STRUCTURAL, TEXTURAL AND METAMORPHIC STUDIES IN THE LEWISIAN OF GAIRLOCH N.W. SCOTLAND

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Woods near Kerrysdale, Gairloch.
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DECLARATION

I declare that this thesis has been composed by myself and that all the work is my own, except where stated in the text.
Acknowledgements

The project was initiated by Dr A. Beach, department of geology, University of Liverpool, and supervised by Dr R.F. Cheeney, department of geology, Edinburgh University, the help and discussion of both, in field and laboratory, are gratefully acknowledged. Sincere thanks also go to Dr D. Sanderson, department of geology, Queen's University of Belfast, for his invaluable help in computing matters, and also to Dr K. Jones of the same department for his comments on parts of the manuscript.

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The bulk of the research was carried out during tenure of a part-time demonstratorship in the department of geology, Queen's University of Belfast, which is gratefully acknowledged. Many thanks are also due to my parents for financial support during the first year of research at Edinburgh, and for the payment of fees thereafter. Some financial help with fees from the Hardship fund, Edinburgh University, is also acknowledged.
ABSTRACT

A quantitative structural analysis and interpretation of the deformation and metamorphic history of the Lewisian complex at Gairloch is proposed, based on finite strain analysis. The Lewisian rocks at Gairloch form part of the southern region of the foreland Lewisian gneisses of mainland Scotland, and are composed of a belt of largely semi-pelitic schists flanked by areas of gneiss. The dominant foliation dips steeply NE, to which lithological layering is parallel throughout most of the area.

From an investigation in two dimensions it is shown that quartz particles in quartzo-feldspathic gneisses can be used as strain markers and, due to the viscosity contrast between quartz and feldspar, overestimate the bulk strain by approximately 10%. Eighteen specimens of quartzo-feldspathic gneisses, along a NE/SW profile across the area have been analysed for three dimensional finite strain.

Using the theoretical superimposition of strain the deformation paths of the gneisses are modelled for rotational and irrotational strain. The models, supported by field evidence, show that the deformation is of dominantly simple shear type and increases progressively towards the schist belt. The Gairloch area is thus interpreted as part of a large scale shear zone some 5 km across with a history of changing shear direction and sense. Estimates of shear strain from comparisons of the strain data and models, are used to construct a strain profile, and imply a total displacement of about 30 km in a moderately NW plunging direction, with a SW-side down sense.

Petrological and textural studies show that the schists on the SW margin of the schist belt are derived from the gneisses by deformation resulting in extreme grain size reduction, and change in metamorphic conditions, including a decrease in CO₂ content of the metamorphic fluid.
The large scale shear zone at Gairloch, active throughout Inverian and Laxfordian times, forms a major tectonic break separating the Scourian, granulite facies gneisses to the NE from the Laxfordian migmatites to the SW.
CHAPTER 1 INTRODUCTION

1.1. Aims and Brief Description of Project

The principal aim of the project is to attempt finite strain analysis of the Lewisian rocks near Gairloch, NW Scotland, see Fig 1-1, as a basis for the construction of the strain history and structural interpretation of the area. Aspects of the metamorphic history are investigated in relation to the structural analysis. Very little strain analysis of basement rocks, such as the Lewisian gneisses has been reported, due mainly to the lack of strain markers. To date, workers have used methods such as the displacement of dykes across shear zones, Beach (1976) or the variation in fold interlimb angles, Waterson (1980). These methods can however only be used to estimate strain under suitable circumstances and are not applicable when, for instance, dykes are absent or subparallel to the foliation or folds of several generations are present.

In the present study, the quartz fabrics in quartzo-felspathic gneisses have been used to estimate strain. This method was first used by Grocott (1977) to measure three dimensional strain in very coarse grained quartzo-feldspathic gneisses in Greenland, using the method of Dunnet (1969). In these rocks, the quartz was easily distinguished from its pink feldspar matrix and quartz particles were large enough to measure directly from the cut surface. The gneisses at Gairloch are finer grained and quartz not easily distinguished from feldspar on fresh surfaces. Careful examination of weathered surfaces shows, however, that the quartz forms particles within the feldspar matrix as in the Greenland gneisses and after chemical treatment and optical enlargement, cut surfaces can be used. The Dunnet method was found to be unsuitable due to the high strains in many of the Gairloch rocks, and so an alternative method based on a statistical analysis
FIG 1-1 Lewisian outcrops of NW Scotland.
by Hext (1963) was used. This method requires more detailed measurement of the quartz particles but provides confidence limits thus giving information on the precision of the technique. The assumptions made in using quartz particles in a feldspar matrix as finite strain markers, are investigated for the two dimensional case, and the effect of the viscosity contrast between quartz and feldspar assessed.

Three dimensional finite strain is estimated for a total of eighteen specimens along a NW/SE profile, perpendicular to the trace of the main foliation, across the Loch Tollie and S. Sithean Mhor gneisses, Fig 1-2. For each area, the strain history is modelled, using the theoretical superimposition of strain for rotational (simple shear), and irrotational strain (pure shear). The strain type is deduced from field evidence, and from the fit of the strain data to the models. Finally a strain profile is constructed for the whole area, and the deformation history throughout Inverian and Laxfordian times deduced.

A subsidiary part of the project is the investigation of the gneiss-schist contact on the SW margin of the schist belt. During the first mapping of the area, Clough (1908) suggested that these schists are derived by deformation from the gneisses, on the basis of texture and the gradational nature of the contact. Later, Park (1964) interpreted them and the adjacent gneisses as sedimentary in origin but recent geochemical studies, Winchester et al (in press) have supported Clough's view.

The gradational contact at An Ard (see Fig 1-2) has been studied in detail to elucidate the nature of the schists, and to investigate the textural and metamorphic processes at the contact.

1.2. Brief description of Geology

The Gairloch area is part of the Southern outcrop of Lewisian Gneiss complex of NW Scotland, see Fig 1-1, and comprises gneisses and schists
FIG 1-2 Geological map of the Gairloch area.

Key

- Torridonian
- Thick amphibolite sheets
- Schists
- An Ard gneiss
- S Sithean Mhor gneiss
- Loch Tollie gneisses
- Ard Iaittaig gneisses

--- major faults

trace of gneissose layering

Scale

0 1 2 3 4 kms.
affected by strong Inverian and Laxfordian deformation. The complex is composed of two groups of gneisses, the Loch Tollie and S. Sithean Mhor groups, separated by the Gairloch schists, a NW-SE trending belt some 2.4 kms broad (see Fig 1-2).

The Loch Tollie gneisses are layered quartzo-feldspathic, hornblende and biotite gneisses, intruded by numerous dykes, now amphibolite, subparallel to the layering. Both the gneissose layering and the dykes define a fold, known as the Tollie antiform, with a gently SE plunging hinge, gently dipping NE limb, and subvertical SW limb. The S. Sithean Mhor gneisses are largely composed of layered quartzo-feldspathic gneisses. In the SW, the gneissose layering dips moderately NW whereas in the NE of the area it dips steeply NE, see Fig 1-2. Dykes are abundant and trend NW/SE throughout the area. The Gairloch schist belt is composed largely of interlayered biotite and hornblende schists, the lithological layering dipping steeply NE. The schist belt includes a discontinuous layer, \( \sim \frac{1}{2} \) km wide, of biotite and quartzo-feldspathic gneisses known as the An Ard gneisses, see Fig 1-2, which have gradational to sharp contacts with the schists.

On the boundary between the S. Sithean Mhor gneisses and the schists lies the augen shaped outcrop of the Ard Ialltaig gneisses, see Fig 1-2. These differ from the Loch Tollie or S. Sithean Mhor gneisses in containing garnetiferous amphibolite and k-feldspar-free quartzo-feldspathic gneisses.

Throughout the central area, including the SW limb of the Tollie antiform, the Gairloch schist belt and the NE part of the S. Sithean Mhor gneisses, the rocks are highly deformed with a steep NE dipping foliation, to which lithological layering is parallel. Lineations plunge variably NW or SE.

1.3. Previous research.

The rocks of the Gairloch district, together with similar rocks
north of Loch Maree, have long generated interest, due to the presence of metasediments such as banded ironstone, marble and graphite schist, which are unique within the outcrop of Lewisian rocks in Scotland. This area was first mapped by Clough et al, and published in 1907 as sheet 92 of the one inch series of the Geological Survey of Scotland. The high standard of the mapping is notable, and sheet 92 remains the best available map of the area. Clough's observations recorded in the Memoir (1908) also give the first indication of minor occurrences of unusual lithologies such as banded quartz-magnetite rocks, cummingtonite-garnet rocks, marbles and graphite schists. He recognised also the principal structures of the area, dominated by the Tollie antiform so named by him.

Later workers have remapped parts of the area with emphasis on structural features. Park (1964) mapped the schist belt, with parts of the adjacent S. Sithean Mhor and Tollie gneisses, (referred to as the Buanichean gneisses) and also later the area of the Tollie antiform (1970).

Bhattacharjee (1964) mapped the area NW of Loch Tollie, and Ghaly (1966), the S. Sithean Mhor gneisses. Their analysis of the structure is based on minor folds and lineation orientations, with the inference of many large scale isoclinal folds, incorporating up to five phases of deformation. Details of the individual analysis of each of the above and comparison with the present interpretation is dealt with in detail in chapter 9. Unfortunately, little attempt is made by any of the above authors to correlate the structures of their different areas, or to relate the deformation history of Gairloch to that of the rest of the mainland Lewisian rocks. One of the main aims of the project is thus to interpret the deformation history of the area as a whole, and discuss its relation to the rest of the mainland Lewisian outcrop.
1.4. Description of Geography

The Gairloch area is covered by the Ordnance Survey, 1:10, 560, sheet NG 87. The grid references used throughout the following chapters follow the convention specified by the Ordnance Survey, i.e. eastings followed by northings, which are either four figures, correct to 1 km, or six figures correct to 100 m. Maps of the area are available on the scales of 1:50,000, 1:25,000, and 1:10560.

The topography of the area is controlled by the geology, the gneisses forming the highlands and the schists low lying ground. Both the Loch Tollie and the S. Sithean Mhor gneisses form moorland similar to Lewisian foreland country throughout NW Scotland, with rocky knolls separated by areas of peat bog, drained by numerous small streams and containing many lochans, especially abundant in the S. Sithean Mhor area. These areas are generally well exposed (around 30%). The average altitude of the Loch Tollie gneisses is ~ 300 m whereas the S. Sithean Mhor gneisses are more low lying at around 150 m.

The schist belt in contrast forms low lying ground at an average altitude of 80 m. The hornblende schists are relatively resistant to weathering and form ridges, the most prominent being that formed by the Aundrary amphibolite, see Fig 1-2. This amphibolite sheet forms the highest hills in the area reaching a height of 419 m (Sidhean Mhor, 836740, and An Groban, 838749). The biotite schists are generally poorly exposed, showing the best exposure in the NE adjacent to the Aundrary amphibolite. Much of the schist belt is wooded or maintained as grazing land for sheep and cattle. It provides a marked contrast to the moorland of the gneisses, and this together with several sandy beaches, has made Gairloch one of the popular tourist areas of NW Scotland.
1.5. Access

Access is easiest in the schist belt, provided by the main road which runs along the length of the belt, and the old road running across the higher country in the NE. Numerous tracks and paths make walking in most parts easy. Access to the areas of gneiss is more difficult. The road from Gairloch to Poolewe crosses the Loch Tollie gneisses on the north side of Loch Tollie itself, and there is a clearly marked path from this road along the east side of Craig Mhor Tollaidh southwards to the shores of Loch Maree. However, large areas of gneiss south of Loch Tollie can be reached only on foot over difficult country. Access to the S. Sithean Mhor gneisses is by a track that runs SW from Shieldaig, past Loch Braigh Horrisdale, across the Shieldaig deer forest to Diabaig at Torridon. However, the track is not fit for vehicles, and access is thus by foot only.

During the present study, transport was largely by bicycle and by foot. Lack of motorised transport has meant that the outlying parts; to the SE of Loch Bad an Scalaig, Craig Mhor Tollaidh and the SE part of the Tollie gneisses (the Buanichean gneisses) have been but briefly visited. Most field work has therefore been concentrated on the Loch Tollie area, the area around S. Sithean Mhor, and the schist belt between Loch Gairloch and Loch Bad an Scalaig.

1.6. Programme of Study.

The project was begun in September 1976 while I was enrolled as a post-graduate student at Edinburgh and was largely initiated by Alastair Beach, department of geology, University of Liverpool, who was leading a field mapping trip in the Tollie area at the time. Points of interest at that time were the possibility of strain analysis across the Tollie antiform using quartzo-feldspathic gneisses, and the nature of the schists on the SW margin of the Gairloch schist belt.
Due to financial circumstances, I moved to Belfast after gaining the appointment as part-time demonstrator in the department of geology, Queen's University and remained registered as a part-time student, with leave of absence, at Edinburgh University. I remained based in Belfast throughout the remainder of my research towards Ph.D, retaining the post of part-time demonstrator until 31 September 1980. During this time the strain analysis was extended to the S. Sithean Mhor area, and with the help of D. Sanderson in computing techniques, theoretical models to the strain data were constructed. Data were collected on rocks across the SW schist-gneiss contact, which includes mineral analyses using the electron-microprobe at Belfast.

Four field seasons ranging from four to seven weeks each were spent in the Gairloch area during which time I was mostly based at Carn Dearg Youth Hostel some five kms west of the village of Gairloch on the north shore of the loch. The field work consisted of sample collecting, large scale mapping of some areas and recording of extensive field notes in others.
CHAPTER 2. THE LOCH TOLLIE GNEISSES

2.1. Introduction

The Tollie gneisses cover an area 3.5 kms broad by 12.5 kms long (Fig.2-1) which is bounded to the NE by the Loch Maree fault, to the SW by the Aundrary amphibolite to the SE by the Flowerdale fault and to the NW by the outcrop of Torridonian sandstones. There is a similar group of gneisses known as the Buanichean gneisses (Park 1970) to the SE of the Flowerdale fault. The area has an undulating land surface with a relief of 270 m and an average altitude of 250 m, with rocks well exposed on the higher parts, typical of much of the Lewisian outcrop of the Scottish mainland. In general, the exposure is good amounting to some 30%.

The principal rock types in the area are quartzo-feldspathic, biotite and hornblende gneisses which contain feldspathic amphibolite and hornblendite pods. These gneisses have been cut by numerous dykes, now amphibolite, which show discordant contacts with the gneisses layering.

The major structure in the area is that of an antiform, known as the Tollie antiform, which affects both the gneisses and dykes. The fold axis plunges approximately 10° SE with a subvertical SW limb and moderately NE dipping NE limb. The hinge zone is broad and flat in the NE and becomes progressively sharper and more steeply plunging to the SE. This structure is bounded to the NE by Creag Mhor Tollie, an area of highly variable fabric type, intensity and orientation and is separated from it by a fault zone, see Fig.2-1.

The fabrics of the quartzo-feldspathic gneisses and dykes vary from largely SE plunging linear in the NE limb and flat belt, to more planar fabrics in the SW steep belt where lineations are variable in
FIG 2-1 Map of the Tollie area.

KEY

- Torridonian cover
- fault zone
- area of dominantly linear fabric
- trend of gneissose layering and dykes
plunge. This variation in fabric type is reflected in the geomorphology. The rocks with linear fabric give rise to a prominantly rodded landscape with steep scarp slopes facing NW and gently inclined rounded slopes facing SE, see Plate 2-1. The more flattened fabrics in the SW give the rocks a platey appearance and a smoother land surface.

The area has received attention from several workers in the past. The geological map of the area was published as Sheet 92 of the Geological Survey of Scotland one inch map series in 1907. The memoir (Clough et al, 1908) gives a description of the rock types and main structures of the Tollie gneisses (op. cit. pp. 191-252). The area around Loch Tollie between the road at 840789 and the Flowerdale fault was remapped and a more detailed description of the structure given by R.G. Park (1970). His interpretation involves four main phases of deformation with a fifth late stage phase of brittle deformation, some of which he correlates with phases of deformation in the South Sithean Mhor gneisses and the Gairloch schist belt to the SW (Park, 1964). The area N of the road at 825781 to the NW end of the LeisNew outcrop was mapped and a structural interpretation given by Bhattacharjee (1968). He invokes three main phases of deformation largely agreeing with the structural interpretation given by Park from the area to the SE.

Some age determinations have been published for the rocks of the Gairloch region by Moorbath and Park (1972) and by Park and Evans (1965). Lead isotope work by Moorbath and Park (1972) gives a date for the end of high grade metamorphism at c. 2900 Ma. K/Ar dates for dykes in the Buanichean gneisses are 1712 ± 35 Ma and 1685 ± 40 Ma. A traverse of specimens across the Aundary amphibolite immediately to the SW of the Tollie gneisses yielded a range of K/Ar dates from 1398± 35 Ma to 2213± 80 Ma. A muscovite-bearing pegmatite from the Buanichean gneisses gave an age of 1426± 35 Ma. The oldest date for the Aundary amphibolite (2213± 35 Ma) is interpreted as a date for Inverian amphibolitisation.
Plate 2-1. Looking NW across the flat belt of the Tollie antiform, towards Loch Tollie (left hand side of background).
and a close approximation to the date of intrusion, and the amphibolite is therefore associated with the Scourie dyke suite in age. The other ages for dykes are interpreted as indicating varying degrees of Laxfordian overprint on original Inverian ages, and the youngest dates, c. 1410 Ma, as due to a late episode of retrogression and metasomatism.

Some K/Ar dates were previously published by Park and Evans (1965) but are somewhat younger, giving an age of 1240±50 Ma for dykes in the Buanichean gneisses and 1400±60 Ma for the Meal Aundrathy amphibolite.

The area studied in detail here spans the broadest part of the Tollie antiform from the SW edge of the Creag Mhor Tollie mass (857787) where deformation is very inhomogeneous to the contact of the gneisses with the Aundrathy amphibolite in the SW (821774). This area was chosen since it shows the greatest range in fabric type across the hinge of the antiform. The purpose of the study was to analyse strain across the area and model the deformation path of the strain in the gneisses and dykes, as an aid to understanding the structural evolution of the area.

Throughout the following chapters the terminology used to describe the type of quartz shape fabric and/or the strain ellipsoid is that outlined by Flinn (1965). L is used to denote prolate ellipsoids and fabrics in which there is a linear but no planar element and S to denote oblate ellipsoids and fabrics in which there is a planar but no linear element. Other fabrics are denoted by combinations of these two symbols, eg. LS implies a fabric in which the planar and linear elements are equally strong.
2.2. Lithologies

The Tollie gneisses are composed of several lithological types. The gneisses themselves range from quartzo-feldspathic to biotite and hornblende gneisses and contain lens shaped bodies, here referred to as pods, of various lithological type including feldspathic, amphibolite, hornblendite, marble and muscovite gneisses. The gneisses and their included pods are cut by feldspathic amphibolite dykes. The mineral constituents of the various rock types are listed in Table 2-1.

2.2.1. The gneisses

The gneisses form the major part of the outcrop acting as a matrix to the pods and as country rock to the dykes. The gneisses appear as three main types: 1) quartzo-feldspathic, 2) biotite, and 3) hornblende gneisses, each of which has gradational contacts with the others.

Quartzo-feldspathic gneisses with minor biotite is the most abundant rock type with more biotite-rich and hornblende-rich gneiss forming included patches. These patches are elongate NW/SE parallel to the structural trend of the rocks. In all gneisses, quartz is subordinate to feldspar, comprising 35% or less of the rock and forming aggregates of grains within a feldspar matrix, most clearly seen in the quartzo-feldspathic gneisses.

The gneisses show layering which is the result of variations in grain size in the case of hornblende and biotite, or aggregate size in the case of quartz, and also by variations in the modal proportions of minerals. The width of layering is variable and is interpreted as being dependent on the deformation of the rock. In general layering width shows a complete range from 25 cms downwards. This layering wraps round the pods which are described below.
Table 2.1. Mineral composition of Lithologies.

<table>
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<tr>
<th>Lithology</th>
<th>Major Minerals</th>
<th>Minor Minerals</th>
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<tbody>
<tr>
<td>Quartzofeldspathic gneiss</td>
<td>plagioclase, k-feldspar quartz</td>
<td>biotite, hornblende</td>
</tr>
<tr>
<td>Biotite gneiss</td>
<td>plagioclase, k-feldspar, quartz, biotite + muscovite</td>
<td>opaques, epidote sphene</td>
</tr>
<tr>
<td>Hornblende gneiss</td>
<td>plagioclase, hornblende quartz</td>
<td>biotite, epidote, sphene, opaques</td>
</tr>
<tr>
<td>Amphibolite</td>
<td>hornblende, plagioclase clinozoisite</td>
<td>quartz, opaques, sphene</td>
</tr>
</tbody>
</table>

Major minerals listed in order of abundance.
2.2.2. The pods.

The gneisses contain pods of four lithological types; feldspathic amphibolite, hornblendite, muscovite gneiss, and marble.

The pods of feldspathic amphibolite are the most common and tend to be concentrated in patches where they constitute up to 20% of the outcrop. These patches are elongate NW/SE in the horizontal projection parallel to the main structural trend and in a vertical projection the concentration of pods is parallel to the gneisses layering showing that these patches are sheet-like in form. Within these sheet-like concentrations pods are themselves elongate by an amount that correlates with the strength of the quartz fabrics in the surrounding gneisses. Pods range in length from 3 m to 10 cms and some show layering, due to variations in the proportions of hornblende to feldspar. The grain size within pods is up to 1 cm but is very variable and often greater than that of the surrounding gneisses.

The second most abundant lithology forming pods is hornblendite, and these pods are found largely in association with the concentrations of feldspathic amphibolite pods. At 841773 there is a large mass of hornblendite some 40 m broad by some 600 m long. The outcrop is lens shaped and trends NW/SE parallel to the gneisses foliation and layering. Another smaller mass exists ~60 m along strike to the NW. These hornblendite masses are coarse grained, up to 1 cm, and contain schleiren of foliated quartzo-feldspathic gneiss elongate parallel to foliation in the surrounding gneisses.

Apart from these two large masses, the hornblendite pods tend to be of smaller size than the associated feldspathic amphibolite pods and in general show a coarser grain size. Hornblendite pods are most commonly found as a succession of strings of pods parallel to the foliation in the rock.
The pods of marble and muscovite gneiss are much less common than those of feldspathic amphibolite or hornblendite. Two localities of marble are reported by the Survey Memoir (1908) at 855722 and at 865731. The marble at 855722 is a fine grained ferroan dolomite (Clough et al., 1908, p. 231) and the marble at 865731 is similar but also contains abundant magnetite. There is also a large elongate mass some 30 m x 200 m of muscovite rich gneiss composed of quartz, plagioclase and muscovite with minor clinozoisite. The abundance of muscovite makes it distinct from the surrounding gneisses which contain very little or no muscovite. The contact with the surrounding gneisses is gradational over approximately 1 m.

2.2.3. The dykes

The dykes are composed of feldspathic amphibolite and vary in size from 20 m across downwards. Their original intrusive nature is shown by the crosscutting relations with the gneissose layering. The proportions of the mineral constituents are very similar throughout the area.

The dykes are well foliated with a fabric reflecting that of the surrounding gneisses but at 830767 lens-like masses, 2 m x 1 m, within a dyke show a preserved ophitic texture. The grain size of the dykes is variable and shows, like the gneisses, a decrease to the SW, from 5 x 1 mm to less than 1 mm long. There is also a range in grain size within individual dykes, many showing fine grained margins and coarser centres.
2.3. Structure

2.3.1. Introduction

The basic structure of the area is that of an antiform, whose axial trace trends NW/SE developed in both the gneissose layering and the dykes. The axial direction of the fold varies in plunge along its length being approximately 10° SE around Loch Tollie (845775) and steepening progressively to the SE to 60° SE at 878735 (Park 1970). The form of the fold also varies along the length of the axial trace from an open box shape with a flat crest in the NW where the axial direction is gently plunging, to a sharper, tighter form in the SE where the axial direction is steeper.

The area chosen for study, immediately to the SE of Loch Tollie, covers a traverse across the fold from NE to SW (858782 to 828764) where the crest is broad and the axial direction gently plunging. Here the greatest variation in fabric types and minor structures is found. In this area, the fold has a moderately dipping NE limb, ~40°NE, a broad subhorizontal crest, "the flat belt" just over 1 km wide and a subvertical SW limb, see Section in Fig 2-2. The fabrics shown by the quartzo-feldspathic gneisses vary from LS with a moderate (~40°) SE plunging lineation in the NE limb to good L fabrics with a gently SE plunging lineation in the flat belt. In the steep SW limb the fabrics go through the sequence LS—S—LS from NE to SW, with lineations ranging from ~10° SE to ~50° NW, through the horizontal.

For the purposes of description the fold has been divided into two parts:

a. the NE limb and flat belt,

b. the SW steep limb.
FIG 2-2 Change in style of folding across the Tollie Antiform

SKETCH SECTION OF THE TOLLIE ANTIFORM SHOWING LOCATION OF FOLD STYLES (a) TO (f)
2.3.2. The NE limb and Flat belt of the Antiform.

a. The Gneisses

The moderately dipping (~40°) NE limb grades into the flat belt by a shallowing of the dip of gneissose layering, a change in fabric type from LS to L and a decrease in plunge in lineation from approximately 40°SE to subhorizontal. These changes take place over a distance across strike of 150 m around 851782.

On the section chosen for study, true LS fabrics are not found on the NE limb due to a fault zone which cuts across the gneisses with L fabric. This fault brings the Creag Mhor Tollie mass, marked by inhomogeneous deformation against the Tollie antiform. However, LS fabrics do occur some 2.5 kms to the SE at 863762 where the gneissose layering dips more steeply ~70° NE. Here, the small folds are similar in style and generally near isoclinal see Fig 2-2(a). The hinges of small folds are parallel to the quartz fabric lineation, and their axial planes parallel to the quartz fabric foliation.

Along strike to the NW, the dip of gneissose layering becomes shallower, ~40° NE, and the quartz fabric becomes more dominantly linear, with a gently SE plunging lineation. Here the dip of the gneissose layering is variable from subhorizontal to 40° NE. The quartz fabric foliation is less variable and dips 60-70° NE, so that in many places this foliation cuts across the gneissose layering. Where this is the case, the quartz fabric lineation is difficult to identify as foliation planes are weak and seldom clearly displayed. Thus the dominant lineation seen here is an intersection lineation between gneissose layering and quartz fabric. This intersection lineation is subhorizontal whereas the quartz fabric lineation plunges 20-30° SE. In this area the planar element of the quartz fabric is sporadic and interspersed with areas of good L fabrics. Folds in
this area range from isoclinal to open with axial planes ranging in
dip from subhorizontal to moderate SW, Fig 2-2(b), Plate 2-2. The tighter
folds are usually similar in style with thickened hinges and thinned
limbs while open folds have more concentric style. The fold hinges
are parallel to the quartz fabric lineation but the axial planes are
generally oblique to quartz fabric foliation.

To the SW, L fabrics with a subhorizontal lineation become
dominant and LS fabrics are seldom seen. Folds become more open with
a greater range in orientation of axial planes from approximately
45° NE through vertical to 45° SW. The majority, however, are subvertical
and trend NW/SE. The folds here are closer to concentric in form,
box-like shapes are common, in contrast to the similar style in the NE,
see Fig 2-2(c). Here the wavelength of folds is dependent on the layer
thickness with thinner layers showing smaller wavelengths, see Plate 2-3.
The folds do not show any development of axial planar foliation but
have good L fabrics throughout, though in a few places a planar element
is seen in hornblendic layers parallel to and folded with the layering.
Approaching the crest of the antiform folds become gradually more open
with interlimb angles commonly greater than 100° and often consisting
of no more than a warping of the gneissose layering, Fig 2-2(a).

Throughout the NE limb and the flat belt of the antiform small
shear zones are developed. These vary in size from 1 to 30 m wide
trending NW/SE and can be traced along strike for several hundred metres.
The gneissose layering has been bent into these zones from subhorizontal
to become steeply dipping (60-80°) NE or SW. A planar element to the
fabric is developed in the quartzo-feldspathic gneisses and the foliation
dip steepens progressively towards the centre of the shear zone up to
85°, see Fig 2-3. In the outer margins of the shear zones the quartz
fabric foliation is oblique to the gneissose layering but becomes
Plate 2-2. Folding in the NE belt at 853784. Gneissose layering dips moderately NE. Vertical face, looking SE.

Plate 2-3. Detail of folding in the flat belt at 853784, showing the relationship between wavelength of folding and layer thickness. Vertical face, looking SE.
FIG 2-3 Examples of Small Shear Zones

a) SKETCH SECTION OF SHEAR ZONE AT 853785.

b) SKETCH SECTION OF SHEAR ZONE AT 851775.
parallel or subparallel to layering in the centre of these zones.
Lineations are constant throughout the shear zones and the surrounding gneisses - approximately 10° SE.

The shear direction as indicated by the bending of the gneissose layering and development of quartz fabrics is always subvertical. The shear sense is variable but mostly NE down, SW side up. Five out of the six examples recorded showed this sense of shear and here quartz fabrics and gneissose layering in the centres of the zones dip 60-80° NE.

At the crest of the antiform itself both gneissose layering and dykes are bent sharply over to dip steeply SW. The transition zone is narrow ~50 m, so that the hinge itself is sharp and the antiform markedly asymmetric. The quartz fabrics in the hinge zone are linear and a weak planar fabric is developed dipping 80° SW as soon as the dip of the gneissose layering becomes overall SW, see Plate 2-4. The rocks maintain a strong lineation dipping gently (~10°) SE, throughout this zone. To the SW the planar fabrics become stronger and the steep SW limb of the antiform is entered. Across the hinge zone the fold style changes from the open box-shaped style of the flat belt to a tighter style with sharper hinges and straighter limbs, see Fig 2-2(e). The hinges of the folds are parallel to the quartz fabric lineation throughout.

b. The Dykes.

Dykes are abundant throughout the area and in many instances show crosscutting relations with the gneissose layering.

In the NE limb of the antiform, the dykes trend NW/SE and have dips ranging from ~10° SE to subvertical. For example, the mapping of one particular dyke shows that its contact relations with the surrounding gneissis are alternately parallel to and less steeply dipping than the gneissose layering. Overall this dyke dips less steeply than the
Plate 2-4. Quartz fabric (a) cross-cutting gneissose layering (b) at 843775, immediately SW of the crest of the Tollie antiform. Vertical face, looking SE.

Plate 2-5. Amphibolite dyke (D) with xenolith of gneiss, cross-cutting gneissose layering at 844777, flat belt of the Tollie antiform. Vertical face looking SE.
gneissose layering, see Fig 2-4. The dyke has LS fabric throughout, the
planar component of which is parallel to the dyke margins. The lineations
are constant in orientation throughout and similar to those in the immediately
adjacent gneisses, ie. 10-20° SE.

In type and orientation, the fabrics and lineation in the dykes
mimic those of the immediately adjacent gneisses. Throughout the NE
limb, therefore, fabrics within the dykes are rather variable but tend
to be dominantly linear, with discontinuous zones of LS fabric dipping
moderately NE.

To the SW dykes become progressively more shallow dipping until
they are subhorizontal. The fabric within the dykes becomes linear
with a gently (∼10°) SE plunging lineation, corresponding to the fabric
change within the adjacent gneisses. Dykes are observed to cut sharply
across the gneissose layering in places, see Plate 2-5. Folds are developed
in the gneisses adjacent to dyke contacts, bringing the gneissose
layering into parallelism with the dyke contact, see Fig 2-5. The folds
tend to be discontinuous, dying out along the dyke contact. Dyke
margins generally cut across folds in the gneisses but in some cases
one or both of the dyke margins has been folded with the gneissose
layering.

Dykes can be traced over the crest of the antiform and into the SW
steep limb. They show considerable decrease in thickness from the crest
towards the SW steep belt, one dyke in particular showing a change
from 20 m in the hinge zone to 8 m in the SW limb. The dykes show good L
fabric in the hinge zone itself but acquire LS fabric as soon as the dip becomes
moderate SW.

2.3.3. The SW Steep Belt

As described in the previous section, at the crest of the antiform
the gneissose layering and dykes are sharply turned over to form the
steeply dipping SW limb of the fold. This limb is broadest in the
FIG 2-4 Relation of Dykes to Gneissose Layering

a) SIMPLIFIED PLAN VIEW OF DYKE AT 868764.

b) INFERRED CROSS SECTION

- dyke dips less steeply than gneissose layering

- gneissose layering

- dyke

50 m
FIG 2-5 Fold in gneiss adjacent to dyke contact in the flat belt at 844775.

Linear fabric in dyke and gneiss plunges 6° SE.
NW, 1.7 kms, and narrows to the SE to become only 0.4 km at 873702.

A portion of the steep belt 0.4 km along strike by 1.4 kms across strike has been mapped in detail at 839773, to illustrate the variation in foliation and lineation in gneisses and dykes across strike, see Fig 2-6. Only dykes greater than 2 m width are represented on the map.
a. The gneisses

The gneissose layering varies in dip from SW near the crest of the antiform through vertical to steep NE in the SW adjacent to the Meall Aundrary amphibolite, see section in Fig 2-1. Quartz fabric foliations are parallel to the gneissose layering except in the NE close to the crest of the antiform as described in the previous section.

The quartz fabric type is variable throughout the SW limb. Near the crest of the antiform it is LS and becomes progressively stronger to the SW grading from LS through S to LS again near the Meall Aundrary amphibolite. The progressive increase in intensity of quartz fabric to the SW gives the gneisses a marked platey appearance.

Lineations are strongly developed in the SW limb and are variable in orientation from gently plunging (∼10°) in the NE through horizontal to moderately plunging (∼50°) NW in the SW. This variation in plunge of lineation across strike is illustrated in Fig 2-7. In the NE and SW the plunge of lineation is gentle SE and moderate NW respectively, with a relatively small range of values. In between these two areas there is a steady gradation in plunge and a much wider scatter of values. Lineations in the dykes are similar to those in the gneisses.

The lineations are parallel to small fold hinges where these are recognisable and, from sections observed parallel and perpendicular to the lineation, they are themselves composed of small fold hinges in individual quartz lenses, see Fig 2-8. These hinges weather out on foliation surfaces giving rise to the marked lineation observed in
FIG 2-6: Detailed Map of a Part of the Tollie Antiform

KEY

LITHOLOGIES

- GNEISS
- DYKES
- HORBLENOITE

SYMBOLS

- Foliation: dip and direction of dip in degrees
- Lineation: dip and direction of dip in degrees
- Hinge trace of the Tollie Antiform
- Lochans and streams

SCALE: 1:50,000

200 m.
FIG 2-7 Variation in plunge of lineation across the SW steep limb

- x: lineation plunge in gneisses
- O: lineation plunge in dykes
FIG 2-8 3D view of block showing form of quartz lineation.

Plane perpendicular to foliation and parallel to lineation, quartz foliae uncrenulated.

Plane perpendicular to foliation and lineation showing crenulation in quartz foliae.

Foliation surface showing lineation formed by crenulations in quartz foliae.
the field, see Plate 2-6.

Minor folds are rarely seen, due to the intensity of deformation but where recognisable are isoclinal and similar in style with sharp hinges, see Plate 2-7.

b. The pods.

The gneisses contain elongate zones of pods of amphibolite and hornblendite ranging in size from a few cms to 3 m long. They are always elongate in the horizontal plane parallel to strike of the foliation and have axial ratios of up to 100, see Plate 2-8. Extremely elongate pods tend to have ragged rather indistinct ends, especially pods of feldspathic amphibolite. The coarser grained hornblendite pods show a tendency to boudinage and all forms from pinch and swell to separate boudins where the spaces between the boudins have been filled with vein quartz are seen, see Plate 2-9. Some pods have an internal foliation oblique to the foliation in the surrounding gneisses, though the pod shape is always elongate parallel to the foliation in the gneisses. A series of boudinaged pods was observed on both subhorizontal and subvertical sections. Boudins appear on both surfaces indicating that the three dimensional form is that of a series of individual sacks, 'chocolate tablet structure' (Wegmann, 1932, Ramsay, 1967, p. 113), see Fig. 2-10.

c. The dykes

Dykes are numerous in the SW steep belt and in general are well exposed. They range in thickness from 13 m down to a few cms, with the exception of one very thick dyke, 30 m wide, close to the crest of the antiform. Dykes are parallel or subparallel to foliation in the gneisses and in a few cases discordances of up to approximately 10° can be seen, see Plate 2-10.

Measurement of dyke thicknesses across strike, from 835768 to 838771, a distance of 450 m across strike shows that the majority
Plate 2-6. Foliation surface in quartzo-feldspathic gneisses showing prominent rodding lineation formed by quartz at 838771, SW steep limb of the Tollie antiform. Vertical surface, looking NE.
Plate 2-7. Isoclinal folds in gneissose layering at 838771, SW steep limb of the Tollie antiform. Subvertical surface, looking SE.
Plate 2-8. Elongate amphibolite pods in highly foliated quartzofeldspathic gneisses at 839773, SW steep limb of the Tollie antiform. Horizontal erosion surface, looking NW.
Plate 2-9. Boudinaged hornblendite pod (h) with quartz filled necks (q), at 837771. Horizontal surface.
FIG 2-9 Histogram of dyke thicknesses in the SW steep belt measured across strike from 835768 to 838771.
FIG 2-10 Boudinage in hornblendite pod.

a) 3-D view of a boudinaged hornblendite pod in gneiss.

b) "Chocolate tablet" structure shown by pod in a.)
plate 2-10. Dyke margin cross-cutting gneissose layering at a shallow angle, 837772, SW steep limb of the Tollie antiform. Horizontal surface, looking NW.
of dykes are 0.75 m wide or less, with a range from 13 cms to 4.65 m, see Fig 2-9. Dykes also vary in grain size of hornblende and feldspar. In general, thinner dykes (∼2 m) have finer grain size (∼1 mm) of both hornblende and feldspar than thicker dykes (∼2 m) which are coarse grained (∼1 mm). Over this area the 30 dykes measured total 8.2% of the outcrop width. All dykes are foliated and lineated, lineations formed by the parallel alignment of hornblende crystals and elongate aggregates of feldspar. One thick dyke at 830767 is considerably less deformed than most. Deformation within this dyke is very inhomogeneous and in the least deformed areas the amphibolite retains its original igneous texture.

Spatial occurrence of the dykes is inhomogeneous throughout the gneisses. They tend to occur in groups covering 50-100 m of outcrop across strike where they make up 50% of the outcrop. Between these groups the gneisses have few or no dykes. Individual dykes vary in thickness and many were observed to die out along strike by the inclusion of schlierens of gneiss on average 5-10 cms wide but also up to 20 cms wide, parallel to the foliation within the dyke. Along strike, over a distance of 20 to 50 m, the proportion of schlierens increases until the dyke is no longer distinguishable.

Dykes also show pinch and swell features and occasionally complete separation into boudins, see Fig 2-11. Pinch and swell is seen most commonly in dykes of 2 m or less, see Plate 2-11, though occasionally occurs in thicker dykes. True boudins were only seen in dykes of less than 1 m width. The distances between necks is highly variable within individual dykes from 2 m upwards as is the breadth of the necks themselves which range from just under dyke width to zero, ie total separation. The foliation in both gneisses and dykes is bent round the pinch and swell features to be parallel to the dyke margins and in some cases a change in dip as well as strike is observed indicating that their shape
FIG 2-11 Sketches of pinch and swell and boudinage in dykes of the SW steep limb.

a) 835770

b) 835772

c) 835770
Plate 2-11. Pinch and swell in dyke with associated deflection of foliation in dyke and adjacent gneiss, at 834770, SW steep limb of the Tollie antiform. Arrows point to dyke contacts. Subhorizontal surface, looking NW.
is lens-like in three dimensions. In one or two cases there is an offset across strike between swellings. Where necks are very thin, ie. 10 cms or less, the region of the neck is often partially occupied by vien quartz. In the case where individual boudins have formed the foliation in the surrounding gneisses is intricately folded within the space.

d. The quartz veins.

The gneisses contain quartz veins which are especially abundant around 8277 in the SW of the area. Veins occur as folded, see Plate 2-12, boudinaged and straight while a few show evidence of both folding and boudinage, the long limbs of folds being boudinaged. Veins range in thickness from ~1 mm to 20 cms and occasionally enechelon lenses of quartz occur.

The trend on subhorizontal surfaces of quartz veins was recorded for 201 veins together with a note of the state of deformation, ie straight, folded, or boudinaged, for six localities from 822777 to 826771, along the strike of the gneisses. The data has been plotted in Rose diagram form in Fig 2-12. This shows that folded, boudinaged and straight veins occupy distinct sectors of the 360° field. For an interpretation of these patterns see the discussion in section 2.4.
Plate 2-12. Folded quartz veins and quartz lenses at 822776, SW steep limb of the Tollie antiform. Subhorizontal surface.
2.4. Discussion

2.4.1. The origins and early history of the Tollie gneisses.

Prior to dyke intrusion the rocks were hornblende, biotite and quartzo-feldspathic gneisses containing pods of varying lithologies (see section 2.32.). The pods occur in sheet like concentrations and their appearance in relatively undeformed sections perpendicular to the lineation in the flat belt, shows numerous blocks separated by veins of quartzo-feldspathic material. It is probable, therefore, that these concentrations represent layers of feldspathic amphibolite which have been broken up and invaded by the quartzo-feldspathic gneisses, implying an igneous or anatectic origin for the quartzo-feldspathic gneisses. The original layers of amphibolite may have been dykes or basic layers in a pre-existing terrain. The gneisses immediately surrounding the amphibolite layers often have 2-3% of hornblende, probably due to ingestion of parts of the amphibolite layer.

The pods are, therefore, the earliest recognisable feature in the Tollie gneisses. Other pods of hornblendite, marble and muscovite gneiss also represent parts of this earlier terrain, but the relationship between the various lithologies is unknown.

The dykes were intruded into layered gneisses containing the above pods. Their shape is step-like ie alternately parallel and crosscutting the gneissose layering. The gneissose layering thus seems to have been a controlling factor of the shape of dyke intrusion. The mode of intrusion of the dykes is seen in the SW steep belt (see section 2.33) where dykes can be followed along strike for several hundreds of metres. The process consists of numerous apophyses at the dyke end coalescing and gradually surrounding blocks of gneiss. Since examples of dyke endings were only seen in highly deformed rocks it is uncertain whether the
gneisses had any tectonic fabric at the time of intrusion. For further discussion see section 2.4.4.

2.4.2. Evidence for deformation type.

All rocks of the Tollie gneisses possess a tectonic fabric and are therefore deformed. The type of this deformation can be inferred from two lines of evidence;

a. the existence of small shear zones throughout the NE limb and the flat belt and,

b. the trend of deformed quartz veins in the SW limb.

a. Small steeply dipping shear zones trending NW/SE occur throughout the NE limb and the flat belt (see section 2.3.2.). The evidence that they are shear zones is shown by the localised and linear nature of the strain zone.

These minor shear zones have a similar trend to the major structure in the area; the Tollie antiform. Thus it is possible that the Tollie antiform was also formed by simple shear, for which the foliation and lineations in the SW of the area indicate a steep NE dipping shear plane and a moderately NW plunging shear direction. The small shear zones have variable sense of shear but are mostly SW side up NE side down, i.e. opposite to the sense implied by the Tollie antiform.

The time relations between the formation of the Tollie antiform and the minor shear zones is uncertain. However, the late stage deformation of the area is brittle in style characterised by small faults, fractures zones and pseudotachylite whereas in the small shear zones deformation is ductile as is the deformation in the steep SW limb of the Tollie antiform. Also the small shear zones are superimposed on linear fabrics and the folding of the gneissose layering, as is the SW steep limb of the Tollie antiform. Thus it is likely that the small shear zones are associated in time with the deformation in the SW steep
limb and it is possible that they have a conjugate relationship with the main shear.

b. As described in section 2.3.3.(d), the orientation and state of deformation (straight, folded, boudinaged) of 201 quartz veins on subhorizontal erosion surfaces were measured in the SW steep limb of the antiform. These have been plotted in Fig 2-12 as Rose diagrams. Straight, folded and boudinaged veins occupy well defined sectors of the diagram. Boudinaged veins are confined to a smaller sector than folded or straight veins and all sectors overlap. The straight veins have been divided into two categories; a) long thin veins with a length/breadth ratio greater than 15 and b) short fat veins with length/breadth ratio less than 15. Category a), long thin veins, occupy a single well defined field whereas the short fat veins, category b) occupy two distinct fields, see Fig 2-12(b).

In two dimensional progressive deformation since the infinitesimal and finite strain ellipses differ in axial ratio and, in the case of rotational strain, orientation, superposition of the two sets of lines of no longitudinal strain divides the strain ellipse into zones with different deformation histories (Ramsay, 1967, p. 114). Thus there are zones of the strain ellipse in which lines have changed their behaviour from contraction to expansion and vice versa.

The position and extent of these zones depends on the magnitude and type of strain involved. For pure shear, the zones in which lines have been contracting but are now expanding are symmetrically arranged about the major axis. In simple shear because the strain is rotational there is only one such zone, see Fig 2-13. These markedly different patterns can be used to distinguish between the two types of deformation (Ramsay, 1967, p. 114-120).

The quartz veins record the history of deformation, folding when in
FIG 2-12 Rose diagrams of quartz vien orientations

a) FOLDED AND BOUDINAGED QUARTZ VIENS

Total number of folded viens 75
Total number of boudinaged viens 58

- Folded viens
- Boudinaged viens

b) STRAIGHT QUARTZ VIENS

Total number of long thin viens 12
Total number of short fat viens 12

- Long thin viens (length to breadth ratio > 15)
- Short fat viens (length to breadth < 15)
FIG 2-13 Zonal arrangement within a strain ellipse developed by progressive deformation (based on Ramsay (1967) figs 3-62 and 3-64)

a) Irrotational strain - pure shear

b) Rotational strain - simple shear

Zone 1a  Boudinage only
Zone 1b  Boudinage in unfolded bands or boudinaged folds
Zone 2  Folds being boudinaged
Zone 3  Folds only

A, B  Principal axes of the strain ellipse
f.l.s.  line of no finite longitudinal strain
i.l.s.  line of no infinite longitudinal strain
the zone of shortening and suffering boudinage in the zone of extension. Veins which have been initially shortened but laterly extended show folds with boudinaged limbs. A plot of the various sorts of quartz vein will thus show the above zones of the strain ellipse.

From Fig 2-12(a) there is a marked overlap of the fields of boudinaged and folded veins from 310 to 360° (and therefore from 130° to 180°) but very little overlap on the other side of the zone of folded veins at around 010°. This pattern indicates simple shear as the type of deformation and from the orientation of the overlap the sense of shear projected onto the horizontal plane is dextral. Assuming that the shear direction is steeply dipping NW from the plunge of lineations in the area, this implies a shear sense that is SW side down, in agreement with the sense required to produce the Tollie antiform.

As described previously in this section, undeformed, ie. straight, veins occur as long thin parallel sided ribbons and as short lenses. The shorter lens-like veins are usually of a simple lens shape but a few show sigmoidal shapes. Occasionally they occur as en echelon rows of 3 or 4 lenses and here the rows are parallel or subparallel to the gneissose layering. Since they show little or no deformation they must have been formed at a late stage in the deformation. The orientations of these two categories of quartz vein have been plotted in Fig 2-12(b); the lenses occur in two main orientations. The main orientation is around 036° and this, together with the fact that veins occur in en echelon rows and show sigmoidal shapes, indicates that these probably originated as tension gashes. Their orientation agrees with the sense of shear indicated by the overlap in orientation of folded and boudinaged veins and is thus associated with the deformation in the steep SW limb.

A second less common orientation of lenses occurs at 154°. These veins also occur in en echelon rows subparallel to the foliation. A
possible origin for these veins is as fractures in shear zones conjugate to the main set. However, in this case they should occur in arrays orientated close to 036°, i.e. subparallel to the first set of lenses (Beach, 1975), when in fact they are subparallel to the foliation, orientation 315°. Another possible origin is that they represent an earlier phase of quartz filled en echelon fractures possibly of the same age as the folded and boudinaged veins that have been rotated through some 110° to become subparallel to the foliation. This implies two phases of quartz veins one early in the deformation and one at a late stage in the deformation history.

The straight veins occur subparallel to the foliation at an average orientation of 165°. These have clearly exploited the foliation plane of the gneisses as a plane of weakness. A possible origin for these veins is that they represent the conjugate set to the major orientation of lenses (036°), but because fractures occurred parallel to the foliation in the gneisses, they developed as elongate parallel sided veins rather than as short lenses.

2.4.3. Comparison of deformation styles in the dykes and gneisses.

There is evidence throughout the Tollie gneisses that the dykes have different rheological properties to the surrounding gneisses during deformation.

In the SW steep limb of the antiform pinch and swell features in dykes are common and occasionally true boudins are developed (see section 233c). The three dimensional shape of these boudins is of individual sacks or what is known as 'chocolate tablet structure' (Wegmann, 1932, Ramsay, 1967, p. 113). These features indicate that the dykes have behaved as relatively competent layers within the gneisses. Evidence for this is also seen in the flat belt where dykes have LS fabrics developed at their margins implying shear along the dyke margins and thus
a competence difference between the dykes and gneisses.

The chocolate tablet structure shown by the dykes and hornblendite pods in the SW steep belt implies that there is extension in all directions, ie. an increase in area within this plane, the plane of the foliation (Ramsay, 1967, p. 112). This also explains the common occurrence of quartz masses between boudins seen especially in boudinaged hornblendite pods, as an increase in area within this plane will lead to the development of voids between boudins in which quartz can be deposited.

The fact that the plane of the foliation is a plane of extension in all directions has implications for the type of strain ellipsoid possible in these rocks. Since $\lambda_1 > \lambda_2 > 1$ within the plane of foliation, the ellipsoid must be of 'pancake' type, ie $1 > K > 0$. $K = \frac{a}{b}$ where $a = \frac{x}{y}$ and $b = \frac{y}{z}$ (Flinn, 1965) and $x$ is the maximum, $y$ the intermediate and $z$ the minimum axes of the strain ellipsoid.

2.4.4. Time relations between deformation and dyke intrusion.

The structures and lineations shown by the dykes have implications for the time relations between dyke intrusion and deformation. Throughout the area the dykes mimic the fabric type and orientation in the immediately surrounding gneisses. The quartz fabrics in the gneisses record the total finite strain and since the fabrics in the dykes are similar to them, the same must be true for the dykes. In the area of lowest deformation, the NE limb, where discrepancies between the state of deformation in the dykes and gneisses would be most easily detected, all dykes have a tectonic fabric similar to the adjacent gneisses. Thus the intrusion of the dykes must precede the deformation.

One dyke in the SW steep belt is less highly deformed and preserves relict ophitic texture (2.2.3.). This dyke is relatively thick (15 m) compared to the majority of dykes in the area (average
equals 0.75m). Deformation within this dyke is inhomogeneous and since dykes are more competent than the gneisses, lenses with relict ophitic texture more probably represent areas that have escaped deformation than inferring a late intrusion age for the dyke.

In the SW limb where deformation is strong, some dykes contain highly elongate and strongly foliated schlieren of gneiss parallel to dyke margins (section 2.3.3.c.). Such numerous gneissose inclusions are not observed in the dykes of the NE limb or the flat belt where planar fabric are less strongly developed and this, together with the regularity of gneissose schlieren in these dykes, suggests that the gneisses were foliated at the time of intrusion. Thus some of the dykes in the SW limb may be syntectonic and therefore later than those in the flat belt or NE limb.

Thus field evidence suggests that the bulk of dyke intrusion occurred prior to the deformation but that it is possible that some dykes in the SW limb are syntectonic.
2.5. Summary

From the evidence of small shear zones and the orientations of traces of quartz veins, the probable deformation type is dominantly simple shear. The major structure of the area, the Tollie antiform, implies a shear direction that is steeply plunging NW and a shear sense that is SW side down, NE side up, which is in agreement with that implied by the trace of quartz veins. The small shear zones in the NE limb and flat belt and perhaps some of the undeformed quartz veins in the SW steep limb imply a sense that is opposite to this, i.e. SW side up and may represent the weakly developed conjugate shear to the main deformation.

The dykes show evidence that they have behaved as relatively competent bodies during deformation and the three dimensional pattern of pinch and swell or boudinage features shown by dykes and hornblendite pods implies a strain ellipsoid of the type $1 > k > 0$ for the central part of the SW steep belt.

Since dykes mimic the fabrics and lineation plunges in the gneisses throughout the area most dyke intrusion is believed to be pre-tectonic though some dykes in the SW steep belt show evidence of syntectonic age.
CHAPTER 3 THE SOUTH SITHEAN MHOR GNEISSES

3.1. Introduction

The South Sithean Mhor gneisses occupy an area trending NW/SE some 14 kms long by 7 kms at widest, bounded to the SE, SW and NW by Torridonian rocks and to the NE by the outcrop of the Gairloch schist belt, see location map, Fig 3-1. These gneisses compose the most southwestern part of the main Lewisian outcrop in the area and can be traced south-westwards to the Lewisian outcrop at Torridon via a series of small inliers in the area of the Sheildaig and Flowerdale deer forests. The area is largely composed of layered quartzo-feldspathic gneisses which are cut by numerous amphibolite dykes.

The area was first mapped by Clough et al in 1907 and published as Sheet 92 by the Geological Survey for Scotland. Part of the area was mapped by R.G. Park on the scale of 1:10,560 in 1964, and part by T.S. Ghaly on the scale of 1:2390 in 1966. For the purposes of this study a small area to the SW of S. Sithean Mhor was chosen at 8070, see Fig 3-1 and mapped in detail on the scale of approximately 1:3520 with the use of ariel photographs, see Fig 3-2. The area is largely well exposed, up to 30% on high ground.

3.2. The Lithologies

The area consists of quartzo-feldspathic gneisses containing minor amounts of hornblende-bearing gneisses, amphibolite dykes and a variety of pods.

3.2.1. Quartzo-feldspathic gneisses.

These gneisses are composed of quartz, plagioclase and k-feldspar. The percentage of feldspar exceeds that of quartz, on average 60% feldspar to 40% quartz, quartz forming separate aggregates within the
FIG 3-1. Map of the S. Sithean Mhor area.

Torridonian cover

Sithean Mhor belt - NW trending gneisses.

Ruadh Mheallan belt - NE trending gneisses.

after Park (1973).
FIG 3-2 Map of part of the S. Sithean Mhor gneisses at 8070.

KEY

- Quartz-feldspathic
- Amphibolite dyke

General dip of gneissose layering.
Dip of foliation.
Dip of quartz fabric and gneissose layering.
Lineation.

- Isotropic fabric
- Foliated fabric
- Amphibolite dyke

Lithological boundaries.
Exposed ± 2 m.
Uncertain.

Streams and lochan shores.

Scale: 20 m.
feldspar matrix. The quartz aggregates give an indication of the intensity of deformation having an overall isotropic fabric in the undeformed state and a range of intensity of parallel alignment and elongation in the deformed state. K-feldspar occurs as large porphyroblasts up to 11 x 6 cms with euhedral shape in undeformed gneiss and forming augen in deformed gneiss, see Plate 3-1.

The gneisses are composed of layers caused by a variation in grain size and/or proportions of quartz to feldspar. The size of individual quartz aggregates ranges from approximately 1 mm to 1 cm across and individual layers from approximately 2 cms to 0.5 m. The gneissose layering is in places, cut across by irregularly shaped patches of a coarse grained homogeneous migmatitic quartzo-feldspathic material up to several metres across. The margins are highly complicated with 'fingers' subparallel to the layering, see Fig. 3-3.

3.2.2. The Pods.

The gneisses locally include abundant pods largely of a layered amphibolite, but also of quartzo-feldspathic gneiss, unlayered amphibolite, and marble. Only two clear examples of quartzo-feldspathic gneiss were found, forming well rounded augen, elongate parallel to the gneissose layering. Layering shown by the pod is sharply cut across by that in the surrounding gneiss, see Plate 3-2.

The pods of layered amphibolite ranging from 10 cms to 1.5 m across show isotropic fabric in the undeformed state. The layering formed by varying proportions of hornblende and feldspar ranges from 1 to 10 cms thick and tends to be parallel to pod length.

These pods occur in elongate aggregates where they make up to 50% of the rock, see Fig 3-4. In general the pods are elongate with their lengths and the trend of aggregates parallel to gneissose layering. Aggregates are 1 to 2 m broad and can be traced for up to 50 m along strike.
Plate 3-1. Large k-feldspar porphyroblast in quartzo-feldspathic gneisses at 804701, S. Sithean Mhor.

Plate 3-2. Pod of quartzo-feldspathic gneiss in quartzo-feldspathic gneisses, at 807702, area of inhomogeneous deformation, S. Sithean Mhor. Subvertical surface, looking NE.
FIG 3-3 Migmatite in layered quartzofeldspathic gneisses at 804703.
FIG 3-4 Sketches of sheet-like aggregates of pods at 806701.

a) sub-horizontal surface.

b) sub-vertical surface.
The majority of pods are of the layered amphibolite type but a few pods of a homogeneous amphibolite resembling undeformed dyke in texture and composition also occur. Aggregates of pods are only sporadically developed and make only a few percent of the outcrop.

In the NE, the gneisses contain pods of marble which constitute a string of outcrops along the strike of the gneisses from 812724 to 835703 as well as two occurrences of quartz-magnetite rock and associated quartz rocks. The marbles are coarse grained and contain phlogopite and abundant foliae of actinolite. They are associated with very coarse grained amphibolites of grain size up to 1 cm. The contact relations of the marbles are very poorly exposed but at 811723, marble is sheathed in a cataclastic chlorite- and hematite-rich schist. It is probable, therefore, that the contacts with the gneisses are tectonically modified.

3.2.3. Quartz Veins.

Quartz veins are not common in the SW of the area but in the NW where deformation is more intense, they are locally very abundant, eg. at 810715. Veins are subparallel to the quartz fabric and layering of the gneisses and have a braided appearance of many intersecting and discontinuous ribbons, from less than 1 mm to 10 cms thick, and in total can account for approximately 30% of the rock, see Plate 3-3.

3.2.4. The Dykes.

The amphibolite dykes are composed of hornblende, plagioclase, + quartz, + garnet. They are normally homogeneous in composition and texture though occasional feldspar rich varieties with up to 30% plagioclase, occur. Grain size ranges from 1.5 mm in the dyke centres to less than 1 mm at dyke margins, and contacts with adjacent gneiss are sharp. In the undeformed state dykes retain their subophitic texture which is progressively destroyed by deformation as foliation develops and the grain size decreases. Garnets are locally very
Plate 3-3. Abundant quartz veins trending NW/SE subparallel to foliation at 810715, area of homogeneous deformation, S. Sithean Mhor. Horizontal surface.

Plate 3-4. Minor shear zone in quartzo-feldspathic gneisses at 805704, area of inhomogeneous deformation, S. Sithean Mhor. Horizontal surface.
abundant forming up to 50% of the rock but their occurrence is confined to foliated amphibolite and they are never seen in amphibolite with subophitic texture. Garnets range from less than 1 mm up to 1 cm, on average.4 mm and are euhedral to subhedral in shape. The garnets tend to be concentrated along joints, most commonly found at dyke margins, but are sporadically developed throughout dykes.
3.3. Structure

3.3.1. Introduction.

Within the South Sithean Mhor gneisses there is an increase in deformation recorded by the quartz fabrics in the quartzo-feldspathic gneisses and intensity of foliation in the dykes from SW to NE. In the SW isotropic fabrics are commonly seen in the gneisses and dykes and deformation is very inhomogeneous. Small shear zones are commonly developed at dyke margins. Elsewhere, quartz fabrics in the gneisses where developed are weak and crosscut the gneissose layering. To the NE quartz fabrics increase in intensity as layering is folded and rotated to become parallel to gneissose layering and deformation becomes more homogeneous. The transition is a gradational one but a division, trending NW/SE through the small lochan at 810702, see Fig 3-1, may be made between the area of largely inhomogeneous deformation from that of homogeneous deformation. The area chosen for large scale mapping spans this division (Fig. 3-2) and so shows the transition between the areas of inhomogeneous and homogeneous deformation.

3.3.2. Area of inhomogeneous deformation.

Isotropic fabrics are common in both the quartzo-feldspathic gneisses and the dykes. Quartz fabrics are rather sporadically developed and crosscut the gneissose layering. Small scale shear zones occur in the gneisses and dykes and are especially prominently developed along dyke margins.

The dykes in the area are variable in shape, some showing constant thickness along their length, and others showing pinch and swell features with a tendency to lens out, see Fig 3-2. The contacts are largely smooth but occasionally have angular irregularities in which the dyke projects into the gneisses, see Fig 3-5. The dykes also include angular blocks of gneiss in places some of which are misoriented relative
FIG 3-5 Sketch of dyke margin at 804703.

\[\text{Layered quartzo-feldspathic gneisses.}\]

dyke showing relict ophitic texture.

\[10 \text{ m.}\]

FIG 3-7 Sketch of minor shear zone in dyke at 804702.

\[\text{Vertical section.}\]
to the adjacent gneisses. In general the dyke contacts are discordant to the gneissose layering.

Where the quartz fabrics are absent or weak the gneissose layering is rather variable in orientation but generally dips approximately 60° NW, see Fig 3-6. The gneissose layering is gently folded, inter-limb angles being generally greater than 90°. The folding is very irregular with rounded hinges dipping generally moderately NW. The layering is wrapped around amphibolite pods where these are present and more intense folding is often developed in the vicinity of the pods. The quartz fabrics weakly developed in these gneisses are variable in orientation but mostly dip ~ 80° NE.

Most of the deformation in this area is concentrated in small shear zones in both the dykes and gneisses. Shear zones within the gneisses are generally small usually of the order of 10 to 20 cms wide and show a displacement of gneissose layering of around 10 cms, see Plate 3-4. Quartz fabrics are developed within these zones at an angle to shear zone margins which are sharp. Most minor shear zones dip subvertically and trend NW/SE but a few examples of shear zones dipping moderately N, around 1 cm wide are also seen. They differ from the other minor shear zones in that no quartz fabrics are developed and the site of the shear zone shows a slight increase in grain size in which gneissose layering is obliterated, see Plate 3-5.

Shear zones within the dykes tend to be smaller, more numerous and more variably oriented than those in the gneisses. There is a gradual development from isotropic to foliated amphibolite over several cms and the shear zone margins are not as clearly defined as those in the gneisses, see Fig. 3-7. As in the gneisses the most intense fabric is often still oblique to the shear zone margins. Individual shear zones tend to show rather variable orientation of foliation and often enclose
FIG 3-6 Foliation orientation in S. Sithean Mhor gneisses and dykes.

- Poles to quartz fabric foliation in gneisses.
- Poles to foliation in dykes.
- Poles to layering in gneiss where quartz fabric is faint or isotropic.
Plate 3-5. Minor Shear zones in quartzo-feldspathic gneisses at 803702, area of inhomogeneous deformation, S. Sithean Mhor. Subvertical surface, looking NE.
pod-like masses of undeformed amphibolite.

These smaller shear zones are discontinuous, rarely extending beyond 2 or 3 cm in length. In the gneisses they either pass laterally into folds or fan out; the fabric becoming less intense until the shear zone is no longer recognisable. Within the dykes they either pass into faults or fan out with a decrease in intensity of deformation similar to that shown in the gneisses.

The largest shear zones which show the most intense fabric are those along the dyke contacts. They generally involve both gneiss and dyke but are largely developed within the gneisses with commonly only a few cms of the dyke margin involved in the deformation, see Fig 3-8 and Plate 3-6. The shear zones trend NW/SE, see Fig 3-1, and dip generally 85° SW parallel to dyke contacts (see Fig 3-9). The lineations shown by the quartz fabrics in the zone of highest deformation within the gneisses show a large range of orientations both within and between individual shear zones, eg. the lineations within a particular shear zone were observed to range from 25° to 60° NW within 4 m of strike. Lineations in general are steep to moderately dipping NW to SE, see Fig 3-9.

Gneissose layering is rotated into the shear zones over a distance of 1 to 3 m. In some cases folds are developed and there is a gradual transition from open folds with rounded hinges outside the shear zone to isoclinal folds with sharp hinges within the shear zone. In the most intensely deformed zones, folds are often no longer distinguishable and a strong quartz fabric is developed to which layering has become parallel, see Fig 3-8.

Intensity of the quartz fabrics and width of the shear zones is variable along their length, being approximately 2 m wide or less. Shear zones occur on one or both sides of the dykes but dykes without marginal shear zones also exist.
FIG 3-8 Sketches of examples of minor shear zones along dyke margins.

a) 807701.

b) 805701.
Plate 3-6. Gneissose layering deflected in minor shear zone along dyke margin, at 806702, area of inhomogeneous deformation, S. Sithean Mhor. Horizontal surface, looking NW.
FIG 3-9 Foliation and lineation orientation in minor shear zones, S. Sithean Mhor.

○ poles to foliation.
× lineation.
3.3.3. Area of homogeneous deformation.

To the NE of the area with minor shear zones, the quartz fabrics in the gneisses becomes stronger and more widespread. Folding changes from an irregular open style with rounded hinges to a tighter more regular style, interlimb angle $90^\circ$ or less, with straighter hinges, see Plate 3-7. Folds often have long limbs in which layering is parallel to the quartz fabric and shorter thickened limbs in which the layering is oblique to the quartz fabric, see Fig 3-10. Hinges dip moderately to steeply NW or SE.

In general the layering in this area is parallel to the quartz fabrics, but in places it is strongly oblique and markedly folded or crenulated, though the quartz fabrics have constant intensity inside and outside these areas. These zones occur up to several metres across and have rather sharply defined margins where layering is rotated into parallelism with the quartz fabric.

Pods are largely elongate and parallel to the quartz fabric and are rather less angular and more augen shaped than those to the SW. Internal folding is seen in some pods especially where gneissose layering is folded, while other pods show signs of boudinage, see Plates 3-8 and 3-9.

The amphibolite dykes are largely foliated though still preserve ophitic textures in isolated patches, see Fig 3-2. The dyke margins are largely parallel to the gneissose layering and quartz fabric, though discordances of up to $20^\circ$ are still occasionally seen. The foliated amphibolite has a sporadically developed mineral lineation. Dyke thicknesses are generally greater than those to the SW, up to 200 m thick. They contain in places numerous inclusions of gneiss in the form of elongate schlieren up to $\sim 10$ m wide by $\sim 100$ m long with well developed quartz fabrics parallel to their length, see Fig 3-2. In general they are aligned to the foliation in the surrounding dyke,
Plate 3-7. Isoclinal folds in quartzo-feldspathic gneisses at 806707, showing thickened hinges and shortened limbs, area of homogeneous deformation, S. Sithean Mhor. Horizontal surface, looking SE.
FIG 3-10 Sketch of folds in area of inhomogeneous deformation.

subhorizontal surface

fold hinges plunge steeply NW, quartz fabric axial planar.
Plate 3-8. Layered amphibolite pod in quartzo-feldspathic gneisses at 811710, area of homogeneous deformation, S. Sithean Mhor. Horizontal surface.
Plate 3-9. Folded amphibolite pod in quartzo-feldspathic gneisses at 806706, area of homogeneous deformation, S. Sithean Mhor. Horizontal surface, looking NW.
but on a small scale discordances are common, in which the amphibolite wraps around the gneiss inclusions parallel to the contact but quartz fabrics within the gneiss is cut across by the contact. To the NW at 809714 amphibolite dyke and gneiss are interleaved with layers 20 cms or less, parallel to the foliation. The proportions of gneiss to dyke here are approximately 1 to 1.

Garnets are locally very abundant in the dykes of this area, especially along joints close to the dyke margins (see section 3.2.4.). Where the gneissose layering and quartz fabrics are parallel, foliation planes are more commonly exposed and a quartz lineation is visible.

Fig 3-11 shows the plunge of lineations in the gneisses and dykes in a traverse from 810721 to 807702. In the NE, lineations plunge from 10 to 70° SE. To the SW the dip and direction of the lineation becomes more variable and in the extreme SW of the area of homogeneous deformation there is a wide range in plunge from 10° SE through directly down the dip of the foliation to 30° NW.
FIG 3-11 Variation in lineation plunge in S. Sithean Mhor gneisses and dykes.
3.4. Discussion

3.4.1. Origin and oldest structures in the gneisses.

The presence of pods of quartzo-feldspathic gneiss and amphibolite suggests that the quartzo-feldspathic gneisses have an intrusive or migmatic relation to an older gneissose terrain represented by the pods. The sheet-like aggregates of amphibolite pods in the areas of low deformation in the SW suggests there were originally layers, perhaps dykes, in a quartzo-feldspathic terrain. The relationship between marble pods and those of amphibolite is uncertain since they occur in the NE, where amphibolite pods are rare.

The oldest structures in the gneisses are the irregular open folds associated with moderately NW dipping layering observed in the SW of the area. Fabrics in the quartzo-feldspathic gneisses are here isotropic and any deformation recorded by quartz fabrics thus postdates these structures. However, angularity and shape of pods suggests that deformation postdating the formation of the quartzo-feldspathic gneisses and predating quartz fabrics is minor.

The variety of dyke shapes in the SW from parallel sided to pinch and swell is probably an original igneous feature since it is not associated with any foliation with the dyke or adjacent gneiss.

3.4.2. The deformation.

The sense of shear in minor shear zones in the area of inhomogeneous deformation can be deduced using the lineations and the deflection of gneissose layering. The lineations are the elongation of quartz aggregates which, since they are shape fabrics, record the direction of the maximum strain axis and not the shear direction. However, since deformation is very strong in the shear zone centres, the lineation provides a close approximation to the shear direction. Lineations within individual shear zones often show a large range of orientations probably due
to variation in fabric intensity and variable shear direction throughout time. Averaging these variations gives a shear direction that is subvertical. The deflection of gneissose layering observed on the subhorizontal erosion surfaces depends on the original layer attitude and the shear direction. Variation in original layer attitude has led to an apparent change in shear direction along the length of many shear zones, see Fig 3-12. Taking the above considerations into account, all minor shear zones indicate a positive sense of shear, ie. southwest side up, with a sub-vertical shear direction.

The small shear zones at 804702 in the gneisses dipping moderately N (Plate 3-5) differ from those described above in that no quartz fabric is developed at their centres. This indicates that they predate any quartz fabrics in the gneisses and therefore also the northwest trending shear zones.

In the northeast deformation is more homogeneous and fabrics are of moderate intensity. The similarity of foliation and minor shear zone orientation and of lineation plunge indicates that both belong to the same tectonic event. The abundance of minor shear zones suggests that the deformation is dominantly by simple shear. In the southwest of the area of homogeneous deformation, fold geometry indicate a positive sense of shear in agreement with minor shear zones supporting the view that deformation is here also by simple shear.

3.4.3. Time relations of dyke intrusion.

In the area of homogeneous deformation, dykes showing only faintly developed foliation contain well foliated schlieren of gneiss, indicating that they were intruded into a gneissose terrain with already well developed fabrics. The dykes are themselves deformed and thus their time of intrusion is syntectonic.

In the area of inhomogeneous deformation three factors indicate the
FIG 3-12 Effect of layer orientation on patterns produced by simple shear.

a) Vertically dipping shear plane.

Layer dip less than shear direction plunge.

b) Layer dip equal to shear direction plunge.

c) Layer dip SE or less than shearing direction.
relationship between shear zones and dyke intrusion:

i) Deformation involves dykes and gneiss, therefore deformation is active after intrusion.

ii) Shear zones are most common and continuous along straight parallel sided dykes suggesting that the shear zones have controlled dyke shape in these cases.

iii) Deformed and undeformed dykes alike trend NW/SE parallel to minor shear zones and to foliation throughout the area.

Thus dyke intrusion and shear zone development are synchronous; shear zones providing favourable places for dyke intrusion and dyke contacts providing weak points for shear zone initiation.
CHAPTER 4. THE USE OF QUARTZO-FELDSPATHIC GNEISSES IN STRAIN ANALYSIS

4.1. Introduction

The quartz aggregates in quartzo-feldspathic gneisses are used as strain markers in the areas of gneiss and an attempt is made in the following chapter to assess their usefulness as strain indicators. The results of four methods of estimating strain are compared and their suitability to these rocks discussed. The treatment throughout is two dimensional.

The rocks used for strain analysis are essentially bi-mineralic, composed of quartz and feldspar (plagioclase and K-feldspar). Rocks with amounts of biotite or hornblende over ~2% were discarded to minimise effects of substantial quantities of these minerals. Quartzo-feldspathic gneisses are abundant throughout both the Loch Tollie and S. Sithean Mhor gneisses as described in Chapter 2 (section 2.1.1.) and Chapter 3 (section 3.3.1.). The modal percentage of quartz ranges from 27% to 40%, on ave 32%, and quartz forms distinct aggregates within a feldspar matrix, see Plate 4-1. In deformed gneiss, the elongation and parallel alignment of these aggregates gives the rock a distinct foliation and lineation, easily observable in the field (see section 2.1.1. and section 3.3.1.).

The quartz aggregates themselves show highly complicated and varied shapes and range in size from 1 mm to 1 cm. Grains within the quartz aggregates show intense strain shadowing, while the feldspar matrix shows near polygonal textures with constant grain size. Overall there is a decrease in grain size and intensity of strain shadows to the SW in the Loch Tollie gneisses and to the NE in the S. Sithean Mhor gneisses, i.e. towards the Gairloch schist belt. As the quartz percentage increases contact between individual particles increases and at a concentration of over 35%, individual quartz aggregates become difficult to identify. In
Plate 4-1. Acetate peel from cut surface of deformed quartzo-feldspathic gneiss, specimen 4, area of homogeneous deformation, S. Sithean Mhor. Dark areas - quartz, pale areas - feldspar.
using the quartz particles as strain markers three assumptions are made. These are:-

1. that the effects of original shape and long axis alignment are negligible or can be taken into account. The possible effects of original shape and orientation were tested for each of the methods investigated using an undeformed specimen from the S. Sithean Mhor gneisses, since no undeformed gneisses exist in the Loch Tollie gneisses. It was found that each method gave a close approximation to a circle for the strain ellipse showing that the quartz particles are randomly orientated and distributed throughout the feldspar matrix.

2. that the quartz aggregates act as passive strain markers. Since the quartz particles are embedded in a matrix of different rheological properties, the strain they record will not be equivalent to the bulk strain. An attempt at estimating the extent of this effect and viscosity ratio between quartz and feldspar has been made in section 4.3.

3. that strain is homogeneous throughout the specimen. To ensure this as nearly as possible, care was taken during sampling to avoid minor folds or fault zones. Throughout the areas of gneiss strain tends to be homogeneous on the scale of a single outcrop, so this was not difficult.

4. that there has been no loss of material from or transfer between individual particles. The question of constancy of volume throughout deformation is difficult to assess. However, right up to the most extreme strains found (strain ratio ≈ 6), individual quartz particles are still easily distinguishable so that this effect cannot be extreme. Also the average volume of quartz aggregates, calculated in the process of estimating three
dimensional strain, (see 5.3.1.) show no correlation with strain. Therefore it is likely that there is little exchange or loss of material from particles throughout deformation.

4.2. Investigation and Comparison of Four Methods of Strain Analysis

4.2.1. The model.

To test some of the methods and to look at the particle-particle interference effects that arise from viscosity difference between particles and matrix, a two-dimensional model was constructed. An area, 4 cms square, of a section through an undeformed specimen of quartzo-feldspathic gneiss, as free as possible from the effects of gneissose layering, was used as a basis for the model, see Plate 4-2. Only particles over a diameter of 0.8mm were used to set a practical limit to the number of particles in the model. The centres of these particles, totalling 101 in all, were marked and each replaced by a circle of approximately the same area, see Fig 4-1. Particles that interfered with others were rejected and those that intruded into the sample area, but whose centres lay outside, were included.

The model was deformed using a computer programme written by D. Sanderson (Belfast) in which different particle and bulk strains may be used, thus simulating the effect of a range of viscosity ratios.

The undeformed specimen was tested by the centre to centre method (see section 4.2.3.) and found to have a small bulk strain, $R = 1.212$, with a major axis at $89^\circ$ to the X axis of the model (see Fig 4-1).

To eliminate the effects of this slight initial fabric the model was used twice for each test, once in its original position and once with X interchanged for Y. Averaging the two results eliminates the effects of initial fabric.
Plate 4-2. Acetate peel from cut surface of undeformed quartzo-feldspathic gneiss, specimen 1, area of inhomogeneous deformation, S. Sithean Mhor. The area outlined is that used in construction of the model.
a) Particle concentration = 28%

b) Particle concentration = 33%

Total number of particles = 101 Model constructed from specimen 1. Model (b) was obtained from model (a) by increasing particle diameters, thus increasing the total area occupied by particles.
The model is used to test methods 3 and 4 (section 4.2.3.) but since all particles are originally circular it cannot be used to test either of the methods 1 or 2. It is also used to investigate the particle-particle interference effect caused by particle-matrix viscosity differences (section 4.3.2.).

4.2.2. Methods of strain analysis.

Four methods of strain analysis have been applied to two-dimensional sections of quartzo-feldspathic gneiss and the results compared. The methods are 1) \(Rf/\phi\) method devised by Dunnet (1968), 2) calculation of the average ellipse, ie. strain ellipse, by the method outlined by Hext (1963), 3) calculation of the strain ratio by counting number of particle matrix interfaces per unit length in directions \(\kappa_1\) and \(\kappa_2\), here called "\(R_1\)", 4) the centre-to-centre method using the distribution of particle centres, devised by Fry (1979).

Methods 1) to 3) measure particle strain whereas method 4) measures bulk strain. Each method and its results are described below:

1) \(Rf/\phi\) Dunnet (1968)

a. The method.

The procedure involves the calculation of \(Rf\) (length/breadth) and measurement of \(\phi\) (long axis orientation) for a number of particles. A plot of \(Rf\) vs. \(\phi\) gives a pear shaped distribution, see Fig 4-2, with particles of the same initial \(R\) lying along pear shaped curves. \(\phi\), corresponding to the apex of the curve, is the orientation of the strain ellipse long axis and the shape of the curve is dependent on the strain ratio (Rs). A series of calibration curves with 50\% data curves, for Rs = 1.25 to 8.0 in intervals of 0.25, produced by the programme 'ONIONS' (Dunnet and Siddans, 1971) was used to estimate strain. The 50\% data curve and a line through the apex of the curve parallel to the \(Rf\) axis divides the distribution into four quadrants, and the curve giving the closest
FIG 4-2 EXAMPLES OF $R_f/\phi$ AND HEXT METHODS

a) $R_f/\phi$ method

- SPECIMEN D
- $R = 3.5$
- 60 POINTS
- $\chi^2 = 1.20$
- $\chi^2 (\text{CRITICAL}) = 11.35$

b) Hext method

- SPECIMEN 6
- $R = 4.2$
- 10 RAYS

- measured diameter

- best fit ellipse

$\phi$ (DEGREES)
to equal number of points in each quadrant is the best estimate of Rs. The method also allows the detection of initial fabrics as they produce assymmetric plots.

b. Application and results.

The Rf/Ø method was applied to nine surfaces of quartzo-feldspathic gneiss with 59-92 particles measured from each. The results are listed in Table 4-1. For each specimen, the number of points in each of the four quadrants was tested for isotropy using the chi-square test. All give values for chi-square less than the critcal value at the 0.01 level of significance, thus the distributions are indistinguishable from one in which the quadrants hold equal number of points, ie. particles had initially random orientation.

Difficulties arise in the application of the method due to highly irregular particle shapes which can make the orientation of the long axis difficult to determine especially at low strains. At high strains (R > 6) all particles are elongate and the range of orientations is small. The resultant plot is long and narrow and the difference between individual calibration curves at high strains is smaller, thus Rs is less accurately determined. Also, for elongate particles, the error in the breadth, and therefore in Rf, is proportionately larger. The method works best at intermediate strains, R \approx 2.5 to 5.0 when both the above difficulties are minimised.

2. Hext Method.

a. The method.

If the particles have initial random orientation, the average particle shape is equivalent to the strain ellipse (Shimanoto and Ikeda, 1970). The method outlined by Hext (1963) fits an ellipse to a number of radii given the length of each radius and its orientation, see Fig 4-2(b). The treatment given in his paper is in three dimensions and the theory has been adapted for use in two dimensions by R.F. Cheeney.
b. Application.

The method was applied to twenty specimens of quartzo-feldspathic gneiss one of which is the specimen of undeformed gneiss on which the model was based. Thirty-three particles per specimen were selected using a grid system and their centres estimated by halving the maximum length and breadth. The diameters of each particle were measured in six to eighteen common directions and the results for each direction summed. The results were fitted to an ellipse using a program written by D. Sanderson (Belfast).

Thirty-three particles per specimen were found to give reproducible strain values; two independent measurements of specimen 1 gave R to within 0.1 and Ø to within 0.6°. Also for this number of particles the average diameters gave a close approximation to an ellipse. The results from specimens A to I, to which the Rf/Ø method was also applied, are listed in Table 4-1 and those for specimens 1 to 11 listed in Table 4-2. Specimen 11, the undeformed specimen on which the model was based gave a result of 1.102, i.e., very close to a circle.

It was found that as strain increases the method tended to overestimate the strain unless extra diameters were measured near the ellipse long axis. The method works best when diameters produce points which are equally spaced around the ellipse circumference and not when they are at regular angular intervals. Thus a rough initial estimate of the strain must be made to choose the most suitable diameter orientation design for analysing strain. This factor affects the strain ratio, R, but has little effect on the orientation of major axis.

The location of particle centres is the most subjective part of the method but variation in choice of long axis orientation has little
Table 4-1. Results: Rf/Ø and Hext Methods

| Specimen | Rs  | $\phi^0$ | No points | $X^2$ | R   | $\phi^0$ | No Rays | $|\phi(Rf/\phi)|^0$ |
|----------|-----|----------|-----------|-------|------|----------|---------|----------------|
| A        | 2.25| 124      | 74        | 3.95  | 2.15 | 136      | 9       | 8              |
| B        | 3.75| 129      | 60        | 6.80  | 2.40 | 137      | 9       | 8              |
| C        | 5.00| 97       | 92        | 2.36  | 5.55 | 82       | 9       | 15             |
| D        | 3.00| 82       | 60        | 1.20  | 2.38 | 64       | 9       | 18             |
| E        | 1.75| 163      | 59        | 4.85  | 1.59 | 160      | 9       | 3              |
| F        | 2.00| 65       | 60        | 0     | 2.06 | 63       | 9       | 2              |
| G        | 4.00| 121      | 70        | 0.286 | 4.42 | 116      | 7       | 5              |
| H        | 4.75| 76       | 60        | 0.400 | 6.33 | 70       | 7       | 6              |
| I        | 3.50| 78       | 68        | 0.706 | 4.61 | 69       | 7       | 9              |

$X^2$ crit at 0.01 significant level = 11.35

angles anticlockwise from strike.

c - c - centre to centre method.
Table 4.2. Results: Hext, \( R_1 \) and centre to centre methods.

| Specimen | \( R(\text{Hext}) \) | \( R_1 \) | \( \phi(\text{Hext})^\circ \) | \( \phi(\text{c-c})^\circ \) | \( \left| \phi(\text{Hext}) - \phi(\text{c-c}) \right|^\circ \) |
|----------|----------------|--------|-----------------|-----------------|-----------------|
| 1        | 3.54           | 2.94   | 87              | 3.07            | 85              | 2               |
| 2        | 1.95           | 2.15   | 102             | 2.31            | 103             | 1               |
| 3        | 2.46           | 2.32   | 81              | 2.48            | 67              | 14              |
| 4        | 3.48           | 3.08   | 166             | 2.83            | 169             | 3               |
| 5        | 4.84           | 3.44   | 97              | 3.22            | 90              | 7               |
| 6        | 4.19           | 3.92   | 88              | 3.78            | 82              | 6               |
| 7        | 4.80           | 4.08   | 82              | 4.42            | 85              | 3               |
| 8        | 5.74           | 5.49   | 69              | 4.30            | 60              | 9               |
| 9        | 5.15           | 3.22   | 80              | 4.44            | 71              | 9               |
| 10       | 2.64           | 3.09   | 129             | 1.93            | 128             | 1               |
| 11       | 1.10           | 1.05   | 86              | 1.21            | 95              | 9               |

\( \text{c-c} \) - centre to centre method

\( \phi \) - angle anticlockwise from reference line in degrees.
effect on the location of the centres and, as long as a convention is kept, on the results. Initial estimates of the strain can be judged by eye before determining the diameter design or extra diameters added after initial results inspected. The method is, therefore, suitable for application to irregularly shaped objects such as quartz aggregates since it does not depend on estimation of long axis orientation or particle length and breadth.

3. The "R₁" Method.
   a. The method.

This method uses the change in the number of particle-matrix interfaces per unit length with direction. Consider two lines; A parallel to \( \lambda_1 \), B parallel to \( \lambda_2 \). On deformation particles originally intersected by line A will tend to move away from it - only particles whose centres lie on line A will not move. Along line B all particles originally intersected will remain so after deformation and others will have moved to intersect the line. Thus, after deformation the number of particles or matrix-particle interfaces per unit length will have decreased in direction \( \lambda_1 \) but increased in direction \( \lambda_2 \), see Fig 4-3. The relationship between the number of interfaces per unit length and the strain ratio, empirically derived from the model is:-

\[
R = \frac{\lambda_1}{\lambda_2} = \frac{\text{No of interfaces per unit length in direction } \lambda_2}{\text{No of interfaces per unit length in direction } \lambda_1}
\]

b. Application

The method was applied to eleven specimens, 1-11, to which the Hext method was also applied. Estimates of the directions found by the Hext method were used, though for a more rigorous application several directions should be measured and fitted to an ellipse independently.

To test the method for accuracy and for the effects of compeptance difference between bulk and particle strains the model described in
**FIG 4-3 R₁ METHOD**

MODEL R = 6

- Length of line A = Total length of lines B to G
- \( N_\text{of interfaces intersected by A} = 8 \)
- \( N_\text{of interfaces intersected by B to G} = 48 \)

Strain Ratio = \( \frac{N_\text{of interfaces per unit length in direction } \lambda_1}{N_\text{of interfaces per unit length in direction } \lambda_2} = 6 \)
section 4.2.1. was used. Here $R$ and the directions of $\lambda_1$ and $\lambda_2$ are known and it was found that some 800 interfaces per direction must be measured to obtain $R$ to an accuracy of $\pm2\%$.

On application to the model it was found that the method gives results close to particles strains and is insensitive to variation in bulk strain, where bulk strain is less than particle strain, see Fig 4-4. The method tends to underestimate $R$ by up to 0.2.

The method was then applied to eleven specimens (1 to 11) on which the Hext method was used. Results are recorded in Table 4-2.

The method was easily applied to the model as all particles are ellipses in the deformed state and each particle can be intersected by a straight line only twice. However, in the case of quartzo-feldspathic gneisses the particles have highly irregular shapes and a straight line intersecting a particle can cut more than two particle-matrix interfaces. This effect is more noticeable in the direction $\lambda_1$ than $\lambda_2$ and will lead to an underestimation of $R$. This effect can, however, be counteracted to some extent by careful 'editing' when counting interfaces.


a. The method.

All centre-to-centre methods use the distribution of particle centres and therefore measure bulk strain independently of particle strain. The method used here is one devised by Fry (1979). A plot is constructed about a central point by placing each particle centre in turn on a central point and marking in all other centres. The result is a plot with an elliptical hole at its centre whose shape is that of the strain ellipse, see Fig 4-6. If the particles are originally homogeneously distributed the resultant plot has a maximum density in a band surrounding the hole which Fry has used to define the strain ellipse, by contouring techniques.
FIG 4-4 $R_I$ VS. $R$(MODEL)

a) $R_p = 4$

$R_I = R_p$

b) $R_p = 6$

$R_I = R_p$
b. Application.

The method has been applied to four examples of the model and to specimens 1 to 11 to which the Hext and \( R_1 \) methods have been applied. The results are listed in Table 4-2. For each specimen the centres of between 107 and 275 particles were located by eye and digitised. From the digitised data a resultant plot was produced and point counted using a programme written by D. Sanderson (Belfast).

On application to both the model and the specimens no density maxima was found to surround the hole in the resultant plot. This is because the particles are randomly distributed and have a large range in original particle separation. The procedure used by Fry to define the strain ellipse cannot therefore be used for these examples. It was found that the best way to define the hole was to use a midway value between the lowest value (zero) and the highest background peaks. This corresponds approximately to the outer edge of the hole representing the strain ellipse, see Fig 4-5. The scale of the plot was chosen so that this value formed a continuous or nearly continuous contour around the hole.

The midway value contour has an irregular shape being made up of perpendicular straight line segments, see Fig 4-6. It was found that these shapes are too 'noisy' an approximation to an ellipse to use the Hext method without measuring an impractical number of diameters (>18). A quicker and easier method was to smooth the contour shape to a curve by eye and to fit this curve to an ellipse by the Hext method. With practice close approximation to an ellipse could be obtained by eye.

The method was tested for accuracy by application to four examples of the model of known \( R \) values, see Fig 4-7. The effects of slight initial fabric in the model (see section 4.2.1.) were eliminated by calculating the logarithmic average of two sets of results produced by rotating the model through \( 90^\circ \). These results show that the method tends to slightly underestimate strain by 0.3 at
FIG 4-5 CENTRE TO CENTRE METHOD: DENSITY VARIATIONS IN PLOT.

a) Point-counted plot.

b) Density vs. distance graphs.

1) Direction \( \lambda_1 \)

2) Direction \( \lambda_2 \)

--- Figures in number of points per 20 mm\(^2\)

--- average density curve.

SPECIMEN H
FIG 4-6 CENTRE TO CENTRE METHOD

a) Plot

b) Point-counted plot

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| 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |

| >2 | 2 | <2 |

c) Contoured plot

SPECIMEN 5.
FIG 4-7 R (CENTRE TO CENTRE) VS. R (MODEL)

+ First position of model
× Second position of model
○ Logarithmic average of results
In practical application of the method to quartzo-feldspathic gneisses, the most subjective part of the process is defining the particle centres as identifying individual particles can be difficult when they overlap. Only particles above a certain size were used, determined arbitrarily for each specimen, to try to eliminate the possibility of including offshoots of larger aggregates as separate particles. However, since the method uses a resultant plot constructed from the original distribution, the danger of influencing the final result by selective choice of aggregate centres is negligible.

The main problem of this method is the definition of the hole representing the strain ellipse. As \( R \) increases the ends of the hole become increasingly difficult to define and variation in the background density at the edge of the hole tends to increase this problem. Use of a midway contour provides an objective method of defining the hole which would be difficult by eye, but must be used with caution especially at high strains, ie. \( R > 4 \).

4.2.3. Comparison of methods.

The methods measure two aspects of strain recorded by the specimen; methods 1) to 3) measure particle strain and method 4) measures bulk strain.

a) Methods measuring particle strain.

i) Comparison of Hext and \( R_f/0 \) methods.

The results of both methods are plotted in Fig 4-8(a). For low strains, \( R > 3 \), the results compare well though above \( R = 3 \), the Hext method tends to give higher values than the \( R_f/0 \) method by up to 1.5 at \( R = 5 \) or 6. The angles of plunge of \( \theta \), relative to strike for both methods are recorded in Table 4-2. Most show good agreement with angular difference of less than 10° except for C and D.
FIG 4-8 COMPARISON OF PARTICLE STRAIN METHODS

a) \( \text{Hext vs. } \frac{R_f}{\phi} \)

b) \( \text{Hext vs. } R_I \)
(15° and 18° respectively). There is no clear correlation between R and angular difference, i.e. at strains greater than ~1.5, angular accuracy does not depend on R. As discussed in section 4.2.2.(1) the Rf/Ø method becomes difficult to apply at high strains and failure to measure very elongate particles will lead to an underestimation of strain. Also, the Hext method tends to overestimate R at high values of strain (section 4.2.2.(2)). The lack of agreement between methods at high strains is therefore due to both these factors. However, at high strains measurement of extra diameters in the Hext method minimises the effect of overestimation so that the discrepancy is probably largely due to underestimation by the Rf/Ø method.

ii) Comparison of Hext and R_I methods.

The results of the Hext and R_I methods have been plotted in Fig 4-8(b). As seen in the comparison of Hext and Rf/Ø, the Hext and R_I methods compare well at low strains, i.e. up to R ~3. Above this the Hext method gives progressively higher values than the R_I method by up to 1.5 at R = 4 to 5.

As discussed previously (section 4.2.2.(3)) the R_I method tends to underestimate strain due to irregular particle shapes and this effect will increase as strain increases. The Hext method tends to overestimate strain emphasising the difference in results between the two methods. The discrepancy in results is therefore a result of inaccuracy in both methods.
iii) Conclusions

For the measurement of particle strain the three methods investigated, $R_{f/\phi}$, Hext and $R_I$, agree well up to values of $R = 3$. Above this strain value differences in the methods become apparent, i.e. Hext tends to overestimate and $R_I$ and $R_{f/\phi}$ to underestimate strain. Angles agree well for the $R_{f/\phi}$ and Hext methods and accuracy does not appear to be affected by magnitude of strain.

The $R_{f/\phi}$ and $R_I$ methods become very difficult to apply at high strains due to low range in long axis orientation and irregular particle shape, respectively. The Hext method can be applied as long as individual particles are recognisable though accuracy decreases as strain increases. This method is also independent of particle long axis orientation and shape.

The Hext method is therefore the most suitable at higher strains and where particle shapes are highly irregular, as long as care is taken to ensure a roughly equal spacing of diameters around the ellipse circumference to minimise the effects of overestimating strain. The $R_{f/\phi}$ and $R_I$ methods are quicker to apply and therefore most suitable if strains are low and particle shape regular.

b. Comparison of particle and bulk strains.

The Hext and $R_I$ methods, measuring particle strain, have been compared to the centre to centre method, measuring bulk strain and the results are plotted in Fig 4-9(a) and (b). The centre to centre method gives values of $R$ which are consistently lower than those for the Hext method, and show little difference from values given by the $R_I$ method. The angles of plunge of $\lambda_1$ anticlockwise from strike are recorded in Table 4-2, for both Hext and centre to centre methods. They compare well, all showing angular difference of less than $10^0$ except one. As in the comparison of Hext and Dunnet methods there is
FIG 4-9 PARTICLE VS. BULK STRAIN

a) Hext vs. Centre to Centre

b) \( R_I \) vs. Centre to Centre
no clear correlation between R and angular difference.

From tests with the model (section 4.2.2.4) the centre-to-centre method underestimates the strain by approximately 0.3 at R = 3 to 5. Since the specimens were analysed using a larger number of centres there is no reason to suppose the specimen results are any less accurate. The R1 method tends to underestimate strain by about 0.2 (section 4.2.2) in the model but this error will be greater for the results for specimens and is probably nearer 0.5 at R = 4 to 6. The Hext method tends to overestimate strain and a figure of 0.5 at R = 4 to 6 for the R1 method implies an overestimation of approximately the same amount for the Hext method. Thus comparison of the centre to centre method with the R1 method will underestimate, and with the Hext method will overestimate, the difference between particle and bulk strains, and thus the viscosity ratio between quartz and feldspar.

4.3. Effect of Viscosity Contrast between Particles and Matrix

4.3.1. Particle versus bulk strain - implications on the viscosity ratio between quartz and feldspar.

Using the estimates of particle and bulk strain of section 4.2.2. upper and lower limits can be placed on the viscosity ratio between quartz and feldspar. Bilby et al (1975) have developed an equation (Eq 13) relating the particle and bulk strains to viscosity ratio for the slow deformation of a viscous material containing an elliptical inhomogeneity:

\[ S + \frac{1}{2} \alpha \tanh S = S_H \]

where \[ 1 + \alpha = \frac{\mu_1}{\mu} \]
and \[ \mu_1 = \text{particle viscosity} \]
\[ \mu = \text{matrix viscosity} \]
\[ S_H = \ln R_T \text{ where } R_T = \text{ratio of bulk strain ellipse} \]
\[ S = \ln R_p \text{ where } R_p = \text{ratio of particle strain ellipse}. \]
Results are tabulated in Table 4-3 and show that the viscosity ratio for quartz to feldspar lies between 0.96 and 0.76. Since the centre to centre and $R_1$ methods underestimate strain, by $\sim 0.3$ and $\sim 0.5$ respectively, and the Hext method tends to overestimate by $\sim 0.5$, the discrepancy between the former two methods is less than that between the Hext and centre to centre methods. Thus the true value of the viscosity ratio lies closer to 0.96 than 0.76.

The deformation is dominantly simple shear (see sections 2.4.2 and 3.4.2.) so one of the effects of a viscosity contrast between particles and matrix is an angular difference between particle and bulk strains. This can be calculated using the theory of Bilby et al. (1975), and Bilby and Kolbuszewski (1977).

Equation 33 of Bilby and Kolbuszewski calculates the angle of particle long axis from the reference axis. Using the maximum viscosity difference $\mu/\mu = 0.75$, and an initially circular particle this equation simplifies to:

$$\cos 2\psi = \left( \frac{1}{G(\sigma)} \right) \left( I(\sigma) - 4.222 \right)$$

where:

$$I(\sigma) = \frac{k_{(\sigma)}^{R+1}}{R \times \sigma \left( \sigma + 1 \right)^2}$$

and

$$G(\sigma) = \frac{(\sigma^2 - 1)}{\sigma} \left[ \frac{k_{(\sigma)}}{(\sigma + 1)^2} \right]^R$$

$\sigma =$ axial ratio of particle

$R =$ viscosity ratio

$\psi =$ angle between X axis and particle major axis

$k =$ constant.

To find the angles relative to the reference axis for the bulk strain, an initially circular particle and a viscosity ratio of 1 were used in equation 33, and results recorded in Table 4-4. These
### Table 4-3. Viscosity Ratios

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<th>Specimen</th>
<th>$\frac{R(\text{Hext})}{R(\text{c-c})}$</th>
<th>$\frac{R_f}{R(\text{c-c})}$</th>
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<td><strong>Average</strong></td>
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<td><strong>0.956</strong></td>
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### Table 4-4. Angular difference between bulk and particle strains under simple shear.

| $R_p$ | $R_B$ | $\theta_p^0$ | $\theta_B^0$ | $|\theta_p - \theta_B|^0$ |
|-------|-------|---------------|---------------|--------------------------|
| 2     | 1.834 | 39.9          | 34.4          | 3.5                      |
| 3     | 2.616 | 31.6          | 31.7          | 0.1                      |
| 4     | 3.364 | 28.6          | 28.6          | 0                        |
| 5     | 4.229 | 25.9          | 25.9          | 0                        |
show that angular differences between bulk and particle strains are greatest at low strains but even at $R = 2$ the difference is only $3.5^0$, too small to be detected by any of the methods used here. Thus there is little error in assuming that bulk and particle strains in quartzo-feldspathic gneisses are coaxial.

The percentage error in using particles to estimate bulk strain can be calculated using equation 13 (Bilby et al (1975)). A viscosity ratio half way between the upper and lower limits (see above), 0.86, gives an error of 5 to 11% overestimation through a range of strains from $R = 2$ to $R = 6$. Since the true viscosity ratio is likely to lie closer to 0.96 than 0.75 this is a maximum error and thus in general quartz particles in quartzo-feldspathic gneiss overestimate strain by a maximum of 10%.

4.3.2. Particle interference - applicability of the model.

One of the effects in the model, of varying bulk and particle strains independently is that particles overlap. This is equivalent to particles in contact with one another in the real rock. The number of overlaps is dependent on:

i. the difference between particle and bulk strains,

ii. particle concentration,

iii. absolute bulk strain.

An attempt to quantify this effect has been made by counting the number of overlaps per hundred particles for varying bulk-particle strain pairs in the model, see fig. 4-10. This was done for two values of particle concentration, 28% and 33%, which correspond to the range found in the specimens. The model of 33% particle concentration was derived from the original (28% particle concentration) by increasing particle areas in proportion to their radii without allowing overlaps to occur. Since particle distributions are an important factor here,
FIG 4-10 PARTICLE INTERFERENCE IN THE MODEL

Particels overlapping
Particels touching

Total number of particles = 101
Number of particle contacts = 10

R (BULK) = 3.00
R (PARTICLE) = 5.00

b)

R (BULK) = 3.00
R (PARTICLE) = 6.00

Number of particle contacts = 15
the effects of small initial fabric in the model were eliminated by averaging the results given by the model in two positions, the second at 90° to the first (see section 4.2.1.). The final results are plotted in Figs 4-11(a) and (b). Factors to be noted from these figures are:-

1. When the viscosity ratio of particle to matrix is less than one, overlaps are of 'end to end' type. There are some overlaps of 'shoulders' of particles but no side to side overlaps occur since particle sides are always moving away from one another. Similarly, when particle to matrix viscosity ratios are greater than one, only side to side or 'shoulder' overlaps should be seen; end to end overlaps being theoretically impossible.

2. From the graph, Fig 4-11, for constant $R_p$ (axial ratio of particle strain) the number of overlaps is proportional to $R_B$ (axial ratio of bulk strain).

3. Proportional difference between particle and bulk strain is more important than the absolute value of the difference.

4. The number of overlaps is to some extent dependent on absolute strains, ie. constant $R_p/R_B$ ratios do not show constant number of overlaps but increases as $R_p$ (and $R_B$) increase.

5. There is a theoretical limit to the number of overlaps possible given by the line $\mu_p/\mu = 0$ (viscosity ratio).

6. Increasing the particle concentration by 5% produces roughly the same increase in number of overlaps, which suggests that it is directly proportional to particle concentration.

7. Contours of equal number of overlaps are slightly oblique to contours of equal viscosity ratio.

The number of overlaps per hundred particles and the particle concentration are recorded in Table 4-5 for the ten specimens (1 to 10)
FIG 4-11 NUMBER OF OVERLAPS: $R_B$ VS. $R_P$

a) Particle concentration = 28%

b) Particle concentration = 33%
Table 4-5. Number of overlaps.

<table>
<thead>
<tr>
<th>Specimen</th>
<th>% particles</th>
<th>No overlaps per 100 particles</th>
<th>Average R</th>
</tr>
</thead>
<tbody>
<tr>
<td>10</td>
<td>26</td>
<td>13.6</td>
<td>2.86</td>
</tr>
<tr>
<td>5</td>
<td>27</td>
<td>9.6</td>
<td>4.14</td>
</tr>
<tr>
<td>6</td>
<td>28</td>
<td>16.3</td>
<td>4.05</td>
</tr>
<tr>
<td>9</td>
<td>29</td>
<td>10.0</td>
<td>4.19</td>
</tr>
<tr>
<td>2</td>
<td>29</td>
<td>14.4</td>
<td>2.05</td>
</tr>
<tr>
<td>3</td>
<td>30</td>
<td>17.7</td>
<td>2.39</td>
</tr>
<tr>
<td>4</td>
<td>30</td>
<td>27.8</td>
<td>3.28</td>
</tr>
<tr>
<td>8</td>
<td>30</td>
<td>17.1</td>
<td>5.61</td>
</tr>
<tr>
<td>7</td>
<td>32</td>
<td>16.3</td>
<td>4.44</td>
</tr>
<tr>
<td>1</td>
<td>33</td>
<td>15.2</td>
<td>3.40</td>
</tr>
</tbody>
</table>
used in the methods of estimating strain. For each an average particle strain (the average of Hext and $R_1$ methods) and the number of overlaps per hundred grains were used to plot the values on Fig 4-11. Results show that the number of overlaps in the specimens exceeds the expected values by up to a factor of two and in some cases exceeds that theoretically possible. There are several possible factors that could account for the discrepancy:

1. The model has all particles initially separate. In the specimens some particles may have been initially in contact, thus increasing the final number of overlaps.

2. The identification of particle centres is done by eye and is therefore subjective. It is possible that some particles are identified as two or more particles due to irregularity of shape, thus artificially increasing the number of overlaps. However, since quartz is less viscous than feldspar (section 4.2.4.b) boudinage is unlikely so that this effect is probably minor.

3. In the model particles change shape in a homogeneous fashion, i.e. circles become ellipses and either end to end or side to side overlaps should be seen depending on the viscosity ratio. According to the model both types of overlap cannot occur simultaneously but in practice both types were frequently observed within a single specimen. In the rocks it is likely that the particles are continuously changing shape throughout the deformation, apart from simply becoming more elongate. Once contact between particles has been established, e.g. between growing apophosae, it is unlikely that they would separate again as this would involve the flow of more viscous matrix into less viscous particle. This will tend to increase the number of overlaps and make side to side overlaps occur in a situation where they are theoretically impossible.

Thus the number of overlaps observed in the specimens is not due solely to the preferential elongation of quartz particles in
relation to their matrix but also to other more complicated factors involving change of shape of particles and flow within the matrix. It is not thought that the choice of particle centres is an important factor though it may have a small effect. The number of overlaps predicted by the model is therefore an underestimation, and cannot be used as a means of estimating the viscosity ratio.

4.4. Summary and Conclusions.

Four methods of strain analysis have been applied to quartzofeldspathic gneisses and tested using models of known strain. The results of the \( \frac{R_f}{\Omega} \) method and analysis of an undeformed specimen show that the quartz aggregates have initial random orientation and can therefore be used as strain markers.

Of the methods investigated the Hext and centre to centre methods are best suited to the material. The \( \frac{R_f}{\Omega} \) and \( R_1 \) methods are quicker to apply but can be used only at low to intermediate strains. The Hext method is slower but applicable to any strain if the diameter design is taken into account. The centre-to-centre method has the advantage of measuring bulk strain but is difficult to use at high strains.

Comparing bulk to particle strains, using the theory of Bilby et al. (1975), gives upper and lower limits to the viscosity ratio of quartz to feldspar of 0.96 and 0.77, with the most likely value lying closer to 0.96. This figure gives a maximum error in using particles to estimate bulk strain of 10% overestimation. Examples of varying viscosity ratio simulated by the model show that the number and type of particle contacts is not due solely to preferential elongation of particles but also to other factors probably involving change of particle shape and flow in the matrix.

The value of viscosity ratio of 0.96 to 0.77 for quartz to feldspar strictly applies only under the conditions prevailing at the
time of deformation, i.e. Amphibolite facies of metamorphism (see 8.3), and will vary according to pressure, temperature, composition of the metamorphic fluid and grain size. However, results compare well with data presented by Gay (1968) who lists viscosity ratios derived from pebble strains in conglomerate. The nearest equivalent to quartz and feldspar is quartzite and granite which give a viscosity ratio (quartzite: granite) of 0.83 to 0.65. This overlaps with the result of 0.96 to 0.77 obtained here and supports the conclusion that quartz is less competent than feldspar.

Quartzoo-feldspathic gneisses are common throughout most gneissose terrains where strain markers of a more conventional type are scarce. They therefore provide abundant material for strain analysis in areas where little or nothing is known of the finite strains. In the following two chapters these rocks are used for the analysis of three-dimensional finite strain as a basis for the interpretation of the structural history of the Gairloch area.
5.1. Introduction

In this chapter the results of strain analysis in NE/SW traverse across the Tollie antiform are presented. The use of quartzo-feldspathic gneisses in strain analysis is assessed in chapter 4 for two dimensions and the treatment is here extended to three dimensions. The major structure of the area, the Tollie antiform, lithologies and minor structures are described in full in chapter 2.

Throughout chapter 5, the following convention for principal axes of strain is used; maximum axis - X, intermediate axis - Y, minimum axis - Z. The lengths of principal axes are presented on a logarithmic deformation plot (Flinn 1965), as ratios where \(a = \frac{X}{Y}\), \(b = \frac{Y}{Z}\) and \(k = \frac{a}{b}\). Ellipsoids and corresponding fabrics which plot in the field where \(K < 1\), i.e., below the \(k = 1\) line are referred to as 'S' and those in the field where \(k > 1\), i.e., above the \(k = 1\) line as 'L'. The orientations of principal axes lineations and planes are plotted on lower hemisphere Lambert Equal-area projections.

5.2. Strain Indicators and Methods

Two types of strain markers were used:-

a. quartz particles in quartzo-feldspathic gneiss and,

b. hornblendite and amphibolite pods.

5.2.1. Quartz particles in quartzo-feldspathic gneiss.

The specimens of quartzo-feldspathic gneiss used are essentially bi-mineralic with less than 2% of biotite or hornblende. The proportion of quartz to feldspar is on average 30:70 and the quartz forms aggregates of grains with highly irregular and complicated shapes within
a feldspar matrix. The particle size is variable and shows a range of average particle volume per specimen from \(0.4 \times 10^{-3}\) to \(12 \times 10^{-3}\) cms\(^3\), see Table 5-2.

The use of quartz particles as strain markers is discussed in detail in chapter 4 and the treatment is here extended to three dimensions using the Hext method (Hext, 1963). Three mutually perpendicular faces were cut from each specimen and from each face 33 grains selected and their diameters measured as described in 4.2.2.(2). Average diameters from each face were normalised using directions common to two faces and all measurements fitted to an ellipsoid by the Hext method, using an unpublished computer programme called 'Paten' compiled by R.F. Cheeney (Edinburgh). The programme gives the lengths and orientations of the maximum, intermediate and minimum axes of the strain ellipsoid, with 99% confidence limits. A more detailed description of the procedure is given in Appendix I.

Twelve rocks collected from a NE/SW traverse from 859777 to 824776 across the Tollie Antiform, see Fig 5-1, have been analysed for strain by the above method, see Appendix II.

5.2.2. Feldspathic amphibolite and hornblendite pods.

The length, breadth and orientation of feldspathic amphibolite and hornblendite pods were measured on sub-horizontal erosion surfaces to estimate 2D strain. Pods showing evidence of boudinage (see 2.3.3.(2)) were avoided. The pods are always elongate parallel to the foliation in the gneisses, deviating from it by only one or two degrees. Because of the narrow range of orientations, the \(R_f/\sigma\) method (Dunnet (1969), see 4.2.2.(1)) is unsuitable so the geometric mean of length to breadth ratios was used to estimate strain ratio though this overestimates true strain by approximately 0.3 at \(R > 2\) (Lisle 1977, Fig 2).
FIG 5-1 Location of specimens—Loch Tollie gneisses.

Key
Location of specimens
O quartzo-feldspathic gneisses
• pods

Scale
1 km.

[Map showing locations of specimens around Loch Tollie, with key annotations and scale.]
Between 15 and 30 pods were measured at each of nine localities; seven in the SW steep belt and two in the flat belt, see Fig 5-1. For each locality the geometric mean with 95% confidence intervals were calculated.

5.3. Results

5.3.1. Quartzo-feldspathic gneisses.

The orientations of strain ellipsoid principal axes are plotted together with the 99% confidence limits on a series of stereographs in Fig 5-2. The lengths of the principal axes are plotted in the form of ratios 'a' and 'b' (see 5.1) on a logarithmic deformation plot.

1. Discussion on confidence limits.

a. Confidence limits on orientations.

Specimens from the NE steep belt and flat belt, Fig 5-2(a) and (b) show principal axes orientation with confidence cone semi-angles of 25° or less with most in the range of 10°-20°. In these cases there is good control on the orientations of the principal axes. Specimens from the SW steep belt, Fig 5-2(c), show similar confidence cones except for two, 8 and 9 for which the maximum and intermediate axis orientations are not distinct at the 99% confidence level. These specimens plot close to the 'b' axis on the deformation plot, Fig 5-3, i.e. the X-Y section through the strain ellipsoid is nearly circular and maximum and intermediate axes orientations are difficult to determine.

b. Confidence intervals on lengths of principal axes.

The confidence intervals on ratios 'a' and 'b' were calculated using natural strain values by the equation given by Topping (1958) p. 82, and are recorded in Table 5-1. These limits are large and in many cases the upper limit to 'a' lies outside the domain of real
FIG 5-2(a) Orientations and 99% confidence intervals of principal strain axes in the NE belt of the Tollie Antiform.

- Maximum axes (X)
- Intermediate axes (Y)
- Minimum axes (Z)

--- X-Y Principal planes
FIG 5-2(b) Orientation and 99% confidence intervals of principal strain axes in the Flat belt of the Tollie Antiform.

- Maximum (X) axis
- Intermediate (Y) axes
- Minimum (Z) axes
- X-Y Principal planes
FIG 5-2(c). Orientations and 99% confidence intervals of principal strain axes in the SW steep belt of the Tollie Antiform.

- Maximum (X) axes
- Intermediate (Y) axes
- Minimum (Z) axes
--- X-Y Principal planes
FIG 5-3 Logarithmic deformation plot and 99% confidence intervals of principal strain axes in the Loch Tollie gneisses.

\[ a = \text{length of maximum axis} \]

\[ b = \frac{\text{length of intermediate axis}}{\text{length of minimum axis}} \]

\[ k = \frac{a}{b} \]

---

99% confidence intervals on 'a' and 'b'.

Confidence interval extends off diagram.
Table 5-1. Strain analysis from quartzo-feldspathic gneisses.

<table>
<thead>
<tr>
<th>Specimen</th>
<th>a</th>
<th>99% conf. limits</th>
<th>b</th>
<th>99% conf. limits</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>lower conf. limit</td>
<td>upper conf. limit</td>
<td>lower conf. limit</td>
</tr>
<tr>
<td>1</td>
<td>4.002</td>
<td>1.018</td>
<td>-</td>
<td>1.905</td>
</tr>
<tr>
<td>2</td>
<td>6.094</td>
<td>0.856</td>
<td>-</td>
<td>2.608</td>
</tr>
<tr>
<td>3</td>
<td>4.484</td>
<td>1.116</td>
<td>-</td>
<td>1.344</td>
</tr>
<tr>
<td>4</td>
<td>2.758</td>
<td>1.298</td>
<td>-</td>
<td>1.656</td>
</tr>
<tr>
<td>5</td>
<td>9.091</td>
<td>1.046</td>
<td>-</td>
<td>1.664</td>
</tr>
<tr>
<td>6</td>
<td>2.114</td>
<td>1.261</td>
<td>6.545</td>
<td>1.284</td>
</tr>
<tr>
<td>7</td>
<td>2.152</td>
<td>1.085</td>
<td>129.870</td>
<td>1.890</td>
</tr>
<tr>
<td>8</td>
<td>1.500</td>
<td>0.980</td>
<td>1.735</td>
<td>3.223</td>
</tr>
<tr>
<td>9</td>
<td>1.537</td>
<td>0.890</td>
<td>5.615</td>
<td>5.333</td>
</tr>
<tr>
<td>10</td>
<td>2.048</td>
<td>0.438</td>
<td>-</td>
<td>9.461</td>
</tr>
<tr>
<td>11</td>
<td>4.608</td>
<td>0.910</td>
<td>-</td>
<td>3.182</td>
</tr>
<tr>
<td>A</td>
<td>1.771</td>
<td>0.406</td>
<td>-</td>
<td>2.731</td>
</tr>
</tbody>
</table>
Table 5-2. Average volume of quartz particles in specimens 1 to 11.

<table>
<thead>
<tr>
<th>Specimen</th>
<th>Average Volume (cm$^3$ x 10$^{-3}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.660</td>
</tr>
<tr>
<td>2</td>
<td>3.563</td>
</tr>
<tr>
<td>3</td>
<td>1.238</td>
</tr>
<tr>
<td>4</td>
<td>0.430</td>
</tr>
<tr>
<td>5</td>
<td>11.665</td>
</tr>
<tr>
<td>6</td>
<td>1.408</td>
</tr>
<tr>
<td>7</td>
<td>3.511</td>
</tr>
<tr>
<td>8</td>
<td>0.528</td>
</tr>
<tr>
<td>9</td>
<td>2.602</td>
</tr>
<tr>
<td>10</td>
<td>10.649</td>
</tr>
<tr>
<td>11</td>
<td>4.809</td>
</tr>
</tbody>
</table>
numbers.

The confidence intervals may be thought of as a smaller ellipsoid inside and a larger ellipsoid outside the best estimate ellipsoid, related to it, along the principal axes, by equal values of \( \frac{1}{X^2} \) where \( X \) is the confidence interval. They are therefore asymmetric about the best estimate.

The Hext method supplies a general quadratic surface for the data, for which an ellipsoid is only one solution. Extreme fabrics of either L or S type have led to the situation where the upper confidence limit surface has one or two imaginary axes, ie. the 'ends' and/or 'sides' have failed to close to form an ellipsoid. In strain analysis such quadratic surfaces are meaningless and in these cases an upper confidence limit on the data is not available. In other cases, though an ellipsoid is achieved, upper confidence intervals are very large.

Lack of, or very large, upper confidence intervals implies poor precision of the data. However, several considerations suggest that the data is more reliable than these confidence intervals would suggest:-

1. The type of strain ellipsoid indicated by the lengths of the principle axes correlates with the confidence interval pattern on the orientations. When 'b' is large and 'a' small, oblate ellipsoids, orientations of maximum and intermediate axes are indistinguishable (specimens 8 and 9) and when 'a' is large and 'b' small, prolate ellipsoids, the maximum axis orientation is precisely defined.

2. The results of strain analysis agree well with the preliminary estimation of fabric type made in the field, ie. specimens plot in the expected region of the deformation plot.

3. The strain ellipsoid type in part of the SW steep belt
around 836771 are supported by field evidence; 'chocolate tablet' structure in the dykes and hornblendite pods indicate an ellipsoid of $k > 1$ type (see 2.4.3.) in agreement with the strain analysis results, specimens 8 and 9.

The reason for large confidence intervals is thought to lie in the design of the measured diameter orientations. Hext discusses this at some length and states that the best design is one in which the points are evenly distributed over the surface of a sphere. However, from application of the method in two dimensions (4.2.2.) it was found that the best results are obtained when points are evenly distributed around the ellipse circumference which, extended to 3D, implies an even distribution over the ellipsoid surface. This discrepancy between theory and practice is probably due to the 'noise' level of the data which increases around the maximum (and intermediate in some 3D cases) of the ellipse (4.2.2.).

In practice, achieving a design in which points are evenly distributed over a sphere or ellipsoid is difficult, since measurements are made on three mutually perpendicular surfaces which results in a markedly uneven distribution of points. From the study of the 2D case (4.2.2.) the effect of using such a design is an exaggeration of the ellipsoid shape, ie, increase in ratios 'a' and 'b'. Thus, points on the logarithmic deformation plot, Fig 5-3, tend to occur closer to the 'a' and 'b' axis and further from the origin than the true values, but always lie on the correct side of the $a = b$ ($k = 1$) line.

Since confidence intervals on the lengths of principal axes are thought to give a misleading view of the precision of the data, they are not plotted on further diagrams. However, better knowledge of the confidence intervals of directions than lengths of principal
axes means that more significance must be placed on directions of axes in the interpretation of the data.

2. Discussion on results.

The results of the strain analysis show variations in ellipsoid type across the antiform as predicted by field observation of quartz fabrics, i.e., from NE to SW, near LS in the NE belt, L in the flat belt and LS to S to LS in the SW steep belt. Orientations of maxima are subhorizontal, N to NW trending in the NE belt and flat belt, and gentle SE to moderate NW plunging in the SW steep belt. The rotation of maxima from gentle SE to moderate NW in the SW steep belt is uncertain as the finite strain is of $k > 1$ type in which maxima orientations within the X-Y plane are not known. Orientations of minima (poles to X-Y planes) range from moderate SW plunging in the NE belt, subhorizontal SW trending in the flat belt to moderate NE - subhorizontal - gently SW plunging in the SW belt. The X-Y plane is equivalent to the quartz fabric foliation in the NE and SW steep limbs and agrees with field observation.

From the data, the average volume of quartz particles for each specimen has been calculated and these are listed in Table 5-2. The rocks show a wide range of quartz aggregate size but no observable trend in volumes from NE to SW. This supports the view that there has been little or no loss or gain of material from the quartz particles during deformation (4.1).

5.2.2. The feldspathic amphibolite and hornblendite pods.

The results of 2D analysis of pods with 95% confidence limits are listed in Table 5-3 and plotted in Fig 5-4. Confidence limits were calculated as for arithmetic means using natural logs of ratios and are therefore asymmetric. Localities of fewer numbers of pods show larger confidence limits.
Table 5-3. 2D strain values for pods.

<table>
<thead>
<tr>
<th>Specimen</th>
<th>geometric mean G</th>
<th>95% conf. limits</th>
<th>number of pods</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>lower conf. limit</td>
<td>Upper conf. limit</td>
</tr>
<tr>
<td>R</td>
<td>69.7</td>
<td>34.8</td>
<td>193.6</td>
</tr>
<tr>
<td>S</td>
<td>24.6</td>
<td>10.1</td>
<td>60.0</td>
</tr>
<tr>
<td>T</td>
<td>12.6</td>
<td>7.4</td>
<td>21.3</td>
</tr>
<tr>
<td>U</td>
<td>25.1</td>
<td>12.5</td>
<td>50.6</td>
</tr>
<tr>
<td>V</td>
<td>11.9</td>
<td>8.0</td>
<td>17.6</td>
</tr>
<tr>
<td>W</td>
<td>32.7</td>
<td>19.7</td>
<td>54.2</td>
</tr>
<tr>
<td>X</td>
<td>38.0</td>
<td>23.4</td>
<td>61.8</td>
</tr>
<tr>
<td>Y</td>
<td>37.7</td>
<td>24.1</td>
<td>59.0</td>
</tr>
<tr>
<td>Z</td>
<td>16.2</td>
<td>9.4</td>
<td>27.8</td>
</tr>
</tbody>
</table>
FIG 5-4  2D Strain values from pods in the Flat belt and S.W. limb of the Tollie Antiform.
Results show a range of 2D strain ratios across the SW steep belt and flat belt. In the flat belt ratios are high and then decrease over the crest of the antiform into the SW steep belt. At 841771 they increase again to a maximum value of \( R = 38 \) at 836770 and then decrease to \( R = 16 \) at 828771.

5.2.3. Comparison of quartzo-feldspathic gneiss and pod strains.

Throughout the flat belt and SW steep belt the minima are subhorizontal or indistinguishable from the intermediate axes. Thus the ratio maximum:minimum provides a rough estimate of the horizontal 2D section through the strain ellipsoid with a tendency to overestimate. These values are also plotted in Fig 5-4 with the results from the measurements of pods.

Comparison of the two sets of results shows that the pods give consistently higher results, five to sixteen times greater than those given by the quartzo-feldspathic gneisses. The only value that gives similar results is specimen 11 and the corresponding pod value of \( R = 15.0 \). The ratio maximum:minimum for the quartz strain ellipse gives a maximum possible value for a 2D section so that the discrepancy between the two sets of results is actually greater than presented in Fig 5-4.

Pods of amphibolite and hornblendite always lie parallel to the foliation with very little deviation in orientation. This implies that either pods were initially equidimensional in section or that long axis of pods were aligned before deformation. Since it is highly unlikely that pods were originally all equidimensional in section suites of pods must have had an initial fabric, ie. parallel alignment before deformation. Evidence of this is to be seen in the flat belt where sections perpendicular to the linear fabric display sheets of pods in which individual pods tend to be elongate parallel
to the trend of sheets and gneissose layering (2.2.2). Similar evidence is seen in the S. Sithean Mhor gneisses (3.2.2).

This initial fabric accounts for the discrepancy in the two sets of strain values and implies initial axial ratios for pods of between 5 and 16 with most between 5 and 10. Pods in undeformed gneiss, S. Sithean Mhor, give a geometric mean of 3.8 which is lower but of the same order of magnitude as the implied initial fabric in the Loch Tollie gneisses.

However, the two sets of results show a similar overall pattern with a decrease in strain ratio on the horizontal plane around the crest of the antiform and in the area immediately to the SW after which strain increases again. The pods can therefore be used as relative strain markers but owing to strong initial fabrics cannot be used as indicators of absolute strain.

5.4. Modelling the Strain Path

5.4.1. Introduction

In the following sections an attempt is made to model the strain path of the deformation using theoretical superposition of strains.

From field evidence, including changes in grain size and textures, dyke and gneissose layering thickness and appearance of quartz viens, deformation in general increases to the SW (see chapter 2). The rocks in the NE of the area therefore represent the least deformed or most primitive fabrics upon which the deformation affecting the rocks to the SW has acted. This was used as a basis for the theoretical models investigated which are of two types:-

a. deformation by rotational strain; field evidence (2.4.2) suggests that the deformation is of simple shear type. In this case the Tollie antiform is produced as a consequence of the deformation.
b. deformation by rotational strain; an alternative hypothesis in which the formation of the Tollie antiform must precede the deformation.

Both models are constructed using an unpublished computer programme called 'Sust', written by D. Sanderson (Belfast).

5.4.2. Rotational strain model.

This model is constructed using the theoretical superposition of simple shear strains. The variables in the model are the shear plane orientation, the initial fabric, shearing direction and sense, and increments of shear. In order to reduce the number of variables to a practical level, field evidence was used to fix the shear plane orientation and shear sense:

1. The shear plane orientation.

Since deformation increases to the SW the average orientation of foliation in the extreme SW of the loch Tollie gneisses, 75° in the direction 040°, was chosen as the shear plane. This orientation is constant throughout the rocks in the centre of the deformed zone which includes the SW part of the Tollie gneisses, the Gairloch schist belt, and the NE part of the S. Sithean Mhor gneisses, and therefore provides an approximation to the shear plane orientation.

2. The initial fabric.

The fabric in the N-eastern most part of the Tollie gneisses, i.e. those furthest from the zone of intense deformation was taken as an initial fabric, specimen 1, see Fig 5-1. This fabric lies in the constrictional field of the deformation plot (Fig 5-3) with a gently N plunging maximum axis (Fig 5-2).

3. The shearing direction and sense.

The shear direction is unknown although there is some field evidence to suggest a direction plunging between 50° and 60° NW. This
includes the NW plunging quartz lineations in the SW part of the Tollie gneisses and similarly oriented hornblende lineations in the NE half of the Aundrary amphibolite. This amphibolite shows evidence of extreme grain size reduction (7.2.1.) which took place as a result of the deformation so that orientation of hornblende laths records the shear direction. The quartz fabric lineations are small fold hinges (2.3.3.(1)) and must therefore be rotated into the shear direction. Theoretically they only reach the shear direction after an infinite amount of shear but will be undistinguishable from it in practice, when the value of reaches the order of 10. Since the quartz lineations have similar orientations to the hornblende lineations in the adjacent Aundrary amphibolite, this implies that the shear strain in the area is at least around ten.

The sense of shear is determined by the geometry of the Tollie antiform and must therefore be SW side down NE side up, i.e. negative with respect to right handed orthogonal axes, in which X points E, Y points N and Z points down.

A range of eight shearing directions were used in the model to test the hypothesis of a NW plunging direction, see Table 5-4.

4. Increments of shear.

In progressive simple shear finite strain changes rapidly under the initial stages and more slowly as the maximum approaches the shear direction and the X-Y principal plane approaches the shear plane. The shear increments (all negative) were therefore chosen as 0.2, 0.5, 1.0, 1.5, 2.0, 2.5, 3.0, 4.0, 6.0, 8.0.

The results.

1. Shearing of initial fabric.

The results of eight different shearing directions are plotted in Figs 5-5(a) and (b). As stated in 5.3.1.(1) the directions of principal
Table 5-4. Shear directions in model

<table>
<thead>
<tr>
<th>Number</th>
<th>Direction (plunge and direction of plunge)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>11 313</td>
</tr>
<tr>
<td>2</td>
<td>38 322</td>
</tr>
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<td>7</td>
<td>29 121</td>
</tr>
<tr>
<td>8</td>
<td>14 125</td>
</tr>
</tbody>
</table>
FIG 5-5(a)1. Rotational strain model, paths described by maximum axes.

- Initial maximum axis orientation (specimen 1)
- Position of axes after increments of shear: 0, 2, 0, 5, 1, 0, 1, 5, 2, 0, 2, 5, 3, 0, 4, 0, 6, 0, 8, 0
FIG 5-5(a)2. Rotational strain model, paths described by minimum axes.

- Initial minimum axes orientation (specimen 1)
- Position of axes after increments of shear: 0, 0.2, 0.5, 1.0, 1.5, 2.0, 2.5, 3.0, 4.0, 6.0, 8.0.
FIG 5-5(b). Rotational strain model: Logarithmic deformation plot of strain paths.

Initial fabric (specimen 1)

Position after increments of strain: 0 2 0 5 10 1.5 2 0 2.5 3 0 4 0 6 0 8 0.
axes are most critical in fitting the models to the strain data described in 5.3.1.(2) with the change in fabric types secondary in importance.

Only NW plunging shearing directions produce the required NW plunging maximum axes; the maximum axes describe a path that passes through horizontal to become SE plunging, then through subvertical or steep SW plunging to become moderately plunging NW. SE plunging shearing directions (models 5, 6, 7 and 8) produce only SE plunging maxima. Also NW plunging directions produce a range of L and S fabrics whereas SE plunging directions produce only S fabrics. Thus only NW plunging shearing directions can explain the range of orientation and fabric types occurring across the Tollie antiform.

Of the NW plunging directions model 2 does not produce S fabrics. Models 3 and 4 are similar but model 3 provides a better fit to the foliation plane orientations and fabric variation, see Figs 5-6(a) and (b). Comparison of Fig 5-6(a) and (b) shows that this direction produces L fabrics as the maximum axis is subhorizontal corresponding to the fabrics in the flat belt of the Tollie antiform. Fabrics enter the S field when the maximum axis is SE plunging and foliation plane (plane X-Y) is moderate to steep SW dipping, corresponding to the SW steep limb adjacent to the crest of the fold. Here the maximum and intermediate axes of the strain data are indistinguishable but the best estimates indicate a SW plunge, in agreement with the model. Finally the fabric becomes close to k = 1 as the maximum axis becomes NW plunging and the X-Y (foliation) plane is subvertical in agreement with the strain data in the extreme SW (specimen 11). The direction of model 3, (55° 333°) also agrees well with the field evidence for the orientation of the shearing direction cited in 5.4.1.(2).

The results of model 3 were tested for stability with respect to the ratios 'a' and 'b' since these parameters are least well known.
FIG 5-6(a). Rotational strain model 3: Comparison with principal strain axes orientation of strain data.  
1) NE and Flat belts.

- Maximum (X) axes
- Minimum (Z) axes

- Strain path of model 3
- Strain path in constrictional field ($k < 1$)
- Strain path in flattening field ($k > 1$)
FIG 5-6(a) Rotational strain model 3: Comparison with principal strain axes orientation of strain data.
2) SW steep belt.
FIG 5-6(b). Rotational strain model 3: Comparison with strain data on logarithmic deformation plot.

- Strain data from quartzo-feldspathic gneisses with specimen numbers.
- Strain path of Rotational strain model 3 with increments of strain; 0.2, 0.5, 1.0, 1.5, 2.0, 2.5, 3.0, 4.0, 6.0, 8.0.
FIG 5-6(c) Rotational strain model: results of variations in ratios a and b for model 3.

Initial fabric (specimen 1)

Position after increments of strain: $\gamma = 0.2, 0.5, 1.0, 1.5, 2.0, 2.5, 3.0, 4.0, 6.0, 8.0$. 
The results are plotted in Fig 5-6(c) and show that variations in 'a' and 'b' of 0.5 have little effect on the results.

2. Unshearing of the initial fabric.

It has been shown from the above that the fabrics SW of 859775 (specimen 1) can be explained in terms of progressive simple shear superimposed on this as an initial fabric. If all fabrics in the area are due to simple shear then the true initial fabric must be of k = 1 type. The initial fabric used (specimen 1) lies in the constrictional field but to the SE of its location at 864763 rocks with LS fabrics and SE plunging maxima exist (specimen A) and these may therefore represent the original initial fabric. To test this hypothesis, specimen 1 has been unsheared, ie. sheared in a positive sense, in the eight directions used in models 1 to 8, see Fig 5-7(a) and (b). Unshearing using direction 3, which gave the best fit to the strain data, does not produce k = 1 (LS) fabrics or SE plunging maxima but does give a path that passes close to specimen 2 which is therefore probably a slightly less highly deformed rock than specimen 1.

Only direction 5 (73° 0690) gives k = 1 (LS) fabrics with SE plunging lineations. A possible explanation for the fabrics in the NE of the area is that they show the result of a progressive change in shearing direction from SE plunging (producing initial fabrics, eg. specimen A) to moderately NW plunging (producing initial fabrics of specimens 1 and 2) though subvertical shear directions lying directly down the dip of the shear plane.

An alternative hypothesis is that specimens A and 1 represent variations in the original fabric prior to the onset of NW plunging shearing. This could explain the augen shape of the area of linear fabrics around loch Tollie, as only some of the initial fabrics would give rise to linear fabrics on subsequent shearing.
FIG 5-7(a). Rotational strain model, 'unshearing' initial fabric.

- Maximum (X) axes
- Minimum (Z) axes
- Strain paths with increments of shear, 1, 2, 4, 6, 8.
FIG 5-7(b). Rotational strain model, 'unshearing' of initial fabric.

○ Strain data from NE belt.

● Strain paths with increments of shear, 0.2, 0.5, 1.0, 1.5, 2.0, 2.5, 3.0, 4.0, 6.0, 8.0.
The true picture is probably a combination of these. Specimens 1 and 2 probably incorporate some deformation by shearing in a NW direction, since specimen 2 is likely to be less deformed than specimen 1, and both lie in the L field. It is also very likely that initial fabrics were originally variable. Possible initial fabrics are seen the area immediately to the NE of the Tollie antiform, Craig Mhor Tollaidh, where deformation is inhomogeneous, fabrics ranging from LS to L with variable lineation orientation.

5.4.3. Irrotational strain model.

An alternative interpretation of the Tollie antiform is that it represents an original fold which has been progressively deformed to the SW. To test this hypothesis a model was constructed in which a fold was subjected to progressive increments of irrotational plane strain.

The original attitude of the fold is unknown so the simplest case of a symmetrical concentric fold with gently SE plunging hinge, was chosen as the basis for the model. The variables in the model are then the pre-existing fabric, the orientation of the principal axis of strain and the increments of strain.

1. Pre-existing fabric.

The fabric in the NE limb (specimen 1) is taken as the original fabric as in the rotation strain model. Folding this concentrically to produce a symmetrical fold with hinge plunging 10° SE, gives fabrics that plunge gently SE on the SW limb, see Fig 5-8. Two sets of initial fabric orientation were used in the model, one representing the crest of the fold and the other the SW limb (A and B in Fig 5-8).

2. Orientation of the principal axes of strain.

The orientation of the X-Y plane of the strain was chosen to coincide with the foliation in the extreme SW of the SW limb (75° 040°).
FIG 5-8. Irrotational strain model, initial fold and fabrics.

- □ Maximum (X) axes
- Δ Intermediate (Y) axes
- ○ Minimum (Z) axes
- A NE limb
- B Crest fabrics on initial fold
- C SW limb
where deformation is most intense. The direction of the maximum (X) axis is unknown so a range of eight orientations, the same eight orientations used for shearing direction in the rotational strain model, were used.

3. Increments of strain.

Ten increments of plane strain were used to outline the strain path. These were chosen as $a = b = 1.1, 1.25, 1.5, 2.0, 2.5, 3.0, 4.0, 6.0, 8.0$.

The results.

The results for the crest of the fold are presented in Fig 5-9(a) and (b) and for the SW limb in Fig 5-10(a) and (b). To satisfy the strain data, the model must produce subhorizontal linear fabrics in the crest of the fold and the sequence LS-S-LS from NE to SW across the SW limb finishing with NW plunging maxima.

a. The crest of the fold.

From Fig 5-9(a) all paths stay within the L field except models 4 and 5. Only 2, 3 and 4 produce stronger linear fabrics with 'a' values that fit the strain data, i.e. between 3 and 6. These models give linear fabrics after 2 to 4 increments of pure shear at which stage maximum axes are horizontal to gently plunging NW, Fig 5-9(a), in agreement with the strain data.

b. SW limb of the fold.

From Fig 5-10(b) the only paths to produce strains in the flattened field are 2, 3 and 4, and of these only 2 and 3 produce NW plunging maximum axes Fig 5-10(a). None of the models produce fabrics that return to the $k = 1$ line (LS fabrics) after entering the flattened field.

Of the models 2 and 3, model 3 provides the closest fit to the strain data. Model 2 produces only gently NW plunging maximum axes.
FIG 5-9(a). Irrotation strain model, paths described by maximum and minimum axes in crest of fold.

- Maximum (X) axes
- Minimum (Z) axes

Initial fabric

Strain path with increments of strain $a = b = 11, 1.25, 1.5, 2.0, 2.5, 3.0, 4.0, 6.0, 8.0$. 
FIG 5-9(b). Irrotation strain model, logarithmic deformation plot of strain paths in crest of fold.

Initial fabric (specimen 1)

Strain path with increments of strain $a = b = 1.1, 1.25, 1.5, 2.0, 2.5, 3.0, 4.0, 5.0, 6.0, 8.0$. 

---

**FIG 5-9(b)**. Irrotation strain model, logarithmic deformation plot of strain paths in crest of fold.
FIG 5-10(a). Irrotation strain model, paths described by maximum and minimum axes in SW limb of fold.
FIG 5-10(b). Irrotational strain model, logarithmic deformation plot of strain paths in SW limb of fold.

○ Initial fabric (specimen 1)

→ Strain paths with increments of strain $a = b = \{1.1, 1.25, 1.5, 2.0, 3.0, 4.0, 6.0, 8.0\}$. 
FIG 5-11(a). Irrotational strain model 3: comparison with principal axes orientation of strain data.

1) crest of fold.

- Maximum (X) axes
- Minimum (Z) axes
- Maximum (X) axes
- Minimum (Z) axes

Initial fabric

Strain data

Strain path with increments of strain $a = b = 1, 1.25, 1.5, 2, 2.5, 3, 4, 6, 8$.

Strain path in constrictional ($k > 1$) field.
FIG 5-11(b). Irrotational strain model 3: comparison with principal axes orientation of strain data.

2) SW limb of fold

Maximum (X) axes

Minimum (Z) axes

Initial fabric

Strain data

Strain path with increments of strain $a = b = 1.1, 1.25, 1.5, 2.0, 2.5, 3.0, 4.0, 6.0, 8.0$.

Strain path in flattening field ($k < 1$).
FIG 5-12(a). Irrotational strain model 3: comparison with strain data.
1) crest of fold.

Strain data from quartzo-feldspathic gneisses.

Strain path of Irrotation strain model 3 with increments of strain:
\[ a = b = 1.1, 1.15, 1.25, 1.5, 2, 2.5, 3, 4, 6, 8, 10. \]
FIG 5-12(b). Irrotation strain model 3: comparison with strain data.

2) SW limb of fold.

Strain data from quartzo-feldspathic gneisses.

Strain path of Irrotational strain model 3 with increments of strain \( a = b = 11, 1.25, 1.5, 2.0, 2.5, 3.0, 4.0, 6.0, 8.0 \).
and only after very high strains. Also model 2 does not produce as markedly S fabrics as model 3. Neither model produces the required $k = 1$ (LS) fabrics after entering the S field. Thus to produce a strain path to completely fit the strain data a more complicated model is required, involving either a variation in initial fabric or variation in orientation of the pure shear strain.

5.4.4. Comparison of the two models and conclusions.

Owing to the uncertainty of the strain data and the assumptions that must be made to construct the models, there is no single factor which conclusively proves the deformation to be by simple or pure shear. However, a comparison of the models shows that in many ways the rotational strain model provides a better fit to the data:

1. The rotational strain model provides all the fabric types and orientations in one model whereas the irrotational model requires additional deformation stages to produce $k = 1$ (LS) fabrics in the SW Tollie gneisses.

2. The irrotational strain model tends to produce fabrics of a more extreme type than the rotational strain model and the strain data. For example the irrotational model crosses the $k = 1$ line at $a = b = 2.5$, the rotational model at 1.7 and the strain data indicates values of around 1.8. Since it is thought that the methods of strain analysis used tend to overestimate strain, the rotational strain model provides the better fit.

3. The rotational strain model provides a mechanism for the formation of the Tollie antiform whereas in the irrotational strain model the fold must exist before the onset of deformation. The rotational strain model therefore explains more of the structure of the area.

As well as providing a better fit to the data there is much better
control of the variables in the rotational strain model than the irrotational strain model; the shear plane orientation is fixed by field evidence whereas there is no information on the original shape of a pre-existing fold.

These factors are also supported by field evidence:

1. Small scale features in the Loch Tollie gneisses such as quartz vein orientations and small scale shear zones indicate simple shear as the dominant type of deformation, see 2.4.2.

2. The interlimb angles of folds as described in chapter 2 show a range from moderate to open with moderately NE dipping axial planes in the NE limb, becoming very open with subvertical axial planes in the flat belt to tight/isoclinal with steep SW to NE dipping axial planes in the SW belt. Assuming that the NE limb represents the original state of folds, the rotational strain model will tend to increase the interlimb angles, whereas the irrotation strain model will result in a progressive decrease in interlimb angles to the SW. Thus, only the rotational strain model is consistent with the open folds found in the flat belt.

Conclusions.

Several factors indicate that the deformation is, at least dominantly, by simple shear, including:

a. from the model - type, range and intensity of fabrics produced, formation of the Tollie antiform and control of the model, and

b. from field evidence - minor features and range of interlimb angles of folds.

Thus the Tollie gneisses lie on the NE margin of a large scale shear zone. Such a large scale shear zone is unlikely to produce a theoretically perfect strain profile and irregularities in
lithology or movement of the two blocks on either margin will cause some component of pure shear in the deformation. This could explain why the planar fabrics in the SW limb of the Tollie antiform are more pronounced than is predicted by either rotational or irrotational models.

5.4.5. Strain profile across the Tollie antiform and comparison with layer attitude and lineation orientations.

a. Shear strain estimated from the model.

The results of strain path modelling and field evidence show that the deformation is by simple shear and that strain progressively increases to the SW. Estimates of the shear strain ($\gamma$) superimposed on the initial fabric have been made by comparing the model with strain data and a strain profile constructed for the Tollie gneisses, Fig 5-13. This shows a steady increase throughout the NE limb and flat belt with a value of $\gamma \approx 2$ at the crest of the antiform and a steeper strain gradient in the SW steep belt with a value of $\gamma = 6$ at the most S-westerly specimen.

b. Shear strain from gneissose layering attitude.

A strain profile can also be constructed from the change in attitude of the layering across the antiform. The attitude of layering in the NE belt was used as the initial orientation and sheared using the shear plane, direction and sense of model 3 (5.4.2.). Three values of dip $45^0$, $50^0$, $55^0$ with constant direction of dip, $052^0$, were chosen to represent the range of dips in the NE belt. These planes were sheared by known amounts using the equation (1) given by Skjernaa (1979):

$$ p^1 = \tan^{-1} (\tan(P - 90) - 1 \cos V1) + 90 $$

where $P = \text{dip of the plane in degrees}$

and $V = \text{strike of the plane in degrees}$. 
Results show the overturning of the plane to form an antiform with hinges dipping 16°, 20° and 24° SE respectively, see Fig 5-14(a), (b) and (c). Comparing the above results with the attitude of layering in the field, the model shows the strike of layering swinging progressively northwards in the SW limb whereas in practice strike swings in the opposite direction, SE. Also in practice the hinge dips more gently (~10°) than that predicted by the model. These discrepancies are probably due to large scale variations in the initial attitude of layering which are not taken into account by the model.

However, the dip of layering can be used to estimate shear strain across the antiform by comparison with the models (Fig 5-14) and the results are plotted on Fig 5-13. Shear strain estimated by dip of layering agrees well with that estimated from finite strains and supports the hypothesis that the deformation is by simple shear. The attitude of layering at the locality of specimen 11, the most S-westerly specimen, giving a shear strain of 6.5, is 85° SW. To the SW of this locality the foliation turns over to dip ~ 85° NE which implies shear strains of approximately 8.0 in the extreme SW part of the Tollie gneisses, see Fig 5-13.

c. Orientation of lineations.

Lineations within the SW steep belt are subhorizontal in the NE adjacent to the crest of the antiform and rotate through horizontal to become moderately dipping NW in the SW; these orientations are plotted in Fig 5-15(a). As described in 2.3.3.(1) they are crenulations in the surface of quartz lenses, ie. are equivalent to small fold hinges. Their formation must have occurred during the shearing deformation from two considerations:-

1. As noted in 5.4.4., folds in the NE belt are progressively opened under simple shear of model 3 to form very open folds in the flat belt after shear strains of between 1 and 2. Under the high

- Estimation of shear strain from comparison of strain data and model 3.
- Estimation of shear strain from comparison of dips of gneissose layering and models A, B, and C.
FIG 5-14(a). Change in attitude of planes in progressive simple shear: initial plane orientation 45° 052°.
FIG 5-14(b). Change in attitude of planes in progressive simple shear: initial plane orientation: 50° 052°.
FIG 5-14(c). Change in attitude of planes in progressive simple shear: initial plane $55^\circ 052^\circ$.

O hinge of fold: $24^\circ 122^\circ$
FIG 5-15(a). Lineation orientations in the SW steep belt.
strains estimated for the SW belt few of these original folds will remain recognisable. Thus folds in the SW limb are more likely to have formed during the shearing deformation than be relict folds pre-dating the deformation.

2. The crenulations are irregularities in the surfaces of quartz lenses. Quartz particles show varying shape from NE to SW and pass through a linear rod-like stage in the flat belt, at which stage there are little surfaces on which such crenulations could form. Thus the lineations must have formed as the quartz fabrics became more planar during the formation of the SW steep limb and therefore have formed during the shearing deformation.

The behaviour of such lineations during deformation is uncertain but one hypothesis is that they behave as passive lines and thus rotate towards the shearing direction. The behaviour of passive lines during simple shear can be modelled using the equations (2), (3) and (4) of Skjernaa (1979):

\[ \lambda = 8^2 \sin^2 \phi - 2 \lambda \sin \phi \cos \phi \sin \beta + 1 \]

\[ \sin \phi' = \frac{\sin \phi}{\sqrt{\lambda}} \]

\[ \cos \beta' = \frac{\cos \phi \cos \beta}{\sqrt{\lambda} \cos \phi'} \]

where \( \phi \) = initial angle between the line and the shear plane,
\( \beta \) = initial angle between shear direction and projection of the line onto the shear plane + 90°,
\( \phi', \beta' \) = final values of these quantities.

Six lineations were chosen to represent the spread of lineations in the extreme SW of the Tollie gneisses. These were unsheared using the shear plane, direction and opposite sense of model 3, to find their original orientations and the results are plotted in Fig 5-15(b).

Note that for constant increments of shear, lineations move little at
FIG 5-15(b). Unshearing of lineations by model 3.

- Representative lineations in SW part of SW limb
- Strain paths with increments of shear: 0, 2, 5, 10, 15, 20, 25, 30, 40, 60, 80.
first, increase the movement per increment of shear as they move away from
the shear plane and decrease again on finally approaching the shear
direction. After a shear strain of approximately eight, lineations
are subhorizontal, NW or SE plunging.

Comparison of the model with the data in Fig 5-15 shows agreement
in the range of lineation orientations but disagrees with the strike
orientation with the model producing original lineations with a more
easterly strike. This is due to the swing in strike to the S that occurs
in the SW steep limb adjacent to the crest of the antiform as already
described in section 5.4.5. The theoretical behaviour of lines as they
rotate agrees with the variation of plunge of lineations described in
2.3.3. and Fig 2-7. From NE to SW across the SW limb, lineations
just show constant orientation with little scatter, then rapid change
of plunge with a wide scatter of points and finally more constant NW
plunging values with less scatter in the extreme SW. As seen in Fig
5-15(b), small variations in the original orientation greatly affects
the rate at which lineations rotate. This accounts for the large scatter
of values where change in plunge is rapid and the smaller range in lineation
plunge as they approach the shearing direction. Thus it seems probable
that the lineations have behaved largely as passive lines during the
shearing deformation.

5.5. Summary

Strain has been analysed in a NE/SW traverse across the Tollie
antiform by two methods:-

a. 2D analysis using quartz fabrics in quartzo feldspathic gneisses

and

b. 3D analysis of pods in subhorizontal sections.

The confidence intervals on the principal axes lengths are large
and are thought to give a misleading view of the accuracy of the
data. From several considerations fabric type (L or S) is thought to be reliably represented though intensity (i.e. ratios 'a' and 'b') tends to be overestimated. Comparison of the two methods of analysis in horizontal sections indicates the pods to have a strong initial fabric but, in general, the two methods show similar patterns.

Three dimensional analysis of quartzo-feldspathic gneisses shows a wide range in fabric types and orientations across the Tollie antiform, which is supported by field observation. Two models of the deformation, based on rotational and irrotational strain were constructed to deduce the probable deformation type and strain path. The rotational strain model has several advantages over the irrotational model;

1. It explains all fabric types within the Tollie gneisses whereas the irrotational model does not.

2. There is better control of variables in the rotational strain model.

3. Field evidence supports a rotational strain model.

The conclusion is that the deformation is, dominantly at least, of simple shear type and that the Loch Tollie gneisses therefore lie on the NE margin of a large scale shear zone. The fabrics in the NE belt thus represent an initial fabric to a main phase of shearing with a steeply NE dipping shear plane, moderately NW plunging shear direction and negative shear sense.
CHAPTER 6. STRAIN ANALYSIS AND STRAIN PATH MODELLING IN THE S. SITHEAN MHOR GNEISSES

6.1. Introduction

The S. Sithean Mhor gneisses lie on the SW edge of the Lewisian outcrop at Gairloch and similar gneisses can be traced southwestwards, via small inliers to the major outcrop of Lewisian at Torridon. The gneisses contain abundant amphibolite dykes and lithological layering, foliation and dykes in general trend NW/SE as in the Gairloch schist belt and Tollie gneisses to the NE. The lithologies and minor structures of the S. Sithean Mhor gneisses are described in chapter 3. Unlike the Tollie gneisses the area shows no major fold structures but the style and intensity of deformation is variable from SW to NE across the strike of the gneisses.

In this chapter, the strain analysis of the S. Sithean Mhor gneisses is presented and interpreted in the light of minor structures and field evidence. The terminology is similar to that used in chapter 5.

6.2. Strain Analysis

6.2.1. Quartzo-feldspathic gneisses.

Six specimens collected from a NE/SW traverse across the S. Sithean Mhor area from 803701 to 813717, see Fig 6-1, were analysed for 3D strain using the Hext method (Hext, 1963). A description of the procedure applied to quartzo-feldspathic gneisses is given in Appendix I. The discussion on the use of the method in 3D and confidence intervals on directions and lengths of principal axes given in 5.2.1. and 5.3.1. also applies to the results here.

The results of the strain analysis are presented in Table 6-1. The direction of the principal axes with 95% confidence intervals are plotted in Fig 6-2(a) and the lengths of the axes plotted on a logarithmic
FIG 6-1 S. Sithean Mhor gneisses — location of specimens.

KEY
Location of specimens
- quartzo-feldspathic gneisses.
- pods.
Scale 500 m.

Area of inhomogeneous deformation

Area in Fig 3-2
Table 6-1. Strain data from quartzo-feldspathic gneisses

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<td>1.085</td>
<td>3</td>
<td>4.210</td>
<td>2.733</td>
</tr>
<tr>
<td>4</td>
<td>2.042</td>
<td>0.353</td>
<td>4</td>
<td>1.987</td>
<td>1.722</td>
</tr>
<tr>
<td>5</td>
<td>1.524</td>
<td>1.299</td>
<td>5</td>
<td>3.440</td>
<td>2.683</td>
</tr>
<tr>
<td>6</td>
<td>1.868</td>
<td>0.966</td>
<td>6</td>
<td>2.187</td>
<td>1.910</td>
</tr>
</tbody>
</table>


deformation plot in Fig 6-2(b). Fig 6-2(a) shows that all specimens show distinct axes orientation except specimen 5 for which maximum and intermediate axes are not distinct. This specimen plots in the k > 1 (S) field. The confidence intervals on ratios 'a' and 'b' are large and similar to those shown by specimens from the loch Tollie gneisses.

Specimens 1 and 2 are collected from the area of inhomogeneous deformation (3.3.3.). Specimen 1 was collected from an outcrop showing isotropic fabric. It gives a strain ellipsoid close to a sphere (see Fig 6-3) with a = 1.58 and b = 1.02 and a shallowly NW dipping maximum axis. Note that in this case the confidence intervals on ratios 'a' and 'b' are relatively small while those on the directions large, compared to those of the more highly deformed specimens. Thus as the strain ellipsoid approaches a sphere, lengths of axes become more precisely determined.

Fig 6-2(b) shows that all specimens lie close to k = 1 line with an increase in strain in specimens 1 to 3. Specimens 4, 5 and 6 show less intense fabrics with a tendency to lie in the k > 1 field. Fig 6-2(a) shows that all specimens have similar orientations of X - Y planes trending NW with a steep NE dip, except specimen 6 which shows a more westerly trend. This corresponds to a swing in the trend of the foliation at 815715 around the Ard Ialltaig gneisses.

Within the area of homogeneous deformation the maximum axis orientations are variable but tend to be either moderately NW plunging or gently to moderately SE plunging.

6.2.2. The pods.

Pods occur within the quartzo-feldspathic gneisses SW of 813711 and these have been measured for length, breadth and long axes orientation in 2D subhorizontal surfaces for ten localities (see Fig 6-1). Strain is estimated for localities 1-8 using the method outlined by Dunnet.
FIG 6-2(a) Orientation and 99% confidence interval of principal strain axes in the S. Sithean Mhor gneisses.

□ Maximum (X) axis
△ Intermediate (Y) axis
○ Minimum (Z) axis
----- X-Y planes
FIG 6-2(b) Logarithmic deformation plot and 99% confidence intervals of principal strain axes in the S. Sithean Mhor gneisses.

\[ a = \text{length of maximum axis} \]
\[ b = \text{length of intermediate axis} \]

\[ k = a/b \]

---

99% confidence intervals on 'a' and 'b'.

Confidence interval extends off diagram.
(1969), the details of which are given in 4.2.2., and show strains from $R = 2.25$ to $R = 5.00$. Localities 9 and 10 show strains greater than $R = 8$, the maximum for which the Dunnet method can be used. For these, the logarithmic mean, $G$, was used as the best estimate of strain. Results are plotted against distance in Fig 6-3.

At locality 1 the pods occur in quartzo-feldspathic gneisses with isotropic fabric and gives a strain value of 2.25 with long axes orientation $024^\circ$. This shows that the pods have an initial fabric upon which deformation is superimposed. Localities 2 to 8 show a small range of strain values from $R = 2.5$ to $R = 5.0$ with average long axis orientation $139^\circ$, consistently greater than that at locality 1 ($R = 2.25$). Localities 9 and 10 show much higher values, $R = 12$ to 15, approximately three times greater than at other localities. The pods of localities 9 and 10 differ from those at localities 1 to 8 in that they are composed of coarser grained amphibolite and occur with layers of amphibolite of similar grain size and composition within quartzo-feldspathic gneisses. They probably therefore have different initial fabrics than those at localities 1 to 8.

6.2.3. Comparison of results.

The two dimensional strains of horizontal sections are estimated for the 3D results of the quartzo-feldspathic gneisses and presented in Fig 6-3, with results of analysis of pods. Since the X-Y planes of the strain ellipsoids are subvertical, ratios of X, Y or an intermediate value, to Z give estimates of $R$ in horizontal sections.

Fig 6-3 shows that the results compare well for localities 1 to 8, giving values of $R$ between 2 and 4.5. Localities 9 and 10 show greater values of $R$ from the analysis of pods than quartzo-feldspathic gneisses. As described in 6.2.2. the pods at localities 9 and 10 differ from those at localities 1 to 8, and it is likely that they have
FIG 6-3  2D strain values from pods in the S. Sithean Mhor gneisses.
strong initial fabric leading to an overestimation of strain by a factor of about four.

Locality 1, where the quartzo-feldspathic gneisses show isotropic fabrics, give a value of $R = 2.25$ which is an initial fabric trending $024^\circ$. This initial fabric has been reoriented to around $139^\circ$ in the localities to the NE, and the effects of such an initial fabric has been minimised by the initial orientation, tending to lead to a slight underestimation of strain.

Results show that different types of pods may have different initial fabrics and thus care must be taken in comparing the results.

6.3. Modelling the Strain Path

6.3.1. Introduction.

Field evidence shows that deformation increases to the NE across the homogeneously deformed area of the S. Sithean Mhor gneisses. This includes the decrease in fold interlimb angles from approximately $90^\circ$ in the SW to isoclinal in the NE (3.3.3.) and the increasing penetrative foliation in the dykes to the NE (3.3.3.).

The strain analysis shows that strain increases along a $k = 1$ path, then decreases again with a tendency to lie in the $k > 1$ field. Thus strains have combined in such a way that initial strains are decreased or 'undone' by a later deformation.

The S. Sithean Mhor gneisses offer less information than those at Loch Tollie to aid the interpretation of the strain data, there being no major structure to parallel the Tollie antiform. However the presence of numerous minor shear zones, parallel to the general trend of foliation, in the SW of the area, suggests that simple shear plays some part in the deformation.

In the following sections the feasibility of obtaining the observed strain data using progressive deformation to the NE by irrotational and
and rotational strain is investigated and discussed.

6.3.2.(1). Rotation strain model.

It is concluded in chapter 5 that the variation in finite strain in the Loch Tollie gneisses is best explained in terms of progressive simple shear superimposed on an initial fabric represented by the NE limb of the Tollie antiform. The gneisses therefore form the NE margin of a large shear zone trending NW/SE with a steep NE dip, moderately NW plunging shear direction and a sense that is SW side down (negative).

The steep NE dip of foliation continues across the whole of the Gairloch schist belt and the area of homogeneous deformation in the S. Sithean Mhor gneisses. The orientation of shear plane used in the model for the Loch Tollie gneisses, $75^\circ$, $040^\circ$, is therefore also applicable here.

A possible interpretation of the fabrics is that they represent a low deformation area within the shear belt across which deformation is constant apart from isolated occurrences of which specimen 3 would be an example. However, the variation in fabrics cannot be explained by a single phase of progressive simple shear for, although the fabrics are constant and close to $k = 1$ in type, their maximum axes orientations show an angular range within the X-Y plane of $104^\circ$. Fabrics produced by a single phase of simple shear should produce similar maximum axis orientations for similar intensities of fabric. The variation in fabric orientation must therefore be a product of superposition of strains in such a way as to produce close to $k = 1$ fabrics with a range of maximum axes orientation.

In the area of inhomogeneous deformation the numerous minor shear zones all indicate a subvertical shear direction with a sense that is SW side up (positive) (3.4.2.). Configurations of folds preserved in the SW of the area of homogeneous deformation also indicate a SW side
One method of effecting a decrease in the intensity of strain is to reverse the sense of shear while preserving the shear plane orientation and shear direction. This will preserve the NW trend and steep NE dip of the foliation in the gneisses while the foliation in the dykes will increase in intensity irrespective of the sense of shear, since it depends on the alignment of hornblende laths to the shear plane.

If superimposed deformation has the effect of 'undoing' the original finite strain then the highest strain observed represents the positive sense of shear and is the most probable initial fabric. Of all specimens analysed, only specimen 3 shows high strain and a NW dipping maximum. However, it is thought that it is a true representative of the strain due to lineations recorded in this area (3.3.3.). Strictly, this fabric is only an initial fabric for the immediately adjacent area since the finite strain need not necessarily have been constant throughout the area. Thus initial fabrics for the rest of the area are unknown. However, specimen 3 is used as a first approximation to the initial fabric.

The foliation plane of specimens 2, 3, 4 and 5 trend slightly more northerly than the shear plane so that the maximum axis of specimen 3 plots on the NE side of it (see Fig 6-4(a)). This would be expected from the initial development of the fabric from a positive sense of shear.

The initial fabric represented by specimen 3 has been theoretically deformed by simple shear using an unpublished programme called "Sust" written by D.J. Sanderson (Belfast). Nine directions of shear listed in Table 6-2 have been applied to the model using a negative sense of shear and the results are presented in Fig 6-4(a) and (b). Fig 6-4(a) shows that strains of $\gamma = 8$ produce only small movements of the principal
Table 6-2. Shear directions of rotational strain model.

<table>
<thead>
<tr>
<th>Number</th>
<th>Plunge and direction of plunge</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>11° 313°</td>
</tr>
<tr>
<td>2</td>
<td>38° 322</td>
</tr>
<tr>
<td>3</td>
<td>55° 333</td>
</tr>
<tr>
<td>4</td>
<td>64° 342</td>
</tr>
<tr>
<td>5</td>
<td>71° 001</td>
</tr>
<tr>
<td>6</td>
<td>73° 069</td>
</tr>
<tr>
<td>7</td>
<td>52° 110</td>
</tr>
<tr>
<td>8</td>
<td>29° 121</td>
</tr>
<tr>
<td>9</td>
<td>14° 125</td>
</tr>
</tbody>
</table>
FIG 6-4(a) Rotational strain model: paths described by maximum and minimum axes.

Initial maximum axes orientation
Initial minimum axes orientation

Specimen 3

Position of axes after increments of shear: 10, 30, 40, 60, 80.
FIG 6-4 (b) Rotational strain model:
Logarithmic deformation plot of strain paths.

- Initial fabric (specimen 3)
- Strain path with increments of strain: 0.2, 0.5, 1.0, 1.5, 2.0, 2.5,
  3.0, 4.0, 6.0, 8.0.
axes of strain and that in all cases the maximum axes rotate towards the SE. This occurs because the maximum and intermediate axes of strain lie very close to the shear plane. In directions 4 and 5, the maximum axes follow paths which cross over the shear plane itself.

Fig 6-4(b) shows the paths taken by the fabric for the shear directions on a logarithmic deformation plot. Only directions 3, 4 and 5 produce appreciable reductions in the finite strain with 3 and 4 staying in the $k > 1$ field and 5 in the $k < 1$ field, see Fig 6-5(b). These directions produce NW plunging maximum axes, see Fig 6-5(a). This directions 3, 4 and 5 fit the strain best with direction 5 showing the best fit to fabric types and intensities, Fig 6-5(b).

Because the X-Y plane of the initial fabric lies close to the shear plane and the maximum axis close to the shear plane, small variations in the orientation of the initial fabric will have a large effect on the strain paths produced. To illustrate this, six variations in the orientation of the maximum axis were applied to the model for shear direction 5 and the results are plotted in Fig 6-6(a) and (b).

Examination of Fig 6-6 shows that initial fabrics whose X-Y planes lies close to the shear plane and whose maximum axes close to the shear direction, (directions 1, 2 and 4) move slowly and require large amounts of shear strain to reduce the finite strain. Those further from the shear plane (directions 3, 5, 6) move relatively rapidly and require less shear to reduce the finite strain. Also the further an initial fabric lies from the shear plane, the further the resulting X-Y planes depart from the shear plane orientation.

Direction 5, the furthest from the shear plane, produces X-Y planes that trend N rather than NW which does not fit the strain data (Fig 6-2(a)). The X-Y planes of the original fabric must have been within $10^0$ to $15^0$ of the shear plane to maintain their NW/SE trend.
FIG 6-5(a) Rotational strain model: comparison with principal strain axes orientation of strain data.

- Maximum (X) axis
- Minimum (Z) axis

Strain path with increments of shear: 0.2, 0.5, 1.0, 1.5, 2.0, 2.5, 3.0, 4.0, 6.0, 8.0.
FIG 6-5(b) Rotational strain model: comparison with strain data on logarithmic deformation plot.

Strain data from quartz-feldspathic gneisses.

Strain path with increments of shear: 0·2, 0·5, 1·0, 1·5, 2·0, 2·5, 3·0, 4·0, 6·0, 8·0.
FIG 6-6(a) Rotational strain model: paths described by maximum and minimum axes for variations in maximum axis orientation.

- Initial maximum axes orientation.
- Initial minimum axes orientation.
- Position of axes after increments of shear: 0.5, 1.0, 1.5, 2.0, 2.5, 3.0, 4.0, 6.0, 8.0.
FIG 6-6(b) Rotational strain model: logarithmic deformation plot of strain paths for variations in maximum axis orientation.

Initial fabric (specimen 3)

Strain paths with increments of shear: 0.2, 0.5, 1.0, 1.5, 2.0, 2.5, 3.0, 4.0, 6.0, 8.0.
Directions 1, 3 and 7 are sufficiently far from the shear plane to produce results fitting the strain data after strains of approximately $\gamma = 6$ to 8, 3 and 7 producing SE plunging, and 1 a NW plunging maximum axis. Thus, even small variations in the maximum axis orientation can have marked effects on the strain paths when the X-Y plane of the initial fabric lies close to the shear plane and maximum axis close to the shear direction.

Note that maximum axes either move rapidly through the horizontal to become SE plunging or move slowly towards the shear direction. Thus after strains greater than $\gamma = 6$, the distribution of maximum axes will tend to concentrate in two directions, one moderately NW plunging and one gently to moderately SE plunging as shown by the strain data Fig 6-2(a).

Thus from a comparison of the model and the strain data it is probable that initial fabrics had a narrow range of X-Y plane orientations within $10^0$ to $15^0$ of the shear plane with a NW plunging maximum axes lying on the NE side of the shear plane. On shearing in a negative sense (SW side down) the initial fabric is effectively undone and the direction of rotation of the maximum axis depends on its initial orientation. Those plunging more shallowly NW than specimen 3 produce SE plunging maximum axes and those plunging more steeply maintaining a NW plunge. Thus after a constant amount of shear strain, fabrics lie close to $k = 1$ with a tendency to lie in the $k > 1$ field and major axes orientation either moderately NW plunging or gently to moderately plunging SE.

6.3.2.(2). Irrotational strain model.

As in the previous model using rotational strains, to fit the strain data strains must be superimposed to produce close to $k = 1$ fabrics with varying maximum axes orientation.
Specimen 3 is the most likely initial fabric for the same reasons as those given in 6.3.2.(1). With specimen 3 as an initial fabric the only way to decrease the strain by irrotational strain is to apply a plane strain with the X-Y plane trending NE-SW, dipping moderately SE. This orientation of strain would tend to produce folds in the gneissose layering and dykes with axial planes trending NW/SE and a parallel foliation in the dykes. No field evidence for these features exists. To agree with the trend of foliation in dykes and gneisses any irrotational plane strain must be orientated with the X-Y plane parallel to the existing foliation, i.e. steeply dipping NE. Any orientation of maximum axis within this plane will lead to an increase in strain or to markedly planar fabrics and cannot 'undo' the original strain. Thus it is not possible to construct an irrotational plane strain model that agrees with both the strain data and the field evidence.

6.4. Conclusions and Discussion

6.4.1. Summary and conclusions from the strain models.
It is shown in 6.3.2. that the strain data and field evidence are compatible with a simple shear model and cannot be explained by any superposition of irrotational strain. Field evidence and the results of the model indicate that there have been two phases of shearing with similar orientations of shear plane and shear direction but opposite sense. This has led to the development of an area of constant strain and range of maximum axis orientations in the NE of the S. Sithean Mhor gneisses. Implications and further aspects of the rotational strain model are discussed in the following sections.

6.4.2. Time relations of phases of shearing.
The order of the two phases of shearing is ambiguous from the fabrics in the gneisses alone but minor features within the dykes
helps to resolve the problem.

The dykes are closely associated with minor shear zones in the zone of inhomogeneous deformation and it is concluded in 3.4.3. that they are syntectonic with this phase of positive shearing. Inclusions of foliated gneiss in largely unfoliated dykes in the SW of the area of homogeneous deformation leads to the same conclusion (3.4.3.).

Lineations within the dykes mimic the lineation orientations within the quartzo-feldspathic gneisses and these parallel the strain ellipsoid maximum axes (see 6.4.3.). Thus, the lineations in the dykes reflect the total finite strain and the dykes themselves must therefore have been subjected to both phases of shearing. If the dykes were intruded during the later phase of shearing, lineations within them would reflect only the effect of one sense of shearing and should approximate the shear direction. In this case lineations should be constant throughout the zone and not parallel to the finite strain maximum axes. Thus, it is more probable that the positive phase preceeded the negative phase of shearing. The positive phase extended further to the SW within the S. Sithean Mhor gneisses and is the deformation represented in the area of inhomogeneous deformation.

It is probable that the phase of negative shearing did not extend as far to the SW because it exploited the deformed zone created by the previous phase of shearing; foliated rocks being easier to deform than unfoliated. Thus, the area of inhomogeneous deformation and the most south-westerly part of the area of homogeneous deformation were left undisturbed throughout the remainder of the active shearing.

6.4.3. Foliation and lineation orientation.

1. Foliation orientation.

The foliations from the four sub-areas A to D marked on Fig 6-1 are plotted in Fig 6-7 and shows that foliation is very constant throughout the area with a tendency to become steeper to the SW.
FIG 6-7 Foliation orientation in the S. Sithean Mhor gneisses.

Poles to foliation:
- □ in subarea A
- ○ in subarea B
- + in subarea C
- • in subarea D
Foliation in the NE is subparallel to the model shear plane, areas C and D, and becomes subvertical to steeply SW dipping in the SW, areas A and B.

Theoretically, the first fabrics that develop from a structurally isotropic terrain for a positive sense of shear should dip approximately 30° NE and for a negative sense of shear, approximately 60° NE. In practice, neither of these first formed fabrics are seen. In the case of the negative sense of shear the deformation is superimposed on a previous fabric which is already close to the shear plane orientation. The result is steep NE dipping fabrics throughout the zone affected by the negative sense of shear. For the positive sense of shear, the start of the zone of homogeneous deformation is abrupt and low intensities of fabric are seldom seen since the deformation progresses by a transition from concentration in minor shear zones to more wide spread quartz fabrics (3.3.1.). However, the dip of foliation here is generally steep SW, instead of the NE dips expected in theory. The reason for this may lie in the lithological layering of the area.

The most prominent layering is composed of alternating dykes and quartzo-feldspathic gneisses and trends NW. On the SW margin of the area of homogeneous deformation around 808704, dykes are abundant and their margins are subvertical or dip steeply SW. The foliation in the surrounding gneisses is parallel to the dyke margins. Field evidence shows that the dykes are less deformed than the gneisses in the area and show evidence of controlling the locality of minor shear zones (3.4.3.). It seems probable, therefore, that they would control the orientation of the foliation in the adjacent quartzo-feldspathic gneisses, especially where dykes are abundant, in the early stages of the deformation. As foliation in the dykes becomes penetrative and deformation stronger the dykes will rotate towards the shear plane orientation and have less effect on the foliation orientation in the surrounding gneisses.
Thus dykes have probably controlled the orientation of foliation in the SW of the area constraining the dip to subvertical or steep SW.

2. Lineations.

The plunge of lineation is recorded in Fig 6-8 for subareas A to D (see Fig 6-1) and progress from moderately plunging NW through subvertical to moderately to gently plunging SW (3.3.3). The quartzofeldspathic gneisses of the area are relatively coarse grained compared with the Loch Tollie gneisses and the lineation measured is the preferred orientation of the quartz fabric, ie. the maximum axis of the strain ellipsoid. These then agree with the general pattern of major axes orientation predicted by the simple shear model. Note that in subarea A most lineations plunge NW and lie on the SW side of the shear plane. These lineations come from the area immediately to the SW of specimen 3, where the dykes are thought to have controlled the foliation orientation and thus the directions of lineations. Specimen 3 comes from the area immediately to the NE where dykes are less abundant and here the foliation orientation is steep NE. Thus in subareas B, C and D lineations plot close to the shear plane.

6.4.4. Construction of a strain profile across the S. Sithean Mhor gneisses.

It has been shown in section 6.3.2. that the range in fabric type and orientation with field evidence is best explained in terms of progressive simple shear with a steeply NE dipping shear plane and moderately NW plunging shear direction. The shear sense was initially positive, indicated by minor shear zones and fold geometry. Superimposed on this is a phase of negative sense of shearing with a similar NW plunging shear direction which has decreased the finite strain of the previous deformation.

Due to small variations in shear direction and fabric orientation
FIG 6-8 Lineation orientations in the S. Sithean Mhor gneisses.

Lineation plunges:
- □ in subarea A
- ○ in subarea B
- + in subarea C
- • in subarea D
it is unlikely that the fabric will travel exactly back along its previous strain path. Even variations in shear direction or fabric orientation of a few degrees can produce markedly different paths. The effect of this is that a greater amount of shear strain is required to 'undo', than was required to create, the initial fabric. This effect can be seen by comparing Fig 6-9, showing the path of an initial sphere, and Fig 6-6. Fig 6-9 implies a shear strain of approximately 5.0 for the development of the strain in specimen 3. Comparison of the strain data and model paths in Fig 6-6 implies a shear strain of approximately 7.0. The total shear strain is therefore approximately 2.0 in a negative sense.

In the model, specimen 3 has been used as an initial fabric though as discussed in section 6.3.2., this can only strictly apply to the rocks immediately adjacent or along strike from specimen 3. In fact it is likely from field evidence that shear strain of the earlier positive sense of shear increases progressively to the SW. The true original fabrics NE of specimen 3 cannot be deduced, and so, to a first approximation specimen 3 is assumed to represent an initial fabric for the whole area. This will lead to an underestimation of the shear strains involved. The constancy of shear strain throughout the area implies that, though the shear sense changed direction, the strain gradient remained constant. A strain profile for the S. Sithean Mhor gneisses is given in Fig 6-10 showing total shear strain within the zone of homogeneous deformation up to the contact with the Gairloch schist belt. There is no information on the finite strain in the area between specimen 6 and close to the schist contact due to outcrop of Torridonian cover and lack of exposure. It is assumed that the strain profile continues across this area at the level of \( \gamma = 2.0 \). This assumption does not appear unreasonable since the gneisses exposed to the SE between Mill na Claise, 833717, and 826704 have constant
FIG 6-9(a) Rotational strain model: path described by maximum axis of an initial sphere during positive simple shear.

- Maximum (X) axis
- Intermediate (Y) axis
- Minimum (Z) axis

Initial fabric.

Strain path with increments of shear: 0, 0.5, 1.5, 2.0, 2.5, 3.0, 4.0, 6.0, 8.0.
FIG 6-9(b) Rotational strain model: logarithmic deformation plot of strain path developed from an initial sphere during positive simple shear.

- Strain path
- ♦️ Quartzo-feldspathic gneiss specimen 3.
FIG 6-10 Strain profile across the S. Sithean Mhor gneisses.

area of homogeneous deformation in gneisses.

Gairloch schist belt.

quartz-feldspathic gneiss specimens.

1 km.

SW

NE
strength of quartz fabric throughout.

The contact with the Gairloch schists is more fully described and discussed in chapter 8. Here fabrics in the gneisses increase markedly in intensity within approximately 1 m of the contact and this together with evidence in the schists (see chapter 8) shows that the shear strain increases markedly at the gneiss/schist contact. This is indicated on the strain profile Fig 6-10.
7.1. Introduction

The Gairloch schists form a NW/SE trending belt some 3 kms wide composed principally of biotite and hornblende schists in approximately equal proportions, see Fig 7-1. Smaller quantities of other rock types (marble, quartz-magnetite schists, garnetiferous amphibolite and graphite bearing schist) occur as lenses within the hornblende and biotite schists. The Ard Ialltaig gneisses form a lenticular outcrop on the SW margin of the schist belt and NW/SE trending belt of augen gneisses, the An Ard gneisses, occurs within the schists. The rocks are foliated throughout with a steep NE dip to foliation and lithological layering, and variable NW or SE plunging lineations.

The topography of the area is largely geologically controlled with gneisses and hornblende schists forming ridges and the biotite schists low lying, poorly exposed ground. Numerous small pockets of Torridonian sedimentary rocks found throughout the area show the present topography to be similar to that of the pre-Torridonian surface.

The area has been mapped by Clough et al (1908) as part of sheet 92, Geological Survey of Scotland, and more recently by Park (1964). A description of the rock types is given by Clough et al (1908) and a structural interpretation by Park (1964). Eight age determinations are recorded by Moorbatch and Park (1971) and indicate an age of 1406 ± 30 M.a. for a biotite schist, interpreted as a Laxfordian metamorphic date, and a maximum age of 2213 ± 80 M.a. for seven specimens from the Aundrary amphibolite, interpreted as Laxfordian overprint on an original Inverian age of intrusion and amphibolitization. Geochemical studies have been carried out by Park (1966) and Winchester,
FIG 7-1 Geological map of the Gairloch schist belt adapted from Park (1964).

KEY

- Loch Tollie and S. Silhean Mhor gneisses.
- An Ard augen gneiss.
- Ard Ialthaig gneiss.
- Hornblende schist.
- Biotite schist.

- Torridonian cover.

- Marble
- Cummingtonite-hornblende-biotite schist.
- Quartz-magnetite schist.
- Garnetiferous amphibolite
- Graphite schist

→ Lineations plunging 30°55' after Park (1964)
→ Lineations plunging 60°85'

SCALE

0 500 1000 metres
Park and Holland (in press) on the hornblende and biotite schists.

The area south of the Flowerdale fault, only, is represented in Fig 7-1, since north of the fault the rocks are largely hornblende schists with the biotite schists being very poorly exposed. Exposure is better to the south of the fault and it is here that exposures of other lithologies (marble, quartz-magnetite schists, garnetiferous amphibolite and graphite-bearing schists) are found. The following description and discussion is therefore largely based on this area.

7.2. The Gairloch Schists

7.2.1. The Lithologies.

1. Schists of semipelitic composition.

Schists of semipelitic composition occupy some 50% of the Gairloch schist belt, see Fig 7-1. Their mineralogy (Table 7-1) is constant throughout and consists largely of quartz, plagioclase (An₁₅), biotite, + muscovite. The mica content is variable and gives rise to a layering composed of mica rich and mica poor lithologies, generally parallel to the foliation, ranging from 1 mm to 2 m in width. Sulphides, largely pyrite, are normally a minor constituent but are locally very abundant causing the schists to weather badly and become rusty brown in colour.

Grain size ranges from 0.5 mm to ~0.1 mm in quartz and feldspar and from 1 cm to 0.2 mm long in the micas. All schists show recrystallized textures with a tendency towards polygonal grain boundaries in the quartz and feldspar components. Mica grains show parallel alignment, generally parallel to the layering, which together with a similar alignment of elongate quartz and feldspar grains gives rise to a marked foliation. The proportion of mica is not great enough to form a true cleavage and the schists have a flaggy parting with quartz-feldspar rich layers tending towards a flinty fracture.
Table 7-1. Mineral composition of lithologies in the schist belt.

<table>
<thead>
<tr>
<th>Lithology</th>
<th>Major minerals</th>
<th>Accessories (&lt;1%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. biotite schist</td>
<td>plagioclase (An$_{15}$), quartz, biotite, + muscovite</td>
<td>epidote, opaque, zircon, + chlorite, + carbonate, + epidote</td>
</tr>
<tr>
<td>2a. marble</td>
<td>ankeritic dolomite, Mg-, Fe-bearing calcite, quartz</td>
<td>epidote, chlorite, white mica, plagioclase, biotite</td>
</tr>
<tr>
<td>2b. foliae in marble</td>
<td>hornblende, biotite, chlorite</td>
<td>quartz, plagioclase, carbonate, opaque, muscovite, epidote, zircon</td>
</tr>
<tr>
<td>3. cummingtonite-hornblende-biotite schist</td>
<td>cummingtonite, + hornblende, plagioclase (An$_{30}$) biotite, quartz</td>
<td>opaques, carbonate, zircon</td>
</tr>
<tr>
<td>4. quartz-magnetite schist</td>
<td>quartz, magnetite</td>
<td>cummingtonite, biotite, + stilpnomelane, + garnet, + chlorite, + carbonate</td>
</tr>
<tr>
<td>5. garnetiferous amphibolite</td>
<td>cummingtonite, garnet</td>
<td>quartz, plagioclase, + biotite, + chlorite, + stilpnomelane</td>
</tr>
<tr>
<td>6. graphite schist</td>
<td>biotite, muscovite quartz, plagioclase, graphite</td>
<td>rutile, pyrite, zircon</td>
</tr>
<tr>
<td>7. hornblende schist</td>
<td>hornblende, plagioclase + biotite, + chlorite</td>
<td>opaques, quartz, epidote, + carbonate</td>
</tr>
</tbody>
</table>
All freshly broken surfaces are dark in appearance due to the fine grain size.

The schists show similar compositions and textures throughout the outcrop of the Gairloch schist belt but on the basis of textural differences can be divided into two groups which are separated by an amphibolite sheet from 810750 to 847708.

a. The SW group of schists.

The schists of this group contain augen of quartz and plagioclase from 2 to 5 mm across, which are detectable by the naked eye on close inspection. The augen are of single grains or composed of aggregates of a small number of grains. In some cases the plagioclase augen show truncated twinning, see Plate 7-1. Foliation in the matrix flows around the augen and many are surrounded by a recrystallised margin of smaller grains.

The grain size of the matrix ranges from 0.5 mm to ∼0.1 mm across in quartz and feldspar and from 1 mm to 0.2 mm long in mica. The coarsest grain sizes occur close to the contact with the gneisses to the SW and in general grain sizes decreases to the NE. This group of schists shows a range of mica content from ∼10% to 30% and they tend to be more finely layered (∼1 mm to 1 cm) than the NE group of schists.

The contact between this group of schists and the gneisses is found well exposed at An Ard, 802754, Druim Ruadh, 823727, Mill na Claise, 838174, and in a road cutting at 832721 (Park, 1964). At An Ard (802754) and at 838714, biotite gneisses grade into biotite schists over a distance of approximately 100 m by a decrease in grain size and the disappearance of k-feldspar. At Mill na Claise, 838714, and at Druim Ruadh, 823727, biotite schists butt sharply against strongly foliated quartzo-feldspathic gneisses. The gneiss-schist contact at An Ard is described in detail in chapter 8.
Plate 7-1. Plagioclase porphyroclast in schist showing truncated twinning.

Plate 7-2. Layering in biotite schist at 826741, showing biotite-rich (B) and plagioclase-rich (P) layers. Note sublayering within biotite layers.
b. The NE group of schists.

This group is composed of interlayered mica-rich and mica-poor schists, individual layers ranging 1 cm to 2 m in width with sharp to gradational contacts, the mica-poor schists tending to form ridges on weathered surfaces. The percentage of mica does not show as large a range as in the schists to the SW. Grain size varies within the range from 0.1 to 0.2 mm across in quartz and feldspar and 0.5 to 0.6 long in the micas. Layering is mostly parallel to the foliation but in one locality, 827742, layering was observed at a high angle to the foliation in the noses of isoclinal folds, 1 mm in width, see Plate 7-2. The layering is distinct and regular; biotite rich layers alternating with plagioclase-chlorite layers. Within many of the biotite rich layers a sub-layering of biotite concentration is observed.

The NE group of schists also differ from the SW group in containing pods and discontinuous layers of a range of lithologies which include hornblende-cummingtonite-biotite schist, marble, garnetiferous amphibolite, quartz-magnetite schist, graphite-bearing schist and others. These rock types are described below.

2. Cummingtonite-hornblende-biotite schists.

These schists occur as discontinuous layers and pods within the NE group of schists where they make up 1-2% of the outcrop, see Fig 7-1. They range from ½ to 10 m wide and from ~2 m to several tens of metres long elongate parallel to the foliation with sharp contacts against the surrounding biotite schists. Their mineralogy is listed in Table 7-1. The grain size tends to be coarser than the surrounding biotite schists, with quartz and plagioclase around 0.2 mm and biotite and amphibole up to 1 mm long. They are characterised by partially to wholly recrystallised euheral laths of andesine plagioclase (An35) up to 1½ mm long, sometimes still showing
twinning (Carlsbad and Pericline), often oriented at high angles to the foliation described by biotite and amphibole grains, see Plate 7-3. Cummingtonite laths have ragged outlines with numerous inclusions and are altered in irregular patches to hornblende. There is a correlation between preservation of plagioclase porphyroblasts and amphibole type; those rocks with better preserved plagioclase porphyroblasts, commonly at high angles to the foliation also contain largely cummingtonite. In hornblende rich types, the plagioclase porphyroblasts are largely recrystallised and subparallel to the foliation.

3. Marbles and associated rock types.

Marbles occur as discontinuous layers and lenses, parallel to the foliation within the NE group of schists and are well exposed at 823739, 837728, 839743, 837725, 845719, see Fig 7-1. The marble is closely associated with amphibolite and commonly contains numerous foliae of hornblende, biotite-chlorite schist and quartz. These foliae and the associated amphibolites show a coarser grain size than that in the surrounding biotite schists with hornblende laths commonly 2-3 mm long. The marble itself is composed of two phases of carbonate, ankeritic dolomite and Mg-, Fe-bearing calcite, see Table 7-2, which can readily be distinguished on the weathered surface, the former weathering a rusty brown and the latter a cream colour. Contacts between the two are sharp and they are interlayered parallel to the foliation described by foliae. Proportion of silicate foliae to carbonate varies greatly from approximately 5% to 90%. Individual foliae range from ~2 mm to ~20 cms on average 1 to 2 cms wide. Locally sulphides and talc are found in association with the marbles, eg. at 837725.

The marbles occur in association with biotite-hornblende schists containing elongate rods of quartz at 823739 and 837728. These quartz rods are elliptical in section and commonly 2 to 5 cms across by
Plate 7-3. Partially recrystallized plagioclase porphyroblast showing twinning, Carlsbad (C) and Pericline (P), in cummingtonite-hornblende-biotite schist at 826741.
Plate 7-4. Marble showing interlayering of carbonate (C) and amphibolite (A) foliae, at 823739.
Table 7-2. Analyses of carbonates recalculated from E.D.S. data.

<table>
<thead>
<tr>
<th></th>
<th>Mg-, Fe-bearing calcite</th>
<th>Ankeritic dolomite</th>
<th>Ankeritic dolomite</th>
</tr>
</thead>
<tbody>
<tr>
<td>% Mg CO₃</td>
<td>2.07</td>
<td>24.25</td>
<td>15.26</td>
</tr>
<tr>
<td>% Ca CO₃</td>
<td>94.64</td>
<td>58.60</td>
<td>61.66</td>
</tr>
<tr>
<td>% Fe CO₃</td>
<td>3.28</td>
<td>17.14</td>
<td>23.07</td>
</tr>
<tr>
<td><strong>TOTAL</strong></td>
<td>100</td>
<td>99.99</td>
<td>99.99</td>
</tr>
</tbody>
</table>
several cms long with marked parallel alignment. Marbles are also found in association with the quartz-magnetite schists described below.

4. Quartz-magnetite schists.

Exposures of quartz-magnetite schists are found at 829727, 837728, 837728, 839743, 846713 and 844713, see Fig 7-1. They occur in discontinuous layers up to 10 m across, on average 1.5 m across, by several tens of metres long and at 839743 a layer can be traced for \( \sim 100 \text{ m} \) along strike. Fig 7-2 shows a sketch map of the locality at 837728.

The schists are composed of alternating quartz-magnetite and dominantly quartz layers ranging in width from less than 1 mm to \( \sim 1 \text{ cm} \) wide, see Plate 7-5. The overall proportion of magnetite ranges from \( \sim 25\% \) to less than 1\%, magnetite-rich rocks passing laterally into quartz schists.

The grain size of these schists is generally coarser than that of the surrounding biotite schists, quartz ranging from \( \sim 0.1 \text{ mm} \) to 1 mm on average 0.5 mm, and tending to be coarser in quartz rich layers.

The quartz-magnetite schists are often found interlayered with, and forming lenses within, the garnetiferous amphibolites (described below). In these cases cummingtonite, biotite and garnet are often found within the quartz-magnetite schists. They are also found in association with marble, eg. at 837728, 839743, and 837725 and with coarse grained amphibolites occasionally containing magnetite porphyroblasts, eg. at 829727.

The major chemical analysis of three typical specimens of quartz-magnetite schist are listed in Table 7-3.

5. Garnetiferous amphibolites.

Garnetiferous amphibolites are well exposed at 829727, 837728 and 846713 and form elongate lenses parallel to the foliation, often
FIG 7-2 Sketch map of the quartz-magnetite schist locality at 837728.
Plate 7-5. Quartz-magnetite schist boulder showing alternating magnetite-rich (M) and quartz-rich layers (Q) at 837728.
Table 7-3. Major chemical analyses of quartz-magnetite schists and banded ironstones.

<table>
<thead>
<tr>
<th></th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>A</th>
<th>B</th>
<th>C</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>46.22</td>
<td>56.29</td>
<td>44.08</td>
<td>35.86</td>
<td>51.66</td>
<td>56.23</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.03</td>
<td>0.03</td>
<td>0.26</td>
<td>0.040</td>
<td>-</td>
<td>0.02</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>0.51</td>
<td>0.48</td>
<td>4.02</td>
<td>1.57</td>
<td>0.08</td>
<td>0.45</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>50.06</td>
<td>40.46</td>
<td>46.96</td>
<td>58.82</td>
<td>45.61</td>
<td>40.63</td>
</tr>
<tr>
<td>MnO</td>
<td>0.13</td>
<td>0.25</td>
<td>0.28</td>
<td>0.16</td>
<td>-</td>
<td>0.07</td>
</tr>
<tr>
<td>MgO</td>
<td>0.43</td>
<td>1.90</td>
<td>2.00</td>
<td>1.74</td>
<td>0.20</td>
<td>1.13</td>
</tr>
<tr>
<td>CaO</td>
<td>1.10</td>
<td>0.91</td>
<td>0.68</td>
<td>0.51</td>
<td>0.02</td>
<td>0.81</td>
</tr>
<tr>
<td>Na₂O</td>
<td>0.11</td>
<td>0.12</td>
<td>0.12</td>
<td>0.02</td>
<td>-</td>
<td>0.15</td>
</tr>
<tr>
<td>K₂O</td>
<td>0.01</td>
<td>0.02</td>
<td>0.12</td>
<td>0.14</td>
<td>0.06</td>
<td>0.05</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.79</td>
<td>0.49</td>
<td>0.15</td>
<td>0.86</td>
<td>97.63</td>
<td>99.66</td>
</tr>
<tr>
<td>Total</td>
<td>99.35</td>
<td>100.95</td>
<td>98.67</td>
<td>98.86</td>
<td>97.63</td>
<td>99.66</td>
</tr>
</tbody>
</table>

1 Quartz-magnetite schist, 839723
2 Quartz-magnetite schist, 829727) Gairloch
3 Quartz-magnetite schist, 846713

A Banded quartz-magnetite rock with minor chlorite, siderite, cummingtonite. Saksagen, Krivoi Rog Series, Ukraine, USSR.
B Banded magnetite chert from the Lower Cherty division of the Biwabuk Iron-Formation, Mesabi district, Minnesota, USA.
C Specularite-magnetite Iron-Formation, Atlantic City district, Wyoming, USA.

Analysis A, B and C from James (1966).
in association with quartz-magnetite schists with which they show sharp contacts, see Fig 7-1. They form the bulk of the outcrop at 829727 and 846713 whereas at 837728 they are dominated by quartz-magnetite schists. Fig 7-3 shows a sketch map of the locality of 829727.

Garnet and cummingtonite amphibole constitute the bulk of the rock and garnet generally forms 25-30% but can be locally very abundant, eg. ~ 80% at 846713. They are manganese rich, see table 7-4, and are dark red in colour, ranging in size from 1.5 cms across downwards, with the majority between 2 and 5 mm. Larger garnets tend to be sub- or anhedral whereas smaller ones show near euhedral shapes. Inclusions of all other minerals in the rock are abundant see Plate 7-6, and occasionally depict S-shaped trails within the larger garnets.

The amphibole is manganan cummingtonite, see Table 7-4, close in composition to the cummingtonite-grunerite divide, grain size approximately 1 mm locally up to 5 mm, which in the field weathers to give a black crust of manganese oxide. It tends to form radiating aggregates of grains within the plane of the foliation. Quartz, plagioclase and biotite form minor constituents and tend to be coarser grained than in the adjacent biotite schists, 0.5 mm or less.

The garnetiferous amphibolites show well developed foliation that wraps around the garnets and is parallel to the foliation in the adjacent schists. Contacts are rarely exposed but at 829727, at the margins of the outcrop garnetiferous amphibolites are interleaved with hornblende schists and there is a decrease in size and abundance of garnets, and grain size of the amphibole. Where interleaved with garnetiferous amphibolites, the biotite schists often also contain large garnets.

The major chemical analysis of three specimens of garnetiferous amphibolite are listed in Table 7-5.
FIG 7-3 Garnetiferous amphibolite locality at 829727.

KEY

Outcrop

- Garnetiferous amphibolite
- Quartz-magnetite schist
- Hornblende schist
- Biotite schist

Exposure

--- Inferred outcrop boundary

\:\:\:\:\\ Steep slope

50 m.
Plate 7-6. Garnetiferous amphibolite. G:Mn-rich garnet, C:Mn-cumming-
tonite, I:ilmenite, Q:quartz.
Table 7.4. Analyses of garnets and amphibole from garnetiferous amphibolites.

<table>
<thead>
<tr>
<th></th>
<th>Cummingtonite</th>
<th></th>
<th>Garnet</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>1</td>
<td>2</td>
<td>3</td>
<td>1</td>
</tr>
<tr>
<td>Na(_2)O</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>MgO</td>
<td>15.32</td>
<td>15.99</td>
<td>15.90</td>
<td>0.79</td>
</tr>
<tr>
<td>Al(_2)O(_3)</td>
<td>0.00</td>
<td>0.17</td>
<td>0.18</td>
<td>19.44</td>
</tr>
<tr>
<td>SiO(_2)</td>
<td>53.48</td>
<td>53.31</td>
<td>52.85</td>
<td>36.51</td>
</tr>
<tr>
<td>K(_2)O</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>CaO</td>
<td>0.39</td>
<td>0.20</td>
<td>3.09</td>
<td>3.63</td>
</tr>
<tr>
<td>TiO(_2)</td>
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<td>0.00</td>
<td>0.00</td>
<td>0.21</td>
</tr>
<tr>
<td>MnO</td>
<td>3.42</td>
<td>3.19</td>
<td>3.17</td>
<td>13.80</td>
</tr>
<tr>
<td>Fe(_2)O(_3)</td>
<td>24.09</td>
<td>23.30</td>
<td>23.21</td>
<td>24.04</td>
</tr>
<tr>
<td>Total</td>
<td>96.70</td>
<td>96.16</td>
<td>98.40</td>
<td>98.42</td>
</tr>
</tbody>
</table>
Table 7-5. Major chemical analysis of garnetiferous amphibolites and manganiferous iron formations.

<table>
<thead>
<tr>
<th></th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>D</th>
<th>E</th>
<th>F</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>44.76</td>
<td>47.94</td>
<td>48.76</td>
<td>42.29</td>
<td>38.37</td>
<td>57.78</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.60</td>
<td>0.44</td>
<td>0.42</td>
<td>0.38</td>
<td>0.68</td>
<td>0.73</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>7.70</td>
<td>5.73</td>
<td>4.78</td>
<td>10.63</td>
<td>14.46</td>
<td>12.01</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>32.64</td>
<td>27.69</td>
<td>30.93</td>
<td>33.25</td>
<td>30.17</td>
<td>12.33</td>
</tr>
<tr>
<td>MnO</td>
<td>7.63</td>
<td>9.54</td>
<td>7.08</td>
<td>6.32</td>
<td>6.88</td>
<td>8.23</td>
</tr>
<tr>
<td>MgO</td>
<td>4.38</td>
<td>6.07</td>
<td>5.08</td>
<td>1.15</td>
<td>1.67</td>
<td>1.79</td>
</tr>
<tr>
<td>CaO</td>
<td>1.22</td>
<td>0.85</td>
<td>1.10</td>
<td>3.67</td>
<td>4.06</td>
<td>2.85</td>
</tr>
<tr>
<td>Na₂O</td>
<td>0.12</td>
<td>0.12</td>
<td>0.12</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>K₂O</td>
<td>0.13</td>
<td>0.00</td>
<td>0.00</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.21</td>
<td>0.19</td>
<td>0.41</td>
<td>1.93</td>
<td>2.11</td>
<td>1.16</td>
</tr>
<tr>
<td>Total</td>
<td>98.79</td>
<td>98.13</td>
<td>98.26</td>
<td>99.98</td>
<td>98.40</td>
<td>96.86</td>
</tr>
</tbody>
</table>

4 Garnet amphibolite (garnet, cummingtonite, opaques, minor quartz) 846714.
5 Garnet amphibolite (garnet, cummingtonite, opaques, quartz) 846713.
6 Garnetiferous amphibolite (garnet, cummingtonite, quartz, biotite) 846713.

D and E Manganiferous Ironstone from Broken Hill, Australia.
F Garnet-magnetite schist, Adelie land, Australia.

Analysis D, E and F from Richards (1966)
6. Graphite schists.

Graphite schists occur scattered throughout the biotite schists from 817744 to 827731, see Fig 7-4, and at 845713. They occur as discontinuous layers and lenses generally 2 m but locally up to 10 m wide by several tens of metres long, parallel to the foliation. Contacts with the biotite schists are gradational over 10 cms to 1 m. The freshly broken surface has a sooty appearance and, due to the abundance of sulphides, the weathered surfaces are a rusty brown colour.

The graphite schists tend to be richer in micas, 20% to 80% (biotite and muscovite), than the biotite schists. Graphite occurs disseminated throughout the micas and pyrite and rutile occur as irregular grains approximately 0.1 mm across, scattered throughout, with pyrite also occurring in irregular veins subparallel to the foliation. Quartz and feldspar have grain sizes of 0.1 mm to 0.5 mm whereas the micas range from 0.5 mm to large poikiloblasts approximately 1 cm across, and often show crenulation cleavage. Opaque content is rather variable ranging from ~20% where disseminated graphite renders thin sections almost opaque, to 5% in types most closely resembling the biotite schists.

7. Hornblende schists and amphibolites.

Layers of hornblende schist and amphibolite are abundant throughout the SE group of the schists but rarer in the NE group which contain only two major sheets, the Aundrathy amphibolite on the NE margin and a smaller sheet at 830731, see Fig 7-1. These layers range from 1.5 kms (Aundrathy amphibolite) down to a few tens of cms wide and all trend NW/SE with a steep NE dip, parallel to the foliation and lithological layering in the schists.

The basic mineral assemblage is hornblende, plagioclase, + minor quartz and biotite and in some of the finer grained hornblende schists, eg. at 807768, chlorite and epidote are also major constituents,
FIG 7-4 Sketch map of graphitic schist locality around 819742.

KEY

Outcrop

Graphitic schist.      Exposure.
Hornblende schist.    Outcrop boundary.
Biotite schist.        Steep slopes.

100 m.
see Table 7-1. Biotite is common in the finer grained layers.

Throughout the area the hornblende schists and amphibolites show a narrow range of the major mineral percentages but in a few localities the rocks are more feldspar-rich. At 768805 (the Little Sand Diorite Mass) and 791816 the amphibolite is \( \sim 50\% \) feldspar, and contains feldspar augen up to 1 cm across. All layers possess a foliation and lineation described by the alignment of hornblende laths and streaks of feldspar. Grain size shows a wide range within individual layers from \( \sim 0.1 \) mm (hornblende schist) to \( \sim 1 \) cm (amphibolite). This is clearly shown within the thickest layer, the Aundrarry amphibolite. Across Meall an Tuill-aoil, 851722, the grain size is constant throughout the sheet at around 1 mm except for lenses and discontinuous layers with gradational boundaries and grain sizes up to 1 cm, eg. at 850721, a lens some 20 by 10 m, and at 849721, a lens 25 x 6 cms, see Fig 7-5. These lenses occasionally preserve original igneous (gabbroic) textures. Similar lenses occur to the SW at 813743 and to the NE at 795795 and 798800.

Towards the NW of the outcrop of Lewisian, layers of hornblende schist increase in width and finally coalesce so that at 780809 the biotite schists have wedged out and outcrop is 100\% hornblende schist. Here the outcrop of Lewisian is restricted by the overlying Torridonian sediments but the only outcrops to the NW are also of amphibolite, the Little Sand Diorite Mass, and hornblende schists.

7.2.2. Minor structures in the schists.

Throughout their outcrop the schists possess a foliation formed by the parallel alignment of micas, hornblende and elongate quartz and plagioclase grains. This foliation is equally strongly developed throughout the biotite schists, but rather more variably developed in the hornblende schists, and is in general parallel to
FIG 7-5 Sketches of lenses of low deformation within the Aundrary amphibolite.

a) 850721.

b) 849721.
lithological layering on all scales. Folds, only occasionally seen, are isoclinal to which the foliation is axial planar. Very intricate folding on a small scale is associated with the outcrops of marble, quartz-magnetite schist and garnetiferous-amphibolite where layering shows strong crenulation and quartz viens ptygmatic folding. Hinges of folds plunge generally SE parallel to mineral lineations. The schists show distinct lineations of the following varieties:

a) alignment of hornblende laths and streaks of plagioclase in hornblende schists.
b) alignment of biotite flakes within the foliation plane in biotite schists.
c) a strong rodding lineation in quartz rich schists and a similar lineation on the surfaces of quartz viens, see Plate 7-7. These are parallel to the mineral lineations in adjacent hornblende and biotite schists.

Two sets of features suggest simple shear as the dominant type of deformation:

a. Isotropic lenses with relict igneous textures within the Aundrany amphibolite show development of fabric whose orientation sweeps into the regional orientation with increase in intensity. The sense of rotation and the steep SE plunging lineation indicate a negative (SW side down) sense of shear, see Fig 7-5.

b. En echelon sygmoidal lenses of quartz on subhorizontal erosion surfaces around 826743 in the biotite schists, see Plate 7-8, indicate a sinistral component of shear. With the moderate SE plunging lineation that occurs throughout this area, this implies a negative sense of shear.

Both sets of features a. and b. give a negative sense of shear which is in agreement with that indicated by the Tollie antiform.
Plate 7-7. Foliation surface in quartzose schist at 824733, showing strong rodding lineation. Subvertical surface looking NE.

Plate 7-8. En echelon sigmoidal quartz lenses in schists at 826743. Horizontal surface.
Quartz veins are common throughout the outcrop of biotite and hornblende schists. The majority are parallel to the foliation and range from 1 mm to 2 m, in width, most being in the range of 15 mm to 1 cm, and up to ~4 m long. Many are highly deformed and have a 'blackened' appearance caused by fine grain size and highly strained grains. These possess a strong rodding lineation on the vein surfaces and boudinage is common. Measurement of quartz vein thicknesses in a traverse from 844714 to 844727 across the SW and NE group of schists shows that quartz veins are evenly distributed throughout the schist belt and account for some 7% of the outcrop width, with a tendency for a greater concentration in the biotite than hornblende schists.

7.2.3. Discussion.

The origin of the schists has been presumed to be sedimentary since the time of the first mapping by the Geological Survey in 1907, because of the presence of lithologies such as marble, quartz-magnetite and graphite schists.

The layering of the quartz-magnetite schists closely resembles that found in Archean ironstones. Comparison of chemical analyses with those of banded ironstones from USA and USSR (Table 7-3) shows that they are very similar in composition, being composed largely of silica and iron. The garnetiferous amphibolites associated with the quartz-magnetite schists are richer in manganese, magnesium and aluminium and poorer in iron. Comparison of chemical analyses with those of pre-Cambrian manganiferous ironstones from Australia (see Table 7-4) shows similar manganese, iron and silica contents, and grunerite-cummingtonite-Mn garnet-magnetite assemblages are described from the Wabush Iron Formation, Labrador, Klein (1966), and N. Michigan, James (1955). It is likely therefore that the layered quartz-magnetite schists and garnetiferous amphibolites are metamorphosed equivalents.
of banded and manganiferous ironstones. A sedimentary origin is also 
most likely for the marbles and graphitic schists, due to their close simila-
ity in composition to known sedimentary rocks. The hornblende-biotite 
schists containing quartz rods associated with marbles do not show 
any regular arrangement of rods that would suggest that they originated 
as boudinaged quartz viens and it is possible that they represent 
deformed quartz pebbles in a conglomerate.

These lithologies of probable sedimentary origin are concentrated 
in a narrow band on the SW margin of the NE group of biotite schists which 
also contain, to the NE, pods of cummingtonite-hornblende-biotite 
schist. From their mineralogy these rocks are richer in Ca, Mg and 
Fe and poorer in alkalis and silica than the biotite schists. They 
preserve no original features to indicate their origin but their 
composition is intermediate and, due to a lack of metasediments of this 
composition, it is probable that they originated as an intermediate 
igneous rock type, eg. andesite, diorite or tuff.

The NE group of biotite schists in which the above lithologies 
occur are uniform in texture and composition throughout their outcrop. 
They show no relict features to indicate their origin, except at 826741, 
where layering is observed oblique to the foliation. Here, the 
regularity in size and composition of the layering suggests a sedi-
mentary origin. Such layering is only clearly seen at this locality 
and if it was once a widespread feature it has subsequently been largely 
destroyed. Thus, due to the presence of probable metasediments and 
rarely preserved layering of probable sedimentary origin in the biotite 
schists, it is likely that the NE group of schists represents a deformed 
and metamorphosed sedimentary sequence.

The SW group of schists, in contrast, do not contain any of the 
probable sedimentary lithologies and possess augen of quartz and
plagioclase not observed in the NE group. These schists border the gneisses to the SW and contain a layer of gneiss as an extension of the An Ard gneisses, see Fig 7-1. As a result, all of these schists lie within 400 m of a contact with gneiss. The augen of plagioclase showing truncated twinning and others of quartz suggest that the rock has undergone grain size reduction. This, together with the transitional gneiss/schist boundaries observed at 802754 and 838714, suggests that these schists are derived from gneisses similar to those to the SW, by processes of deformation and metamorphism.

A tectonic origin for these schists due to the presence of augen was first tentatively suggested by Clough et al (1908), see 8.1. The schist-gneiss contact are more fully described and discussed in chapter 8.

The abundant hornblende schists have been interpreted variously as sills, Clough et al (1908), lavas, Bhattacharjee (1968) or as a combination of sills and lava flows, (Park (1964). Rare undeformed lenses showing relict gabbroic textures (7.2.1.(7)) indicate an igneous intrusive origin, ie. as either dykes or sills, and since all the layers show constant composition similar to the dykes in the gneisses (Holland and Lambert, 1973) and similar textures, it is likely that they all belong to the same intrusive episode. Because of deformation their time of intrusion is unclear but must lie some time before the end of the deformation.

b. Deformation in the schists.

Strain analysis in the schists is not possible because of the lack of strain markers but several features suggest that strain is high. These include 'blackened' quartz viens due to fine grain size and highly strained grains, highly elongate possible quartz pebbles (7.2.1.(3)) parallel foliae within the marbles, constancy of foliation, isoclinal nature of folds, uniform grain size and recrystallised textures
The exposures of quartz-magnetite and graphite schists, marbles and garnetiferous amphibolites lie on the flanks of two thick amphibolite sheets, see Fig 7-1. Assuming from the above that the hornblende schists are intrusive in nature, removing them to restore the metasediments to their original relative positions shows that they occupied a narrow layer in the biotite schists only some 200 m wide. These lithologies occur in pods and commonly show coarser grain sizes that the surrounding schists, suggesting that they are less highly deformed. It is possible, therefore, that the separate pods are the boudinaged remnants of one, or possibly two, narrow layers of metamorphosed limestones, banded ironstones, manganiferous ironstones and carbonaceous sediments. The pods of cummingtonite-hornblende-biotite schists probably also represent the boudinaged remnants of one or more layers.

7.3. Ard Ialltaig and An Ard Gneiss

7.3.1. Ard Ialltaig gneisses.

a. Lithologies.

The Ard Ialltaig gneisses are composed of two basic types; hornblende and quartzo-feldspathic gneisses.

i) The hornblende gneisses consist principally of hornblende, plagioclase, + garnet which can form up to 50% of the rock. The detailed mineralogy is listed in Table 7-6. In the coarse grained unfoliated gneisses there are two varieties of hornblende; an actinolitic variety associated with garnet and a brown hornblende with abundant inclusions of opaques. Pyroxenes are reported by Park (1964) but not found in the present study. These rocks occasionally show layering produced by varying proportions of feldspar. With the development of foliation, the two varieties of hornblende are recrystallised to give a single green hornblende, epidote, chlorite and biotite appear, scapolite disappears and garnets are altered to feldspar and chlorite. The grain
Table 7-6. Mineral composition of Ard Ialltaig and An Ard gneisses.

a. Ard Ialltaig gneisses.

<table>
<thead>
<tr>
<th>Lithology</th>
<th>Major minerals</th>
<th>Accessories (&lt;1%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>hornblende gneiss - unfoliated</td>
<td>actinolitic hornblende</td>
<td>quartz, scapolite,</td>
</tr>
<tr>
<td></td>
<td>brown hornblende,</td>
<td>opaques.</td>
</tr>
<tr>
<td></td>
<td>garnet, plagioclase (An20)</td>
<td></td>
</tr>
<tr>
<td>hornblende gneiss, foliated</td>
<td>common hornblende,</td>
<td>quartz, opaques,</td>
</tr>
<tr>
<td></td>
<td>garnet, plagioclase</td>
<td>epidote, + biotite,</td>
</tr>
<tr>
<td>quartzo-feldspathic</td>
<td>quartz, plagioclase (An20)</td>
<td>+ chlorite, sphene,</td>
</tr>
<tr>
<td>gneiss</td>
<td>biotite</td>
<td>carbonate, zircon.</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

b. An Ard gneiss.

<table>
<thead>
<tr>
<th>Lithology</th>
<th>Major minerals</th>
<th>Accessories (&lt;1%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>biotite gneiss</td>
<td>plagioclase (An15)</td>
<td>sphene, zircon, opaques</td>
</tr>
<tr>
<td></td>
<td>quartz, biotite,</td>
<td>+ carbonate</td>
</tr>
<tr>
<td></td>
<td>k-feldspar, epidote</td>
<td></td>
</tr>
</tbody>
</table>
size in unfoliated gneiss is 0.5 to 2 mm with garnets up to 1 cm across. In foliated gneiss the grain size is finer, \( \sim 0.1 \) mm to 1 mm, and garnets form porphyroclasts round which the foliation flows.

ii) The quartzo-feldspathic gneisses.

These gneisses are composed principally of quartz and plagioclase (oligoclase, An\(_{20}\)). Variable proportions of minor hornblende and biotite produce layering. Grain size in unfoliated gneisses is \( \sim 2 \) mm, decreasing to 0.2 mm and less in foliated gneisses. Quartz and plagioclase show good polygonal textures in unfoliated gneiss which becomes recrystallised at grain boundaries as foliation develops. Finally the rock is totally recrystallised except for relict porphyroclasts of plagioclase up to 1 mm across. Other minerals show a similar grain size reduction and garnet shows fracturing and alteration to feldspar and chlorite.

Hornblende gneisses form a core at the centre of the Ard Ialltaig peninsula and are flanked on either side by quartzo-feldspathic gneisses, see Fig 7-6. At the contact of the two, there is a transition zone in which numerous pods angular to rounded of hornblende gneiss up to 1 m across are set in a matrix of quartzo-feldspathic gneiss. Pods are often closely packed separated by a framework of quartzo-feldspathic material, see Fig 7-7. Similar patches of pods occur throughout the quartzo-feldspathic gneisses.

b. Structure and relation to the S. Sithean Mhor gneisses.

The Ard Ialltaig gneisses form a lens shaped outcrop, the SE position of which is exposed on the Ard Ialltaig peninsula. The outcrop trends NW/SE and is bounded to the SW by S. Sithean Mhor gneisses and to the NW by the Gairloch schist belt. Along their SW margin the Ard Ialltaig gneisses are interlayered with the k-feldspar bearing S. Sithean Mhor gneisses. Individual layers of k-feldspar bearing gneisses occur up to \( \sim 2 \) m wide and are parallel to the foliation.
FIG 7-6 Geological map of Ard Ialltaig and adjacent areas.

KEY

- Hornblende gneiss
- Quartzo-feldspathic gneiss
- Augen gneiss
- Hornblende schist
- Biotite schist
- S. Sithean Mhor gneiss

Legend:

- Torridonian cover
- Minor shear zones
- Strike and dip of foliation
- Direction and plunge of lineation
- Lithological boundary

SCALE

500 metres
FIG 7-7  Pods of hornblende gneiss in quartzo-feldspathic matrix, Ard Ialltaig. Sketches of examples.

(a) Angular pods.
807736
vertical surface facing N.

(b) Subrounded pods.
803735
subhorizontal surface

(c) Rounded pods.
805737
subvertical surface facing NW.
A core of unfoliated gneisses occupy the centre of the outcrop and is surrounded by a mantle of deformed gneisses whose foliation follows the trend of the outcrop margin, see Fig 7-6. The state of deformation in the gneisses is clearly displayed by the shape of pods of hornblende gneiss, quartz fabrics in quartzo-feldspathic gneisses and foliation in the hornblende gneisses. In the central part of the outcrop roughly equidimensional pods of hornblende gneiss occur within a quartzo-feldspathic matrix with weak and variably oriented quartz fabrics. In the main outcrop of hornblende gneiss, fabrics are isotropic and layering dips variably SE. Outwards from this undeformed core the pods become elongate parallel to the simultaneously developing foliation in the pods themselves and the quartzo-feldspathic matrix. The gneissose layering is rotated to dip steeply NE or SW. This occurs with the change in grain size and mineralogy described in 7.3.1.(a). The lineations associated with the quartz fabric and those described by the alignment of hornblende laths in the hornblende gneisses plunge moderately to steeply NW throughout the margins of the outcrop, see Fig 7-8.

The increasing deformation outwards from the core of the outcrop is also reflected in the shape of the quartzo-feldspathic hornblende gneiss boundaries: on Ard Ialltaig itself, the boundaries have lobate forms whereas to the NE at Torr Salach (815727) outcrops of quartzo-feldspathic gneiss form elongate lenses within the hornblende gneisses, see Fig 7-6.

The relatively undeformed core contains numerous minor shear zones trending NW/SE, ranging in width from 1 cm to 1.5 m, which deflect the NE dipping layering. Width of gneissose layering and grain size is greatly reduced within the zones which frequently have mylonitic textures. The shear zones dip subvertically and have moderately NW plunging lineations in the zone of strongest deformation, see Fig 7-9. The deflection of gneissose layering and the lineations indicate that
FIG 7-8 Foliation and lineation orientations in the Ard Ialltaig gneisses.

- Poles to foliation
- Lineations.
FIG 7-9  Foliation and lineation orientations in minor shear zones.

- Poles to foliation.
+ Lineation.
all have a negative sense of shear (SW side down). Individual shear zones can be traced for up to 370 m along strike, and at their terminations fan out with a decrease in the intensity of deformation until they are no longer distinguishable.

7.3.2. An Ard gneisses.

K-feldspar-bearing biotite gneisses containing layers and patches of quartz-feldspathic gneisses outcrop at An Ard, 803751. Their mineralogy is listed in Table 7-6. These gneisses contain abundant augen of plagioclase and k-feldspar up to ~1 cm across, round which quartz and biotite foliae flow. Occasional euhedral porphyroblasts of k-feldspar up to 1.5 cms long occur. They are thus very similar to the S. Sithean Mhor gneisses in mineralogy and texture but contain a higher percentage of biotite (10 to 15%).

These gneisses continue to the SE, becoming more quartz-feldspathic, to 832721. From 813737 to 819730, they form layers of augen gneiss subparallel to the foliation within the biotite and hornblende schists and Ard Ialltaig gneisses, see Fig 7-6. To the SE the outcrop becomes wider and is flanked on both sides by biotite schists, see Fig 7-1. The quartz-feldspathic gneisses wedge out to the SE becoming progressively finer grained and finally grading into biotite schists at 823721.

At An Dun, 802753, part of the An Ard peninsula, the biotite gneisses are seen as crosscutting viens within amphibolite, see Fig 7-9 and Plate 7-9. These viens are deformed with the amphibolite and many show folding. Similar crosscutting relationships are seen on the coast section to the SE at 813737.

7.3.3. Discussion.

The Ard Ialltaig gneisses differ from both the Loch Tollie gneisses and the S. Sithean Mhor gneisses in that they contain an abundance of garnetiferous amphibolite and k-feldspar-free quartz-feldspathic
FIG 7-9 Cross-cutting relations between augen gneiss and amphibolite; sketches of examples.

(a) An Ard, 803754 subhorizontal surface
(b) An Ard, 802754 subvertical surface facing NW.
(c) Loch Kerry shore, 812738 subvertical surface facing SE.
(d) Loch Kerry shore, 813736 subvertical surface facing SE.
Plate 7-9(a). Augen gneiss (A) veins cutting amphibolite (B), 805750, An Ard. Subvertical surface, looking NW.
Plate 7-9(b). Detail in Plate 7-9(a) showing folded augen gneiss vein (A), 805750, An Ard. Subvertical surface, looking NW.
gneisses. They, therefore, belong to neither of these groups of
gneisses and represent a third group of different origin and possibly
age.

The foliated amphibolite mass at Torr Salach, 815728, is, in this
account, grouped with the Ard Ialltaig gneisses, contrary to Park
(1964) and Park and Bowes (1966). They correlate these rocks with the
hornblende schists common to the NE and postulated a major thrust
trending NW and SE through 810734, separating the Ard Ialltaig gneisses
from the schists. However, the amphibolite at Torr Salach contains lenses
of k-feldspar-free quartzo-feldspathic material especially in the SE
around 812730, and also has relict and partially altered garnets
throughout. Park and Bowes (1966) interpreted the lenses of quartzo-
feldspathic material as due to metamorphic segregation. However, no
such segregation is observed in any of the other hornblende schist or
amphibolite sheets in the area and it seems more probable that these
rocks are deformed and retrogressed Ard Ialltaig gneisses. This places
the NE boundary to the Ard Ialltaig gneisses at the contact with biotite
schists or augen gneiss, see Fig 7-6.

The shape of the outcrop and the low state of deformation in the
centre indicates that the gneisses are an augen of relatively low
deformation. The change in mineral assemblage; appearance of a single
variety of hornblende, the disappearance of pyroxene (Park, 1964) and
scapolite, the appearance of epidote and the alteration of garnets
to feldspar and chlorite, associated with the margins of the outcrop
and minor shear zones, indicates a decrease in metamorphic grade
associated with the deformation. The Ard Ialltaig gneisses have
therefore suffered a phase of higher grade metamorphism predating the
deformation.

The small shear zones common in the centre of the outcrop all have
a negative sense of shear (SW side down) in agreement with the sense implied by the deformation in the Loch Tollie and S. Sithean Mhor gneisses. Lineations throughout the Ard Ialltaig plunge NW as in the minor shear zones implying a similar age for the deformation. Lineations in the schists to the NE, in contrast, plunge SE, implying a different age for their formation.

The amphibolites at Torr Salach correlated with the Ard Ialltaig gneisses are cut by the An Ard gneisses showing that the An Ard gneiss post-date and are intrusive into the Ard Ialltaig gneisses. The An Ard gneisses are very similar in mineralogy and texture to the S. Sithean Mhor gneisses and since S. Sithean Mhor gneisses show similar relations with the Ard Ialltaig gneisses as the An Ard gneisses (at 817725) it seems probable that they are a NE extension of the S. Sithean Mhor gneisses. Thus it is probable that the Ard Ialltaig gneisses pre-date and are intruded by the S. Sithean Mhor gneisses.

7.4. Summary and Conclusions

The schist belt is composed largely of biotite schists of semipelitic composition which are divided into two groups:—

a. The NE group of schists are less conspicuously layered than the SW group and contain lenses and discontinuous layers of other lithologies of probable sedimentary origin, eg. marble, quartz-magnetite schists, garnetiferous amphibolites, graphite schists, and cummingtonite-hornblende-biotite schists of probable igneous origin. Occasionally preserved layering also suggests that the biotite schists are sedimentary in origin.

b. The SE group of schists, in contrast, contain none of the above lithologies and show a greater range in grain size. Widespread augen of plagioclase and quartz, not present in the NE group of schists, indicate that these rocks have suffered grain size reduction and this,
together with the presence of transitional gneiss-schist contacts, suggests that these schists are derived from the gneisses by the processes of deformation and metamorphism. The origin of the schists and the nature of the gneiss-schists contact are investigated in chapter 8.

The biotite schists contain abundant layers of hornblende schist which have probably originated as intrusive igneous bodies (dykes or sills) as indicated by occasionally preserved gabbroic igneous textures in lenses of low deformation.

The Ard Ialltaig gneisses form an augen shaped outcrop at the boundary between the S. Sithean Mhor gneisses and the schists. The centre of the outcrop shows a low state deformation with minor shear zones trending NW/SE with subvertical dip and NW plunging lineations. The shear zones show a negative sense of shear in agreement with that indicated by the Tollie antiform. Outwards from this core, deformation increases with the development of steep NE or SW dipping foliation and NW plunging lineations.

The An Ard gneisses are seen to be intrusive into amphibolites correlated with the Ard Ialltaig gneisses. These gneisses and SE extension are correlated with the S. Sithean Mhor gneisses, on the basis of similar textures, mineralogy and relationship with the Ard Ialltaig gneisses. Thus the S. Sithean Mhor gneisses are thought to post-date and be intrusive into the Ard Ialltaig gneisses.

The relationship between the S. Sithean Mhor gneisses and the schists is uncertain. It is probable that the SW group of schists are derived from the gneisses and, if this is the case, the true contact with the schists of sedimentary origin lies to the NE. Due to the similarity of tectonically derived schists and those of probable sedimentary origin and modification of the contact by deformation, the relationship between the gneisses and metasedimentary schists is unclear.
CHAPTER 8 THE TRANSITION BETWEEN S. SITHEAN MHOR GNEISSES AND SCHISTS

8.1. Introduction

In this chapter the southwestern gneiss-schist contact described in 7.2.1. is investigated in detail. The area of An Ard is chosen for the study, as here the transitional nature of the contact is clearest and best exposed.

The gradational contact with progressive disappearance of augen textures, and the presence of sporadic feldspar augen in the schists, suggests that the schists are derived by the process of grain size reduction from the gneisses, an idea which was first tentatively proposed by Clough et al (1908, p. 215) who wrote,

"It is noteworthy that one of the commonest of what are taken to be sediments is a flaggy brown mica schist, resembling granulitic bands, which in Sutherland have in many places been formed along Pre-Torridonian shear-lines from more massive gneisses."

and ibid, p. 216;

"In several places .... a fine flaggy augen schist, which lies close to and is sometimes intermixed with the mica schists, might conceivably represent sheared pebbly sediments, but is more probably a sheared form of massive augen gneiss like that a little further SW."

Winchester, Park and Holland (1980) have analysed chemically the gneisses and schists and separated the schists into two divisions:-

a. Schists chemically similar to the gneisses.

b. Schists with higher total iron, MgO, TiO$_2$, Nb, Y, Zr, Zn, Cu, Cr and Ni and lower Al$_2$O$_3$, Na$_2$O, Ba and Sr. On the basis of TiO$_2$/SiO$_2$ ratios, Zr/TiO$_2$ vs Ni, and Na$_2$O/K$_2$O vs SiO$_2$/Al$_2$O$_3$ plots, they are designated as nonvolcanogenic greywackes.

The schists adjacent to the S. Sithean Mhor gneisses fall largely in group (a) with a minority of samples falling in group (b). They, therefore, conclude that these schists are 'sheared gneisses' with
included bands of metasediments.

The above evidence strongly suggests that the most southwesterly schists are derived from the gneisses. The purpose of the present study is to investigate the textural and mineralogical evolution and the metamorphic conditions of the gneiss-schist transformation.

8.2. Gneiss-Schist transition

8.2.1. Field description.

The transition between gneisses and schists is most clearly seen at An Ard, 803750, see Fig 8-1. Biotite gneisses outcropping in the southwestern part of the peninsula, have marked augen textures in which biotite and quartz foliae wrap around augen of plagioclase and k-feldspar. Large porphyroblasts of k-feldspar similar to those in the S. Sithean Mhor gneisses (see 3.2.1.), up to 2 cms across, occur throughout the gneisses. The content of biotite is variable and almost pure quartzofeldspathic gneisses are found irregularly interleaved with biotite gneisses, often showing folding. The foliation throughout the gneisses dips steeply SW and lineations plunge moderately to steeply NW or W.

To the NE the gneisses become finer grained and progressively lose their augen textures over distance of $\sim 100$ m across strike. This transition is clearly exposed along the SE coast of the peninsula from 807750 to 805750 and across An Dun, a small peninsula, on the NW coast of An Ard at 802754. The gneisses finally grade into biotite schists which range from quartzofeldspathic-rich types with a flinty fracture to biotite-rich types with a flaggy parting. In some localities eg. at 808753 and at 808750, augen of feldspar up to 4 mm across are visible to the naked eye. The foliation throughout the schists dips steeply NE and lineations are moderately to steeply plunging SE. In the schists quartz veins are common and occasional concordant veins of carbonate, which grade along and across strike into calcareous schists,
FIG 8-1 Location of specimens- An Ard gneisses.

Cross-section section showing location of specimens.
also occur, eg at 807750.

Similar gneiss-schist relations are seen to the SE at Coill a' Ghlinne from 809749 to 813742, but the transition here is rather more poorly exposed.

In the following sections, the mineralogy and textures of the rocks over the gneiss-schist transition are examined to investigate the processes and conditions under which the transition takes place. Seventeen specimens were selected for the study and their locations are marked on Fig 8-1. The gneiss-schist boundary was arbitrarily chosen on the basis of field examination as the stage at which augen textures are no longer distinguishable, excluding the presence of sporadic plagioclase augen in the schists. The position of specimens on cross-sections were located relative to this boundary, see Fig 8-1.

8.2.2. Mineralogy.

a. The gneisses.

The biotite gneiss occupying the bulk of the An Ard peninsula,

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In order to distinguish feldspar from quartz, and plagioclase
from K-feldspar in the finer grained rocks, all thin sections
were etched with hydrofluoric acid and stained with sodium
cobaltinitrite. K-feldspar is then stained yellow and plagioclase
white while quartz remains clear.

Fig 74, Deer et al (1966). The green colour is probably due to a relatively high Fe$_2$O$_3$ to TiO$_2$ ratio, Deer et al (1966). Epidote shows constant composition throughout, Ps 20 to 27. Microprobe analyses of plagioclase, biotite and epidote for all specimens are listed in Appendix III.
Table 8-1. Modal percentages of minerals in An Ard gneisses and adjacent schists.

<table>
<thead>
<tr>
<th>Specimen</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
<th>8</th>
<th>9</th>
<th>10</th>
<th>11</th>
<th>12</th>
<th>13</th>
<th>14</th>
<th>15</th>
<th>16</th>
<th>17</th>
</tr>
</thead>
<tbody>
<tr>
<td>Quartz</td>
<td>21.1</td>
<td>18.5</td>
<td>25.0</td>
<td>22.9</td>
<td>27.3</td>
<td>18.7</td>
<td>26.2</td>
<td>19.5</td>
<td>18.0</td>
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FIG 8-2 Modal mineral analyses, An Ard.

- **Gneiss**
  - Plagioclase
  - K-feldspar
  - Quartz
  - Biotite
  - Epidote
  - Opaques
  - Carbonate

- **Transitional zone**

- **Schist**

**NOTE:** Points have been joined for visual convenience only.
b. Transition to schist.

In the transition zone two sets of reactions take place:

i) Microcline is replaced by plagioclase of the same composition as the plagioclase augen (An 10-17), containing inclusions of quartz as tiny platelets in common orientation and irregular rod-like inclusions, and also of biotite. Microcline does not appear in any of the rocks to the NE.

ii) Epidote disappears simultaneously with an increase in the anorthite content of plagioclase, $\frac{Al_2O_3}{(FeO \text{ (as total Fe)} + MgO + MnO)}$ ratio in biotites, and a general increase in the modal proportions of carbonate and opaques, see Fig 8-3 and 8-2. Plagioclase increases in anorthite content from 16.2 to 32.5. The range of plagioclase compositions with individual specimens is similar to those in the gneisses but it is seen that there is a greater variation between specimens, see Fig 8-3. Plagioclase augen are unzoned and show similar compositions to matrix plagioclase.

The ratio of $\frac{Al_2O_3}{(FeO + MgO + MnO)}$ in biotites increases from 0.47 to 0.60 in the gneisses to 0.54 to 0.67 in the schists, ie. composition moves closer to the siderophyllite-eastonite compositions. The change is accompanied by a colour change to red-brown probably reflecting a decrease in $Fe_2O_3$ in response to an increase in $Al_2O_3$ so that the colour here is due dominantly to the $TiO_2$ content (Deer, et al, 1966).

There is considerable variation in $\frac{Al_2O_3}{(FeO + MgO + MnO)}$ ratio between specimens which correlates well with the variation in anorthite content of plag; specimens with higher ratios also show higher anorthite content, see Fig 8-3.

The schists to the NE retain similar modal proportions and compositions of minerals throughout their outcrop. Reactions i) and ii) above, both occur within the gneiss-schist transition but since epidote is always found in rocks still containing alkali feldspar it is probable that reaction i) slightly preceeds reaction ii).
FIG 8-3 Variation in plagioclase and biotite composition across the An Ard gneisses.

a) Plagioclase

b) Biotite
Discussion.

a. Reaction i).

The replacement of alkali feldspar by myrmekitic intergrowths of plagioclase and quartz has been reported by several workers and more recently by Phillips, Ransom and Vernon (1972) and Phillips (1964) in both metamorphic and igneous rocks. There are two possible modes of origin for their formation; (a) exsolution from high temperature alkali feldspar of plagioclase and quartz, Phillips (1964), and (b) destruction of alkali feldspar under retrograde conditions, (Phillips et al, 1972). Since alkali feldspar is totally replaced here the reaction must be of type (b).

Phillips et al (1972) give the following reaction:-

\[ 3 \times \text{KA}_1\text{Si}_3\text{O}_8 \]
\[ y \times \text{NaAlSi}_3\text{O}_8 \]
\[ Z \times \text{Ca(AlSi}_3\text{O}_8)_2 \]

alkali feldspar

\[ + X \times \text{H}_2\text{O} \]

\[ \iff \]

\[ X \times \text{KA}_1\text{(Al, Si)}_3\text{O}_{10}(\text{OH})_2 \]

muscovite

\[ + Y \times \text{NaAlSi}_3\text{O}_8 \]
\[ + \text{Z CaAl}_2\text{Si}_2\text{O}_8 \]

plagioclase

\[ + (6X+4) \times \text{SiO}_2 \]

quartz

\[ + X \times \text{K}_2\text{O} \]

In general muscovite does not occur in any major quantity in these rocks, so it is likely that the reaction here is more complicated involving the formation of biotite instead of muscovite, and including reaction with the mafic minerals in the surrounding rock. This view is supported by small biotite inclusions common within myrmekitic plagioclase augen.

The above reaction is generally referred to as a retrograde reaction, i.e. one due to a decrease in temperature and/or pressure, but since water is involved its availability could also control the reaction under constant pressure and temperature conditions. In the
gneisses, biotite is common and no anhydrous mafic minerals, such as pyroxene, are found anywhere in the area. This, together with the abundance of quartz veins throughout the gneisses suggests the presence, during deformation, of a metamorphic fluid containing at least some water. Also, the stability of epidote is strongly dependent on the composition of the metamorphic fluid and indicates a H$_2$O-rich composition, Lindh (1978), Storre and Nitsch (1972). Since epidote is abundant in the gneisses, the regression of alkali feldspar cannot therefore be due to restricted availability of water, but must therefore be due to a decrease in temperature and pressure.

Myrmekite formation from alkali feldspar in gneisses from Broken Hill, Australia, of similar mineralogy to the An Ard gneisses (quartz-plagioclase-garnet-biotite) is reported to occur on retrogression to the lower Amphibolite facies or Staurolite-almandine facies of Turner and Verhoogen (1968), by Vernon and Ransom (1971). This is in agreement with the garnet grade conditions prevailing in the schists (8.3) and infers that the gneisses are of higher grade.

b. Reaction ii).

The major changes brought about by this reaction are the disappearance of epidote and the increase in anorthite content of plagioclase. Epidote-plagioclase equilibrium has been extensively studied and the dependence of anorthite content of plagioclase co-existing with epidote on pressure, temperature, oxygen fugacity and composition of the metamorphic fluid, notably XH$_2$O, has been clearly demonstrated by Liou, (1972), Storre and Nitsch (1972), Lindh (1978) and others.

The temperatures and pressures of the garnet grade of pelitic rocks, indicated by other assemblages in the schist belt (8.3) are well within the stability range of epidote, Strens (1965), Liou (1972). Also the occurrence of magnetite and quartz in the quartz-magnetite schists (Table 7-1), chlorite schists and hornblende
schists and the occurrence of haematite in the biotite schists indicates that oxygen fugacities are well within the stability range of epidote, Liou (1972).

Epidote is only stable in the presence of $H_2O$-rich metamorphic fluids. The occurrence of carbonate veins in the biotite, chlorite and hornblende schists indicates that the metamorphic fluid does contain $CO_2$. In the presence of more $CO_2$-rich fluids, epidote is eliminated by a reaction of the type:

$$2\text{ zoisite} + 1CO_2 \rightarrow 3\text{ anorthite} + 1\text{ calcite} + 1H_2O \quad (A)$$

(Storre and Nitsch, 1972).

The increase in modal proportion of carbonate in the schists and increase in anorthite content suggests that a reaction of this type has taken place. Due to the presence of other minerals, here, plagioclase and biotite, the reaction is not as simple as (A) above, but involves a more complicated redistribution of elements in which Ca and Al are incorporated as anorthite content of plagioclase and the remaining Al is accepted by biotite. Since the gneisses and schists are essentially isochemical (Winchester et al, in press) this leaves Si and O from epidote, Na and Si from plagioclase, and Fe from biotite which are accommodated by tourmaline, quartz and opaques. This explains the correlation between the anorthite content of plagioclase and the $Al_2O_3/(Fe + MgO + MnO)$ ratio in biotite, see Fig 8-3, since both are linked to the decomposition of epidote. Varying proportions of epidote in the original rock also explain the greater range of plagioclase and biotite compositions found in the schists than in the gneisses. The increase in carbonate in the schists is probably not due to its formation as a result of the reaction but a reflection of the increased $XCO_2$ of the metamorphic fluid.

Nitsch and Storre (1972) show that the reaction (A) is independent
of temperature in the range they investigated (400-720°C). Thus the
distribution of epidote indicates H₂O-rich fluids in the gneisses
and a greater CO₂ content of the metamorphic fluid in the schists.

8.2.3. Textures.
   a. The gneisses.

   The gneisses are composed of plagioclase and k-feldspar augen
surrounded by foliae of quartz and biotite, see Plate 8-1. The
plagioclase augen, up to 1 cms across, have cores of single grains or
aggregates of large (approximately 5 mm) grains surrounded by a
mantle of small grains, on average 0.2 mm across, see Plate 8-2. Core
grains show slight undulose extinction and deformation twins. Large
grains within cores have optically close orientations and grain boundaries
are either straight or composed of straight line segments. In the mantles,
grains show greater optical misorientation and grain boundaries are
highly irregular with numerous bulges, ~ 0.02 mm across, and 'fingers'
extending into the core grains, see Plate 8-2. Occasionally, small
grains, around 0.05 mm across, occur along grain boundaries and there
is a complete gradation from bulges to small grains. The core grains
have numerous small inclusions of epidote, around 0.1 mm, which, in
the mantle and adjacent to grain boundaries in the core, are replaced
by fewer larger inclusions around 0.05 mm, see Plate 8-2.

   k-feldspar forms large augen up to approximately 2 cms across
of single grains with simple and crosshatched twinning, surrounded by
mantles of smaller grains similar to that shown by the plagioclase
augen.

   Quartz form foliae that wrap round the feldspar augen, with a
grain size range from 0.2 to 0.8 mm, largely determined by the thick-
ness of the foliae. Quartz grains show strong banded undulose extinction,
the bands of which are orientated between 90 to ~ 45° to the foliae
Plate 8-1. Biotite gneiss showing foliae of quartz (Q) and biotite (B) wrapping around augen of plagioclase (P), specimen 1.
Plate 8-2. Part of a plagioclase augen in biotite gneiss, specimen 2, showing old core grains (C) with abundant small epidote inclusions (E) and new inclusion-free mantle grains (M). Note indented grain boundaries with bulges (b).
margins, see Plate 8-3. Grain boundaries are straight or smoothly curving with bulges occurring only rarely. In the larger quartz aggregates textures are subpolygonal.

Biotite foliae, similar to those of quartz, wrap around the feldspar augen, see Fig 8-1, and individual grains are, on average, 1.5 mm long. Few grains show any bending of the cleavage, variations in the orientation of the foliae being accommodated by the change in orientation of grains. Epidote, with subhedral shape, is associated with biotite and often cuts across biotite grains.

Interpretation of textures.

Textures in the plagioclase and k-feldspar augen closely resemble the core and mantle structures described by Vernon (1975) in deformed plagioclase grains. Two features indicate that the smaller mantle grains have developed at the expense of the core grains.

i) Bulges and indentations on grain boundaries project towards, and indicate grain boundary migration into, core grains. The feldspar augen are therefore undergoing grain size reduction by the development of low angle grain boundaries, which become high angle boundaries as the misorientation between adjacent grains grows. Such grain boundaries form as dislocations created by strain, arrange themselves in stable low energy arrays; a process known as recovery, White (1973, 1976, 1977). Highly irregular shapes with numerous bulges indicates mobile grain boundaries whereas straight or smooth shapes indicate stable boundaries. The sequence observed here in which low angle grain boundaries in the core are immobile while high angle boundaries in the mantle are mobile is in agreement with observations by White (1977). The combined processes of subgrain formation by development of low angle boundaries in strained grains, and new grain formation at bulges in mobile grain boundaries, therefore leads to recrystallisation by recovery and a
Plate 8-3. Foliae of quartz (Q) showing banded undulose extinction, specimen 1. Grains span the width of foliae and boundaries are smoothly curving (b).
subsequent grain size reduction.

ii) Epidote inclusions have recrystallised to form fewer larger inclusions in mantle grains and adjacent to core grain boundaries. Thus, newly formed grain boundaries have promoted diffusion and recrystallisation of the inclusions. Similar behaviour of inclusions has been noted in plagioclase by Vernon (1975).

Quartz grains have smoothly curving, and therefore stable, grain boundaries. Quartz has accommodated strain by the development of banded undulose extinction and since the grain boundaries show no relation to the extinction patterns it is probable that these grain boundaries are inherited from a previous textural state. Thus, quartz is not at this stage undergoing significant grain size reduction.

Biotite grains show little sign of strain and must therefore have recrystallised syntectonically. The behaviour of epidote within plagioclase augen and in biotite foliae, shows that it too is recrystallising during the deformation.

b. Transition to schists.

Towards the schists the augen texture of the gneisses is gradually destroyed. Plagioclase augen become more elongate and are largely recrystallised to a grain size of 0.2 mm on average, with few old core grains remaining, see Plate 8-4. All grain boundaries are irregular with bulges and small grains, 0.02 mm upwards, abundant on grain boundaries. All epidote inclusions have now recrystallised.

Microcline is here wholly or partially replaced by plagioclase with quartz inclusions. Smaller microcline grains have altered to a polygonal mosaic of plagioclase and minor quartz grains, 0.3 to 0.01 mm across, see Plate 8-5, and larger unstrained grains in the cores of large porphyroblasts have altered to large single grains, see Plate 8-6.

Quartz foliae are thinner and shorter with a grain size of 0.2 mm
Plate 8-4. Recrystallized and strung out augen of plagioclase surrounded by biotite (B), transition zone, specimen 9.
Plate 8-5. Strained K-feldspar grains (K), partially replaced by fine-grained mosaic of plagioclase and quartz (PQ), from transition zone, specimen 7.

Plate 8-6. Large unstrained K-feldspar grain (K) partially replaced by plagioclase (P) with quartz inclusions (Q), transition zone, specimen 2.
controlled, as in the gneisses, by foliae width. Undulose extinction is stronger than in the gneisses and deformation bands occur, see Plate 8-7. Grain boundaries are straight or smoothly curved, subperpendicular to foliae walls.

Biotite still forms foliae but these are now less distinct and biotite is more evenly distributed throughout the rock, see Plate 8-7. The grain size has now reduced to around 1 mm.

Interpretation of Textures.

Plagioclase augen are now almost completely recrystallised. All grain boundaries are irregular and mobile and this together with the increased incidence of bulges indicates that new grains are forming primarily from bulges in mobile boundaries and no longer by the development of low angle boundaries.

The survival of single grain microcline augen in contrast to plagioclase augen is probably due to their greater size in the original gneiss.

The grain size of quartz has reduced in response to a reduction in foliae width and the development of deformation bands shows that new grains are forming, resulting in grain size reduction.

Since biotite has become more evenly distributed throughout the rock new grains must have nucleated within the recrystallised plagioclase augen. This implies that diffusion, as opposed to recovery in plagioclase and quartz, plays a large part in the response of biotite to deformation.

c. The schists.

In the schists all augen textures of the gneisses have been destroyed except for the preservation in some schists of plagioclase augen up to 1.5 mm long, see Plate 8-8. Platelets and rod-like inclusions of quartz within the augen indicates that they originated from the plagioclase augen replacing original microcline porphyroblasts.
Plate 8-7. Quartz foliae (Q) showing subgrains with smoothly curving boundaries (b) and banded undulose extinction, transition zone, specimen 6.
Plate 8-8. Plagioclase augen (P) with abundant quartz inclusions (Q) recrystallizing at margins, in schist, specimen 16.

Plate 8-9. Plagioclase augen (P) with recrystallized zone of very fine grained plagioclase and quartz (R), in schist, specimen 12.
Plate 8-10. Biotite-rich and biotite-poor layers in schist showing the effect of biotite concentration on the grain size of quartz and plagioclase, schist at 839720.
Plate 8-11. Texture in schists. Note elongate plagioclase grains, aggregates of quartz grains, and smoothly curving grain boundaries with subpolygonal textures.
These augen range from single grains to totally recrystallised examples. Partially recrystallised augen have linear zones of fine grain size, 0.02 to 0.2 mm, with good polygonal textures, see Plate 8-9.

The grain size of quartz and plagioclase is related to the percentage of biotite and this results in a range from 0.15 to 0.06 mm, the more biotite rich rock have the finer grain size of quartz and plagioclase, see Plate 8-10. Biotite grain size is 0.1 to 0.15 mm. Textures of all schists are similar, regardless of grain size.

Both plagioclase and quartz tend to form elongate grains whose boundaries are in part determined by neighbouring biotite grains, see Plate 8-11. This is most marked in plagioclase in which length to breadth ratios of 1:4 are common. Plagioclase grains are generally strain free and deformation bands and twinning are rare. Quartz occurs as aggregates of several grains which have elongate augen shapes. Individual grains range from strain free to possessing strong undulose extinction, though banded undulose extinction is rare. Textures within quartz aggregates approach polygonal and grain shapes, equidimensional. Quartz-plagioclase grain boundaries are commonly smoothly curving with subpolygonal texture.

Concentration of micas is variable and causes layering on the scale of 1 mm to 1 cms, most easily detectable in thin section (Plate 8-7). Micas are evenly distributed with each layer and most commonly occur along quartz and plagioclase boundaries. Only the larger grains of 1 mm or more, largely muscovite, show bending and kinking of the cleavage, and the smaller grains are strain free. Mica grains commonly have small opaque grains decorating their boundaries, and trails of opaques within quartz and plagioclase grains probably mark the position of former mica grains.
Interpretation of textures.

All augen, except some plagioclase augen derived from original microcline, have been destroyed. The remaining plagioclase augen are undergoing recrystallisation which has progressed by differing amounts in individual augen. Recrystallisation results in linear zones of a fine mosaic of grains which then grow to the grain size of the matrix. The lack of large subgrains in augen and progression from low to high angle boundaries as seen in the gneisses, together with the development of small grains, shows that recrystallisation now progresses by nucleation of large numbers of small grains in strained areas. Grain growth follows nucleation to the grain size of the matrix.

Grain shapes are partly controlled by the presence of micas which encourage elongate shapes. Quartz aggregates of augen shape are probably relict from original foliae. Undulose extinction and the development of subgrains indicates that the processes of recovery are still proceeding through the decrease in the strength of undulose extinction, lack of deformation bands and subpolygonal textures shows that the grain configuration approaches stability. Lack of deformation features and smooth grain boundaries indicates that plagioclase has reached a stable grain size and texture. This, together with evidence of recrystallisation by nucleation in plagioclase augen indicates that the deformation mechanism in plagioclase has changed from dominantly by dislocation movement (recovery), to diffusion processes. Because the change has occurred in response to a decrease in grain size, ie. increase in grain boundary area, the probable deformation mechanism is by grain boundary diffusion, ie. Coble creep, White (1976). Temperature estimates for the schists of \( \sim 420^\circ C \) (8.3.3.) are consistent with Coble creep as a possible deformation mechanism in quartz, for the grain size range 0.1 mm to 0.01 mm (which includes the grain size of the schists) from deformation maps of quartz, White (1976). In quartz, the presence of subgrains and strain extinction
shows dislocation creep is still active, though the decrease in intensity of strain extinction suggests that, here too, Coble creep is active.

The even distribution of micas and the trails of inclusions of opaques marking previous location of mica grains in quartz and plagioclase indicates that mica is syntectonically recrystallising and preferentially nucleates at grain boundaries of other minerals. This indicates that mica deformation is controlled largely by diffusion processes.

Such changes in deformation mechanisms in response to grain size reduction have been shown by White (1976, 1977) who suggest that the process is accompanied by strain softening and that, as a result, rocks of very fine grain size, i.e. mylonites, are effectively superplastic. Thus, the schists have possibly accommodated very high strains.

d. Discussion.

The textural observations above allow comparisons to be made between the behaviour of plagioclase, quartz and biotite, during deformation resulting in grain size reduction.

In the coarse grained rocks, i.e. the gneisses, plagioclase is more competent than quartz which forms foliae wrapping around the feldspar augen. The dominant mechanism in these rocks is dislocation creep for both plagioclase and quartz. Dislocation creep comprises two processes:-

a) the build up of dislocations within grains expressed optically as strain extinction, which leads to strain hardening and,

b) arrangement of dislocations into low energy configurations expressed optically as grain boundaries and walls of deformation bands, which leads to strain softening.

In steady state deformation the two processes are balanced, White (1976).

The greater development of strain extinction in quartz than feldspar
indicates that plagioclase becomes strain hardened at a lower density of dislocations. This limits the strain rate in plagioclase to that at which recovery processes, ie. the development of subgrains, can keep pace. In contrast, quartz shows little sign of forming subgrains indicating that dislocation densities are not high enough to cause significant strain hardening. These relations may in part, explain the competence difference between quartz and feldspar. At this stage in the deformation, plagioclase is undergoing grain size reduction whereas quartz is not.

As the deformation proceeds, represented by the rocks transitional from gneiss to schist, the reduction in grain size allows recovery to reduce the effects of strain hardening. Plagioclase augen are strung out and eventually destroyed. Only larger grains of microcline (now replaced by plagioclase and quartz) survive as augen by virtue of their size. Grain size reduction occurs in quartz indicating that the maximum density of dislocations before strain hardening occurs has now been reached, and that recovery processes are active.

Grain size of quartz and feldspar continues to decrease by dislocation creep until sizes of around 0.1 mm are reached. Plagioclase and quartz both tend towards polygonal textures, indicating stable grain boundary configurations. Plagioclase grains show no sign of strain, indicating that dislocation creep has been replaced by grain boundary diffusion, ie. Coble creep. Such a change in mechanism is made possible by the increase in grain boundary area enhancing the effect of grain boundary diffusion. Quartz, in contrast, still exhibits strain extinction, though to a lesser extent than in coarse grained rocks, showing that dislocation creep is still active. The increase in grain boundary area affects quartz also, so it is likely that here also some grain boundary diffusion takes place, as in plagioclase.

Micas show little sign of strain throughout the deformation,
indicating that recovery balances any dislocation development. Micas readily nucleate and recrystallise so that they become gradually more evenly distributed throughout the rock as grain size reduces. Grain boundaries of quartz and feldspar provide favourable sites for mica nucleation and growth. Thus they resorb from sites within grains to leave trails of opaques.

Within the schists there is a relationship between grain size of quartz and feldspar and mica content, the more mica-rich rocks having finer grain size. Since grain boundaries of quartz and feldspar are partly controlled by mica grains, increasing the percentage of mica has the effect of reducing the average width between mica grains and therefore the grain size of quartz and feldspar.

Textures and grain size within the range 0.15 to 0.05 mm are constant throughout the schist belt, indicating that they are stable. Thus the processes leading to grain growth (recovery, grain boundary diffusion) are balanced by grain size reduction processes (strain hardening) and the deformation has reached a steady state condition.

8.3. Metamorphic Conditions.

8.3.1. The Pelites.

Rocks of pelitic composition are found at two localities. At 826741 pelitic layers occur in a layered biotite schist described in 7.2.1.(b), see Plate 7-2, and are composed of biotite, chlorite, quartz and minor opaques and tourmaline. At 837713 the mica-rich schist is composed of biotite, muscovite, quartz, garnet with minor opaques, tourmaline and chlorite.

In the schist at 837715, garnet occurs as large grains with abundant inclusions of quartz and biotite often forming S-shaped trails, see Plate 8-12. The garnets have generally ragged outlines, with 'atoll' forms occurring, suggesting that they are resorbing. However, abundant
Plate 8-12. Garnet in mica schist at 837715. Note small euhedral garnets (a) and smaller euhedral shapes making up large garnet outline (b).
small garnets at the margins and small euhedral shapes that make up large grain outlines (see Plate 8-12) are consistent with equilibrium between the margins of garnets, at least, and the surrounding biotite, muscovite and quartz. The garnet is an almandine-rich variety with minor quantities of Mg, Ca and Mn, see Appendix III. A temperature estimate is obtained from these schists using the geothermometer of Perchuk (1977) from the distribution of Mg, Fe and Mn in coexisting garnet and biotite. Analyses were taken as close to the garnet-biotite contacts as possible, normally within 20\mu m, and are listed in Appendix III. \( \ln K_D \) values were calculated for each of fifteen garnet-biotite pairs using equation 32 of Perchuk (1977):

\[
\ln K_D^{Mg} = x^{Gnt \cdot Mg} (1 - x^{Bi \cdot Mg})/x^{Bi \cdot Mg} (1 - x^{Gnt \cdot Mg})
\]

where
\[ x_{Mg} = \frac{Mg}{Mg + Fe + Mn} \]

for garnet

Results are listed in Table 8-2.

\( \ln K_D \) is related to temperature by:

\[
\ln K_D = \frac{A}{T} + B \quad (\text{Perchuk, 1977, Eq. 33})
\]

where \( A = 3650 \) and \( B = -2.57 \)

and \( T = \) temperature in °K

This gives a temperature of \( 456^\circ C \pm 56^\circ C \) (95% confidence limit).

Corrections for non ideal distribution of elements between garnet and biotite (Perchuk, 1977, Fig 5) indicates that this is an overestimation by approximately \( 25^\circ C \), and a better estimate is around \( 430^\circ C \).

The area immediately surrounding garnet is composed of quartz, biotite and minor chlorite and garnets contain inclusions of quartz and biotite only. This suggests that the garnet grew at the expense of muscovite and chlorite by a reaction similar to that quoted by Winkler (1974) p. 210:-
Table 8-2. $K_D$ for coexisting garnet-biotite pairs.

<table>
<thead>
<tr>
<th>Gnt $X_{Mg}$</th>
<th>Bio $X_{Mg}$</th>
<th>$\ln K_D$</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.0585</td>
<td>0.481</td>
<td>2.702</td>
</tr>
<tr>
<td>0.0707</td>
<td>0.458</td>
<td>2.408</td>
</tr>
<tr>
<td>0.0703</td>
<td>0.490</td>
<td>2.501</td>
</tr>
<tr>
<td>0.0601</td>
<td>0.450</td>
<td>2.549</td>
</tr>
<tr>
<td>0.0695</td>
<td>0.405</td>
<td>2.210</td>
</tr>
<tr>
<td>0.0757</td>
<td>0.488</td>
<td>2.455</td>
</tr>
<tr>
<td>0.0870</td>
<td>0.413</td>
<td>2.000</td>
</tr>
<tr>
<td>0.0748</td>
<td>0.500</td>
<td>2.515</td>
</tr>
<tr>
<td>0.0776</td>
<td>0.502</td>
<td>2.484</td>
</tr>
<tr>
<td>0.0798</td>
<td>0.494</td>
<td>2.422</td>
</tr>
<tr>
<td>0.0634</td>
<td>0.495</td>
<td>2.673</td>
</tr>
<tr>
<td>0.0845</td>
<td>0.507</td>
<td>2.410</td>
</tr>
<tr>
<td>0.0775</td>
<td>0.485</td>
<td>2.417</td>
</tr>
<tr>
<td>0.0749</td>
<td>0.482</td>
<td>2.442</td>
</tr>
<tr>
<td>0.0882</td>
<td>0.504</td>
<td>2.352</td>
</tr>
</tbody>
</table>

Mean $\ln K_D = 2.436 \pm 0.370$ (95% Confidence limits)

All data calculated from E.D.S. analyses listed in Appendix III, from published thin section MNC 17, location 837713.
Fe chlorite + muscovite + quartz → almandine garnet + biotite + H₂O

Lack of muscovite in the layered schists prevents the growth of garnet and their assemblage implies that temperatures are not high enough to allow the reaction (ibid p. 210):

 chlorite + biotite₁ + quartz ←→ almandine garnet + biotite₂ + H₂O

This suggests that temperatures are close to the lower limit for stability of almandine garnet. The appearance of garnet cannot be used to estimate temperature due to solid solution in, and the varying reactions that produce, garnet, but appears just above the 'stilphnomelane + muscovite out/biotite + muscovite in isograd, ibid, p. 203. The presence of biotite and absence of stilnomelane with chlorite, quartz, + muscovite implies that temperature is above this isograd. Temperatures of around 420°C estimated from garnet-biotite equilibrium implies that the total pressure is at least below 5 kbars and probably less than 4 kbars, ibid, Fig 14-1.

8.3.2. Marbles.

a). Marbles in the gneisses.

Pods of silicate bearing marble occur along strike from 812724 to 835703 see 3.2.2. The mineral assemblage in decreasing modal abundance is dolomite, actinolite, phlogopite, muscovite, chlorite, sphene, calcite with minor rutile, ilmenite, hematite, + quartz, + k-feldspar, see Plate 8-13. Actinolite and k-feldspar commonly show resorbed margins. Phlogopite occurs in association with fine grained aggregates of sphene and partially resorbed actinolite which occasionally pseudomorph actinolite. Rutile occurs as rounded grains within sphene and rounded to euhedral grains in carbonate. Chlorite has patchily developed apple green to pale green pleochroism, sometimes containing patches of muscovite. The above relations indicate that actinolite is being replaced by sphene.
Plate 8-13. Marble at 826713 in S. Sithean Mhor gneisses, showing tremolite (Tr) altering to phlogopite (Ph) and sphene (Sp) in a dolomite (Do) matrix.
and phlogopite and that quartz, k-feldspar and rutile are also involved in reaction.

Hunt and Kerrick (1976) describe the equilibria of sphene-bearing assemblages in impure dolomitic rocks and the above relations indicate their reaction 10:-

\[
\text{rutile + calcite + tremolite + k-feldspar + } H_2O \rightleftharpoons \text{phlogopite + sphene + } CO_2
\]

This reaction explains the resorption textures of actinolite, k-feldspar and rutile and low content of calcite in these rocks.

Other relationships limit the possible \( T - X_{CO_2} \) space, see Fig 8-4 of this assemblage on Fig 8 of Hunt and Kerrick (1976):-

i) Phlogopite, sphene and quartz are stable, thus \( X_{CO_2} \) must be greater than that required for reaction 9:-

\[
\text{Phlogopite + sphene + quartz \rightleftharpoons tremolite + k-feldspar + rutile + } H_2O
\]

ii) Muscovite, calcite and quartz occur whereas zoisite is not present. Therefore temperature must be lower than reaction 15:-

\[
\text{muscovite + calcite + quartz \rightleftharpoons zoisite + k-feldspar + } H_2O + CO_2
\]

Hunt and Kerrick present a review of absolute \( T - X_{CO_2} \) values for the equilibria in the range 0.5 to 5 kbars pressure. Since pressure is not known this produces a range of possible temperatures and \( X_{CO_2} \) values for this assemblage. The data is summarised in Table 8-3.

The reaction involving the resorption of tremolite, by reaction 10 of Hunt and Kerrick in these rocks has gone from left to right which implies either an increase in temperature or decrease in \( X_{CO_2} \) or combination of both.

b) Marbles in the schists.

These marbles are composed of hornblende and biotite foliae interleaved with carbonate. The carbonate portions of the rock are composed of ankeritic dolomite, Mg-Fe-bearing calcite, quartz, with
FIG 8-4  T-X CO₂ relations of marble assembles, S. Sithean Mhor gneisses.

Reactions 9, 10 and 15 after Hunt and Kerrick (1976), Fig 8.

Abbreviations:
Cc calcite
Ksp k-feldspar
Musc muscovite
Phl phlogopite
Qtz quartz
Rut rutile
Sph sphene
Tr tremolite

Possible T-X CO₂ space occupied assemblages of marbles.
Table 8-3. P-T-$X_{CO_2}$ ranges from Hunt and Kerrick (1976)

<table>
<thead>
<tr>
<th>Pressure Kb.</th>
<th>$X_{CO_2}$</th>
<th>T °C</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.5</td>
<td>0.29 - 0.46</td>
<td>400 - 420</td>
</tr>
<tr>
<td>1</td>
<td>0.20 - 0.31</td>
<td>405 - 435</td>
</tr>
<tr>
<td>3</td>
<td>0.19 - 0.30</td>
<td>455 - 490</td>
</tr>
<tr>
<td>5</td>
<td>0.13 - 0.27</td>
<td>475 - 550</td>
</tr>
</tbody>
</table>
minor biotite, muscovite, chlorite, intermediate plagioclase, ilmenite and sphene. The hornblende foliae are composed largely of hornblende, biotite, quartz, intermediate plagioclase, carbonate, epidote, chlorite and ilmenite. Biotite-rich foliae are composed of biotite, quartz, chlorite, muscovite, carbonate and ilmenite. The hornblende and biotite foliae are especially abundant at marble pod margins, and their most probable origin is as infolded biotite schist and amphibolite.

The marble from the schist belt at 843719, composed of ankeritic dolomite and Mg-, Fe-bearing calcite, was used to estimate temperature. The method used is that of Bickle and Powell (1977) which includes a correction for Fe content.

Twenty-five grains of Fe-Mg-bearing calcite were analysed on a microprobe using E.D.S., see Appendix III, and plotted with calibration lines given by Bickle and Powell (1977), Fig. 1, see Fig 8-5. All points lie between 400°C and 480°C lines with most between 420 and 450°C lines, indicating temperatures around 435°C.

The Mg and Fe content of calcite gives an estimation of temperature only when in equilibrium with dolomite. The specimen used here was found to be composed largely of calcite with only a small percentage of dolomite. However, using Fig 2 of Bickle and Powell (1977) and an average value for the MgCO₃ content of calcite, the FeCO₃ content of dolomite indicates temperatures of around 420°C in close agreement with the estimate from calcite. This indicates that calcite is in equilibrium with dolomite and the temperature estimate is valid.

The temperature calibration presented by Bickle and Powell (1977) is for 5 kbrs total pressure. Pressure in the schists is unknown but is inferred to be less than 5 kbrs (see 8.2.2.). Goldsmith and Newton (1969) have estimated that the solubility of MgCO₃ in CaCO₃ increases by 0.12 mole% per 10 kbrs. From Bickle and Powell Fig 1., 0.1% mole MgCO₃ is equivalent to 6°C, thus the maximum error possible in the temperature
FIG 8-5  Mg CO₃ vs. Fe CO₃ in calcite for marble at 842719.

25 points.
Calibration lines after Bickle and Powell (1977), figures in degrees centigrade.
estimate is 30°C, and a more likely error is around 15-20°C. A probable temperature of metamorphism is therefore around 420°C.

Marbles in the gneisses were found to be unsuitable for temperature estimation as they are largely composed of dolomite and the calcite contains little or no MgCO₃. The marbles often show brecciated textures and the marble at 811723 is associated with cataclasites, see 3.2.2. It is therefore probable that calcite has re-equilibrated at low temperatures during late stage brittle deformation.

Since Fe is abundant in these rocks it is not possible to apply the equilibria of Hunt and Kerrick (1977). However the presence of epidote in these rocks indicates that the metamorphic fluid is more H₂O rich than that in the biotite schists. This point will be discussed later, see 8.4.1.

Ferry (1977) describes the assemblages of Fe-bearing metacarbonate rocks with increasing grade, from South Central Maine. Similar assemblages here place them between his biotite-chlorite isograd and amphibolite-anorthite isograd, at approximately the garnet grade for pelites of the same region. Temperature estimates for the area are 370 to 550°C and for pressure, 2.5 to 3.8 kbars. This is in agreement with temperature estimates of around 420°C and pressures of less than 5 kbars, i.e. garnet grade, for these marbles.

8.3.3. Quartz-magnetite schists and garnetiferous amphibolites.

a. Quartz-magnetite schists.

These are composed of quartz, magnetite, cummingtonite, ± biotite, ± stilpnomelane, ± chlorite, ± carbonate, ± garnet, see Plate 8-14. Amphiboles of the cummingtonite-grunerite series are reported in ironstones from medium to high grade metamorphism, Klein (1973), and in the garnet and staurolite zones by James (1955). Formation of amphibole can occur by two means.
Plate 8-14. Quartz-magnetite schist showing quartz-rich layers (Q) and magnetite-rich layers (M) with cummingtonite (C) and stilpnomelane (S), from 837728.
i) ferrodolomite + quartz + H₂O ⇌ grunerite-cummingtonite + calcite + CO₂
(Klein, reaction 4), and
ii) minnesotaite ⇌ grunerite-cummingtonite + quartz + H₂O
(Klein, reaction 8).

The assemblage ankeritic dolomite + quartz is stable in the marbles even though the presence of epidote indicate an H₂O-rich fluid. However carbonate occurs as only a minor constituent in quartz-magnetite schists. Thus it is more probable that cummingtonite originated from a silicate mineral such as minnesotaite and that the ironstones therefore originated as quartz silicate facies.

b) Garnetiferous amphibolite.

These are composed of Mn-garnet, Mn-cummingtonite, (for analyses see Table 7-4), quartz, ilmenite, ± chlorite, ± biotite. Mn-bearing carbonate survive to medium and high grades of metamorphism, Klein (1973), thus as in the quartz-magnetite schists, these probably originated as a silicate facies of manganiferous ironstone. Spessartine garnet can occur at lower temperatures and pressures than almandine garnet and is therefore not an indication of the same grade of metamorphism as shown by the pelites.

The assemblage quartz-magnetite-cummingtonite-garnet indicates at least the garnet zone of pelites which is consistent with assemblages found in the pelites and marbles.

8.3.4. Amphibolites and hornblende-chlorite schists.

Assemblages in these rocks fall into three main groups:-

a. hornblende, plagioclase (andesine), quartz, garnet, epidote, sphene, ilmenite, opaques.

b. hornblende, plagioclase (oligoclase), quartz, opaques,
+ biotite, + garnet, + carbonate.

c. chlorite, carbonate, epidote, plagioclase (albite to oligoclase), quartz, + opaques, + hornblende.

Assemblage (a) occurs in the deformed amphibolites in the gneisses. In undeformed dykes with relict ophitic texture, the assemblage is hornblende and plagioclase of labradorite or bytownite composition. These rocks have therefore changed little since intrusion and only the pyroxenes have altered to hornblende. On deformation, the plagioclase composition changes to andesine and other silicates listed in assemblage (a) appear. Garnets are locally abundant, see 3.2.4., but are unstable and partially altered to hornblende, epidote, quartz and plagioclase. This assemblage belongs to the lower to middle part of the medium grade of Winkler (1974) or to the Staurolite and Kyanite zones of Turner (1968).

Assemblage (b) is found in the amphibolites and hornblende schists within the schist belt. Plagioclase is more sodic than those in the gneiss (oligoclase). Garnets are rare and where found are partially replaced by biotite, plagioclase and quartz. This assemblage belongs to the upper part of low grade or lower part of the medium grade of Winkler (1974) or to the Almandine facies of Turner (1968).

Assemblage (c) is found within the hornblende schists and in association with carbonate viens up to 10 cms wide by up to 3 m long, parallel to the foliation. In schists containing both hornblende and chlorite, chlorite occurs in foliae associated with epidote, carbonate and quartz. The abundance of carbonate in chlorite schists and the occurrence of carbonate viens suggests a metamorphic fluid rich in CO₂. These factors suggest that hornblende is converted to chlorite schist by the reaction.

\[ \text{hornblende} + \text{H}_2\text{O} + \text{CO}_2 \rightleftharpoons \text{chlorite} + \text{epidote} + \text{calcite} + \text{quartz}. \]

reported by Harte and Graham (1975). Chlorite schists are always found
in association with finer grained, well foliated hornblende schists, and textures indicate that the reaction is syntectonic. The increase in carbonate suggests that the reaction is caused, in part, by an increase in the proportion of \( \text{CO}_2 \) in the metamorphic fluid and the decrease in anorthite content of plagioclase from mid-oligoclase to albite or lower oligoclase suggests that a slight decrease in temperature is also involved.

8.4. Summary and Conclusions

8.4.1. Metamorphic conditions.

Assemblages in the schists indicate garnet grade conditions with temperatures around \( 420^\circ \text{C} \) and pressure less than 5 kbrs. Assemblages in the amphibolites and the retrogression of microcline to plagioclase and quartz indicates that the gneisses are of slightly higher grade. Amphibolite assemblages indicate lower Amphibolite facies for the gneisses, equivalent to the Staurolite-Kyanite grade of pelites. The schists are therefore retrograde with respect to the gneisses.

The disappearance of epidote in the gneiss-schist transition indicates \( \text{H}_2\text{O} \)-rich fluids in the gneisses and more \( \text{CO}_2 \)-rich fluids in the schists. The hornblende schists in the schist belt show local alteration to chlorite schist due to influx of \( \text{CO}_2 \). This implies that fluids in the hornblende schists were \( \text{H}_2\text{O} \)-rich relative to the surrounding biotite schists at the time of metamorphism. Influx of \( \text{CO}_2 \) from the biotite schists then led to the formation of chlorite schists. Influence of the metamorphic fluid composition by neighbouring rocks can also be seen in the marbles. Adjacent to amphibolites the marbles contain epidote. Since the \( \text{CO}_2 \) content of the fluid in the schists is too high to stabilise epidote, this implies influx of \( \text{H}_2\text{O} \) from the adjacent amphibolite. These relations indicate that the composition of the metamorphic fluid varies with lithology and also
that these differences are, to a large extent, maintained during metamorphism, migration occurring only over distances of the order of a metre across strike, when aided by deformation and fine grain size. Similar variation in metamorphic fluid composition and limited fluid mobility is noted by Harte and Graham (1975) in the southwest Highlands of Scotland.

8.4.2. Relationship between metamorphism and deformation.

The simultaneous occurrence of metamorphic reactions, i.e. regression of microcline and resorption of epidote, and a change in texture and grain size suggests that the two events are interdependent.

It has been shown that the metamorphic fluid is more CO₂ rich in the schists and also that deformation facilitates migration of fluid across strike. Mobility of the fluid will also be aided by an increase in grain boundary area, a consequence of grain size reduction, which provides more pathways for fluid migration. A decrease in grain size will therefore allow the more CO₂ rich fluids of the schists to penetrate the gneisses promoting the resorption of epidote. K-feldspar is metastable at the time of deformation affecting the gneiss-schist contact. The deformation triggers off the retrograde reaction and k-feldspar is replaced by plagioclase and quartz.

While the deformation has promoted the reactions in the rocks, it is also possible that the reactions have promoted the deformation. It has been noted that recrystallised plagioclase and biotites have different compositions from old grains, Vernon (1975), Vernon (1976) p. 199, and they suggest that release of chemical free energy may help nucleation and recrystallisation. Though no chemical difference is noted between old and new grains in individual specimens, the change in anorthite content of plagioclase may promote recrystallisation. Evidence of such processes is seen in the replacement of smaller strained
k-feldspar grains by mosaics of fine grained plagioclase (Plate 8-5) whereas cores of large porphyroblasts are replaced by single grains of plagioclase.

From textures in the gneisses, the processes of grain size reduction start before metamorphic reactions occur. At a critical grain size the strain energy triggers the retrogression of k-feldspar and fluid compositions are modified enough to start the reaction involving epidote. These reactions in turn aid grain size reduction and the attainment of a stable texture in the schists.

Since the deformation centred on the schists downgrades the gneisses, it must partly post date the deformation represented in the gneisses. Little or no deformation in the gneisses allows them to keep their mineral assemblages metastably. Thus the SE dipping lineations occurring throughout the schist belt post date the NW dipping lineations though minor structures in the schists indicate that the sense of shear is unchanged, i.e. negative or SW side down, see 7.2.2.

The processes involved in altering gneisses to schists will cause the gneiss-schist boundary to migrate into the gneisses with time. Such a process would have started at the true gneiss-sedimentary schist boundary and migrated to the SW as deformation and metamorphism cause grain size reduction in the gneisses adjacent to the contact.
CHAPTER 9 CONCLUSIONS

9.1. Introduction

The major part of the present work has been to attempt a quantitative analysis of the deformation history in the Lewisian rocks of the Gairloch area. It has been shown that finite strain can be analysed using quartz shape fabrics in quartzo-feldspathic gneisses. The method here is an initial attempt at the technique, and the difficulties of adequately measuring highly irregular quartz particles make themselves apparent in the size of the confidence limits of the results. The procedure has much scope for improvement and further investigation, such as, the effects of sample plane orientation in relation to the principal planes of finite strain, and the improvement of measuring techniques, eg. possible use of optical and computerised methods. Lack of markers has previously inhibited quantitative analysis of strain in basement areas. However, quartzo-feldspathic gneisses are widespread throughout these areas and thus provide much scope for the analysis of finite strain.

In the following sections, the results from the application of the above technique in the two areas of gneiss are compared and correlated, and a strain profile for the whole area constructed. Finally, the deformation and metamorphic history of the Gairloch area is discussed in relation to the rest of the Lewisian gneisses of mainland NW Scotland.

9.2. Synthesis of Structural Analysis

9.2.1. The gneisses.

The main points of the deformation histories as deduced by field evidence, strain analysis and strain path modelling of the Loch Tollie gneisses and S. Sithean Mhor gneisses are summarised at the end of chapters
5 and 6. These histories differ in detail but both end with a main phase of negative northwesterly shearing that is responsible for the prominent steeply NE dipping foliation continuous throughout the central part of the Lewisian outcrop at Gairloch. These phases are therefore probably time equivalent and are the last major tectonic event recorded by the gneisses. The area thus represents a large steeply NE dipping shear zone, the boundaries of the central zone of intense deformation being marked in the NE by the hinge of the Tollie antiform, and in the SW by the division between the areas of homogeneous and inhomogeneous deformation.

A phase of NW positive shearing predating the main negative shearing event and contemporaneous with dyke intrusion, is implied for the S. Sithean Mhor gneisses, but no evidence for a similar phase of shearing was found in the Tollie gneisses. The northern extent of this phase therefore presumably terminates within the schist belt. In the Tollie area the gneisses possessed an initial fabric prior to negative shearing. It was probably inhomogeneous in nature, as seen on Craig Mhor Tollie and the NE limb of the Tollie antiform, and is post-dyke in age. This initial fabric is not necessarily connected with the shear zone, but must be contemporaneous with, or post date, the phase of positive shearing, since it postdates dyke intrusion.

9.2.2. Deformation in the schist belt.

No strain measurements have been attempted in the schist belt due to lack of suitable material but several minor features and textures indicate high strains (see 7.2.2.). Also the strain profile for the Loch Tollie gneisses shows strain progressively increasing towards the schist belt (Fig 5-13), and quartz fabrics in the S. Sithean Mhor gneisses increase sharply in intensity, within one to two metres of
the gneiss-schist contact, suggesting that schists are a site of high deformation. Since schists occupy the centre of the shear zone, strains are likely to be as least at high as those in the immediately adjacent gneisses; \( \varepsilon \approx 8 \) for the Loch Tollie gneisses, and \( \varepsilon \approx 2 \) in the S. Sithean Mhor gneisses. However, the lower grade of metamorphism shown by the schists (garnet grade) compared to the gneisses (stauolite-kyanite grade), 8.4.1., and the textural and mineralogical transformation of gneisses to schists (8.3.), indicates that the last deformation in the schists is younger than that in the gneisses. The preservation of the steeply NE dipping foliation and minor features within the schists (7.2.2.) indicates that deformation is also by simple shear, and that the sense is negative, ie. SW side down. The moderately SE plunging mineral lineations throughout the area imply a change in shearing direction from NW to SE. The extent of the shearing deformation postdating the deformation in the gneisses is unknown and indicates that estimates of shear strain for the schists based on the strains in the adjacent gneisses are likely to be lower than the true values.

9.2.3. The dykes.

The dykes in the Tollie gneisses are interpreted as largely pre-tectonic, predating the formation of the initial fabric of the NE limb, with evidence of some possible syntectonic dyke intrusion confined to the SW steep limb of the Tollie antiform (2.4.4.). In the S. Sithean Mhor gneisses dyke intrusion is syntectonic with the earlier phase of positive shearing (3.4.3.). The occurrence of dykes with abundant schlieren of gneiss in the highly deformed region of the S. Sithean Mhor gneisses (3.3.3.) similar to those in the Loch Tollie gneisses suggests that here too some dyke intrusion is syntectonic.

Amphibolite sheets and hornblende schists within the schist belt are shown to be of igneous intrusive origin (7.2.3.). Since all sheets are deformed, the time relations of intrusion are unclear but must at
least be syn- or pre-tectonic. Since the majority of dykes throughout the zone of steep foliation in gneisses and schists are well foliated, syntectonic intrusion cannot have been extensive and probably only occurred in the earlier stages of deformation.

Since strain in the schist belt is high and the shear direction subvertical, dykes in the Loch Tollie gneisses and those in the S. Sithean Mhor gneisses will have been separated in depth by some tens of kms at the time of intrusion. This presents the possibility that there is more than one phase of dyke intrusion. However, several factors suggest a single major phase:-

1. All dykes have similar trend and dip except those in the crest of the Tollie antiform where their orientation can be shown to be tectonically dictated (5.4.4.). Though, in the zone of strong deformation, this trend could be tectonically induced, it also persists in the marginal areas, i.e. the NE limb of the Tollie antiform and the area of inhomogeneous deformation in the S. Sithean Mhor gneisses.

2. Dykes in both sets of gneisses show similar time relations to the main, negative phase of shearing suggesting they are of similar age.

3. The dykes and hornblende schists show similar chemistry throughout the area (Holland and Lambert, 1973).

4. Nowhere in the area was evidence found of crosscutting relations between dykes.

Thus the simplest and most probable interpretation is that all dykes throughout the gneisses and schists are of a single major intrusive phase. This is in agreement with observation by Park (1970), Evans and Tarney (1964) and Crane (1978). Dyke intrusion occurred largely before the main phase of shearing during an earlier phase of positive shearing, and the last stages of intrusion may have occurred in the early stages of the main phase of shearing, but if so, was
confined to the zone of high deformation.

That the dykes are syntectonic with the earlier phase of shearing may explain the abundance and thickness of amphibolites in the schist belt, since foliated semipelites would offer favourable weak planes for dyke intrusion. Since pods of quartz-magnetite schist and marble occur in strings parallel to dyke margins, dykes appear to have been intruded along original bedding planes suggesting that deformation had already rotated bedding into a near vertical attitude at the time of dyke intrusion.

9.2.4. Construction of a strain profile across the shear zone.

Strain profiles for the gneisses have already been constructed (Fig 5-13 and Fig 6-10). A strain profile for the schists is more tentative and has been constructed in the following manner. The schists adjacent to the Loch Tollie gneisses are assumed to have strains of $\gamma \approx 8$. The belt of gneisses within the schists are assumed to have similar strains to the S. Sithean Mhor gneisses, to which they are closest geographically, and in petrography and texture. The schists adjacent to the S. Sithean Mhor gneisses have higher strains than the gneisses, so a figure of $\gamma \approx 6$ is chosen as a conservative estimate. The total profile is shown in Fig 9-1.

The displacements for the component parts of the shear zone, and thus the total displacement, can be estimated by calculating the respective areas under the curve in Fig 9-7, (Ramsay and Graham, 1970). This yields displacements of 7.6 kms for the Loch Tollie gneisses, 1.9 kms for the S. Sithean Mhor gneisses and 20 kms for the schists, giving a total of 29.5 kms, along the shear direction, ie. moderately to steeply dipping NW. Resolving the displacement into vertical and horizontal components using shear direction plunges of $55^\circ$ in the Tollie gneisses, $71^\circ$ in the S. Sithean Mhor gneisses, and an average $64^\circ$ in the schist belt, yields a vertical displacement of 26 kms of negative sense and a horizontal dextral displacement of 14.0 kms.
FIG 9-1 Strain profile across the Gairloch area.

- SW S. Sithean Mhor Gneisses
- area of homogeneous deformation.
- Gairloch Schist Belt
- Loch Tollie Gneisses
- SW steep limb
- flat belt
- NE limb

Scale
0.5 km.
The uncertainties in the calculations of these figures, eg. fitting of data to models and the assumptions in the schist belt, means that these figures may be taken as rough guides only, and due to later deformation in the schists are probably underestimates. They do, however, indicate that displacements are large and that the shear zone is therefore a major tectonic break between the rocks to the NE and SW.

9.3. Age of the Deformation

Age determinations for the Gairloch region are presented by three authors; Park and Evans (1965), Holland and Lambert (1963) and Moorbath and Park (1971). A summary of their data is given in Table 9-1. The data is discussed in the light of the present structural interpretation and the time scale of the Lewisian as a whole.

9.3.1. The dykes.

Three dates for amphibolitized dyke material preserving relict ophitic texture are given by Moorbath and Park (1971) and one by Park and Evans (1965), see Table 9-1. Dates of around 2,100 and 2,200 Ma for the Aundrary amphibolite is in close agreement with the age range of Scourie dykes found in Assynt by Evans and Tarney (1964) of 2,200 to 1,900 Ma. Moorbath and Park have assigned this to the Inverian age and interpreted the date as a minimum age for intrusion. Ages for dykes on S. Sithean Mhor give 1700 ± 40 and 1685 ± 40 Ma. Ages of amphibolitized dykes in Assynt give ages down to 1400 Ma and have been shown to be modified by later Laxfordian metamorphism. However, the Inverian age for undeformed parts of the Aundrary amphibolite has remained unaffected by two later deformations of most of the sheet, and it is possible that the ages for dykes on S. Sithean Mhor are truely ages of amphibolitization and therefore close to the age of intrusion. It is possible therefore that dyke intrusion continued longer at Gairloch than at Assynt.
Table 9-1. Summary of age determinations in the Gairloch area.

<table>
<thead>
<tr>
<th>AUTHOR</th>
<th>METHOD</th>
<th>ROCK</th>
<th>LOCATION</th>
<th>AGE</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>C Sr isotopes</td>
<td>undeformed pegmatites</td>
<td>NE limb of Tollie antiform</td>
<td>1720 ± 1630 Ma</td>
</tr>
<tr>
<td>2</td>
<td>B k/Ar on hornblende</td>
<td>weakly foliated dyke</td>
<td>Buanichean gneisses</td>
<td>1712±35, 1685±40 Ma</td>
</tr>
<tr>
<td>3</td>
<td>C Sr isotopes</td>
<td>deformed pegmatites</td>
<td>SW limb of Tollie antiform</td>
<td>c. 1500 Ma</td>
</tr>
<tr>
<td>4</td>
<td>B k/Ar on muscovite</td>
<td>deformed pegmatite</td>
<td>Buanichean gneisses</td>
<td>1426 ± 35 Ma</td>
</tr>
<tr>
<td>5</td>
<td>A k/Ar, whole rock</td>
<td>deformed dyke</td>
<td>Buanichean gneisses</td>
<td>1240 ± 50 Ma</td>
</tr>
<tr>
<td>6</td>
<td>B k/Ar, hornblende</td>
<td>unfoliated dyke</td>
<td>S. Sithean Mhor gneisses</td>
<td>1700±40, 1686±35 Ma</td>
</tr>
<tr>
<td>7</td>
<td>B k/Ar, hornblende</td>
<td>foliated dyke</td>
<td>S. Sithean Mhor gneisses</td>
<td>1472 ± 40 Ma</td>
</tr>
<tr>
<td>8</td>
<td>A k/Ar, whole rock</td>
<td>unfoliated dyke</td>
<td>S. Sithean Mhor gneisses</td>
<td>1120 ± 45 Ma</td>
</tr>
<tr>
<td>9</td>
<td>B k/Ar, hornblende</td>
<td>garnetiferous amphibolite pod</td>
<td>Ard Ial</td>
<td>taig gneisses</td>
</tr>
<tr>
<td>10</td>
<td>A k/Ar, whole rock</td>
<td>hornblende-pyroxene gneiss</td>
<td>Ard Ial</td>
<td>taig gneisses</td>
</tr>
<tr>
<td>11</td>
<td>B k/Ar, hornblende</td>
<td>acid gneiss</td>
<td>Ard Ial</td>
<td>taig gneisses</td>
</tr>
<tr>
<td>12</td>
<td>B k/Ar, biotite</td>
<td>biotite schist</td>
<td>Kerry river</td>
<td>1406 ± 30 Ma</td>
</tr>
<tr>
<td>13</td>
<td>A k/Ar, whole rock</td>
<td>hornblende schist</td>
<td>Aundrary amphibolite</td>
<td>1400 ± 60 Ma</td>
</tr>
<tr>
<td>14</td>
<td>B k/Ar, hornblende</td>
<td>fine grained hornblende schist</td>
<td>Aundrary amphibolite</td>
<td>1409 ± 35 Ma</td>
</tr>
</tbody>
</table>
Table 9-1 (continued)

<table>
<thead>
<tr>
<th>AUTHOR</th>
<th>METHOD</th>
<th>ROCK</th>
<th>LOCATION</th>
<th>AGE</th>
</tr>
</thead>
<tbody>
<tr>
<td>15 B</td>
<td>k/Ar, hornblende</td>
<td>fine grained hornblende schist</td>
<td>Aundrary amphibolite</td>
<td>1499 ± 40 Ma</td>
</tr>
<tr>
<td>16 B</td>
<td>k/Ar, hornblende</td>
<td>medium grained hornblende schist</td>
<td>Aundrary amphibolite</td>
<td>1543 ± 40 Ma</td>
</tr>
<tr>
<td>17 B</td>
<td>k/Ar, hornblende</td>
<td>medium grained, weakly foliated hornblende schist</td>
<td>Aundrary amphibolite</td>
<td>1952 ± 55 Ma</td>
</tr>
<tr>
<td>18 B</td>
<td>k/Ar, hornblende</td>
<td>coarse grained, weakly foliated amphibolite</td>
<td>Aundrary amphibolite</td>
<td>2104 ± 110, 2080 ± 120 Ma</td>
</tr>
<tr>
<td>19 B</td>
<td>k/Ar, hornblende</td>
<td>coarse grained, weakly foliated amphibolite</td>
<td>Aundrary amphibolite</td>
<td>2213 ± 80 Ma</td>
</tr>
<tr>
<td>20 B</td>
<td>k/Ar, hornblende</td>
<td>fine grained hornblende schist</td>
<td>Aundrary amphibolite</td>
<td>1398 ± 35 Ma</td>
</tr>
</tbody>
</table>

Authors:
A - Park, R.G. and Evans, C.R. (1965)
C - Holland, J.G. and Lambert, R.St.J. (1973)
9.3.2. The gneisses.

In the NE limb and crest of the Tollie antiform, undeformed pegmatities give ages from 1720 to 1630 Ma and indicate that the deformation, and thus the formation of the Tollie antiform, is older than about 1700 Ma. The antiform is post-dyke in age and its formation is therefore placed between 2,200 and 1700 Ma.

In the steep SW limb of the antiform, highly deformed dykes and pegmatites give ages from 1240 ± 50 Ma to 1500 Ma. The age of 1240 ±50 Ma (Park and Evans, 1965) is much younger than the rest, which lie in the range 1426 to 1500 Ma, and is probably therefore anomalous. Deformation in the SW limb therefore ended between 1425 to 1500 Ma.

Less strongly foliated, coarser grained dykes give older ages of 1712 ± 35 and 1685 ± 40 Ma. These ages are probably relict from the initial formation of the Tollie antiform with some overprint by later deformation.

A well foliated dyke from S. Sithean Mhor gneisses gives an age of 1472 ± 50 Ma (Moorbath and Park, 1971) which is in agreement with dates from the SW limb of the Tollie antiform and supports the hypothesis that the phases of negative shearing in both sets of gneisses is contemporaneous.

9.3.3. The schists.

One date from a biotite schist and five dates from the finest grained parts of the Aundrary amphibolite give ages in the range 1398 ± 35 to 1409 ± 35, dating the end of metamorphism and deformation in the schist belt at around 1400 Ma, slightly later than the end of deformation and metamorphism in the adjacent gneisses. This is in agreement with field and petrological evidence that deformation in the schists is later, and grade of metamorphism lower than that in the gneisses (8.4.1.).

9.3.4. Summary of deformation history.

The major structural events and their time relations are summarised in Table 9-2, and show that the zone of active deformation and metamorphism
### Table 9-2: Summary of the structural history of the Gairloch area.

<table>
<thead>
<tr>
<th>S. SITHEAN MHOR GNEISSES</th>
<th>SCHIST BELT</th>
<th>LOCH TOLLIE GNEISSES</th>
<th>TIME (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Area of inhomog. defm.</td>
<td>Area of homog. defm.</td>
<td>Tectonic schists</td>
<td>Metased. schists</td>
</tr>
<tr>
<td>dyke intrs. and +ve NW shear</td>
<td>dyke intrs. and +ve NW shear</td>
<td>dyke intrs. and fm. of initial fabric</td>
<td>dyke intrus. and inhomog. defm.</td>
</tr>
<tr>
<td>-ve NW shear</td>
<td>-ve NW shear</td>
<td>-ve NW shear</td>
<td>-ve NW shear</td>
</tr>
<tr>
<td>? late dykes ?</td>
<td>? late dykes ?</td>
<td>fm. of Tollie antiform</td>
<td>end of dyke intrs. dykes?</td>
</tr>
<tr>
<td>end of defm. and dyke intrs.</td>
<td>end of defm. and dyke intrs.</td>
<td>end of defm. and dyke intrs.</td>
<td>end of defm. and dyke intrs.</td>
</tr>
<tr>
<td>END OF DYKE INTRUSION</td>
<td>END OF DEF.</td>
<td>END OF DEF.</td>
<td>END OF DEF.</td>
</tr>
<tr>
<td>-ve NW shear</td>
<td>-ve NW shear</td>
<td>-ve NW shear</td>
<td>-ve NW shear</td>
</tr>
<tr>
<td>end of defm.</td>
<td>end of defm.</td>
<td>end of defm.</td>
<td>end of defm.</td>
</tr>
<tr>
<td>fm. of tectonic schists</td>
<td>fm. of tectonic schists</td>
<td>fm. of tectonic schists</td>
<td>fm. of tectonic schists</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Abbreviations:** fm. - formation, defm. - deformation, intrs. - intrusion, Amph. - Amphibolite, homog. - homogeneous, Metased. - Metasedimentary
narrowed towards the schist belt with time.

The first deformation is the formation of the initial fabric represented by the NE limb of the Tollie antiform and the phase of positive shearing in the S. Sithean Mor gneisses, which is syntectonic with dyke intrusion. Dykes of the region of inhomogeneous deformation have not been dated but doleritic, undeformed dykes of similar trend from the Rhuadh Meallan belt, approximately 7 kms southwards, give typical Scourie dyke ages of 2213 ± 55 and 2246 ± 50 Ma (Moorbath and Park, 1971). Assuming similar ages for the dykes south of S. Sithean Mor the phase of positive shearing is dated at around 2,200 Ma, i.e. in the later stages of the Inverian episode.

The phase of negative shearing was initiated between 2200 and 1700 Ma and ended between 1425 and 1500 Ma, spanning the Laxfordian episode. The later deformation within the schists, of around 1400 Ma, is a late Laxfordian event. Periods of non-activity must have occurred between the phases of differing shear sense or direction, and the duration of these periods is unknown. The break between NW plunging shearing in the gneisses and SE plunging shearing in the schists is marked by a slight decrease in metamorphic grade (8.4.1.) and from the dates of Table 9-1 lasted between ~50 and 100 Ma. The earlier change from positive to negative shearing is not marked by any notable change in metamorphic grade and thus it is unlikely that this period of inactivity was any longer. Thus, shearing activity was largely continuous from late Inverian (~2200 Ma) through to late Laxfordian times (~1400 Ma), a period of some 800 Ma.

9.3.5. Original ages of gneisses and schists.

Lead isotope data from the Torridon, Gairloch and Gruinard Bay gneisses yields a single isochron and a date of 2890 ± 180 Ma (Moorbath and Park, 1971). They conclude that the basement gneisses were already
in existence at about 2900 Ma ago and underwent high grade metamorphism at this time.

The age of the Gairloch metasediments is unclear, but they are cut by the dykes dated at 2,200 Ma and therefore must at least predate this age. Park (1970) suggested a post-Badcallian (Early Scourian), pre-Inverian age for the metasediments in Gairloch on structural grounds. Crane (1978) reached a similar conclusion for the age of the Loch Maree series (2,400 - 2,300 Ma) based on structural and microfabric analysis. Bickerman, Bowes and Van Breemen (1975) obtained a data of sedimentation of 2.2 byrs from Rb-Sr isotope isochrons. However, two factors suggest that this date is unreliable:

1. The samples yielding this age came from the Ard gneisses and adjacent schists on the assumption that the gneisses were originally derived from sedimentary schists, as proposed by Park (1964). Later work, Park (1965), shows that in fact these schists are tectonically derived from the gneisses. The date obtained thus represents, if anything, an original age for the gneisses and not for the metasediments.

2. The gneisses belong to the S. Sithean Mhor group for which Moorbatch and Park (1971) obtained the age of \(~ 2900 Ma\) from lead isotopes, far older than that proposed by Bickerman et al (1975). The schists and gneisses come from the zone of high deformation with the shear zone where Rb and Sr are likely to be mobile. This would account for the anomalous age. Similar conclusions are reached by Crane (1978) who suggests that muscovite and microcline porphyroblasts in the gneisses and metasediments N. of Loch Maree are evidence that the system was open with respect to k, Rb and Sr during Laxfordian metamorphism.

9.4. Comparisons with Previous Research

Structural analysis of the area has been published by Park (1970) for the Loch Tollie gneisses, Bhattacharjee (1964) for the area NW of
Loch Tollie, Ghaly (1966) for the S. Sithean Mhor gneisses and Park (1964) for the schist belt.


Park separates the deformation into four main phases with a fifth late stage brittle phase. Essentials of his analysis are presented in Table 9-3.

His interpretation implies superimposition of successively younger deformation phases to the SW. The Tollie antiform itself is interpreted as a D$_3$ structure modified on the SW limb by D$_4$. The lineations in the SW limb are interpreted as intersection structures, and their variation in plunge due to variation in the relative orientation of S$_3$ and S$_4$ planes.

In comparison with the present interpretation, the initial fabric of the NE limb is equivalent to the first and second deformation phases and the phase of negative shearing equivalent to the third and fourth phases. However, the present interpretation differs from Park's on several points:-

i) Park's model of the deformation as a D$_3$ fold flattened by D$_4$ is similar to the irrotational model of 5.4.3., here. However, it was found that this model fails to explain the range of fabrics and is incompatible with the variation of fold interlimb angles across the antiform. Here, the Tollie antiform is interpreted as a structure produced by simple shear superimposed on an initial fabric.

ii) Park's analysis involves four main phases of deformation whereas the present interpretation implies only two.

iii) Lineations in the SW limb are interpreted by Park as intersection structures. However, the lineations are seen to be composed of rodding structures on the surfaces of quartz aggregates, i.e. are a property of a shape fabric and cannot therefore be an intersection

<table>
<thead>
<tr>
<th>Phase</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>D₁</td>
<td>Gneissose layering ($S₁$)</td>
</tr>
<tr>
<td>D₂</td>
<td>NE limb of Tollie antiform, isoclinal folds ($F₂$), axial planes subvertical trending NW/SE, axial planar foliation, $S₂$</td>
</tr>
<tr>
<td>D₃</td>
<td>Crest of antiform, recumbent folds ($F₃$), gently dipping foliation, $S₃$</td>
</tr>
<tr>
<td>D₄</td>
<td>SW limb of Tollie antiform, isoclinal folds ($F₄$), with associated lineations ($L₄$), vertical NW/SE trending foliation.</td>
</tr>
<tr>
<td>D₅</td>
<td>Brittle folds, faults, crushbelts, and pseudotachylite development.</td>
</tr>
</tbody>
</table>
feature. Their variation in plunge across the antiform is thought to approximate to the passive rotation of lines during simple shear.

iv) In Park's analysis the dykes are intruded during D₃. This is incompatible with the present interpretation which implies that the dykes were intruded prior to the formation of the initial fabric (equivalent to Park's D₂).

9.4.2. Area NW of Loch Tollie, Bhattacharjee (1964).

This area is the northwestwards continuation of the Tollie antiform and Bhattacharjee's analysis is similar to that of Park. He divides the deformation into three phases; the Main phase, the Mid phase and the Late phase which are broadly equivalent to Park's D₁ and D₂, D₃ and D₄, and D₅ respectively. The same correlations and differences to the present interpretation therefore also apply here.

9.4.3. The South Sithean Mhor gneisses, Ghaly (1966), Park (1964).

Ghaly mapped the area on the scale of 24" to 1 mile. In his structural analysis, he divided the deformation into three phases; the main phase, producing the steep NE dipping foliation present throughout much of the area, the mid phase, parallel to and folding the main phase foliation causing a change in plunge of lineation across strike, and a late phase, causing variation in the trend of the main phase foliation in the SW.

The present interpretation is at variance with Ghaly's analysis on two main points:-

i) The change in lineation orientation is interpreted as due to a change in shearing direction affecting the quartz shape fabrics, and not due to major isoclinal folds.

ii) The late phase deformation postulated in the SW (here known as the area of inhomogeneous deformation) occurs where isotropic quartz fabrics are common. Folds in the banding do not fold strong quartz fabrics.
as they should if superimposed on Ghaly's main phase deformation. It is therefore more probable that the area is one of low deformation relatively unaffected by the main deformation.

Park (1964) mapped the NE part of the gneisses and postulated a major isoclinal fold, the axial trace of which lies within the discontinuous bands of marble from 812724 to 835703 (3.2.2.). This seems unlikely as the fold would have to have a hinge orientated to within a few degrees of horizontal, whereas all minor folds in the area and the major Mill na Claise folds plunge moderately SE. A fold structure is not necessary if the outcrops of marble are regarded as distinct pods within the quartzo-feldspathic gneisses as their discontinuous outcrop suggests.

Both Park and Ghaly divided the amphibolites into two suites on the basis of degree of deformation. However, deformation throughout the amphibolites tends to be inhomogeneous in nature and they retain occasional pods of relict ophitic texture even when largely converted to hornblende schist. The constant orientation of all dykes suggests a single phase of intrusion and that variations in texture are due to inhomogeneous deformation rather than significantly differing ages of intrusion.

9.4.4. The schists, Park (1964).

The Gairloch schist belt SW of the Flowerdale fault has been mapped and structurally analysed by Park (1964) who divided the deformation into three phases, two of major folding and a third of minor folding and brittle deformation. Major folds are recognised from changes in lineation plunge, on the assumption that the variation is due to folding, and on the shape and thickness of the amphibolite sheets.

The present interpretation views the schists to occupy the centre
of a large scale shear zone, where the mineral lineations are thought to represent the shear direction. In this case, though major folds possibly exist, the changes in lineation plunge indicate changes in shear direction and not the location of major fold axes. The amphibolites are interpreted as dykes of the Scourie suite and only one undoubted fold closure outlined by an amphibolite is seen in the area, at Mill na Claise 833714. This fold plunges steeply SE from lineations at the fold nose (Park, 1964). The reason for its formation within the shear zone is uncertain but is possibly connected with the Ard Ialltaig mass, situated along strike to the NW. This mass is here interpreted as a remnant of older gneisses, intruded by the S. Sithean Mhor gneisses, that has acted as a resistant augen throughout the history of deformation (7.3.2.). The mass is free of dykes and a dyke deflected to the SW of the mass at the time of intrusion would undergo folding to form a fold similar to that at Mill na Claise during the phase of positive shear, see Fig 9-2, with the second closure located under the Torridonian cover to the SE.

9.5. Relation of the Gairloch Area to the Rest of the Lewisian Foreland

9.5.1. Structure and chronology.

It has been shown the Lewisian outcrop at Gairloch is part of a large scale shear zone some 5 kms broad, with a displacement of around 30 kms in a subvertical, negative sense. Such a large structure constitutes a major tectonic break between the rocks on either side and its significance in relation to the rest of the Lewisian is examined in the following section.

The first major subdivision of Lewisian gneiss was made by Sutton and Watson (1951) on the basis of metamorphic assemblages, age and structure. The ages of the Lewisian subdivisions used here are those given by Sutton and Watson (1965), ie. Early Scourian at c. 2600 Ma,
FIG 9-2 Schematic diagram showing possible mode of formation of the Mill na Claise fold.

a) before deformation.

b) after deformation.

fold closure under Torridonian cover?
the Inverian at c. 2200 Ma and the Laxfordian at c. 1650 to 1450 Ma. With the aid of these divisions they resolved the Lewisian of mainland Scotland into a central block of ancient Scourian gneisses flanked to the N and S by areas of extensive reworking in the Laxfordian period, the Laxfordian gneisses, see Fig 9-3. These two ages of gneiss where separated by intrusion of the Scourie dyke suite, present throughout the mainland Lewisian gneisses which have been used as time markers. Later work established that the Scourian of Sutton and Watson could be subdivided into an early Scourian (or Badcallian) event of granulite facies metamorphism, and a later Inverian event of amphibolite facies metamorphism toward the end of which the Scourie dykes were intruded (O'Hara, 1961).

This view of the Lewisian complex is substantiated by the geochemical work of Holland and Lambert (1973). Their work shows the existence of two main geochemical assemblages corresponding to the structural divisions of Sutton and Watson. The Scourian gneisses are typically low in SiO2, H2O, U, Th, Rb and K and richer in ferromagnesian minerals relative to the Laxfordian gneisses.

The Gairloch shear zone is itself largely Laxfordian in age and lies on the northern boundary of the area of Laxfordian reworking. The Loch Maree series and the Gruniard gneisses are separated from the Gairloch schists and the Tollie gneisses by the Loch Maree fault, a large transcurrent structure affecting both Lewisian and Torridonian. The Loch Maree series and the Gairloch schists have been correlated on the basis of their similar lithology since their original mapping, Clough et al (1908), and more recently on the basis of structural analysis and dating by Park (1973), Crane (1978) and Bikerman et al (1975). The Loch Maree series possess a slightly higher grade of metamorphism than the Gairloch schists as indicated by the growth of microcline porphyroblasts and the occurrence of kyanite, Clough et al
FIG 9-3 Major areas of Laxfordian reworking in NW Scotland (after Sutton and Watson, 1965).
(1908) and Crane (1978). It is therefore probable that they occupy a structural and metamorphic position equivalent to the SW limb of the Tollie antiform in Gairloch which is equated to the Carnmore antiform immediately to the NE of the Loch Maree series. In this case, it seems that the most highly deformed and lowest grade part of the shear zone has been faulted out by the Loch Maree fault.

North of the Loch Maree series are the Gruinard gneisses which have been structurally analysed by Davies (1977). He records early Scourian and Inverian structures of pre-dyke age and low Laxfordian strain. Though he records metamorphic assemblages as amphibolite facies, Field (1978) records unambiguous granulite facies assemblages in acid gneiss, and other reports are given by Park (1970), Moorbath and Park (1971), Cresswell and Park (1973), and Holland and Lambert (1973) in the N of the area. Also, geophysical evidence shows that dense and highly magnetic rocks interpreted as granulites underlie the northern part of the Gruinard gneisses, Watson (1973). These rocks yield ages around 2600 Ma (Moorbath and Park, 1971) and are therefore early Scourian in age. Southwards NE/SE trending subvertical shear zones and foliation, and amphibolite facies assemblages, affecting the dykes progressively intensify. The rocks are folded by the Carnmore antiform, a structure equated with the Tollie antiform.

A similar pattern is shown in the Gairloch area with pre-dyke early Scourian and Inverian structures, and occurrences of granulite facies assemblages, recorded on Craig Mhor Tolliadh, Park (1973), passing into the Tollie antiform, a post-dyke, early Laxfordian structure, and finally into the late Laxfordian SW limb of the antiform, see 9.3.

The geochemistry shows that the Gruinard gneisses have similar chemical patterns to the Scourian, though with increasing Si, alkalis
and decreasing ferromagnesian elements to the south as Laxfordian reworking increases in intensity. Similar patterns exist in the Loch Tollie district, Holland and Lambert (1973).

To the south of the Gairloch shear zone lies the 5 km wide Ruadh Mheallan belt of NE/SW trending pre-dyke structures where early Scourian granulite facies assemblages are preserved in the core of an ultrabasic mass (Park and Cresswell, 1973), the rest of the gneisses being in the amphibolite facies. To the south of this belt, post-dyke Laxfordian structures, migmatization and lower Amphibolite facies assemblages progressively predominate (Fig 5, Park and Cresswell, 1973, Sutton and Watson, 1965). Holland and Lambert (1973) have shown that these rocks are chemically indistinguishable from the type Laxfordian at Loch Laxford, Sutherland.

The Gairloch shear zone is therefore a major tectonic break between the Scourian type gneisses to the N, affected by Inverian deformation adjacent to the shear zone, and the predominantly Laxfordian gneisses with patchily preserved Scourian structures to the S.

Similar NW/SE trending shear zones are developed at and near the 'Laxford front' (Coward, Graham and Beach, 1974, Beach, 1974), at Tarbet and Stoer, Sutherland. The Tarbet shear zone lies at the junction between the Scourian granulites in the S, downgraded to amphibolite facies in the shear zone, and the Laxfordian migmatites to the N. The shear zones are the first post-dyke phase of deformation (F\textsubscript{2} of Coward, Graham and Beach) and therefore contemporaneous with the shear zone at Gairloch. The Tarbet shear zone consists of numerous individual shear zones some 50-100 m apart which dip \(\sim 70^\circ\) SW, have a shear direction plunging \(40^\circ\) SE and a SW side up sense, Beach (1974). Coward, Graham and Beach (1974) have established that the granites and migmatites were on the same tectonic level as the granulites prior to the shearing. They and Beach (1974) have postulated that the granulites were emplaced
above the migmatities by regional overthrusting. Later shearing brought the granites and migmatites adjacent to the granulites at the present level of erosion.

A displacement of 25 kms has been calculated for the Tarbet shear zone using the dykes as marker horizons by Beach (1974) which resolves into a 16 km vertical, SW side down, and 18 kms sinistral horizontal displacements. This is comparable to the total displacement of 29.5 kms on the Gairloch shear zone, resolving into 26 kms vertical, SW side down, and 14 kms dextral displacements.

Coward, Graham and Beach (1974) have shown that the first appearance of successive phases of deformation from Inverian to Laxfordian move progressively northwards. This is a mirror image of the pattern found at Gruinard bay and in the Tollie region.

The Gairloch structure is therefore similar to the Tarbet shear zone in that both separate the central stable block of granulites from zones of extensive Laxfordian reworking and both have comparable displacements and sense of shear. Differences occur in the shear directions, which are moderately plunging NW at Gairloch and moderately SE plunging at Tarbet, and in the width of the zone of granulites affected by Inverian and Laxfordian reworking, which is much more extensive at Gruinard bay than at Tarbet. There is also a difference in style, the Tarbet shear zone consisting of numerous small shear zones whereas the Gairloch shear zone is a single structure. This is probably related to the presence of metasediments at Gairloch, whose lower competence compared to the gneisses may have served to concentrate the deformation.

Thus during active shearing in late Inverian to late Laxfordian times these shear zones have acted as mobile boundaries to a central stable block of Scourian granulites. Deformation N and S of the central block is contemporaneous with that of the shear zones and thus the Laxfordian gneisses show similar metamorphic assemblages (lower amphibolite)
to those in the shear zones themselves. The similarity in lithology, petrology and chemistry of the Laxfordian gneisses at Torridon and Laxford suggests that they originated at similar levels in the crust. Thus the Tarbet and Gairloch shear zones have served to replace the Torridon gneisses to their original tectonic level with respect to the Laxford gneisses, effectively reversing the upward component in the earlier regional thrusting of granulites over granites and migmatites, see Fig 9-4. The aggregate vertical displacement on the Tarbet and Gairloch shear zones is 42 kms, a figure which is an underestimation since it does not include the displacement on the Stoer shear zones for which an estimate does not exist. This provides an indication of the magnitude of the earlier over-thrusting event.

9.5.2. Metamorphism and fluid composition.

Studies of the metamorphic reactions that took place during the transition from gneiss to schist have shown that the metamorphic fluid is CO$_2$-rich compared to fluids in the gneisses (8.3.2., 8.4.1.).

Metamorphic studies of shear zones at Tarbet in which granulite facies gneisses are downgraded to amphibolite facies have shown that the changes are overall isochemical except for the introduction of Na and H$_2$O and loss of Ca and Mg, (Beach, 1974, 1978). CO$_2$-rich fluid inclusions in granulite, and H$_2$O-rich inclusions in amphibolite facies rock together with a strong correlation between deformation and amphibolitization has suggested that mantle derived H$_2$O-rich fluids have travelled up tectonically determined paths, ie. shear zones, Beach (1978, 1974).

At Gairloch, the most highly deformed part of the shear zone, the schists, are surrounded by gneisses of amphibolite grade at the present level of erosion. Since the shear zone is a major structure with considerable displacement, it is likely to extend to significant depths in the crust, perhaps to its base, and thus is likely to cut through granulite facies rocks at depth. If the processes described
FIG 9-4 Schematic cross-section of the mainland Lewisian gneisses of NW Scotland. (incorporating Beach, 1974, fig 3)

Torridon Gairloch Gruinard Bay

Stoer, Lochinver Tarbet Laxford

present day erosion surface

Migmatites

Granulites

Migmatites

Granulites

thrust

0 10 20 30 kms.
in shear zones at Tarbet by Beach (1974) occurred at depth in the Gairloch shear zone, amphibolitization would preferentially remove $\text{H}_2\text{O}$ from the fluid in its upward migration, and produce more $\text{CO}_2$-rich fluids at higher levels of the crust. Such an effect would be most noticeable in the centre of the shear zone where higher strain rates and more prolonged deformation would allow the migration of fluids over larger distances. This may explain the more $\text{CO}_2$-rich fluids in the schist belt since they occupy the centre of the shear zone.
APPENDIX I Procedure for preparation and measurement of quartz particles in quartzofeldspathic gneisses.

1) Three mutually perpendicular slabs are cut from each specimen, see Fig I - 1. The remaining block is then set up in its original position using the orientation marked in the field. The strike is marked on each cut face and the dip and direction of dip measured directly from the specimen. Strike directions are then transferred to the slabs.

2) Marked faces of the slabs are etched with 40% HF by exposure to the fumes, a few mm from the acid surface, for approximately 4 mins. They are then stained by immersion in a saturated solution of sodium cobaltinitrite for approximately 30 secs. After washing in cold water and drying, the k-feldspar has a yellow stain, the plagioclase is coated with an opaque white deposit, and the quartz remains clear. Quartz particles are clearly distinguishable from their feldspar matrix.

3) Acetate peels are made from the etched slab in the manner conventional for carbonates. Up to three peels may be taken from each slab before re-etching is necessary. Acetate peels give a rapid and easy method of producing a transparent replica of the quartz-feldspar textures.

4) Peels are optically enlarged using a "Shadow Master" which projects the image onto a glazed glass screen from which quartz particles can be traced. Enlargements of x 10 were found to be adequate for most specimens, with x 32.25 for the finer grained...
FIG I-1. Cut specimen with acetate peels.
specimens. Thirty-three quartz particles together with the strike orientation are traced from each peel. Particles are selected using a grid of dots, to avoid bias.

5) Each face is plotted on a stereonet, using a Lambert equal area lower hemisphere projection, as a great circle and six equally spaced directions plotted on each, so that three of the directions coincide with the face intersections. The six directions for each face are then transferred to the tracing of quartz particles. Note that if the face is downwards facing on the original specimen, directions appearing in clockwise sequence on the stereograph, will appear in anticlockwise sequence on the tracing.

6) The centre of each particle is located by halving its maximum length along the long axis, located by eye, and then halving the maximum breadth perpendicular to the long axis.

Example:

Note that it is possible in some particles for a centre thus located to lie outside the particle perimeter.

7) The maximum diameter, i.e. from the first to the last appearance of quartz, for each particle is measured through its centre in each
of the six directions of step (5). For each direction the diameters are summed to give an "aggregate diameter".

8) The directions common to two faces are used to adjust the measurements from one face to another. The second face is adjusted to fit the first, and the third to fit the second. The discrepancy between the third and first after this procedure is the "closure error" and is shared out between the faces.

Example, Specimen 5, S. Sithean Mhor gneisses.

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Closure error = 50.9 - 44.6 = 6.3

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9) Angles to $N(X_1)$, $E(X_2)$ and down $(X_3)$ are measured from the stereograph for all directions.

10) Corrected aggregate diameters and their angles to $X_1$, $X_2$ and $X_3$ are then processed by the unpublished computer programme "PATEN" compiled by R.F. Cheeney, Edinburgh. This gives the best fit ellipsoid in terms of the lengths of the principal axes and their directions, with 99% confidence limits. The results are listed in Appendix II.
APPENDIX II  Finite strain data.

Tables 1(a), (b), (c) and 2(a), (b), (c) contain the results of the processing of measurements on quartzofeldspathic gneisses by the computer programme "PATEN", which produces lengths and orientations of the principal axes of finite strain with 99% confidence limits and cones.

The programme "PATEN" is based on the theory of Hext(1963) and was written by R.F. Cheeney, Department of Geology, Edinburgh.

Contents

1. Loch Tollie gneisses - 12 specimens.
2. S. Sithean Mhor gneisses - 6 specimens.
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Angles to N(X₁), E(X₂) and down (X₃), in degrees.
Table 1(c) Semi-angles of 99% confidence cones for principal axis orientations, Loch Tollie gneisses.

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Figures refer to angles, in degrees, about principal axes within the principal planes. Negative figures imply that confidence cones of principal axes overlap, i.e. the orientations of the two principal axes are indistinguishable.
Table 2(a) Lengths of principal axes with 99% confidence limits, S. Sithean Mhor gneisses.

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Table 2(c) Semi-angles of 99% confidence cones for principal axis orientations, S. Sithean Mhor gneisses.

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Figures refer to angles, in degrees, from principal axes to their 99% confidence cones within the principal planes. Negative figures imply that the confidence cones of the respective principal axes overlap, i.e. the orientations of the two principal axes are indistinguishable.
Table 2(b) Orientations of principal axes, S. Sithean Mhor gneisses.

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Angles to $N(X_1)$, $E(X_2)$ and down ($X_3$) in degrees.
APPENDIX III  Electron Microprobe mineral analyses.

Operating conditions

The Cambridge Geoscan electron microscope at the department of Geology, Queen's University, Belfast was used with 20 keV accelerating potential and $4 \times 10^{-9}$ A specimen current. Analyses were produced by a Link Systems E.D.S. using 0-10 keV range, 40 µsecs processing time and 100 livetime seconds.

Contents

1. Mineral analyses for the An Ard gneisses and adjacent schists.
   a) Plagioclase analyses
   b) Biotite analyses
   c) Epidote analyses

2. Analyses of garnet-biotite pairs used in temperature estimation.

3. Analyses of calcite and dolomite grains used in temperature estimation.
1. Mineral analyses of An Ard gneisses and adjacent schists.

a) Plagioclase analyses

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2. Analyses of garnet-biotite pairs used for temperature estimation (see 8.2.1).

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|XMg | 0.0585| 0.481| 0.0707| 0.458| 0.0730| 0.490|

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All analyses from polished thin section "MNC 21" of marble at 843719.
BIBLIOGRAPHY


Storre, B., Nitsch, K. (1972). Die reaktion, $2 \text{zoisit} + 1 \text{CO}_2 \overset{\leftrightarrow}{\Rightarrow} 3 \text{anorthit} + 1 \text{calcit} + 1 \text{H}_2\text{O}$. Contr. Mineral. Petrol., 35, 1-10.


