm) quartz and feldspar segregations (including lensoid plates of pegmatite) to produce a platy fabric (Plate 5.7). Garnet within this blastomylonitic schist shows evidence of polyphase growth. Individual garnets may possess a core indicating early to syntectonic growth in relation to the D1 shear deformation surrounded by a rim that postdates the deformation (Plate 3.5).

The presence of fractured and pulled apart porphyroblasts of garnet at the base of the Coire nan Laogh Formation (Plate 5.8) gives a qualitative indication of the strain they suffered during the shearing. Note that whilst the lower garnet has been pulled apart parallel to the foliation, the upper one has been pulled apart at angle almost perpendicular to it. Such features have been reported by Gregg (1978), White et al (1980) from mylonitic rocks and by Law et al (1984) from the upper and central levels of the Upper Arnaboll thrust sheet in Sutherland and are indicative of non-coaxial deformation under progressive simple shear. Note, too, that this brittle deformation recorded by the garnets is overprinted by a later biotite fabric, and new quartz growth underscoring the general observation that the shear fabrics have been largely annealed by post-tectonic recrystallisation.

The blastomylonitic semi-pelites also contain garnets elongated parallel to the D1 lineation (Plate 5.10) which are inferred to have formed under intense shear deformation, probably comprising a pure shear component. The section was cut parallel to the D1 lineation and the large porphyroblast gives an X/Z ratio of about 2.7.

Plate 5.8 :   Fractured and pulled-apart garnets in semi-pelite
              at the base of the Coire nan Laogh Formation.
              (NN 482925).
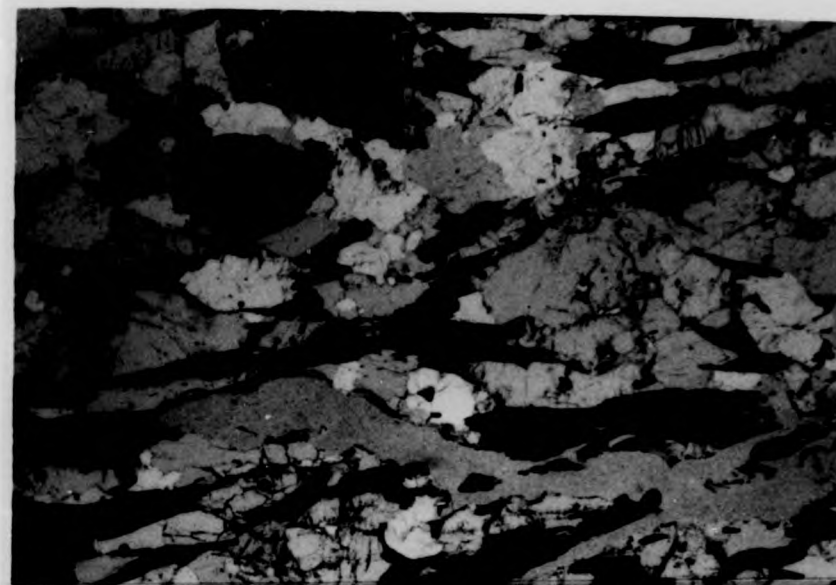   Plane polarised light. Scale bar represents 0.2mm.
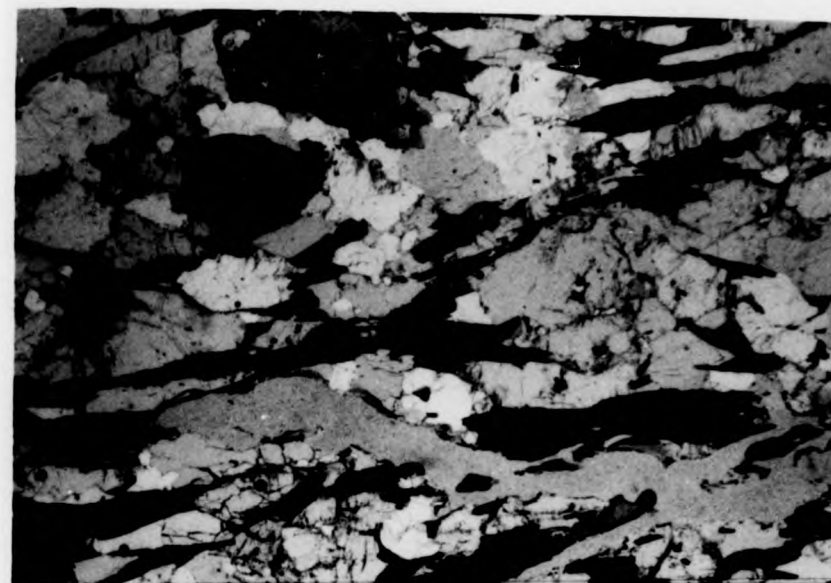
Plate 5.8 : Fractured and pulled-apart garnets in semi-pelite
at the base of the Coire nan Laogh Formation.
(NN 482925).
Plane polarised light. Scale bar represents 0.2mm.

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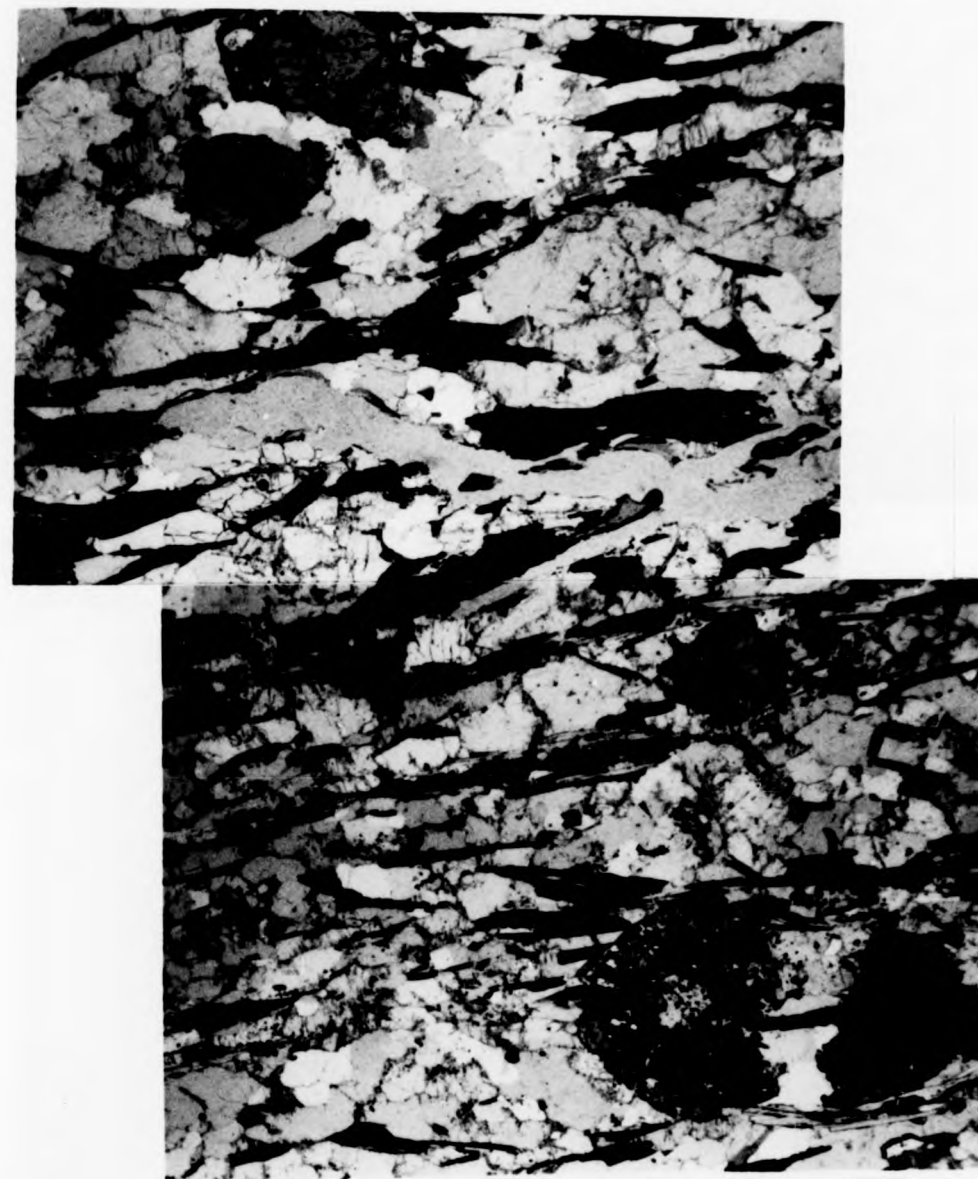   Plane polarised light. Scale bar represents 0.2mm.

Plate 5.9 : Psammite from the Garva Formation 550m
from the slide showing less strained and granulated
quartz and feldspar grains. NN 514944. (Crossed
nicols. NN 514944 . Scale bar represents 0.1mm)

Plate 5.10 : Elongate garnets in semi-pelite from the base of
the Coire nan Laogh Formation. NN 520939.
(Plane polarised light. Scale bar represents 0.2mm).

Plate 5.9 : Psammite from the Garva Formation 550m
from the slide showing less strained and granulated
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Plate 5.10 : Elongate garnets in semi-pelite from the base of
the Coire nan Laogh Formation. NN 520939.
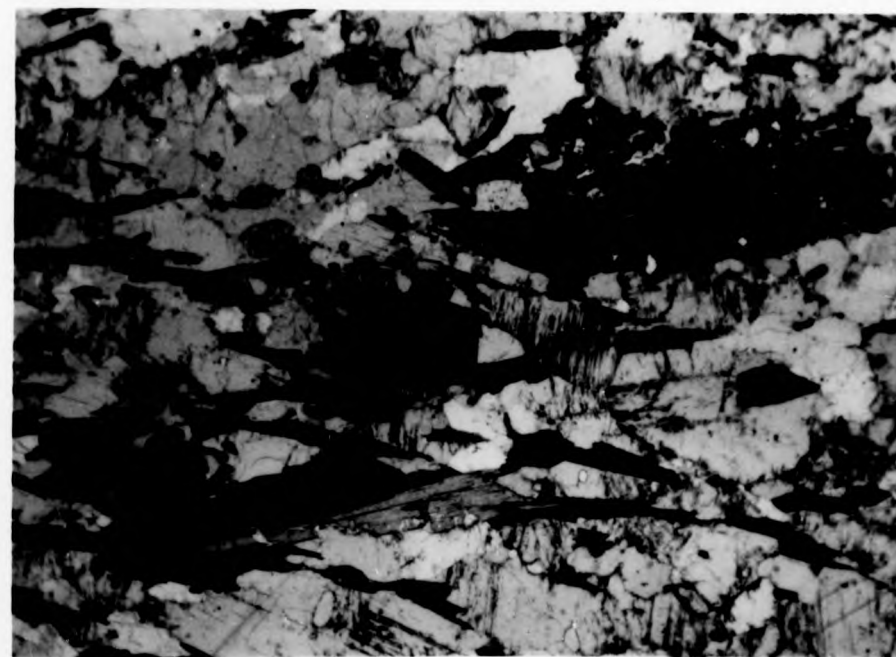(Plane polarised light. Scale bar represents 0.2mm).

Plate 5.9 : Psammite from the Garva Formation 550m
from the slide showing less strained and granulated
quartz and feldspar grains. NN 514944. (Crossed
nicols. NN 514944 . Scale bar represents 0.1mm)

Plate 5.10 : Elongate garnets in semi-pelite from the base of
the Coire nan Laogh Formation. NN 520939.
(Plane polarised light. Scale bar represents 0.2mm).

The semi-psammites and psammite bands in the upper part of the Garva Formation close to the tectonic contact contain highly strained quartz and feldspar grains (probably porphyroclasts) in a matrix of smaller grains which are commonly elongated parallel to the stretching lineation (Plate 5.1₁). Quartz grains also occur as recrystallised aggregates of long ribbons (Plate 5.12). They are occasionally associated with granular epidote, ilmenite, and sphene strung-out along the foliation (Plate 5.12).

Since the climax of the regional metamorphism postdated the shear deformation, resultant mineral fabrics have undergone extensive recrystallisation and annealing, so that grain shape fabrics are likely to underestimate the amount of strain suffered by the rocks. This late recrystallisation has generally produced small subgrains from original highly strained coarse grains.

A qualitative indication of the strain heterogeneity in the Garva Formation may be obtained from a comparison of the microstructures in samples collected within the high strain zone, close to the Gairbeinn Slide (Plate 5.11) with those in a sample collected in the lower section of the formation (Plate 5.9) which has a weaker grain shape fabric and little grain size reduction and a more decussate fabric.

iii   Transport direction

The tectonic transport azimuth on the Gairbeinn Slide is given by the stretching direction and may be inferred from the D1 hinges and mineral lineations which have been rotated towards the transport direction in the high strain zone adjoining the slide. These lineations have been rotated from the regional NE-SW trends towards the NW-SE quadrants (Fig 5.8bii).

The sense of movement on the slide may be inferred from the overall regional vergence of both D1 minor folds and recumbent D1 medium scale folds mapped on Meall Ghoirleag (NN518907) and above the slide, south of Stac Buidhe (NN520938) Fig 5.2 and Enclosures). Both indicate tectonic transport to the NW.

Plate 5.11 : Highly elongate quartz plate and "ribbons" which have
undergone syntectonic grain size reduction and probably
some post-tectonic recrystallisation in psammite from
the Garva Formation just below the Garbeinn Slide.
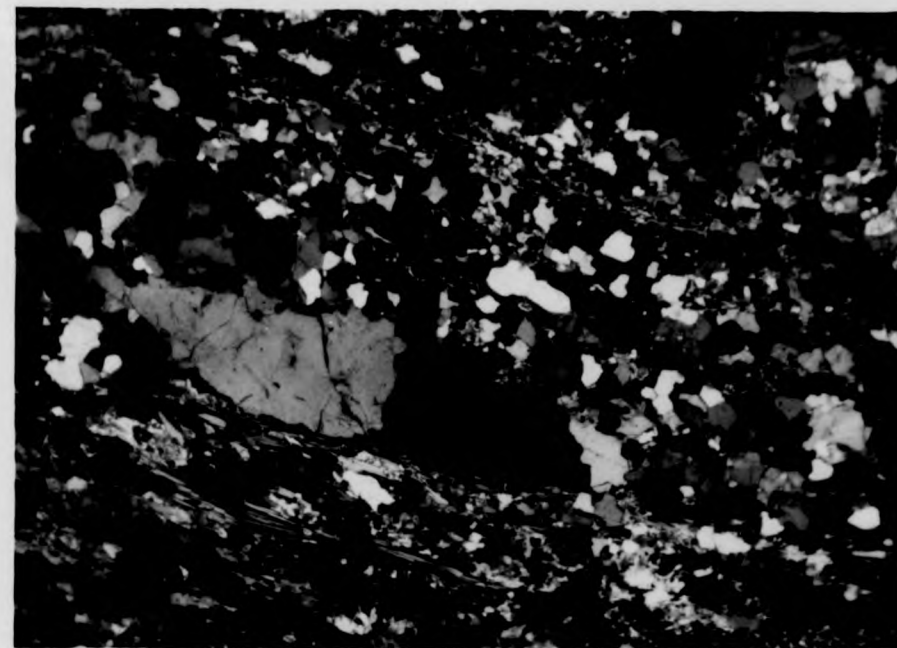NN 467926.

Crossed nicols.   Scale bar represents 0.2mm.

Plate 5.12 : Quartz ribbons and granular epidote, sphene and ilmenite
"strung-out" parallel to S1 in Garva semi-psammite 1m
below the slide. NN 482926.

Crossed nicols.   Scale bar represents 0.1mm.

Plate 5.11 : Highly elongate quartz plate and "ribbons" which have
undergone syntectonic grain size reduction and probably
some post-tectonic recrystallisation in psammite from
the Garva Formation just below the Garbeinn Slide.
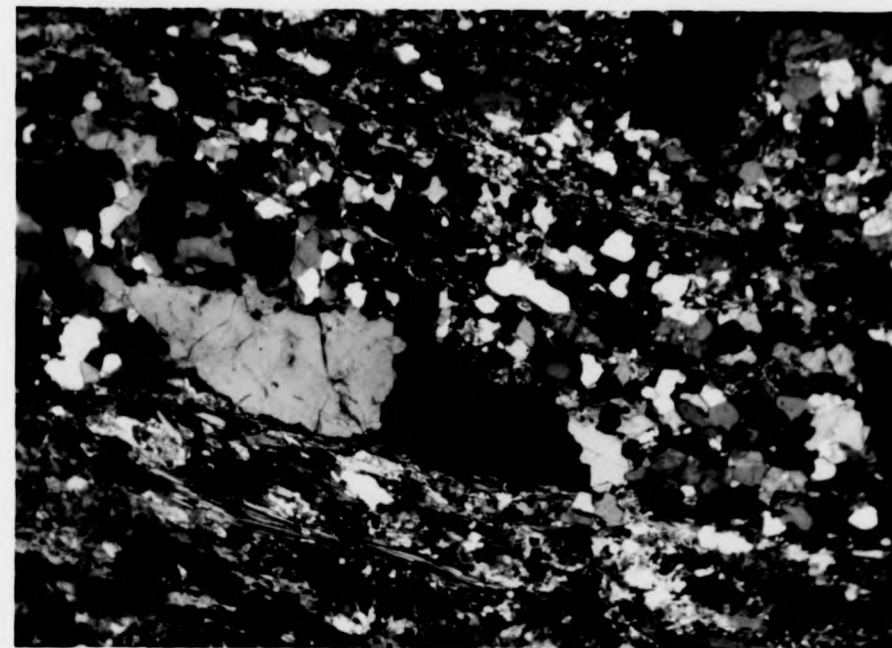NN 467926.

Crossed nicols.   Scale bar represents 0.2mm.

Plate 5.12 : Quartz ribbons and granular epidote, sphene and ilmenite
"strung-out" parallel to S1 in Garva semi-psammite 1m
below the slide. NN 482926.

Crossed nicols.   Scale bar represents 0.1mm.

Plate 5.11 : Highly elongate quartz plate and "ribbons" which have
undergone syntectonic grain size reduction and probably
some post-tectonic recrystallisation in psammite from
the Garva Formation just below the Garbeinn Slide.
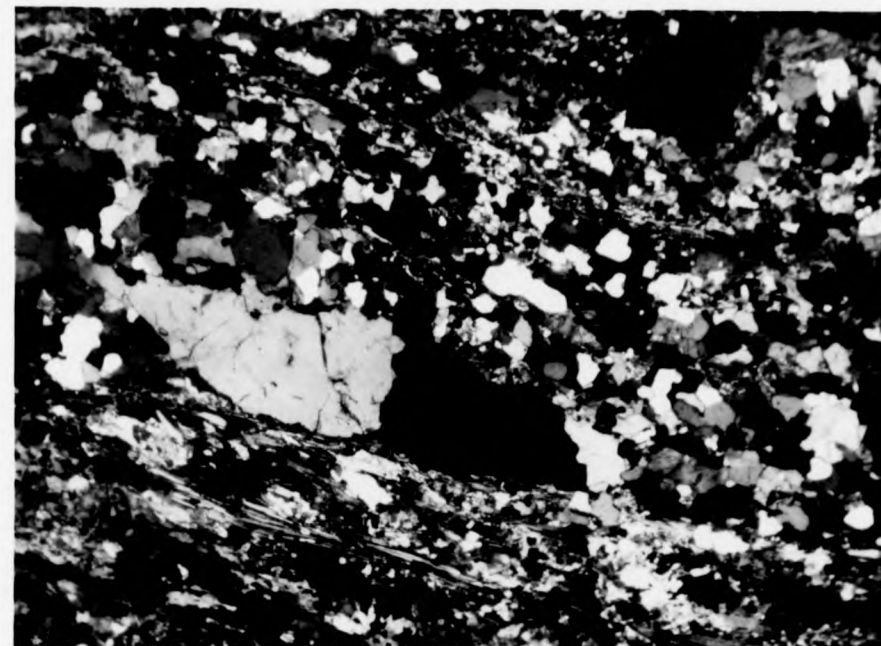NN 467926.

Crossed nicols.   Scale bar represents 0.2mm.

Plate 5.12 : Quartz ribbons and granular epidote, sphene and ilmenite
"strung-out" parallel to S1 in Garva semi-psammite 1m
below the slide. NN 482926.

Crossed nicols.   Scale bar represents 0.1mm.

The implications of the absence of the Gairbeinn Pebbly Semi-psammite

The absence of the Gairbeinn Pebbly Semi-psammite (Haselock et al 1982) in the present study area may be important with regard to the nature of the movement on the Gairbeinn Slide and the geometry of the decollement surface. The absence of this pebbly unit may be explained by three possible hypotheses:

a. as the slide propagated towards the NW it cut up the stratigraphic succession, ramping towards the NW (Fig 5.5)

b. the pebbly unit has been wedged out to the SE by extension and thinning associated with the D1 deformation.

c. the absence of the unit is due to a primary, sedimentary wedging out of an alluvial fan (Fig 5.6) (see also Chapter 4).

Apart from the reported thickening of the pebbly unit towards the west (Haselock et al 1982) and an increase in clast size in the same direction, little sedimentological information is available for this formation. Paleocurrent indicators from the Chathalain and Garva Formations all show a trend to the NNW (Chapter 4 and Fig 4.7). If this trend is also assumed for the pebbly formation, then its thickening towards the west cannot be attributed to western-derived provenance. However, it is possible that the reported increase in clast size towards the NW is as a result of sediment transport from that direction, and if so this formation may have had a different sediment transport direction from the Chathalain and Garva Formations.

While some flattening and thinning associated with D1 deformation may have occurred it is not likely that this would cause the total disappearance of the pebbly unit in the present study area. Hence this alternative cannot provide an adequate explanation.

Available evidence suggests that models involving thrust propagation over a ramp and primary sedimentary wedging-out are both likely. Haselock et al. (1982) have observed that the Gairbeinn Slide is overlain by progressively
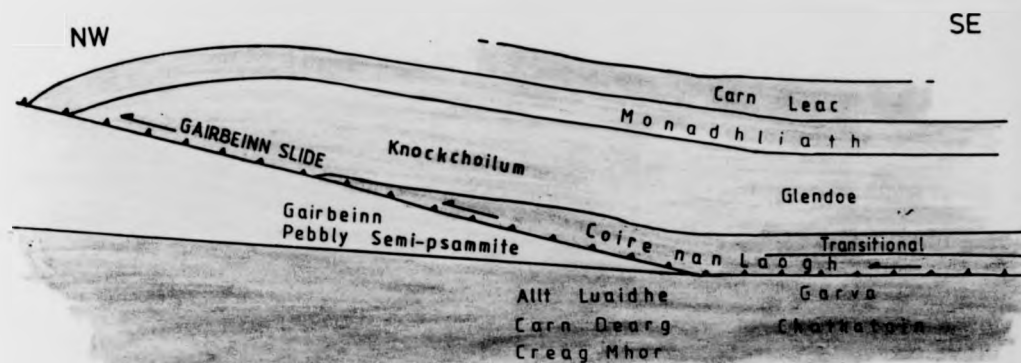
Fig.5.5a: Schematic model of the structure of the Loch Laggan-Upper Strathspey and Corrieyairack Pass areas prior to D2 deformation
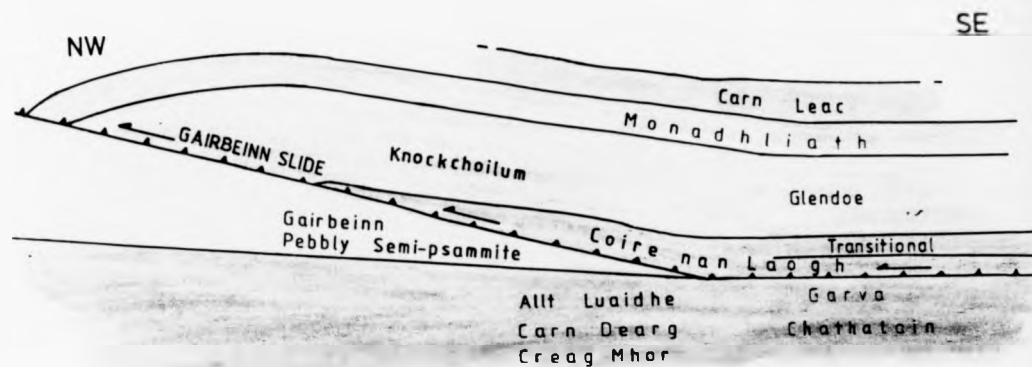
Fig.5.5a : Schematic model of the structure of the Loch Laggan-Upper Strathspey and Corrieyairack Pass areas prior to D2 deformation
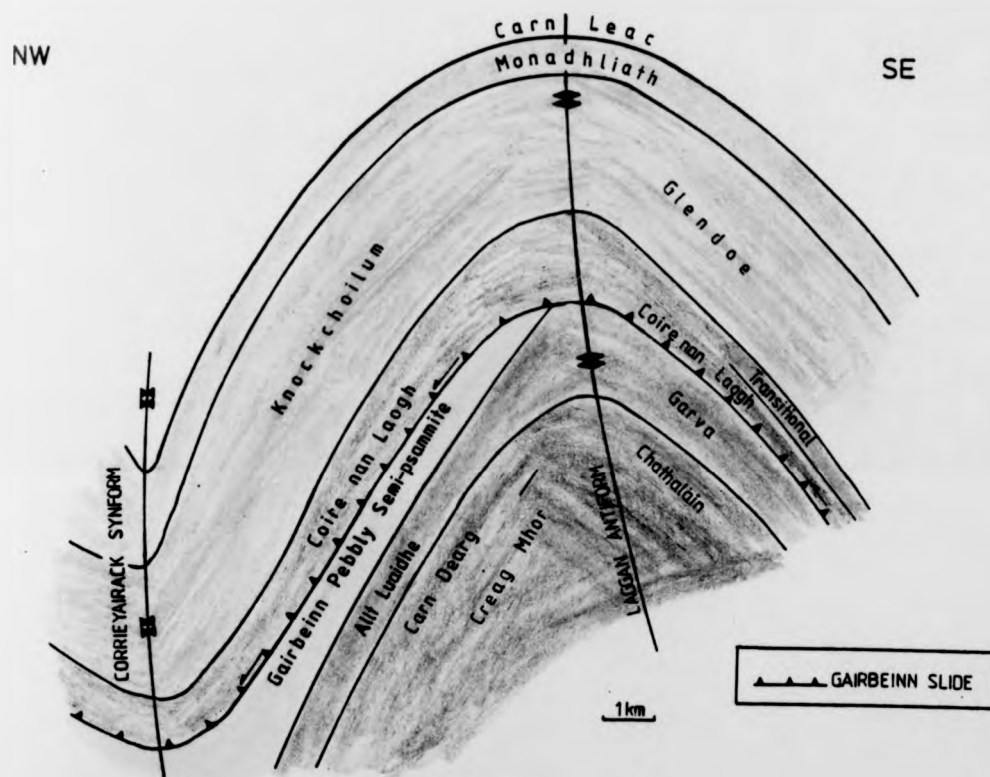
Fig.5.5b: Schematic section across the Corrieyairack Pass and Loch Laggan-
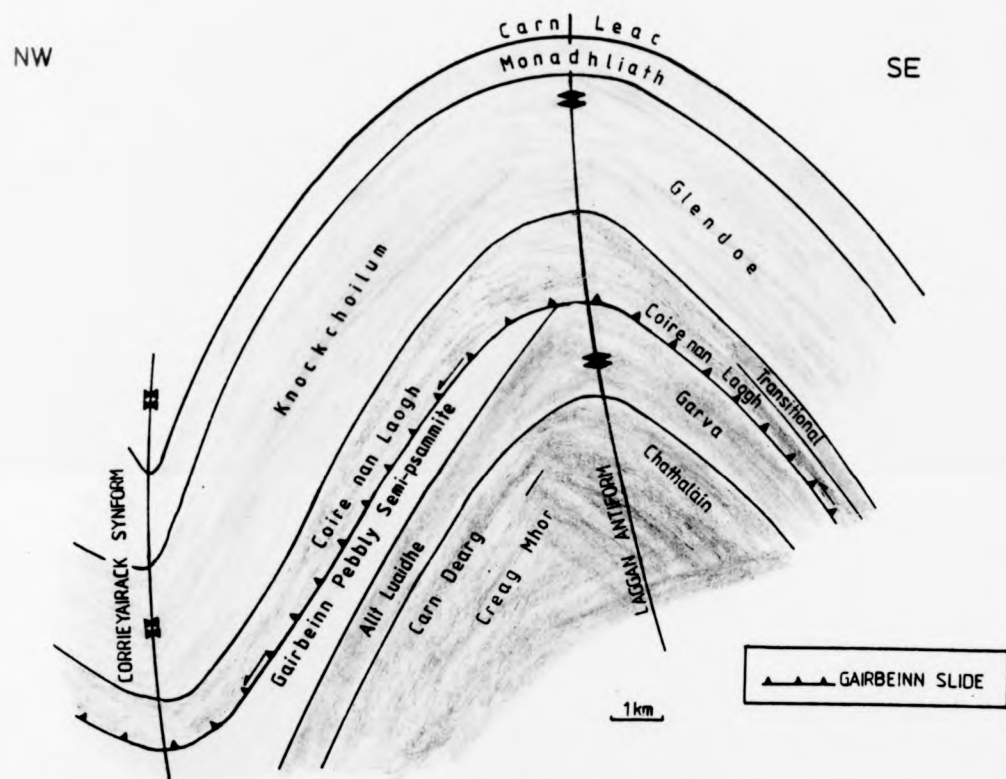Upper Strathspey·areas

Fig.5.5b: Schematic section across the Corrieyairack Pass and Loch Laggan-
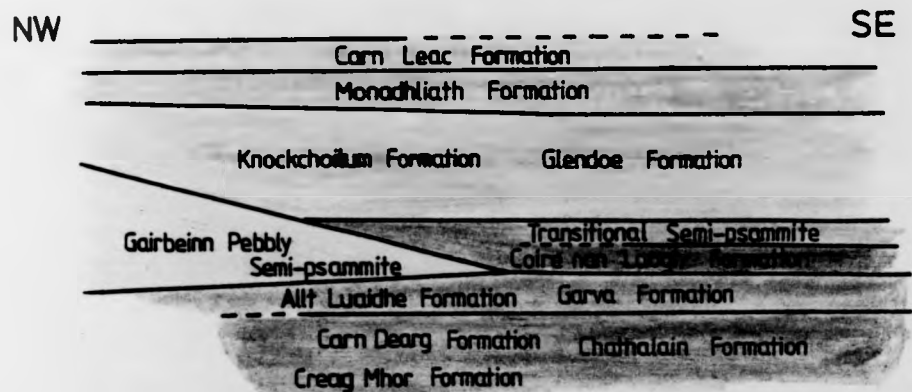Upper Strathspey·areas

Fig.5.6: Schematic model of the sedimentary disposition of the various stratigraphical units

NW ———————————— – – – – – – – – – SE

Carn Leac Formation

Monadhliath Formation

Knockchoilum Formation   Glendoe Formation

Gairbeinn Pebbly   Transitional Semi-psammite
         Semi-psammite   Coire nan Laogh Formation

Allt Luaidhe Formation   Garva Formation

Carn Dearg Formation   Chathalain Formation
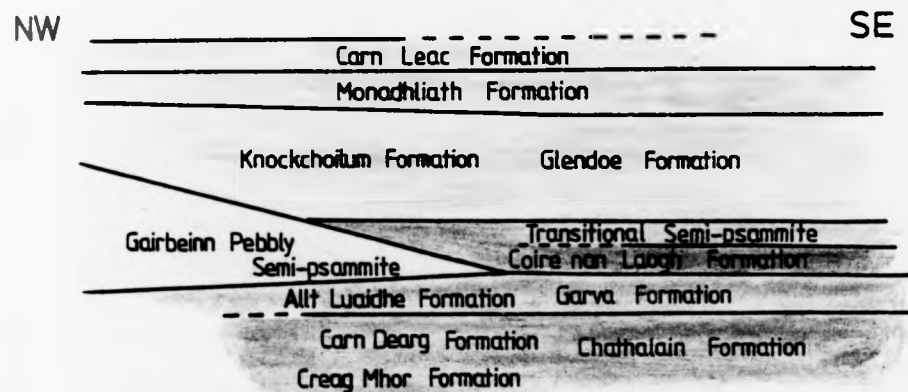
Creag Mhor Formation

Fig.5.6: Schematic model of the sedimentary disposition of the various stratigraphical units

higher formations of the Corrieyairack Succession in the NW. This geometry
is consistent with thrust propagation and ramping towards the NW. This model
may involve a significant displacement on the slide and appreciable
transport of the Corrieyairack rocks. The poor stratigraphical constraints
and lack of appropriate markers, however, make determination of the displacement
impossible.

On the other hand, the top of the Gairbeinn Pebbly "alluvial wedge" could
have provided a surface on which the less competent and possibly onlapping
dominantly semi-pelitic Coire nan Laogh Formation slid during the D1 deformation

5.4 The Second Deformation, D2

5.4.1 Minor Structures

Minor F2 folds deforming So/S1 are found mainly on the hinge of the
Laggan Antiform in the Creag Liath and Creag Chathalain area (subarea 1)
and on its limbs in semipelitic and thinly-bedded semi-psammites.

The style of D2 folding varies according to the lithology with inter-limb
angles varying from tight in pelitic rocks to open in thickly-bedded semi-
psammites.

They are commonly asymmetric and possess a lower amplitude to wave length
ratio than F1 minor folds (Plates 5.13 & 5.14) South of Creag Liath (NN494937)
they verge consistently to the north, whereas north of Creag Liath and Creag
Tharsuinn (NN446856) they verge consistently to the south, indicating the
relative location of the D2 major fold, the Loch Laggan Anticline.

The dips of the axial surfaces of the F2 minor folds are variable, ranging
from 40° to 80°. In some subaraeas i.e. 1,2, 6,7, poles to F2 axial planes
lie on the same great circle as the poles to the So/S1, reflecting the tight
nature of most of the folds in these subareas, and the parallelism of S2 to So/
S1, locally.

An L2 intersection lineation parallel to the fold axis is locally measurable
and shows variable orientation. This variability in the hinge

Plate 5.13 : D2 minor fold of platy semi-psammite. Garva

Formation. NN 520940.

Plate 5.14 : D2 minor fold of semi-psammite.

Glendoe Formation. NN 447854.

Hammer is 33.5cm long.

Plate 5.13 : D2 minor fold of platy semi-psammite. Garva

Formation. NN 520940.

Plate 5.14 : D2 minor fold of semi-psammite.

Glendoe Formation. NN 447854.

Hammer is 33.5cm long.

Plate 5.13 : D2 minor fold of platy semi-psammite. Garva

Formation. NN 520940.

Plate 5.14 : D2 minor fold of semi-psammite.

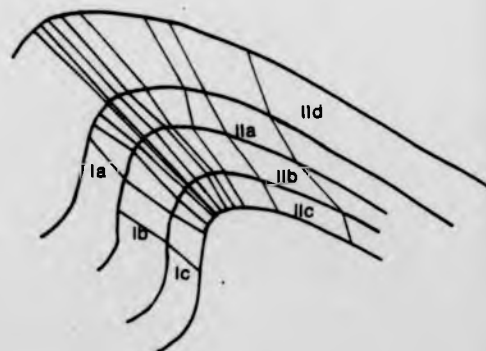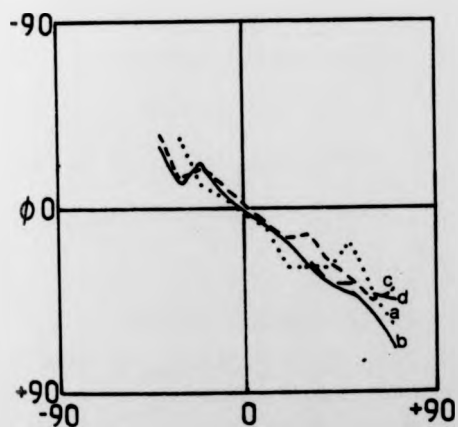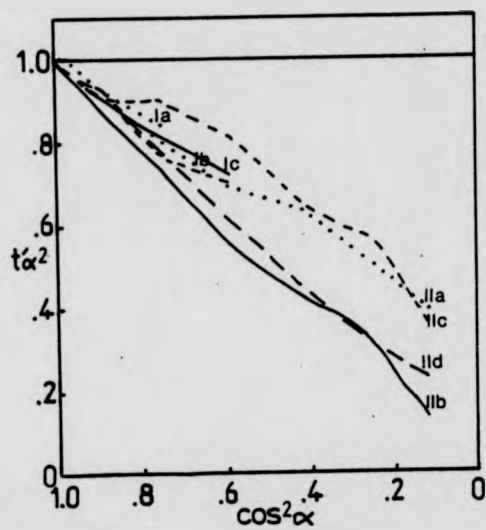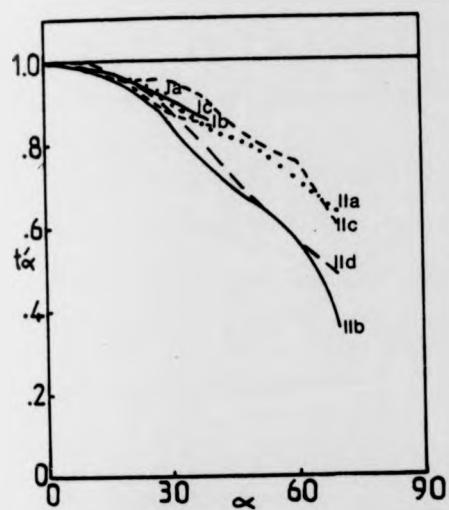Glendoe Formation. NN 447854.

Hammer is 33.5cm long.

Fig.5.7: Analysis of the geometry of a D2 minor fold

directions of minor F2 folds arise partly from their having been superposed on variably oriented D1 surfaces and partly due to deformation during later events, D3 and D4. In subareas 1 and 2 this lineation plunges about $20^{\circ} - 30^{\circ}$ to the ENE and occasionally to the WSW, whereas in subareas 5,6,8 and 9, it plunges $10^{\circ} - 30^{\circ}$ to the SW. In the Coire Ardair area, subarea 7, there is a wide scatter of F2 lineation in the southern quadrants.

The high axial direction stability of F2 folds in partsof subareas 1 and 2 is due to the coaxial nature of D1 and D2 folding, the interference structure of which has been described in section 5.5.

F2 minor folds are associated with a crenulation cleavage which is tight in the semi-pelites.

5.4.2 F2 minor fold layer analyses

12 minor folds comprising 29 individual layers were analysed using the methods already outlined with regard to F1 minor folds. The results for F2 minor folds are presented in Table 5.2. There is generally a good agreement between the results from the $t'\alpha$ against α and $\emptyset$ against α plots.

Most of the fold layers have Class 1C geometry, but a few fall into Class 2 and 3; these are invariably layers underlying those with Class 1C geometry. The various layers of F2 minor folds which were analysed are, generally, of the same rock-type, semi-psammite and so it is not possible to determine the effect of rock type on fold layer geometry.

5.4.3 D2 Major Structures

The structure of the Loch Laggan-Upper Strathspey area is dominated

Table 5.2 : D2 minor fold profile analyses

Fold No	Location and stratigraphic position	Lithology	Fold class from tα vs α	Fold class from φ vs α	Flattening strain ratios: $\sqrt{\ }$ from tα vs cos²α	from φ vs α
1a	NN 493946 Chathalain Formation	a)thinly-)bedded	1C-2	2 - 3	0.2	-
		b)semi-psammite	3	2 - 3	-	-
1b	NN 493946 Chathalain Formation	a)thinly-)bedded	1C-2	2	0.2	-
		b)semi-psammite	1C-2	2	0.3	-
2	NN 489948 Chathalain Formation	a:thinly-bedded semi-psammite	1C-3	3	0.4	-
3	NN 520940 Garva Formation	a)thinly-	1C	1C	0.2	0.3
		b)bedded	1C	1C	0.55	0.5
		c)semi-	1C	1C	0.3	0.3
		d)psammite	1C	1C	0.5	0.4
4	NN 520940 Garva Formation	a)thinly-)bedded	1C	1C	0.65	0.63
		b)semi-psammite	1C	1C	0.5	0.5
5	NN 497927 Coire nan Laogh Formation	a:thin semi-psammite band in semi-pelite	1C	1C	0.65	0.67
6	NN 487927 Coire nan Laogh Formation	a:quartzo-feldspathic band in semi-pelite	1C	1C	0.4	0.35
7	NN 528938 Coire nan Laogh Formation	a)thinly-)bedded	1C	1C	0.72	0.75
		b)semi-psammite	1C	1C	0.52	0.45
8	NN 445852 Glendoe Formation	a)thinly-	1C	1A-1C	0.35	0.5
		b)bedded	1C	1A-1C-3	0.9	0.5
		c)semi-	3	1A-1C-3	-	-
		d)psammite	3	1A-1C-3	-	-
9	NN 445852 Glendoe Formation	a)thinly-	1C	1C	0.7	0.6
		b)bedded	3	1C-3	-	0.3
		c)semi-	1C-3	1C-3	0.2	0.6
		d)psammite	1A-3	1A-3	-	-
		e)	1C-2	1A-1C	0.4	0.6
10	NN 448868 Glendoe Formation	a)thin semi-)pelite	1C	1C	0.6	0.6
		b)thin semi-)psammite	1C	1C	0.75	0.7
11	NN 436913 Monadhliath Formation	a:thin band of semi-psammite	1C-3	1C-3	0.2	0.2
12	NN 435878 Monadhliath Formation	a)calc-silicate band	1C	1C	0.55	0.5
		b)semi-psammite band	1C	1C	0.55	0.6

Fig.5.8(i):Equal area stereographic projections of poles to So//S1

_____ Great circle(girdle) defined by poles to So//S1

o   Pole to the great circle

(ii):Equal area stereographic projections of minor fold axes
and other lineations, poles to minor fold axial planes

x : D1 lineation

⋇ : D2 lineation
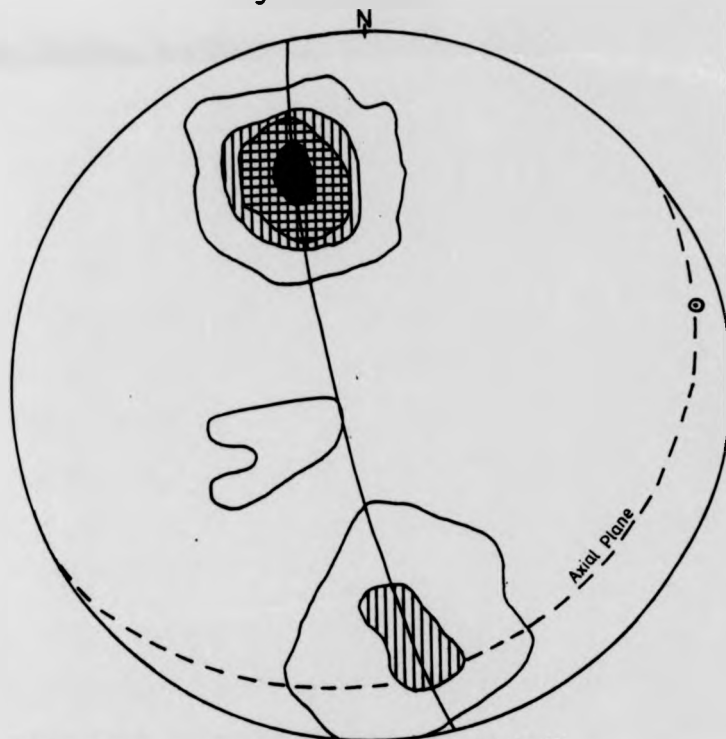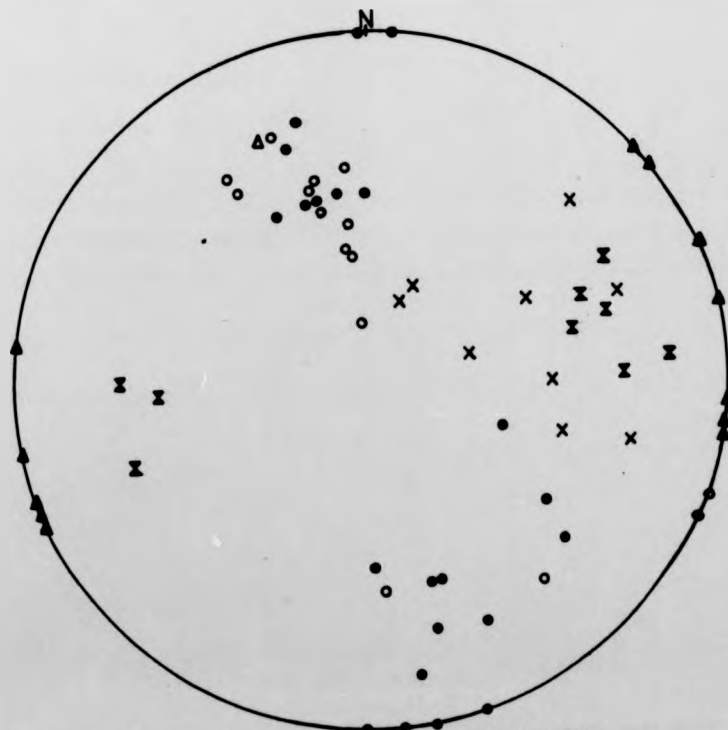
△ : D3 lineation

Poles to axial planes

o : D1

● : D2

▲ : D3

Fig.5.8a: Subarea 1

(i) Poles to So/S1   Contours at 1,5,10,20%   No of points 271

(ii) Minor fold axes and other lineations, poles to minor fold axial planes

by D2 major folding. One major D2 fold, the Laggan Antiform, with an axis
generally trending NE-SW has been recognised from the attitudes of the So/S1
surfaces and the vergences of the D2 minor folds.

Anderson (1956) recognised this fold and termed it the Loch Laggan
Anticline. He traced it from its southern extension into the Ossian Granite,
through the ground north-west of Loch Laggan to south of Carn Liath (NN470902),
where he reported that its north-western limb steepens and is overturned.
This led Anderson (op. cit.) to consider it a complex structure in the Carn
Liath area.

This study confirms that the geometry of the structures in the area south
of Carn Liath is a complex one which resulted from the interference of three
generations of folding: D1, D2 and D3.

In its south-western extension, north-west of Loch Laggan,
(subareas 7 and 8), the Laggan Antiform is a tight fold with a subvertical
axial surface dipping 72°NW. The fold axis of this section of the antiform
plunges shallowly at about 10° to the south-west (235). The cylindrical
nature of the fold is indicated by the good concentration of the poles to
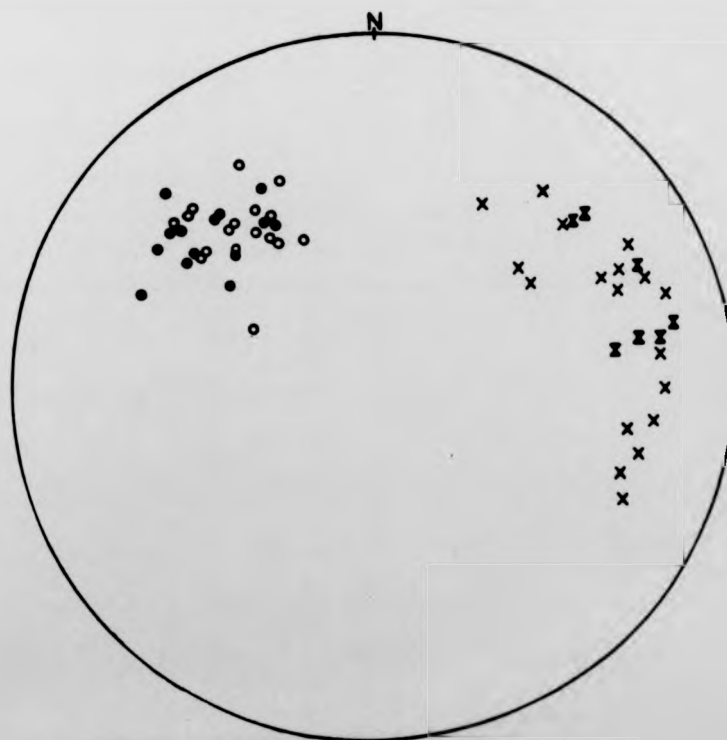the So/S1 surface (Fig 5.8a).

Just south of Na Cnapanan (NN477888), the strike of the So/S1 surfaces
of the northwestern limb of the Laggan antiform swings round from NE-SW to
almost N-S (Fig 5.1). This rotation is a result of later D3 folding and is
discussed later.

The effect of this D3 refolding is illustrated by a stereographic
analysis of poles to So/S1 in subarea 5 (Fig 5.2). The stereogram (Fig 5.8c)
produces a cross-girdle; the pole to the girdle in the northern quadrants
(continuous line) gives the trend and attitude of the rotated axis of the
Laggan Antiform as plunging $30^{o}$to the SSW.

Fig. 5.8b : Subarea 2
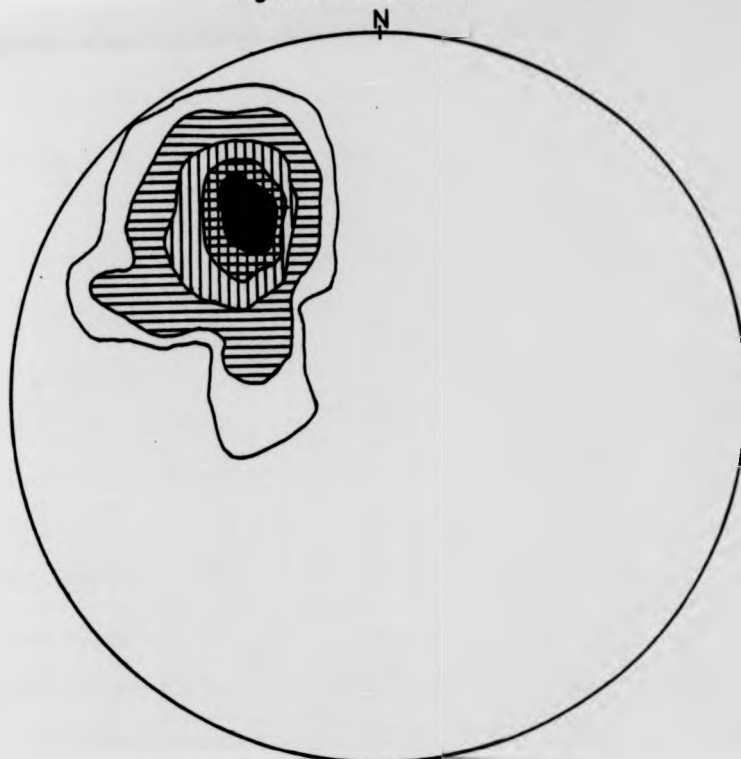
(i) Poles to So∥S1   Contours at 1,3,9,15,20%   No of points 274

(ii) Minor fold axes and other lineations, poles to minor fold axial planes

North of Na Cnapanan, the dip of the So/S1 surfaces steepens to 75° becoming almost vertical at NN 472892. The dip then changes abruptly to the SE indicating a synformal fold hinge, in this area, inferred to be D2.

A local reversal of younging at NN482887 suggests a D2 refolding of a minor or medium scale D1 isoclinal fold pair (Encl. 3).
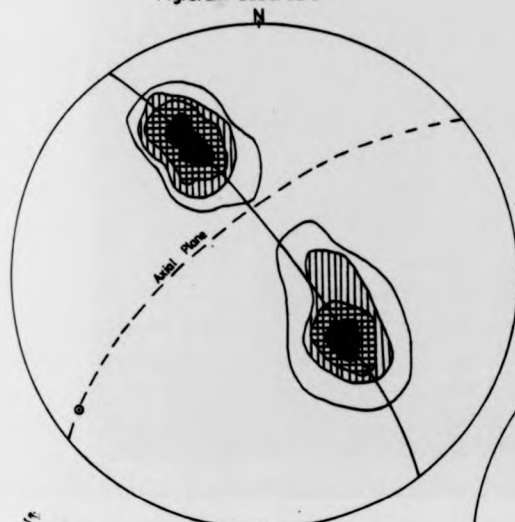
Immediately extending east from Coille Coire Chrannaig (NN 485908) the strike of the So/S1 surfaces abruptly become ENE-WSW and dip SE at approximately 45°. A major dislocation is therefore inferred at Coille Coire Chrannaig considered a late fault.   This fault juxtaposes the highly inclined, subvertical north-western limb of the Laggan Antiform in this area with the more gently dipping southern limb of the easterly section of the antiform.

The northerly swing of the fold axis and the tightening of the limbs of the antiform north of Na Cnapanan is attributed to the interference of D2 folds with later D3 folds which resulted in its rotation. The location of the fault in this area suggests that it could be related to brittle movements following the D3 ductile phase of deformation.

The axis of the easterly extension of the antiform, beyond the fault, in the An Doire (NN470920) area, subarea 6 (Fig 5.1) trends at N2OE and plunges 44° SSW (Fig 5.8e). The stereographic plot of the axes of minor folds and crenulations in the Coire nan Laogh Semi-pelite is not congruous with the β -axis determined from the analysis of S1 surfaces. These axes on the average plunge 22° at 312. This observation suggests that the major fold and the minor folds considered were not coeval or related. Whereas the major structure is a section of the D2 Laggan Antiform, the minor folds are D3. These minor folds are congruous with the major F3 synform, the Allt Chrannaig synform, described later.
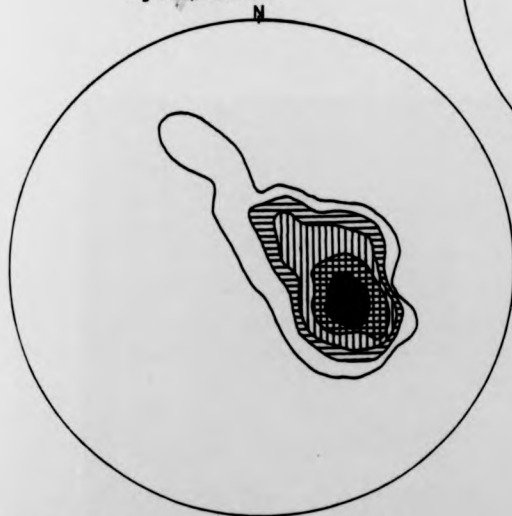
The north-easterly extension of the Loch Laggan Anticline in the Creag

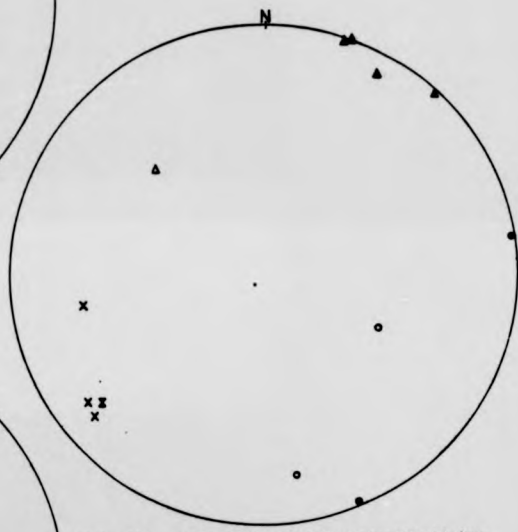Fig. 5.8c: Subarea 8

(i) Poles to So/S1   Contours at 1,5,10,20%   No of points 133

(ii) Minor fold axes and other lineations, poles to minor fold axial planes

Fig. 5.8d: Subarea 9

(i) Poles to So/S1   Contours at 1,3,6,12,24%   No of points 105

Plate 5.15 : Interference structures between D1 and D2 folds of
quartzo-feldspathic segregations in semi-psammite.
Chathalain Formation. NN 504947.
Hammer handle (black) is 17.5cm long.

Plate 5.16 : Interference structure between D1 and D2 folds of
semi-psammite.
Chathalain Formation. NN 489948.
Lens cap is 5.5cm in diameter.

Plate 5.15 : Interference structures between D1 and D2 folds of
quartzo-feldspathic segregations in semi-psammite.
Chathalain Formation. NN 504947.
Hammer handle (black) is 17.5cm long.

Plate 5.16 : Interference structure between D1 and D2 folds of
semi-psammite.
Chathalain Formation. NN 489948.
Lens cap is 5.5cm in diameter.

Plate 5.15 : Interference structures between D1 and D2 folds of
quartzo-feldspathic segregations in semi-psammite.
Chathalain Formation. NN 504947.
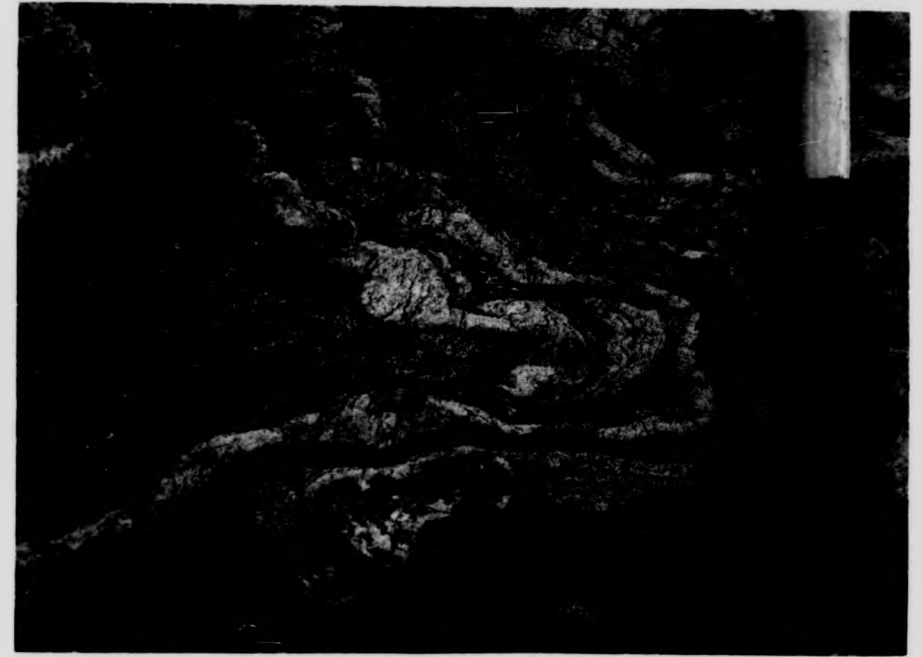Hammer handle (black) is 17.5cm long.

Plate 5.16 : Interference structure between D1 and D2 folds of
semi-psammite.
Chathalain Formation. NN 489948.
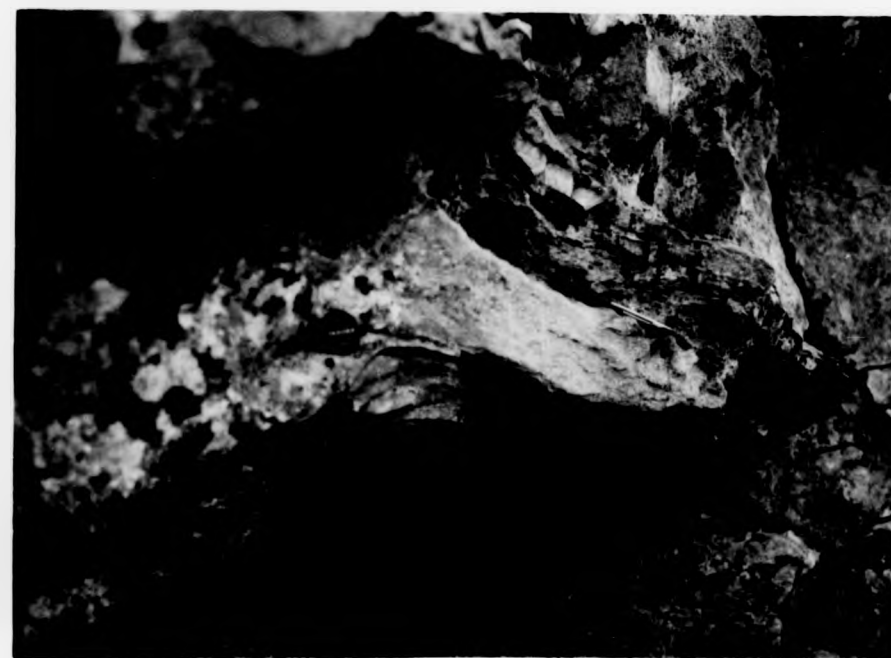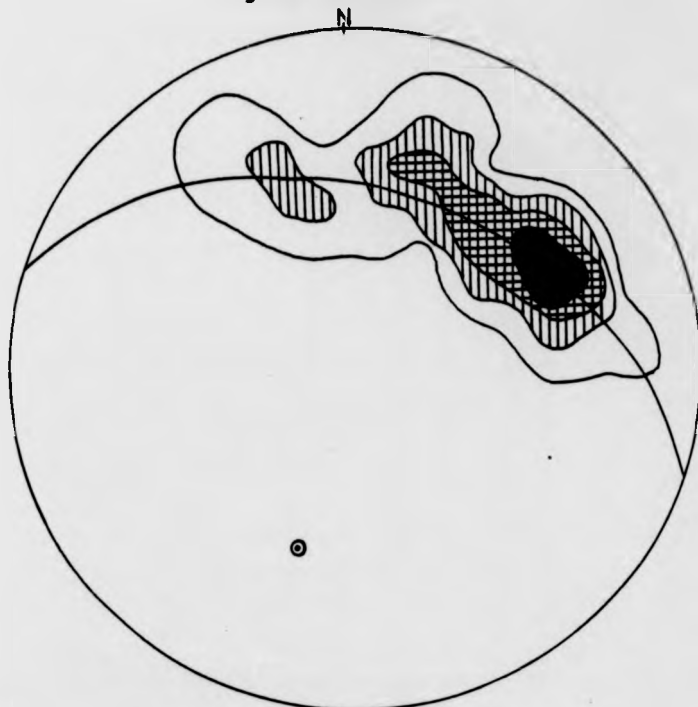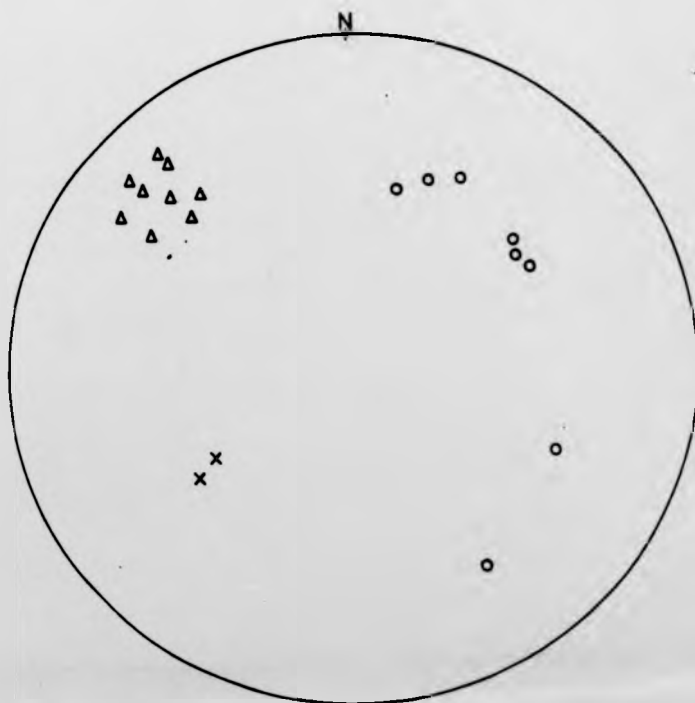Lens cap is 5.5cm in diameter.

Fig.5.8e: Subarea 6

(ii) Contours at 1,5,10,20 %   No of points 181

(i) Poles to So/S1

(ii) Poles to minor fold axial planes, fold axes and other lineations

Liath (NN492942) area , subarea 1, is also a tight fold with a subvertical axial surface, dipping 70-80°SE. Its axial trace trends 76° and plunges approximately 10° towards the Allt Crom Granite. The changes in the direction of the plunge of the axis of the antiform as it is traced from the southwest is caused by later crossfolding during D3.

On Creag Liath and Creag Chathalain, the hinge zone of this fold is occupied by the rocks of the lowest stratigraphic unit mapped in the Loch-Laggan-Upper Strathspey area, the Chathalain Formation.

5.4.4. The relationship between D1 and D2 structures

Where overprinting of D2 minor folds on D1 minor folds have been observed, as at NN 489948 on Creag Chathalain (Plates 5.15 and 5.16) coaxial refolding is the result. Subvertical D2 minor folds refold subhorizontal D1 minor folds producing type 3 interference structures of Ramsay (1967)

The general coaxiality of F2 and F1 is also illustrated by the various stereographic equal area plots of both F1 and F2 fold axis (Fig 5.8). which are concentrated in the same quadrant of the net.

However, in the vicinity of the Gairbeinn Slide, D1 minor fold axes are oriented at high angles to those of D2 folds (Fig 5.8b ii). This local non-coaxiality may be attributed to the rotation of D1 minor fold axes towards the transport direction during the sliding (Escher and Watterson, 1973; Ramsay 1980).

5.5   The Third Deformation, D3

5.5.1 Minor structures

Minor F3 folds are common in most parts of the Loch Laggan-Upper Strathspey area, especially near the hinges of F3 major folds. These F3 folds deform So, S1, S2, granitic and pegmatite bands and possess an axial planar crenulation fabric, S3, the intensity of which varies according to the deformed lithology. It tends to be strong in the pelites and semi-pelites and weak in the psammites and semi-psammites.

F3 minor folds are generally gentle to open folds with interlimb angles of 120° - 160° (Plates 5.17 and 5.18). In the semi-pelites, however, the folds become tight. They generally possess subvertical axial surfaces and variably plunging axes. The S3 crenulation axis is generally measurable as a lineation (L3) and is strongly developed in the semi-pelitic lithologies.

Only a few F3 minor folds were available for fold profile analyses. Those analysed showed Class 1B to Class 1C geometries.

5.5.2 Major D3 Folds

Three major F3 folds have been recognised from the analyses of the variations in the geometry of So/S1 surfaces, the study of the geometry of D3 minor folds and the modifications of D1 and D2 linear structure

a.   The Coire Ardair Synform

This is a large open fold, with variably dipping limbs and a subvertical axial plane dipping 75° SE. Stereographic analyses of the poles to the So/S1 surfaces in subarea 9 (Fig 5.8f) show that this synform plunges by about 30° to the west (272°). The great scatter shown by the poles to the

Fig.5.8f: Subarea 7
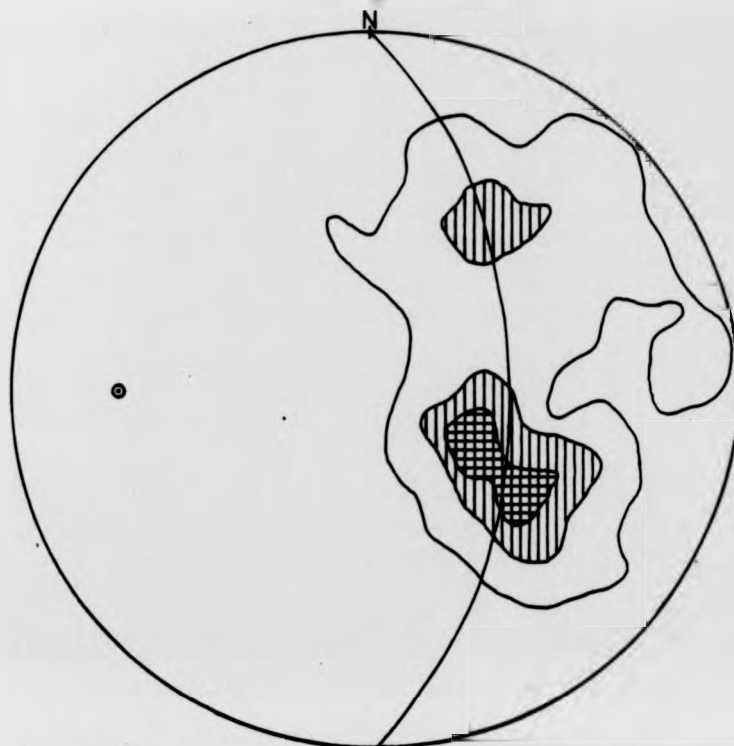
(i) Poles to So/S1   Contours at 1,5,10%   No of points 392
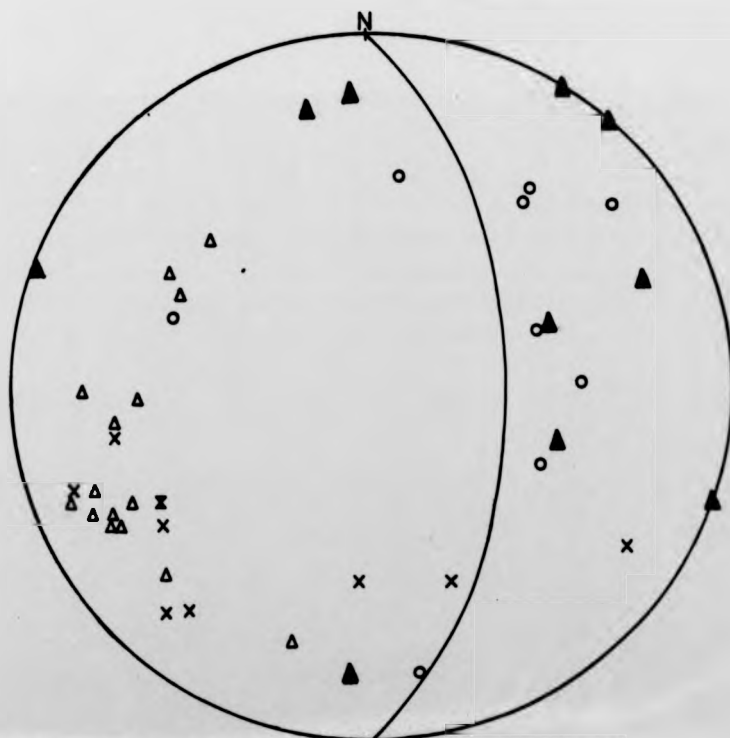
(ii) Minor fold axes and other lineations, poles to minor fold axial planes

Plate 5.17 : D3 minor fold of a thin calc-silicate band in

Glendoe Formation. NN 483895.

Plate 5.18 : Early pegmatitic granite folded during D3.

NN 482887.

Hammer is 33.5cm long.

Plate 5.17 : D3 minor fold of a thin calc-silicate band in
Glendoe Formation. NN 483895.

Plate 5.18 : Early pegmatitic granite folded during D3.
NN 482887.
Hammer is 33.5cm long.

Plate 5.17 : D3 minor fold of a thin calc-silicate band in
Glendoe Formation. NN 483895.

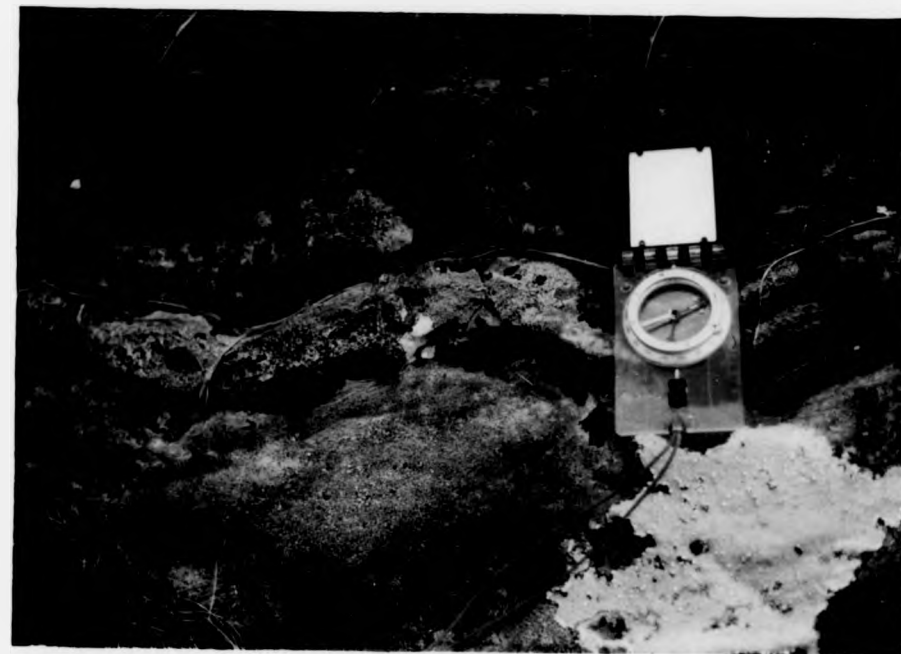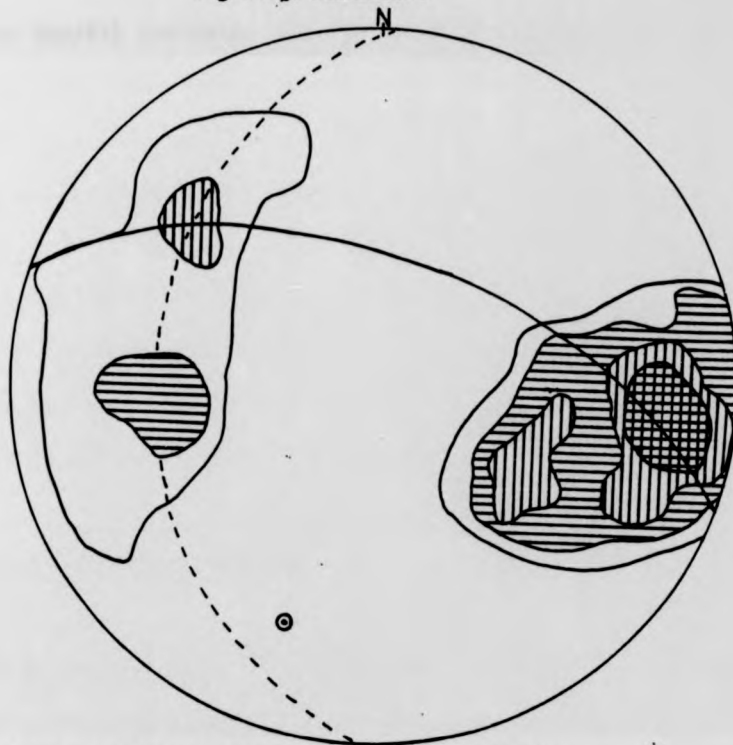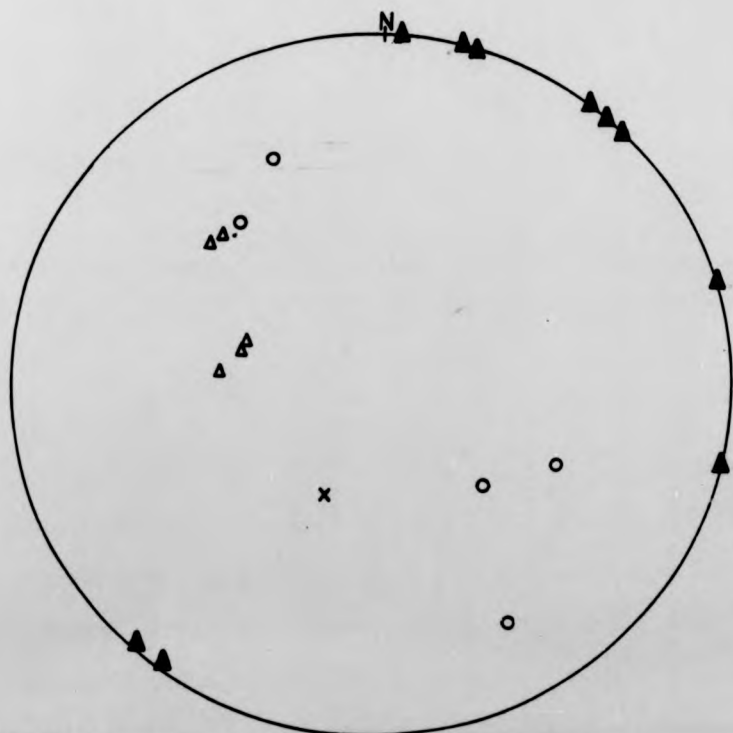Plate 5.18 : Early pegmatitic granite folded during D3.
NN 482887.
Hammer is 33.5cm long.

Fig. 5.8g: Subarea 5

(i) Poles to So/S1   Contours at 1,2,5,10%   No of points 123

(ii) Minor fold axes and other lineations, poles to minor fold axial planes

foliation (So/S1) indicates the non-cylindrical geometry of the Coire Ardair Synform. This geometry may either be a result of post-D3 deformation or of the fact that the synform has deformed previously folded and therefore variably oriented So/S1 surfaces. This scatter is also shown by F3 minor fold axes (L3). The great circle (girdle) formed by L3 coincides with the mean orientation of S3 (the axial plane of F3). Also, the pole of the L3 (lineation) girdle coincides with the maximum for the poles to So/S1. This would suggest that the D3 lineations (i.e. axes of minor F3 folds) have been superposed on earlier folded surfaces (Turner and Weiss, 1963).

b.   The Cam Liath Antiform

The axial trace of the complementary antiform to the Coire Ardair and the Allt a Chrannaig synforms runs approximately WNW-ESE, south of Carn Liath (NN470902). This antiform is named the Carn Liath antiform.

Stereographic analysis of the poles to the So/S1 surfaces in subarea 5 (Fig 5.1) show a cross-girdle (Fig 5.8g). The axis of this antiform is given by the pole to the girdle in the western quadrants (dashed curve), which shows a plunge of $49°$ towards 095.

The other girdle in the northern quadrants (continuous line in Fig 5.8c) belongs to the refolded Laggan Antiform which has been described earlier.

c.   The Allt a Chrannaig Synform

This is an open fold with an interlimb angle of about $150°$, and moderately dipping limbs (50-60 NE & SW). The synform is indicated by the analyses of the variations in the So/S1 surfaces in subarea 3. It possesses a vertical axial plane striking at 312 and an axis that plunges to $59°$ to the SE (134) (Fig 5.8h).

The Allt a Chrannaig synform is a non-cylindrical, fold as a result of

Fig. 5.8h : Subarea 3
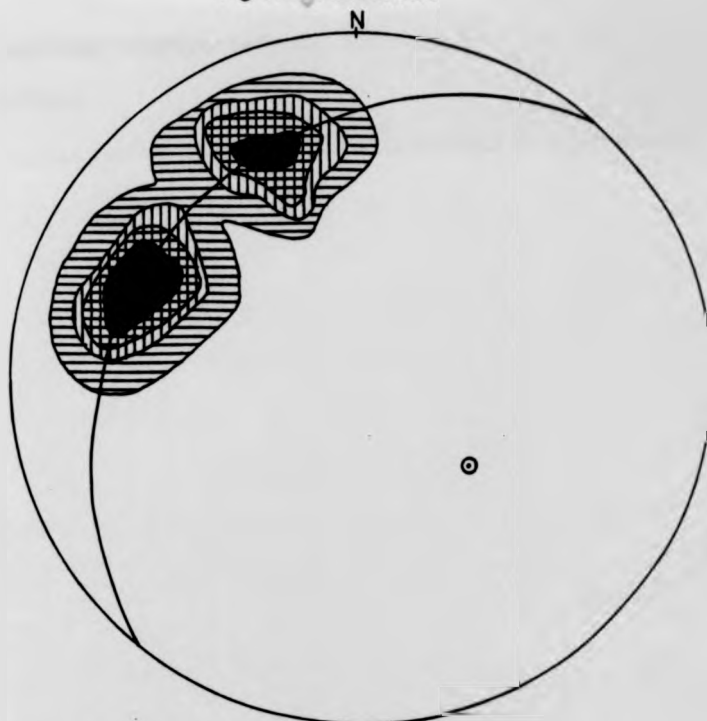
(i) Poles to So//S1   Contours at 1,5,10,20%   No of points 74

Fig. 5.8 i : Subarea 4

(i) Poles to So//S1   Contours at 1,5,10,20%   No of points 93

its superposition transversely on the southeastern-dipping limb of the D2 Laggan Antiform.

5.5.3 The interference patterns of D2 and D3 folds

Where overprinting of D2 structures by D3 folds have been observed, as at NN464932 and Creag Chathalain (NN488947) the resulting interference pattern is a mushroom shape similar to type 2 of Ramsay (1967). (Plates 5.19 and 5.20) . This pattern is the result of the superposition of upright D3 folds on slightly recumbent D2 folds, with the axial plane of the D3 fold intersecting that of the D2 fold at a high angle.

Plate 5.19 : Mushroom interference structure resulting from the
superposition of subvertical D3 fold on an earlier
D2 fold.
Coire nan Laogh Formation. NN 464931.

Plate 5.20: Upright D3 crenulation refolding earlier
subhorizontal D2 crenulation in semi-pelite in
Monadhliath Formation. NN 450901.
Plane polarised light. Scale bar represents 0.2mm.

Plate 5.19 : Mushroom interference structure resulting from the
superposition of subvertical D3 fold on an earlier
D2 fold.
Coire nan Laogh Formation. NN 464931.

Plate 5.20: Upright D3 crenulation refolding earlier
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Plane polarised light.   Scale bar represents 0.2mm.

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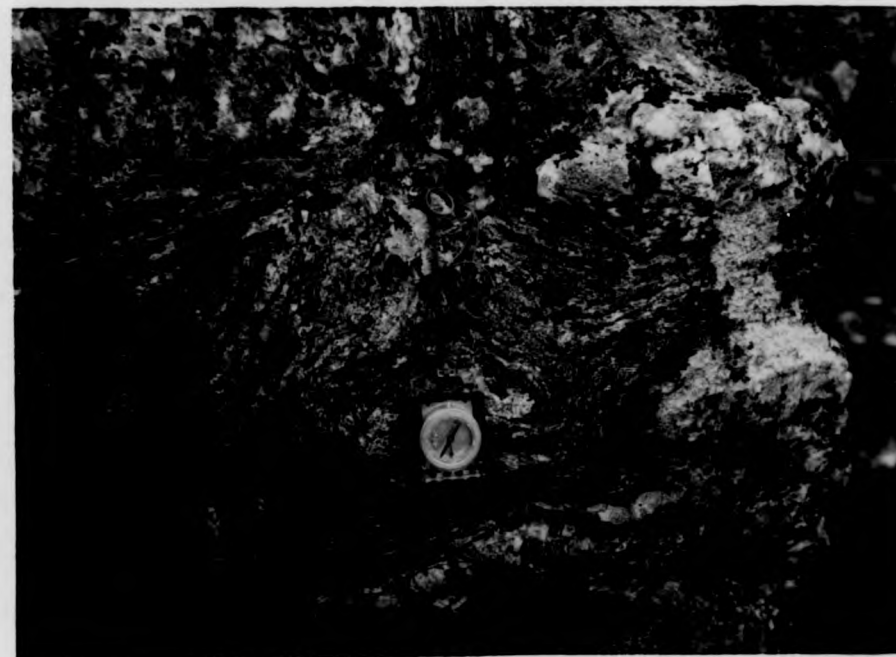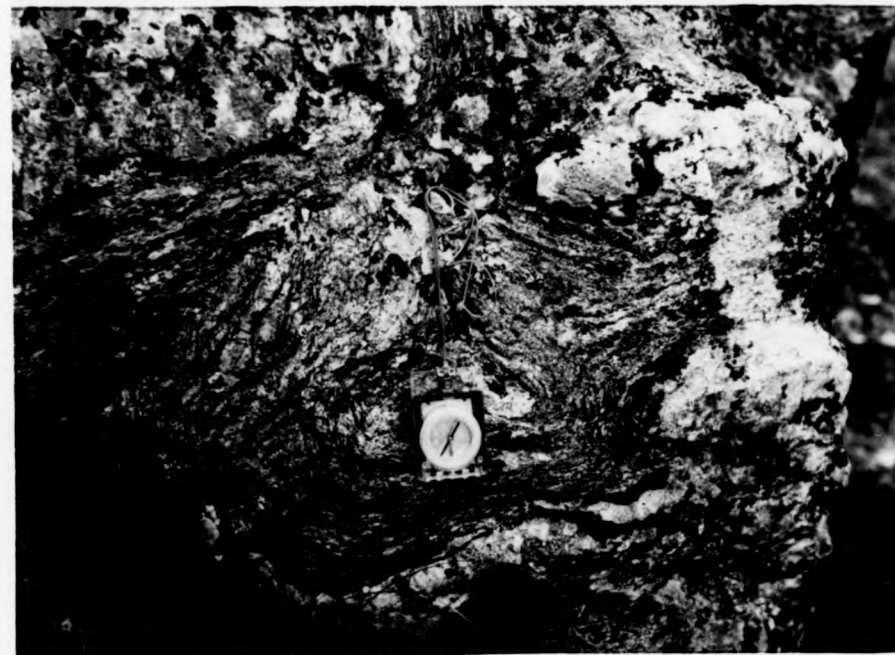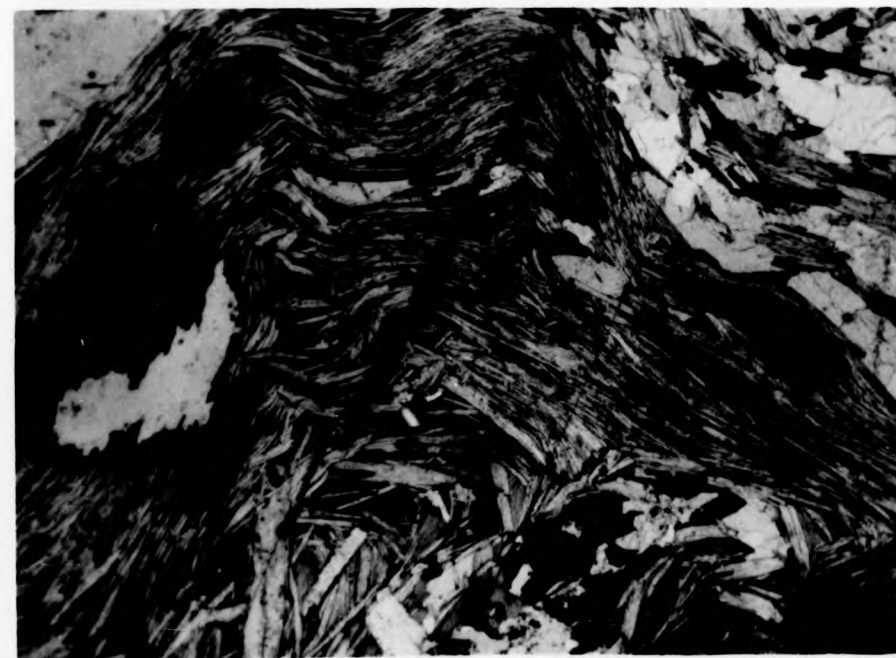Coire nan Laogh Formation. NN 464931.

Plate 5.20: Upright D3 crenulation refolding earlier
subhorizontal D2 crenulation in semi-pelite in
Monadhliath Formation. NN 450901.
Plane polarised light. Scale bar represents 0.2mm.

5.6   The Fourth Deformation, D4

This consists of brittle deformation (faults and fractures) which have variably displaced earlier structures. These faults generally trend ENE or NW and are invariably marked by narrow gorges commonly occupied by streams. They include:

a.   The Glen Markie Fault

The present study is bounded in the east by the Glen Markie Fault (Anderson, 1956) who reported a sinistral displacement of about 7.2km on the related Ericht-Laidon Fault. Definite statements about the effects of the Glen Markie Fault must await the results of current research in the area immediately to the east and correlations across the fault.

b.   A NNW-SSE trending transcurrent dislocation which displaced the axial trace of the Laggan Antiform sinistrally for about 1km. The trace of this fault which was discussed in connection with the Laggan Antiform follows the Coille Coire Chrannaig (NN 485895) into the ravine beyond Carn Liath (NN475905).

c.   Another fault trending NE-SW sinistrally displaces the Monadhliath Formation by about 200m south of Meall Ptarmigan (NN 432912). The fault zone is marked by a narrow gorge now occupied by a tributary stream to the Allt Coire Bhannain.

d.   A fault subparallel to the Meall Ptarmigan Fault also sinistrally displaces the slide contact between the Garva and the Coire nan Laogh Formations.by about 60-100m along a narrow ravine occupied by Allt na Doire (NN 464924).

e.   A deep gorge, now partly occupied by a stream, Caochan na Dirce, separates Creag Chathalain from Creag Liath (NN 491940). This fracture trends ENE but the sense of movement on it cannot be determined because of the absence of markers.

f.   A fracture trending NW-SE is defined by Coill a Choire and the gorge

of Uinneag Min Choire. The sense and amount of displacement on this fracture

is however, not known.

g.   Another fracture also trending NE-SW which was traced by Anderson (1956)

runs through Coire Garbh (NN430843) and the valleys of Allt a Choire

Chomharsain and Allt a Chrannaig. The sense of displacement on this fracture

cannot , however, be determined because of the absence of any markers.

The general NE-SW trend of most of the fractures and faults in this

study area suggests that they could be related to the Glen Markie and

subparallel faults mapped by Anderson (op. cit.).

This late deformation has also given rise to a widespread kinking of

micas and other minerals in the rocks of the present study area (see

Chapter 3 Petrography).

5.7   Strain Analysis

Traditionally several methods involving objects of known original shapes

and orientations have been used to determine the amount of strain suffered

by deformed rocks.

In the rocks of the Loch Laggan - Upper Strathspey area useful strain

markers are very rare. Since extensive recrystallisation under amphibolite

facies conditions postdated the important deformation episodes, D1 and D2,

mineral grain shape fabrics which could have been used for semi-quantitative

finite strain estimation are annealed. However, as discussed under D1 shear

zones, it is possible to show qualitatively the high strains suffered by the

rocks of the Garva and the Coire nan Laogh Formations within this zone.

The only semi-quantitative method of strain analysis applicable in this

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finite strain estimation are annealed. However, as discussed under D1 shear
zones, it is possible to show qualitatively the high strains suffered by the
rocks of the Garva and the Coire nan Laogh Formations within this zone.

The only semi-quantitative method of strain analysis applicable in this

study area is the determination of the degree of flattening of concentric

fold layers (Ramsay 1967, Hudleston, (1973). Flattening, however, gives a

minimum estimate of the total strain suffered by the rocks because it

excludes the strain components caused by the initial buckling and any layer

parallel shortening.

Ramsay (1967) suggested that in environments of ductile deformation,

the shape of the buckled layers becomes modified by the superposition of

a homogeneous strain. De Sitter (1958) suggested that there is a limit

to the amount of shortening, up to 36%, which can be achieved by buckling

alone. Further compression, he suggested, would lead to the modification

of the shape of the folded layers to flattened parallel folds (Ramsay, 1962)

or to their rupture.

Superimposed flattening strains have been worked out by Ramsay

(1962, 1967) in terms of values of $\sqrt{\lambda_2/\lambda_1}$ from a graph of *tα against *α

(Fig 5.3) (*These terms have already been defined.).

Hudleston (1973) has also devised several graphs for evaluating

superposed flattening in terms of $\sqrt{\lambda_2/\lambda_1}$ from plots of $t\alpha^2$ against $\cos^2\alpha$

(Fig 5.3) and $\emptyset$ against α (Fig 5.3.)

By the application of these curves flattening strains for folds can

be determined in two dimensions within the profile plane, provided that the

initial fold axis was parallel to one of the principal directions of three--

dimensional strain. This is not always the situation (Flinn, 1962).

For a complete determination of strain, flattening must be measured in

three profile planes of different orientations given that one of the principal

axes of strain lies in the axial plane of the fold and that the orientation

of the fold axis changes within the axial plane (Mukhopadhyay, 1965).

Since, in this study, flattening $(\sqrt{\lambda_2/\lambda_1})$ was determined in only one profile plane, only apparent flattening was obtained.

Hudleston (1973) has suggested that where the ductility contrast between the layers is low the fold layer will deform by simultaneous buckling and flattening. This combination of both processes gives rise to a different set of curves for the determination of flattening strain, a difference which is accentuated at very high limb dips but is negligible at low dips.

The apparent flattening of D1 and D2 folds from various subareas as evaluated from the curves may enable a qualitative comparison of the overall variation of flattening strain in the rocks of the Loch Laggan - Upper Strathspey area.

Hudleston (1973) has compared the relative merits and advantages of using either the $t'\alpha$ vs α or the $\emptyset$ vs α plot. Since the errors in $t'\alpha$ are less than those in $\emptyset$ the $t'\alpha$ vs α plot gives more accurate results especially when the datum is correctly located. On the other hand, the $\emptyset$ vs α plot is best used for multiple folds or folds with obscure hinges.

Application of both plots, in most cases, gave similar results (Tables 5.3 and 5.4). These results show that D1 folds have suffered, on the whole, more apparent flattening (mean $\sqrt{\lambda_2/\lambda_1}$ = 0.30) than D2 folds (mean $\sqrt{\lambda_2/\lambda_1}$ = 0.5) (See Fig 5.7) It should be noted, of course, that the D1 folds contain a component of flattening caused by D2 deformation.

Most of the minor folds analysed came from the Chathalain, Garva, Coire nan Laogh formations and the transitional semipsammite to the Glendoe Formation, and there is very little areal variation in the values of apparent flattening strains obtained. However, as already observed, the methods which have been used may describe an empirical property of the fold not related to flattening (Hudleston, op cit). Also, only folds with classes 1C and 1B
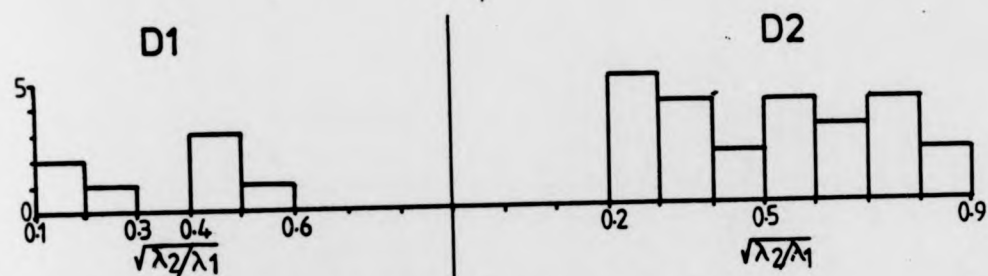
Fig. 5.9: Results of fold flattening analysis

geometries can be used in these methods, and most F1 folds which have classes 2 and 3 geometries are inevitably omitted.

5.8   Regional Correlations

The earliest phase of deformation, D1, recognised in the Loch Laggan - Upper Strathspey area produced minor recumbent isoclines which were refolded by D2 close to tight subvertical folds. These are similar to the early phases of deformation recognised elsewhere in the Grampian Division and the Dalradian of Lochaber by Treagus (1964) Roberts and Treagus (1977, 1979), Hickman (1978), Thomas (1979, 1980), Piasecki (1975, 1980), Whittles (1981), Haselock (1982), Haselock et al. (1982) and Smith (1968).

Medium scale   overturned D1 folds recognised in the present study area may be coeval with the major recumbent D1 folds mapped by Roberts and Treagus (op. cit), Treagus (op. cit), Hickman (op. cit) and Thomas (op. cit) in the Grampian Division rocks of the Ben Alder Forest, north of the Schiehallion Complex and the Lochaber areas.

Complementary to the Laggan Antiform are the Corrieyairack Syncline to the N.W. and the Kinlochlaggan Syncline to the S.E. (Anderson, 1947, 1956; Smith 1968; Piasecki, 1975; Hickman, op. cit; Whittles, op. cit; Haselock, op. cit, Haselock et al, op. cit.)

The zone of high strain, including the slide, between the Garva and the Coire nan Laogh Formations is correlated with the Garbeinn Slide of Haselock (1982), and Haselock et al (1982). The repetition of the slide in the present study area is a result of D2 folding by the Laggan Antiform.

Correlations with the areas to the east and the southeast are hampered

by the intervening presence of the steep belt (Thomas, 1979, 1980) and the Glen Markie Fault (Anderson, 1956). Until the nature and extent of these structural dislocations are fully established no specific correlations are possible although as already noted The Grampian Division rocks of these areas share comparable deformational histories and lithological similarities.

To the northeast the Loch Laggan Anticline of Piasecki (1975) may be regarded as the northern extension of the structure recognised in the present study area.

Although it is attractive to correlate the Laggan Antiform with the Mamore Anticline (Treagus, 1974) or the Inverlair Antiform (Hickman, 1978), such long-range linkages must await the results of detailed work in the intervening area, especially since the axial trace of the Laggan Antiform has been deflected to the east, south of Loch Laggan by the later deformation, presumably by an antiform complementary to the Coire Ardair Synform as shown by reconnaisance traverses east of Moy Lodge (NN 441834) and Loch Spean.