THE STRUCTURAL AND METAMORPHIC GEOLOGY OF SOUTH UIST, OUTER HEBRIDES .

A thesis submitted for the degree of Ph.D.,

University of London

Michael Peter Coward B.Sc. A.R.C.S..

June 1969

ABSTRACT.

This thesis is taken from a field study of an area of Lewisian basement rocks which have suffered varying amounts of polyphase deformation. The rocks consist of a series of quartzo-feldspathic gneisses, metasediments and amphibolites which were intruded by a suite of tholeiitic dykes and later deformed. Four phases of post-dyke deformation (f_1, f_2, f_3 and f_4) have been recognised. Each phase was accompanied by high grade (amphibolite facies) metamorphism and the last phase was accompanied by localised migmatisation and granitisation. The deformation is heterogeneous, and for each phase, areas of low deformation and areas of high deformation can be recognised.

In the areas of low deformation, structures of f_1 and f_2 age have been examined in great detail and the strong linear fabric which is a characteristic feature of these areas is considered to have been produced by the intersection of the bulk strain ellipsoids of these deformation phases. The variation in extent of folding is examined and it is concluded that the relative competence of the rocks played an important part in determining the structural pattern.

Structures of f_3 age are characterised by large antiformal regions in which the deformation is low, separated by tight pinched-in synforms. It is considered that this fold shape was caused by the large scale folding of rocks of different competences at different structural levels. During the f_4 phase, the deformation was largely concentrated within the zone of most intense migmatisation.

The structural history of the rocks above the Outer Isles Thrust Zone is discussed. These rocks, which were largely derived from a large igneous intrusion, have suffered a similar deformation phase history to the rocks to the west, though the style of deformation was very different.

Structures and rock types associated with the Outer Isles Thrust are examined and the various thrust and crush zones are described. The age of the thrust may be late Laxfordian or it may be Caledonian.

CONTENTS

Chapter I General Introduction

1.1 General 11

1.2 Previous work on the solid geology of South Uist 12

1.3 Present work 13

1.4 The area 17

1.5 The geology of South Uist 17

1.6 Structural terminology 21

1.7 Localities 21

Chapter II General structure of South Uist

2.1 General 25

2.2 Measurement of rock fabrics 25

2.3 General notes on the structural history 26

2.4 F_1 phase 28

2.5 F_2 phase 31

2.5A F_2 small scale structures 34

2.5B F_2 large scale structures 35

2.5C F_2 fabric 35

a) Scourie-dykes 35

b) Early-amphibolites 37

c) Acid gneiss 37

2.5D Deformation of early structures 38

2.6 F_3 phase 38

2.6A General 38

2.6B Small scale structures 43

2.6C F_3 interference patterns 43

2.6D Plunge of f_3 structures and orientation of the axial plane 43

2.6E Variation in dip of the f_3 axial plane 44

2.6F F_3 fabric 45

2.7 F_4 phase 45

2.7A Trned of f_4 axial plane 46

2.7B F_4 small scale structures 46

2.7C F_4 fabric 47

2.7D Folding of earlier structures 47

2.7E Folding of earlier linear fabrics 48

Chapter III The metamorphic history

3.1 General 51

3.2 Problems in fabric analysis 52

3.3 The metamorphic history of the individual rock types 53

3.3A The metamorphism of the Scourie-dykes 53

3.3B The metamorphism of the early basic rocks 61

3.3C The metamorphism of the metasediments 65

3.3D The metamorphism of the quartzo-feldspathic gneisses 69

3.4 Synthesis of the Laxfordian metamorphic history 72

Chapter IV Methods of strain analysis

4.1 Directly defining the shape of the strain ellipsoid using retrogressed garnets. 86

4.2 Defining the surface of no finite longitudinal strain within the strain ellipsoid 90

4.3 Using folded or boudinaged veins to calculate the amount of strain 92

4.3A Using folded layers 92

4.3A1 Fold classification 92

4.3A2 Strain measurement from folded single layers 93

4.3A3 Strain measurement from mutilayer folds 99

4.3B Using boudinaged layers 101

4.4 Using rock fabrics 102

4.5 Using the deformation of discordant relationships such as those between dykes and gneissose foliation 104

Chapter V Areas of low deformation

5.1 General 107

5.2 Previous work 108

5.3 General structure 108

5.4 The f_1 phase of deformation 113

5.4A General 113

5.4B F_1 deformation zones 117

5.4C The f_1 fabric 121

5.4D Deformation outside the deformation zones 121

5.5 The f_2 deformation 121

5.5A General 121

5.5B The effect of f_2 deforamtion on the discordant structures 124

5.5C The f_2 fabric 133

5.6 The sequence of early strains 135

5.6A The f_1 strain 135

5.6B The f_2 strain 138

5.6C Sequence of strains 138

5.6D Superimposition of two strains 139

5.6E The effect of buckle folding on the strain path 149

5.6F Application to the Uists 154

5.7 Variation in amounts of f_2 deformation 156

5.8 The viscosity ratio between dykes and gneiss 156

5.9 The f_3 deformation phase 160

5.10 The f_4 deformation phase 166

5.11 The structures at the edge of the areas of low deformation 169

5.12 Low deformation elsewhere in the Uists 175

5.13 The origin of the low deformation areas 178

Chapter VI The areas of high deformation

6.1 The early Laxfordian deformation phases 181

6.1A North-west Benbecula 181

6.1B Sgeir Dremisdale 184

6.1C South Uist as a whole 189

6.1D Conclusions 192

6.2 The f_3 phase of deformation 193

6.2A North-west Benbecula 193

6.2B Sgeir Dremisdale 196

6.2C South Uist as a whole 199

6.2D The variations in f_2 and f_3 plunge 209

6.3 The f_4 phase of deformation 210

6.3A F_4 axial plane trajectories 210

6.3B Variations in fold geometry and in the amount of flattening on the folded layers 211

Chapter VII The pre-Laxfordian history

7.1 The Scourie-dykes 230

7.2 Pre-Scourie-dyke intrusives 234

7.3 Early basic rocks of indeterminate origin 237

7.4 Scourian fold history 238

7.5 Scourian metamorphic history 239

7.6 Synthesis of the Scourian history 240

7.7 Origin of the gneissose foliation 242

7.8 Origin of the gneiss 242

Chapter VIII The Eastern Gneisses

8.1 General 244

8.2 Lithology 245

8.3 Structural history of the Eastern Gneiss 246

8.3A Main phase deformation 247

8.3B Structures earlier than the main phase deformation 248

8.3C The f_3 phase of deformation 250

8.3D Beinn nan Caorach shear zone 250

8.3E The f_4 phase 251

8.3F Glen Liadale shear zones 251

8.4 The structural history of the other bodies above the Outer Isles Thrust Zone 252

8.4A The Mashed Gneisses 252

8.4B The Usinish Mylonites 253

8.5 General structure of the Eastern Gneiss 258

8.6 Metamorphism of the Eastern Gneiss 260

8.7 Origin of the Eastern Gneiss 263

Chapter IX Outer Isles Crush Zone

9.1 Introduction 265

9.2 The rocks of the crush zones 268

9.2A Crushed rock and pseudotachylite 269

9.2B Metamorphic rocks associated with the crushing 271

9.3 Small scale structures associated with the fault deformation 272

9.3A Visible faults and thrusts 272

9.3B Folds 272

9.3C Crush breccias 273

9.3D Pseudotachylite veins and intrusions 275

9.3E Small scale fractures in the gneiss and pseudotachylite 276

9.4 Large scale structures 278

9.4A Folds 278

9.4B Large scale crush belts 279

9.4C Large scale fractures 288

9.4D Late faults 289

9.5 The sequence of large scale fault deformation 289

9.6 The problem of pseudotachylite production 291

9.7 Discussion 294

Chapter X Summary of results and discussion 296

References 302

Maps in back pocket:-

Map 1 (two sheets) General geology of South Uist

Map 2 Topographic map of South Uist (Ordnance Survey map, sheet 23).

Map 3 Structural profiles through South Uist

Map 4 F_2 lineations and lineation sub-areas

Map 5 Axial planes of f_4 minor folds

Map 6 The Eastern Gneiss and the main crush zones

Map 7 Detailed map of Ardivachar Point, South Uist

Map 8 Detailed map of Garry-a-siar, Benbecula

Map 9 Rib orientations in South Uist

Acknowledgments

I would like to express my gratitude to Dr. J.V.Watson for her supervision of my work in the field and in the laboratory and for her criticism of the manuscript.

I would also like to thank my fellow post-graduate students and the various members of staff at Imperial College who have helped me at one time or another. In particular I would like to thank my colleagues in the Hebridean research group, Mr.P.W. Francis, Mr.R.H.Graham and Dr.J.S.Myers and also Mr.P.J.Huddleston for his help in the sections on fold analysis.

I am deeply indebted to the MacMillan family of Bornish, South Uist, for their generous hospitality both to myself and to visiting colleagues.

I should like to thank Miss Caroline Bowling for typing this thesis and Dr.R.Dearnley and Mr.F.Dunning for permission to reproduce their map of the Uists (Fig. 1.1)

The work was carried out during the tenure of a N.E.R.C. research studentship.

CHAPTER I

GENERAL INTRODUCTION

1.1 General

The island of South Uist lies in the south of the Outer Hebridean archipelago and is some 40 kilometres long and 10 kilometres broad. Much of the western side of the island, except for a few coastal patches and sporadic loch-side outcrops, has very poor exposure. Most of the geological work has therefore to be confined to the eastern half of the island, to a chain of wild and heather-ridden hills rising to 650 metres, the lower slopes of which have been scoured of all soil and loose rock by the advancing Pleistocene glaciers. On the south-facing slopes of these hills, where the ground is exposed to the full fury of the Atlantic weather, the rock outcrops can be clean and fresh and large areas are stripped of their vegetation. On the north-facing slopes, the state of the outcrop is much more variable, depending on the abundance of bog-moss, lichen and heather. Map 2 gives the general physiography and the main localities on this island.

With the exception of the suites of Tertiary dykes, emanating from the Tertiary volcanic centres of the Inner Hebrides, the solid geology is almost entirely of Lewisian gneiss and its derivatives. This gneiss, like its counterparts in the North West Highlands of Scotland, was deformed and metamorphosed almost entirely within the Pre-Cambrian. There are no Proterozoic or Palaeozoic rocks and the only possible Palaeozoic structures are groups of pseudotachylite thrusts and faults which could be related to Caledonian thrusts on the mainland.

1.2 Previous work on the solid geology of South Uist

The first geologists to make a detailed study of the solid geology of South Uist were Jehu and Craig, in the course of their survey of the whole of the Outer Hebrides. Their report on South Uist (1925) gives a good description of certain aspects of the petrology, but makes practically no attempt to describe the structure or pre-Cambrian stratigraphy. Their most important contribution was the description of the most striking geological feature of South Uist, the Outer Isles Thrust, a zone of intense crushing and pseudotachylitised rocks which forms the backbone of the chain of hills. They noted both structural and mineralogical differences between the rocks above and below this thrust zone.

In 1957, Kursten dabbled in Hebridean geology and produced a small but worthwhile paper in which he discussed in more detail some of the small scale structures associated with the thrusting. He mapped a small area of South Uist, on both sides of the thrust, close to Lochboisdale, and recognised the existence of more than one phase of folding, two phases of which he demonstrated from a petrofabric analysis of gneisses from the hill Arnaval.

By far the most important contribution to Hebridean geology was made by Dearnley in 1962 and 1963 when he recognised that the history of the Hebridean Lewisian could be divided into two orogenic periods separated by a period of dyke intrusion (the "Scourie-dykes"). He correlated these older and younger orogenic periods with the Scourian and Laxfordian orogenic periods on the mainland of Scotland, as set up by Sutton and Watson (1951), Gilletti et al (1961). Most of Dearnley's petrological and geochemical work was done on South Harris, but in the early '60s (published Q.J.G.S. 1968) along with Dunning, he analysed in

detail the structure of six pre-Laxfordian dykes; on the west coast of Benbecula, at Ardivachar Point and at Pollachar, South Uist (see Fig. 1.1). From such work on these dykes and from his own general reconaissance mapping, Dearnley divided the Hebridean Lewisian into three structural zones; a central zone with mainly Scourian deformation, showing practically no Laxfordian tectonic effects, and northern and southern zones where the Laxfordian reworking was much more intense. The boundary between Dearnley's Central and Southern Zones ran through the middle of South Uist. This subdivision, according to Dearnley (1962), would correlate with a similar zonation set up by the Geological Survey (Peach et al, 1907) and by Sutton and Watson (1951) on the mainland, though offset by about 100 kilometres. This offset was considered to be due to a large dextral fault, the "Minch Fault".

The Lewisian structural and metamorphic history of South Uist, as set up by Dearnley (1962) and Dearnley and Dunning (1968) is given in Table 1.1, and the field relationships suggested by them for the Hebrides as a whole in Fig. 1.1.

1.3 Present work

I first began work for this thesis in Summer 1966, as a follow-up to Dearnley's work. My original idea was to study the structural and metamorphic variations which occured at the edge of an orogenic belt. Dearnley (1962) had described such a boundary running through the centre of South Uist, and so this seemed to be the ideal place for such an investigation. I found, however, that the geology was by no means as simple as Dearnley had suggested and that there was no actual boundary between Scourian and Laxfordian zones. Instead, I found a few small pockets of weak Laxfordian deformation sitting in the middle of an area

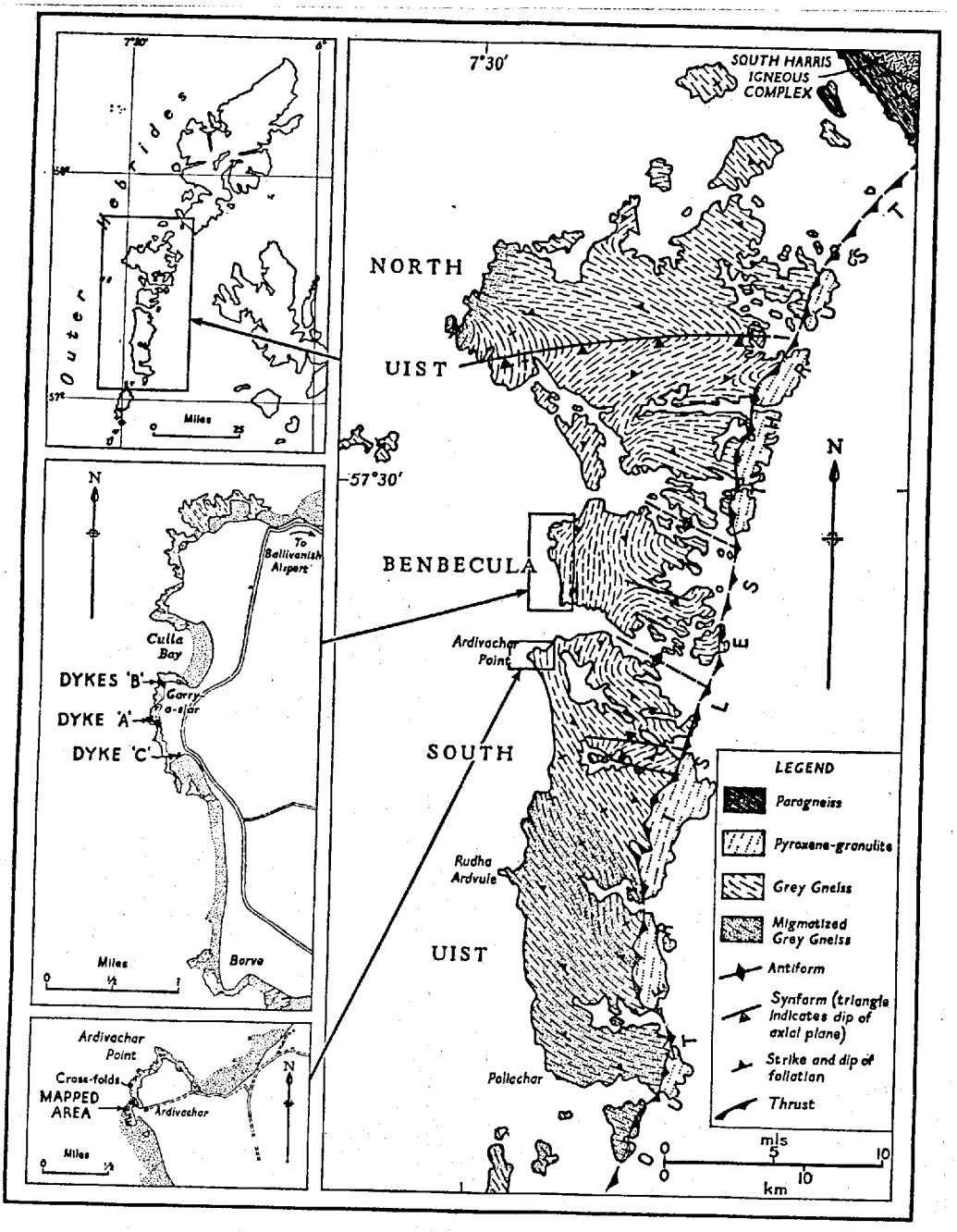
Table 1.1

The Lewisian chronology of the Uists as set up by Dearnley (1962, 1963) and Dearnley and Dunning (1968)

PERIOD	STRUCTURE	METAMORPHISM
Late Laxfordian c.1600 my	NW-SE folding and some cross-folding (NE-SW)	amphibolite facies retrograde metamorphism; migmatisation and formation of granites and pegmatites
Early Laxfordian (age not determined	Main NW-SE folding	Granulite facies metamorphism of basic intrusions
2190-1900 my	Intrusion of tholeiitic sills and dykes	
Late Scourian (Inverian) c.2200 ± my		Local pegmatites at Garry-a-Siar cut the main foliation of the early complex of gneisses
Early Scourian c.2600 ± my	Main Scourian folding formation of palingenetic veins. Deformation of earlier pegmatites	Main Scourian metamorphism, generally in the granulite facies
Pre-Scourian (Katarchaean?) 3000 my?	Earliest complex intruded by banded basic and ultrabasic rocks and microcline pegmatites	

Fig. 1.1

The structure of the Uists and Benbecula, according to Dearnley and Dunning (1968). The region which Dearnley and Dunning considered to have suffered Late-Laxfordian migmatisation is shown stippled and the edge of the stippled area marks the northern edge of Dearnley's (1962) southern Laxfordian zone.


7°30'
SOUTH HARRIS IGNEOUS COMPLEX
NORTH
UIST
57°30'
BENBECULA
Ardivachar Point
SOUTH
UIST
Rudha Ardvule
Pollachar
N
LEGEND
Paragneiss
Pyroxene-granulite
Grey Gneiss
Migmatized Grey Gneiss
Antiform
Synform (triangle indicates dip of axial plane)
Strike and dip of foliation
Thrust
mls
km
Outer Hebrides
Miles
To Ballivanish Airport
Culla Bay
DYKES 'B'
DYKE 'A'
DYKE 'C'
Garry a-sior
Borve
Ardivachar Point
Cross-folds
MAPPED AREA
Ardivachar

of strongly deformed Laxfordian rocks. I found no justification for considering any of the rocks west of the thrust planes as simply "Scourian"; all had felt some tectonic imprint of a Laxfordian period of reworking.

The topic of my thesis was thus changed to a general study of the structural and metamorphic history, especially the actual deformation or strain history of the Laxfordian. Fifteen months were spent in the field area, spread over three summers, 1966 to 1968, with a further twenty months in London.

1.4 The area

An area of about 100 square kilometres, from the "South Ford" to the A865 Daliburgh - Lochboisdale road, was mapped on various scales up to 1 : 10560 . Only about half of this area contained enough rock exposure for very much detailed work. Field mapping was carried out with the use of aerial photographs; the topographic 1 : 10560 maps were almost useless, consisting of only the 19th century stream and tidal patterns with an occasional spot height. Much more detailed mapping, on scales of up to 1 : 4220 was carried out along parts of the west coast of Benbecula and South Uist, using enlarged aerial photographs.

1.5 The geology of South Uist and the terminology used in this thesis

Dearnley (1962) and Dearnley and Dunning (1968) recognised the heterogeneous nature of the Laxfordian deformation. More detailed work has fully substantiated this observation. To a certain extent, South Uist can be considered as consisting of areas of high deformation and also small contrasted areas of low deformation, and a discussion of this variation in amount of deformation forms a large part of this thesis.

Table 1.2

Lewisian chronology as given in this thesis

PERIOD	STRUCTURE	METAMORPHISM
?	Outer Isles Thrust Zone, period of crushing, pseudotachylite production and locally late alteration	
?		late Laxfordian retrogression - low amphibolite facies metamorphism
1600 my		post-kinematic f_4 migmatisation
	f_4 - folding on east-west axial planes	syn-kinematic f_4 metamorphism
	f_3 - folding on NW - SE axial planes	syn-kinematic f_3 amphibolite facies metamorphism
		post kinematic f_2 amphibolite facies metamorphism
?	f_2 - folding on N - S to NE - SW axial planes, low dip to the west.	syn-kinematic f_2 amphibolite facies metamorphism
	f_1 - folding on NW - SE axial planes	syn-kinematic f_1 amphibolite facies metamorphism
2190-1900 my	Intrusion of Scourie-dykes	
2200-2400 my	late-Scourian, fs3 and fs2, deformation in the west, no deformation in the east	amphibolite facies metamorphism in the west
2600 my	Intrusion of early basic rocks	
?	Early gneiss complex	

Four phases of folding have been recognised, following a period of Scourie-dyke intrusion and these are given along with a general summary of South Uist geology in Table 1.2. In each of these Laxfordian phases the rocks appear to have responded to the deformation in various ways, depending on:

(i) the homogeneityor heterogeneity of the rock

(ii) the pre-tectonic composition and competence of the rock

(iii) the syn-tectonic metamorphic conditions

The aims of this thesis are therefore:

(i) To examine the Laxfordian deformation sequence and also the form of the Laxfordian structures on South Uist - Chapter II

(ii) To examine the Laxfordian metamorphic history in relation to the deformation history and to examine any variation in metamorphic grade or intensity both in time and space - Chapter III

(iii) To examine the Laxfordian deformation history, the variation in amount of Laxfordian deformation and, where possible, the progressive finite strain history for various groups of rocks in the areas of low deformation - Chapter V

(iv) To follow the structural history into the areas of high deformation, to examine the variation in amount of deformation within these areas and to examine in more detail the later Laxfordian phases which are more fully developed here - Chapter VI

In addition to the above, there has also been a more general survey of the geology of the island. The island of South Uist is divided into two unequal regions by a major crush zone, "the Outer Isles Crush Zone". This zone separated two groups of rocks, the "Western Gneiss" below, and

the "Eastern Gneiss" above. The term "Western Gneiss" (after Dearnley 1962) covers a whole group of rock types including basic rocks of several ages, ultrabasic rocks and acid gneisses. These acid gneisses include bands or strips of rock of a definite sedimentary origin and it may be that much of these "Western Gneiss" were once supra-crustal. The terminology used in this thesis for the "Western Gneiss" is as follows:

<u>Scourie-dykes</u> These rocks are quartz-amphibolites which when seen in their least deformed state, cross-cut the gneissose foliation and the earlier basic rocks.

<u>Early amphibolites</u> These are a group of intermediate, basic and ultrabasic rocks, ranging in composition from peridotites to leuco-diorites, which all date from the pre-Scourie-dyke period. The term "Early amphibolite" is strictly a field term and is used to distinguish them from basic rocks of the Scourie-dyke age. The more correct petrological terminology is used in the appropriate places where necessary.

<u>Acid Gneisses</u> These include biotite and hornblende-bearing quartzo-feldspathic rocks. They exhibit a "gneissose foliation" (or "foliation"), a mineral layering in the rock, caused by -

a) variations in proportions of leuco- to mafic minerals
b) " " " " quartz " feldspar
c) " " " " biotite to hornblende
d) " " " " homogeneous to inhomogeneous rock
e) " " " " coarse grain size to finer grain size

Where the layering is caused by different lithological units, for instance where there are parallel layers of Scourie-dykes and acid gneiss, this layering is termed "banding".

Metasediments Amongst the acid gneisses there are bands of brown-weathering gneiss, rich in biotite, garnet and quartz, with also minerals such as sillimanite, staurolite and anthophyllite. They are considered to be of sedimentary origin. A more detailed discussion of these rocks and their relations to the acid gneiss can be found in a separate publication (Coward, Francis, Graham, Myers and Watson, 1970).

The "Eastern Gneisses", (after Dearnley, 1962), consist largely of rocks belonging to what appears to be a metamorphosed igneous mass underlying a zone of intensely phyllonitised rocks and also of crushed acid gneiss of the "western gneiss" type. The petrology of the Eastern Gneisses will be discussed along with their general structures in Chapter VIII. Similarly, the rocks associated with the crushed zone will be discussed along with a general description of the thrust tectonics in Chapter IX.

1.6 Structural terminology

The structural terminology used in this thesis follows exactly that of Ramsay (1967), and a list of the structural abbreviations is given in Table 1.3. The amount of strain is given in the form of strain ratios $1 + e_1 / 1 + e_2$, $1 + e_2 / 1 + e_1$ and in percentage shortening. In order that these strain values can be compared more easily, their relationship is given graphically in Fig. 1.2.

1.7 Localities

The localities throughout South Uist can be found using the National Grid references given in the brackets following the locality name. The localities at Ardivachar and Garry-a-siar have been given a

FIG. 1-2

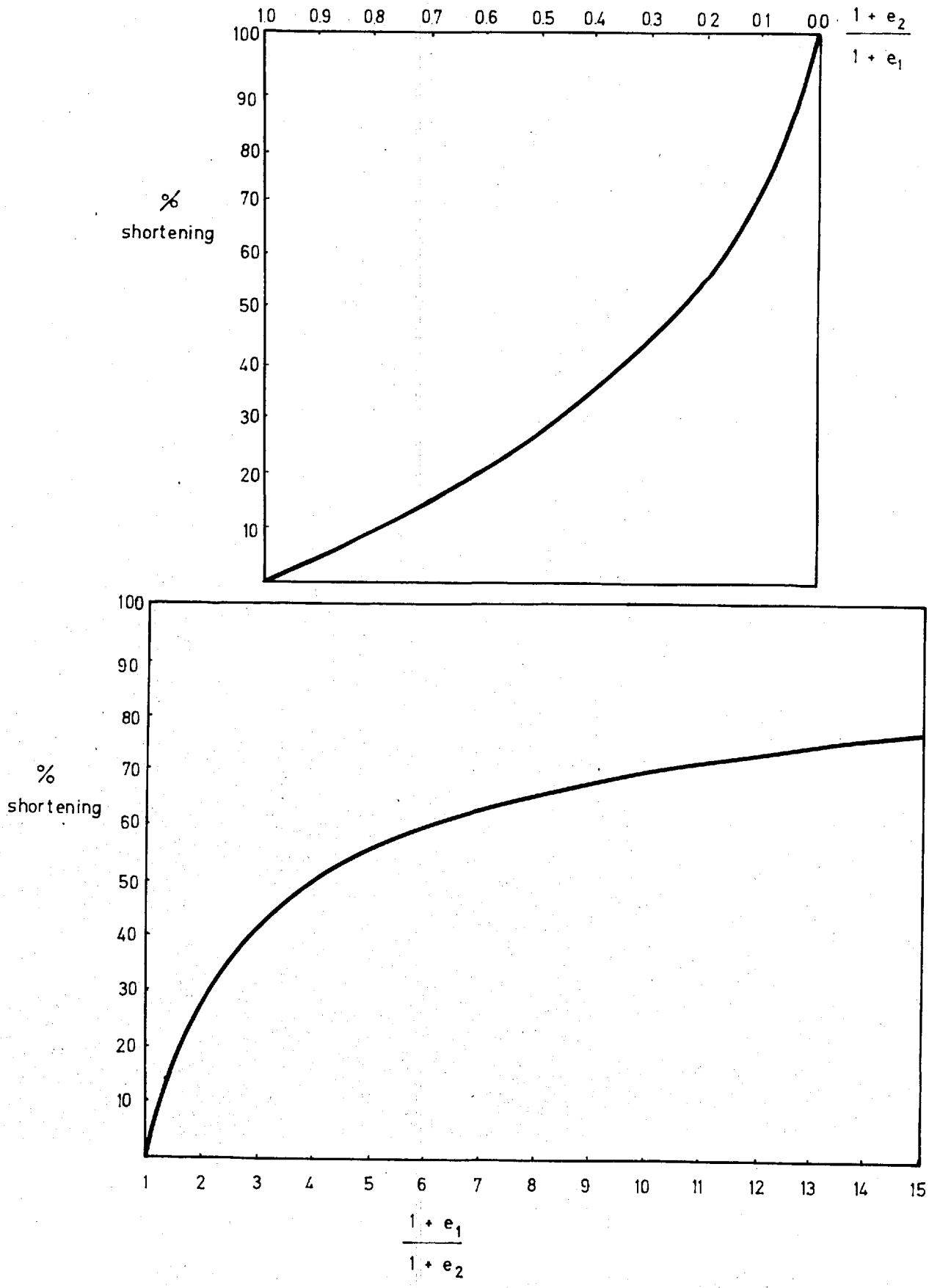

1.0 0.9 0.8 0.7 0.6 0.5 0.4 0.3 0.2 0.1 0.0
1 + e2 / 1 + e1
100 90 80 70 60 50 40 30 20 10
% shortening
100 90 80 70 60 50 40 30 20 10
% shortening
1 2 3 4 5 6 7 8 9 10 11 12 13 14 15
1 + e1 / 1 + e2

Table 1.3 Structural Nomenclature

x, y, z	cartesian coordinate axes
e	extension
$e_1 e_2 e_3$	principal extensions
λ	quadratic extension
$\lambda_1 \lambda_2 \lambda_3$	principal quadratic extensions
$\omega, \omega_1 \omega_2 \omega_3$	rigid body rotation, and rigid rotations round x,y and z axes
X, Y, Z	measured ellipsoid axes from a deformed sphere of radius r, $X = r(1 + e_1)$, etc.
a	ratio of principal strains $(1 + e_1)/(1 + e_2) = X/Y$
b	ratio of principal strains $(1 + e_2)/(1 + e_3) = Y/Z$
ε	logarithmic or natural strain $= \log_e(1 + e)$
ψ	angular shear
γ	shear strain
$\sigma_1 \sigma_2 \sigma_3$	principal stresses
μ	viscosity shear modulus

number and a pre-fix, A for Ardivachar and G for Garry-a-siar. These refer to localities marked on Maps 7 and 8.

Except where otherwise stated, in this thesis north should be taken as the top of a map or diagram. Similarly, pegmatites are shown as stippled areas, Scourie-dykes are either clear or black areas and Tertiary dykes are shaded brown. The metric system is used in all maps and diagrams.

CHAPTER II

THE GENERAL STRUCTURE OF SOUTH UIST

2.1 General

The aims of this chapter are to describe some of the large and small scale folds and other structures, which are associated with the various deformation phases, and to indicate some of the evidence which leads to the establishemnt of a fold-phase history. In a later chapter I shall analyse in as much detail as possible the variations in amounts of deformation in time and space, and try to trace the history of deformation in South Uist. This chapter deals only with the general description of the deformation effects and I have tried to limit it to field observations and the basic assumptions which these often entail.

The most important deformation phases, which are those labelled f_2, f_3 and f_4 produce major and minor folds as shown on Map 1. These large structures with their associated small folds and fabrics require a certain amount of rather tedious description.

Map 1 shows the main structures of South Uist. The trend lines and fold axial planes shown on this map are as mapped in the field (allowing for simplification because of the map scale). During the three field seasons over 40,000 structural measurements were taken, some of which are now given in maps 4, 6, 7 and 8.

2.2 Measurement of rock rabrics

In areas where the deformation phases produced a good axial plane foliation or a lineation this often acts as a parting in the rock and measurement of the orientation of the fabric is fairly easy. However,

in many areas (see Chapter III) the f_4 deformation phase and the associated phase of late recrystallisation obliterated much of the earlier fabric built up during the previous Laxfordian phases and also tended to remove any tendency for the rock to weather or break along a regular parting plane. Accurate measurements of any steeply or gently dipping components are made very difficult in these annealed quartzo-feldspathic gneisses. One cannot place much trust on plunge measurements, especially of the f_4 folds, in the eastern part of the area, and so these have been neglected as far as possible in this discussion.

2.3 General notes on the structural history

Table 1.2 gives a summary of the structural history of South Uist.

The period of dyke intrusion before the commencement of the Laxfordian fold-sequence was the most important event as far as the recognition of a structural history is concerned. In areas such as Garry-a-siar individual dykes show varied amounts of deformation along their length. It can be seen that the presence of highly deformed dykes and far less deformed dykes in close proximity is not due to differences in age of intrusion, but to very rapid variations in the amounts of Laxfordian deformation. There are however amphibolites of a generation earlier than these Scourie-dykes and in many areas care has to be taken when assigning an age to a basic rock.

The most important character of the Scourie-dykes is that they give an indication of the age of the gneissose foliation and the folds. The foliation was in existence long before the dyke intrusion, as can be seen from the discordant relationships that the dykes make with the

gneiss. However, this gneissose foliation is often modified and the rock may be turned into a new banded gneiss by the intense Laxfordian deformation. Folds can sometimes be dated by the way they deform either previously deformed or undeformed dykes, but this method of fold dating is by no means infallible, as f_3 or f_4 folds are often the first to deform the basic dykes intensively.

Folds can generally be "dated" by using combinations of

(i) the general properties of the fold, i.e. whether it has a certain type of cleavage or linear fabric, or whether it folds an earlier banding, etc.

(ii) The orientation, i.e. strike, dip and plunge of the fold

(iii) the form of the large-scale structure (if any) to which the fold is related

(iv) the effect of the fold on the earlier fold structures

(v) the effect of the fold on basic dykes

If a group of folds of unknown age cannot be traced into a neighbouring and more easily understood area and if the fold orientation (as is often the case) is ambiguous, then certain stock generalisations can be made. Note, these apply to South Uist only!

a) F_2 folds are usually very tight and bear a strong axial plane fabric and an intersection lineation. (These characters do not help to identify f_2 folds in the south-west of the area, where f_3 folds appear the same).

b) f_3 folds in the north and north-east are usually more open structures than those of f_2 age. They often fold an earlier schistosity as well as the lithological banding. They are usually reclined and asymmetrical.

c) f_4 folds vary in tightness. They often have heterogeneously

deformed limbs and show accompanying migmatisation. Granite or pegmatite often occurs along the axial planes or the more deformed limbs.

d) F_1 folds are very difficult to identify.

e) Scourian folds can only easily be recognised when

(i) in areas of low Laxfordian deformation

(ii) cross-cut by Scourie-dykes

or (iii) folded by obvious f_1 or f_2 structures.

The most important deformation phase, the "Main Phase" or "F_2 Phase" produced and/or reorientated the tectonite fabric of the gneiss. This f_2 fabric is re-orientated round large open folds of f_3 age in the north and round much tighter structures of the same age in the south. It is these large f_3 structures which set the Laxfordian structural trends in the Hebrides. The f_4 folds produce large scale structures only in the east where they locally make large scale fold interference patterns with the f_3 folds. They are best developed in areas of most Laxfordian migmatisation.

2.4 The f_1 phase of deformation

Folds or other signs of intensive deformation of Laxfordian age, which formed prior to the f_2 main phase deformation can only be demonstrated with any certainty at a few localities, notably Ardivachar and Garry-a-siar, where the Laxfordian deformation as a whole is low.

Elsewhere (e.g. at Sgeir Dremisdale) round the hinges of some f_2 folds, the Scourie-dykes show an earlier tectonite fabric folded by the f_2 phase. This fabric can be interpreted as being due to the f_1 deformation phase.

Fig. 2.1

A branching dyke, folded during the f_2 phase, from south of Loch Teanga (817378).

FIG. 2-1

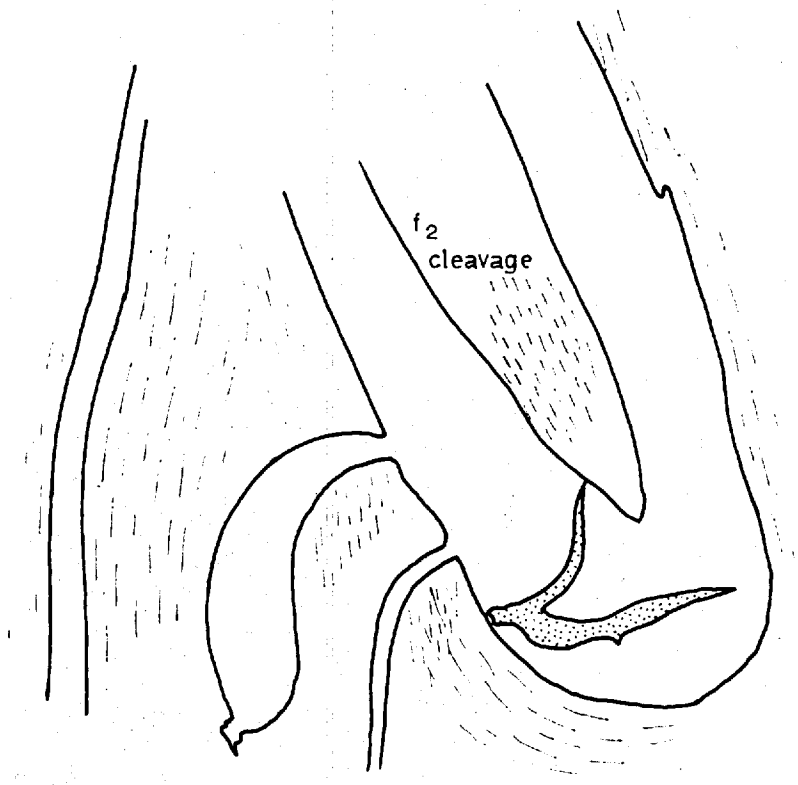


Fig. 2.1 shows a dyke with branching apophyses folded by f_2 structures. In some cases such branching dykes could easily be mistaken for f_1 folds. I have never been able to identify with certainty f_1 folds in basic dykes outside the low deformation areas and I doubt very much that the f_1 deformation phase was of much importance in South Uist as far as large-scale structures are concerned.

2.5 The main phase deformation, f_2

Many of the strong penetrative mineral fabrics and the tight isoclinal folds in South Uist belong to this deformation phase. When well developed. it induces a strong planarity and may produce a new banding in the rocks. It isthis strong planar fabric which is folded and reorientated by the f_3 and f_4 structures.

Over much of South Uist the f_2 strain is intense, but there are a few exceptions (see Chapter V). In these low deformation areas the less developed stages of f_2 deformation and the early structures can be observed. At Ardivachar Point and Garry-a-siar (Maps 7 and 8), both of which will be described in detail in Chapter V, folds of f_2 age are often the first to deform the Scourie-dykes. The same situation also occurs in other low deformation areas. East of Trinival, where locally the f_2 structures were poorly developed, no earlier phase of folding affected the dykes. Fig. 2.2 shows the variation in f_2 intensity within a small area south of Trinival (780265) where within a few feet the f_2 structures vary from isoclinal folds to gentle warps which hardly affect the discordant nature of the Scourie-dykes.

When deformation is more intense the f_2 phase again may appear to be the first to modify the transecting nature of the dykes. Discordant

Fig. 2.2

A map to show the variation in amount of f_2 deformation within a small area, south-east of Trinival (780265). The f_2 structures vary from tight folds to open structures which have little effect on the discordant nature of the Scourie-dyke (shown in black). The east - west trending folds in the north are of the f_4 generation. A late pegmatite is shown stippled.

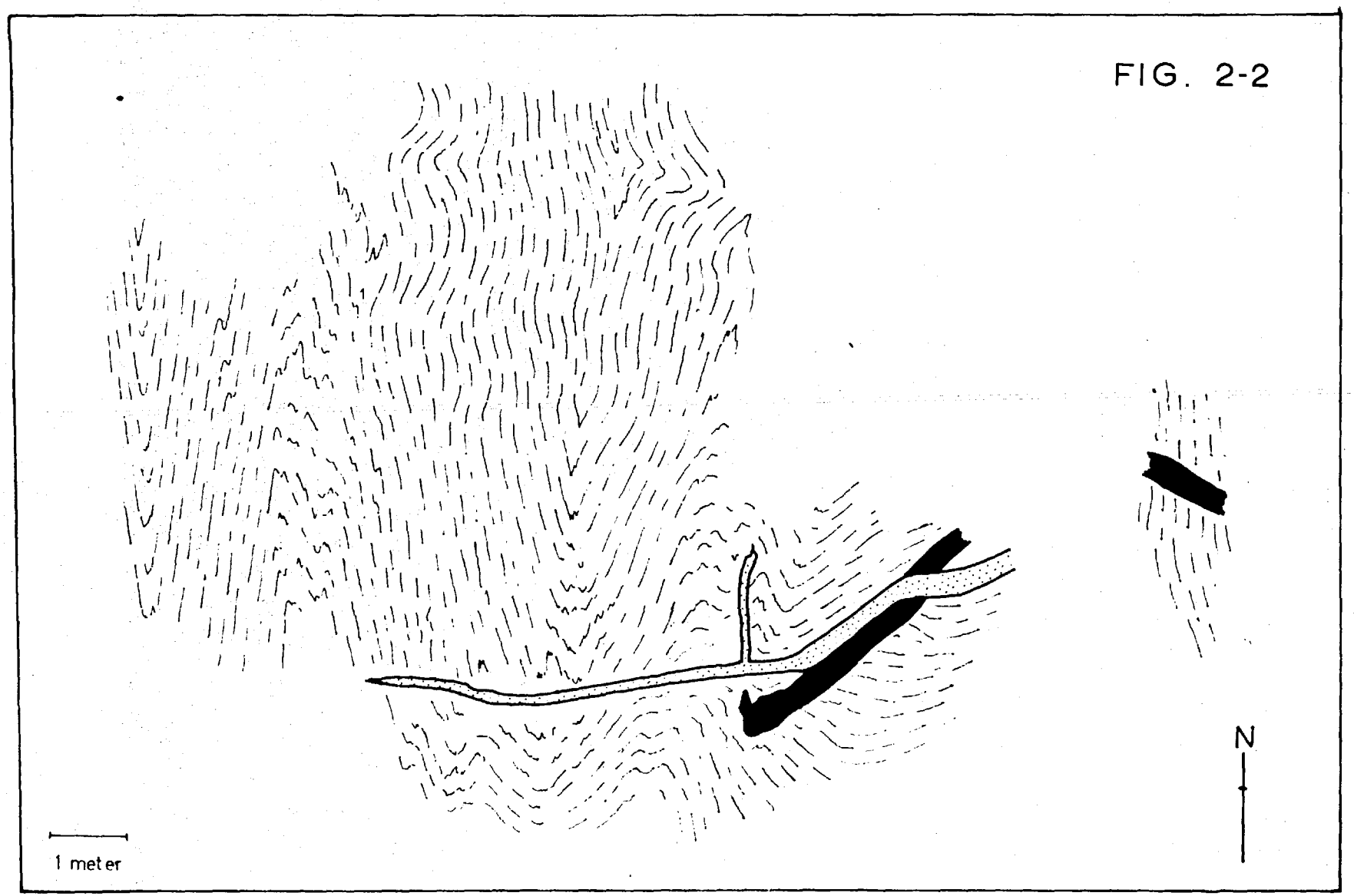


FIG. 2-2

structures are still visible close to the fold hinges and an f_2 fabric may be the first that the dyke sustained.

From the orientation of axial planes and f_2 foliation planes in the areas of low deformation and from the general trend of the enveloping surfaces of later large scale structures, the f_2 X - Y plane would appear to strike approximately N.10° - 20°, with a dip of about 40° to the west. The plunge of the f_2 lineations is to the north-west or south-east as shown in Map 4. This plunge variation is probably due to a later re-orientation (see sections 2.6D and 6.2).

2.5A F_2 small scale structures

Folds of this generation are common throughout South Uist. The fold shape in the regions of north-westerly plunge is dominantly of "Z" or "M" shaped folds; relatively few "S" shaped folds occur.

In some western areas it is commonly only the dykes which are folded, and the foliation remains axial planar to the f_2 structures, but elsewhere both dykes and gneissose foliation are folded. One can therefore infer that there was a variation in the orientation of the gneissose foliation prior to f_2.

On the limbs of the large f_2 folds, or even in zones of intense f_2 "M" folding, the gneiss may attain a banded appearance. Within these areas f_2 folds show a high degree of flattening most early structures are almost obliterated and the dykes which now occur as basic sheets may suffer pinch and swell or boudinage deformation. The boudins are usually of the blocky or barrel shaped types and there is often a rapid transition across the foliation from basic bands which have only suffered normal extension or pinch and swell to those which are intensively

boudinaged. This boudinage and variation in amount of boudinage is most conveniently exposed in a roadside quarry near the Central Filling Station (766365) but is best developed in banded gneisses extending from Beinn a' Chara (781318) to the thrust at Ben Eallan (791265). On the south-west of Spin (795304) and on Bennn Bheag Dheas these boudins amply show the effects of the later structures of f_3 and f_4 age.

2.5B F_2 large scale structures

Fig. 2.3 shows the distribution of "S", "Z" and "M" shaped folds of f_2 age and the few larger order structures which may be inferred from these. These large order folds appear asymmetrical, with a "Z" profile (or "S" in the area where the plunge is to the south). Certain lithological units trace out f_2 sttuctures, e.g. the banded old basic gneisses on Bheinn Bheag Fuath (785306), Spin (795314) and Loch Fada (826366).

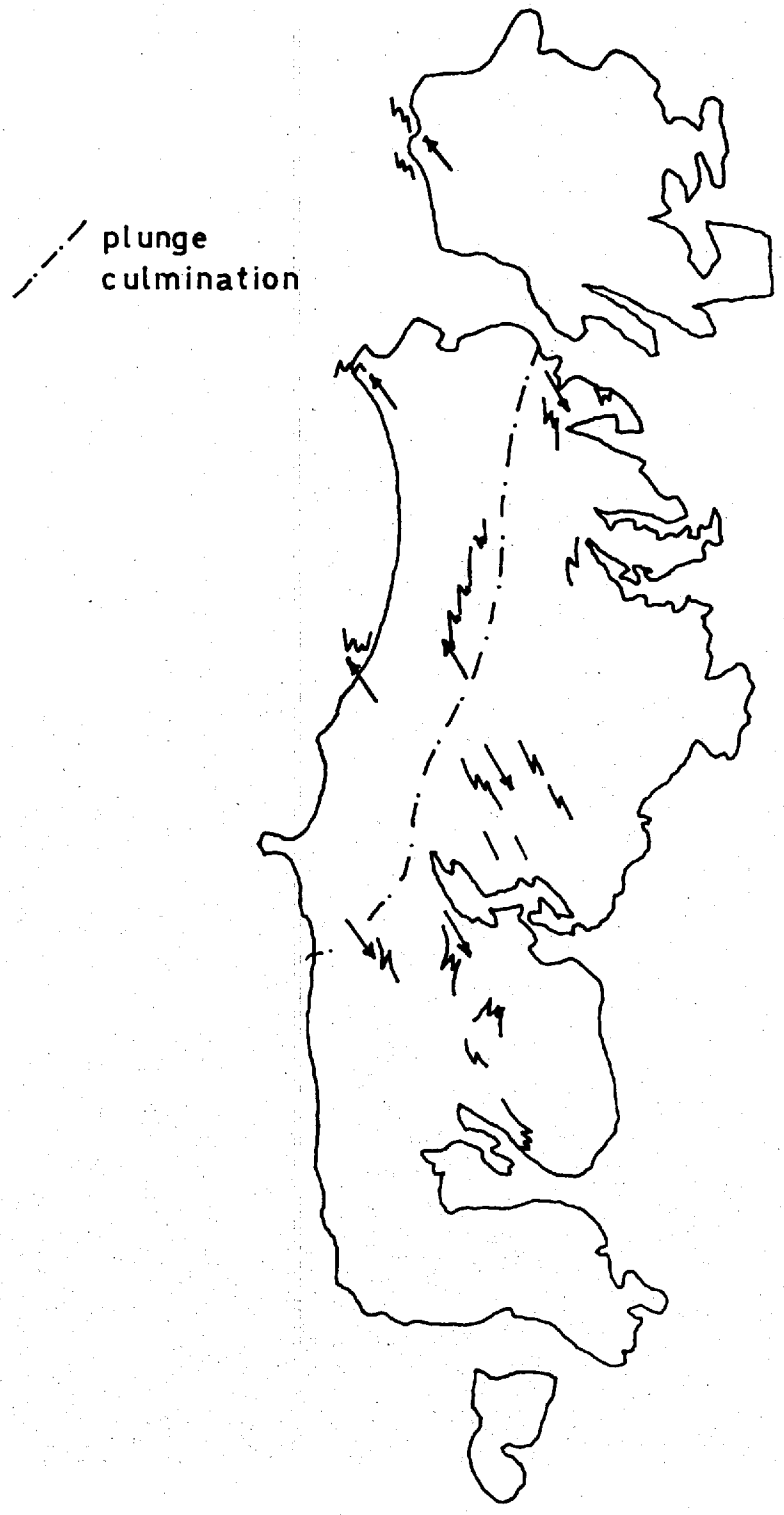
2.5C The f_2 fabrics

It is worthwhile here considering the gneissose fabrics in some detail, though fabrics from the low deformation areas will be described in a later chapter.

a) Scourie-dykes

In the Scourie-dykes where there had presumably been no strong earlier planar fabric, a good schistosity developed, demonstrated by the random orientation of hornblende and feldspar on preferred planes and by mineral concentration and segregation along these planes. Thin feldspathic or garnetiferous horizons formed. These internal lithological layers may themselves have suffered boudinage or pinch and swell during the f_2 phase. This strong planar fabric is best developed on the

FIG. 2-3



F_2 Fold Profiles

limbs of the f_2 folds, and at the cores of the folds or at boudin separations the fabric is more linear. The lineation, usually a hornblende mineral lineation, is parallel to all the local f_2 linear fabrics in the country rock and in the old basic gneiss.

Some of the large basic bodies only exhibit a good f_2 fabric close to their edges. From the edge inwards, the fabric intensity often decreases and the central portions retain a mimetic original igneous texture. On the west side of Beinn Bheag Dheas a basic sheet of Scourie-dyke age, 30 metres wide, shows this relict igneous texture, and south of Loch Eynort it is also shown in the centres of a series of large pods, presumably formed from one boudinaged thick basic layer.

b) in early amphibolites

In these gneisses it is often difficult to be certain that one is dealing entirely with a Laxfordian fabric and not with one partially of Scourian age. If a basic gneiss is highly deformed, then one often cannot be sure whether it is of Scourie-dyke generation or older, and thus one cannot be certain to what extent the fabric is Laxfordian or Scourian.

Within the areas of high f_2 deformation, the Scourian mineral banding, lithological layering, metamorphic segregations and migmatitic veins within these old basic gneisses all become transposed into parallelism with the f_2 planar fabric. Generally, the mineral fabric may be planar, where hornblendes show a random orientation on the schistosity plane, or linear with a preferred orientation of all linear minerals.

c) in acid gneiss

The apparent intensity of the fabric shown by these rocks depends not only on the strain history but also on the mineral composition of the rocks; biotite-, silliminite- or hornblende rich rocks show much better

fabrics than the normal quartzo-feldspathic rocks. The mineral fabric may be planar but is more usually linear.

2.5D F_2 deformation of earlier structures

Often the f_2 deformation was of sufficiently high intensity to obliterate most of the earlier structures, leaving only ghost structures or floating intrafolial folds.

There are however occasional fold interference patterns with folds of Scourian age. These may be of the closed or open types. Large scale Scourian folds probably exist; the lithological pattern (Map 1) often does not entirely fit the proposed f_2 structures and probably results from the superimposition of f_2 strucutres on a Scourian structural pattern.

2.6 The f_3 phase of deformation

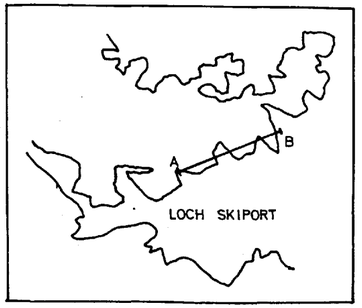
2.6A General

This is one of the more difficult fold phases to describe as the folds are extremely variable in size and shape. Maps 1 and 3 show the distribution of the main f_3 fold structures and some of their effects can be seen in the sections on Map 3. The axial planes trend generally north-west - south-east with a dip of 0° to 90° to the north-west.

In the north, between Carnan and Skiport, the f_3 folds appear as fairly open structures (section C-D, Map 3); the f_2 fabric and folds, the gneissosity and banding are reorientated round large open reclined structures with a gentle limb dipping from 0° to 20° to the west, and a steep limb either vertical or relatively overturned, dipping steeply to the east. In the core of some of the large reclined folds, at East

FIG. 2-4

Section through f_3 structures, East Geirinish



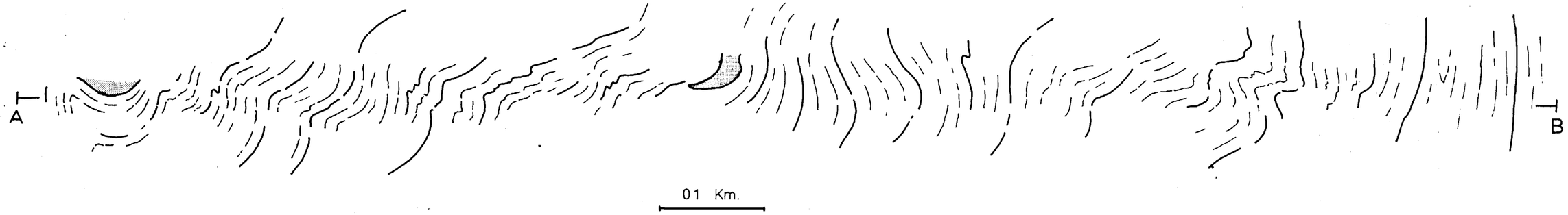


Fig. 2.5

Section through the f_3 structures on Rudha Caolas Luirsay, north of Loch Skiport (see inset diagram). Metasediments are shown by a coarse stipple, basic rock by a fine stipple (see also Fig. 2.4).

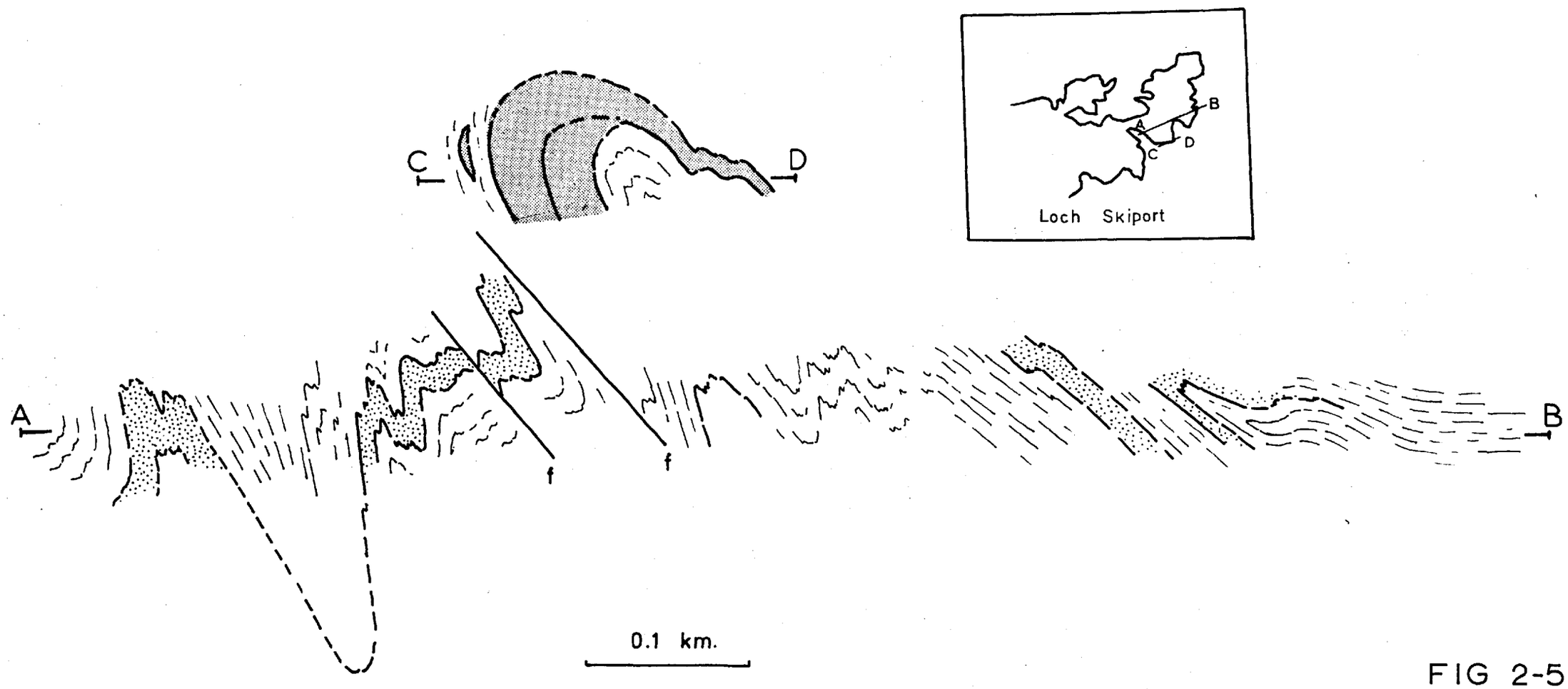


FIG 2-5

Geirinish (Map 1 and Figs. 2.4 and 2.5) the f_3 structures are much tighter, almost isoclinal, and there is a much greater variation in size and number of fold orders.

To the south, round Skiport (e.g. round Loch Teanga 817383), the large f_3 folds are relatively open but are associated with many small scale structures.. This area has suffered extensive f_4 deformation with the result that the f_3 fold plunge varies considerably, and the f_4 plunge varies from steep on the f_3 steep limb to gentle on the f_3 gentle limb. The large fold structure on Ben Tarbert could possibly be of f_3 age, but because of its varied plunge to the north of Loch Airigh (808389) where it refolds earlier recumbent features I consider it to be of f_4 age.

From Hecla to Beinn Mhor, the foliation swings round in a large open reclined synform, the flat limb dipping at 25°, to the south-west at Bealach Hellisdale (814321) or to the north-west at Hoarsal (775359). The hill Spin and the southern slopes of Beinn Mhor are occupied by a large, moderately inclined antiformal fold (section I-J, Map 3), the overturned limb dipping at between 40° and 60° to the north-east. To the south-west of this structure there is a steeply dipping planar zone interspersed with a few f_3 folds of all sizes. On Arnaval a large upright tight f_3 fold deforms a group of early basic gneisses (section K-L, Map 3).

To the south-west of Ben Corary and on Sheaval, the small scale folds have generally an "M" shape, and to the south-west, folds of various sizes show an "S" profile (when plunging to the north-west). This would indicate that the area from Ben Corary to Glen Labisdale is occupied by an upright isoclinal f_3 synform as shown on Map 3 and sections K-L and

M-N.

2.6B F_3 small scale structures

Small-scale folds are poorly developed in the areas where the large scale folds are open structures, but at East Geirinish, locally round Loch Skiport and south-west of Spin they are common. Their variations in profile can be used to trace out the form of the major structures noted above. In many areas, folding of f_2 boudins is common, but it is difficult to identify with certainty many boudins of f_3 age. No f_3 boudins are to be expected or are found in the north, where the folds remain as open structures, and hence the foliation remains in the contraction field of the strain ellipsoid.

2.6C F_3 fold interference structures

The most obvious refolding by the f_3 phase occurs in the north round Gashernish and Calternish, where the f_2 major structures and minor folds are re-orientated round the large open f_3 folds. Fig. 2.5 shows isoclinal f_2 folds being refolded by this phase, and other similar examples are very common in this eastern area. On Sheaval and Arnaval small-scale interference patterns are fairly common. All the fold interference patterns are of the open type, as f_3 is co-axial with the earlier structures.

2.6D Plunge of f_3 structures and the orientation of the f_3 axial planes

In most localities the plunge of the f_3 structures is co-axial with those of f_2 age, and in a very few localities only (e.g. Sgeir Dremisdale) is an earlier lineation folded by f_3 folds.

The variation in f_2 and f_3 fold and lineation plunges is shown in Map 4. The f_3 folds are periclinal structures trending broadly north-west - south-east, but the line joining the plunge culminations of the major f_3 periclines has a north-north-east - south-south-west trend. Where the f_3 plunge is to the north-west, the f_3 axial plane dips to the north-east with a west-north-west strike. Where the f_3 plunge is to the south-east the f_3 axial plane dips to the east with a strike of N. 20° - 30°. The f_3 vertical axial planes and the vertical foliation show no change in strike. It is therefore obvious that as both the f_2 and f_3 plunges show this variation and as there is a corresponding change in orientation of f_3 fold axial planes, then this variation cannot be due to the interaction of an f_3 strain of uniform orientation on an originally irregular surface, but must be due to a re-orientating mechanism. The "re-orientation" occured prior to the development of the f_4 folds and is probably and expression of the heterogeneity of the f_3 deformation (section 6.2).

2.6E Variation in dip of the f_3 axial plane

Disregarding for the moment the change in strike of the axial planes, which is due to heterogeneous f_3 strain, there is also a variation in dip of the axial plane. In the northern area, where the folds are open, the axial plane generally has a dip of between 10° and 50° to the east. As the f_3 structures become tighter to the south, the axial plane steepens to the vertical, and on the southern limb of the main synform it has a steep dip to the south-west. This large axial plane fan is shown in the sections in Map 3.

The small scale folds may also show fanning axial planes, the fan

sometimes being almost through 180°. In all cases the fan is about the f_3 fold plunge and is probably related to the formation of larger order structures (section 6.2).

2.6F F_3 fabric

In the north, where the f_3 phase formed large open folds, an f_3 fabric is rarely seen. The strength of the fabric increases to the south-west to the tight synformal area, where a new schistosity was developed in the acid gneiss axial planar to the f_3 folds. The intersection of the f_3 X - Y strain plane on the f_2 or pre-f_2 fabric produced a good rodding or crinkle lineation. This f_3 schistosity and linear fabric occurs locally in the north, notably around Loch Skiport and East Geirinish.

In the basic gneisses, the f_3 deformation phase induced a preferred hornblende orientation parallel to the intersection of the f_3 X - Y plane on the earlier schistosity. The basic gneiss then became a homogeneous but lineated amphibolite.

2.7 F_4 phase of deformation

The most intense late phase deformation is concentrated along the east side of South Uist, where there are tight to isoclinal folds of all sizes and orders. On the west coast, the f_4 effects are usually seen only as local warps or monoclinal structures which fold an earlier lineation. From west to east, there is a gradual increase in the amount of f_4 deformation and presumably f_4 strain (section 6.3), with an increase in tightness and number of fold orders. This increase culminates in the Skiport and Glean Mhor structures (Map II). The change in geometry of the f_4 folds from west to east will be examined in Chapter VI.

2.7A Trend of the f_4 axial plane

Map 5 shows the general trend of the f_4 structures. This varies from N. 140° round Loch Bee to N. 90° - 95° round Loch Skiport and N. 70° south of Loch Eynort. The axial planes usually have a very steep southerly dip, often vertical, but occasionally as low as 30° - 40°. Areas where the dip is not vertical are shown in Map 5.

The f_4 fold plunge is far from co-axial with the earlier fold axes, and is dependant on the previous foliation orientation. Many f_4 folds (e.g. the Ben Tarbert - Skiport structures) change from steeply plunging antiforms to neutral folds to gently plunging synforms as the axial plane crosses the f_3 structures.

2.7B F_4 small scale structures

F_4 small-scale structures are most common in the east, especially round Loch Skiport and Glean Mor and in these areas they are by far the most common and easily understandable of the fold groups.

Within the Skiport area, the f_4 large order folds become tight and the foliation therefore enters the extension field of the strain ellipsoid, with the result that boudins form. Outside this area the large order folds are more open and presumably the foliation remained outside the extension field as boudins are rare. The form of the boudins at Skiport is usually of a ragged pinch and swell or irregular tear boudin.

Boudins of earlier phases suffer further deformation. Boudinage infolds become flattened and attenuated well into the boudin necks, and the boudins themselves may be folded or show evidence of a rotation of the foliation relative to themselves.

Many f_4 folds appear to be modified by an inhomogeneous shear component resulting in a flattening or stretching of certain limbs. Most of these "shears" are parallel to the f_4 fabric but some appear unrelated to any fold structure. They will be examined in more detail in Chapter VI.

2.7C F_4 fabric

The f_4 deformation phase was accompanied by a phase of amphibolite facies metamorphism and migmatisation which continued in the areas of highest deformation as a period of post-tectonic static recrystallisation. Therefore in the Skiport region and South of Beinn Mor, most syn-tectonic f_4 fabrics have been obliterated; but locally an f_4 axial plane schistosity is preserved and in the basic gneiss there is often a strong hornblende mineral lineation. In the west and south-west the f_4 fabric had more chance to survive and there is often a good penetrative schistosity within the acid gneiss.

In a large Scourie-dyke basic mass on the west side of Beinn Bheag Dheas (781293) an f_4 fold produces an f_4 schistosity through a previously non-tectonised rock. This fabric is significant in that it is more planar than linear. It appears that the first tectonite fabrics to affect Scourie-dykes are always more planar than linear, no matter what the age of the fabric, and that linear fabrics are produced by the superimposition of two or more strains at a high angle to each other. This phenomenon will be examined again in more detail in Chapter V.

2.7D Folding of earlier structures

The shape of the interference patterns depends on the orientations of the component fold systems. On the f_3 gentle limbs or close to the

hinges of large order f_3 folds (as at Arnaval, East Geirinish, North Skiport) closed interference patterns (eyes - mushrooms) develop between f_3 and f_4 folds. Large dome and basin structures are produced; the largest of these occurs on Arnaval where f_2 and f_3 folds are refolded by the f_4 phase. Similarly, when f_2 folds occur on the gently dipping limbs of large order f_3 folds, then they also make closed interference patterns with the f_4 folds.

When the f_4 plunge is steep, i.e. on the f_3 steep limbs or in the tightly folded regions, then open interference patterns result. Between Loch Teanga and Loch Airigh (approximately 814384) these interference patterns are well developed between folds of the f_2 and f_4 phases.

2.7E Folding of early linear fabrics

The distribution of the f_2 and f_3 lineation directions is the result of two deformation phases. Prior to the f_4 phase the heterogeneities in the f_3 strain caused a re-orientation of the linear fabrics to give plunges to the north-west and south-eastas alreaddescribed (see Map 4). The f_4 phase then re-folded this lineation distribution. South Uist can therefore be split up,into several sub-areas (Map 4) which represent areas in which the early lineation is modified in different ways.

a) Plunge of early linear fabric re-orientated by heterogeneous f_3 only.

Steregrams I, III, VI and VIII (Map 4) give the distribution of early fold plunges in areas relatively unaffected by f_4 folding. The plunge varies from 45° to 50° to the north-west to 40° to 50° to the south-east, though locally it may increase to over 80° in either direction. Some of the extreme plunge values have been possibly accentuated by f_4 deformation.

b) Plunge of early linear fabrics affected by f_4 folds

(i) When the f_4 plunge is gentle

The f_2/f_3 lineation usually lies on a foliation plane and when this plane is folded by the f_4 phase, the lineation traces out a curved locus in space about the f_4 axis (assuming non-similar folding). When the plane had a low dip, as on the gentle limb of an f_3 fold, then the f_4 axis has a gentle plunge, usually at a high angle to the original orientation of the f_2/f_3 lineation,,and thus the locus traced out by the deformed lineation lies on a polar great circle (Stereograms VI, IX). The deformed lineation pattern is therefore very similar to that given by the heterogeneous f_3 deformation, though of a very different origin.

(ii) When the f_4 plunge is dominantly steep

When the f_3 fold limb was steeply dipping, then the deformed lineation traces out a locus about a steeply plunging f_4 axis (Stereograms II, IV). When the f_4 folds are open structures the lineation traces out part of a helix, and on the stereogram the lineation plot traces out part of a small circle (or partially modified small circle). When there was a higher amount of f_4 deformation, and the f_4 folds were tighter, the trace of this lineation became modified. Stereogram X shows the trace of the early lineation rotated about a steeply plunging f_4 axis, but now the trace is more nearly a great circle than a small circle. Similarly in the area of Loch Skiport the lineation traces out a great circle and the f_4 folds in this region are tight structures wn which there has been a large amount of superimposed homogeneous strain. However this area is more complex than would at first appear as the area has

highly variable f_4 plunges and in the most critical areas the early lineation is almost obliterated by f_4 migmatisation and recrystallisation.

CHAPTER III

THE METAMORPHIC HISTORY OF THE WESTERN GNEISSES

3.1 General

Dearnley (1962) and Dearnley and Dunning (1968) considered the history of South Uist to have been marked by an alternation of granulite and amphibolite facies metamorphisms:

Late Laxfordian (1600 ± m.y.)	amphibolite facies metamorphism
Early Laxfordian	granulite facies metamorphism
Late Scourian (Inverian) (2200 ± m.y.)	amphibolite facies metamorphism
Early Scourian	granulite facies metamorphism

I found, however, that the tectonic and metamorphic history could not be so simply split up into "Early" and "Late" events, not did I find it as easy as the above authors to place the rocks in a specific metamorphic facies, and at many times I was led to question the actual idea of a "metamorphic facies".

I do not consider myself a metamorphic petrologist, and with fifteen months spent analysing rocks in the field, there has been very little time for any detailed mineral studies. The petrological part of this thesis must by modern standards, therefore, appear inadequate, but I hope to give some broad generalisations of the metamorphic history as seen from the field, from hand specimens or from microscope thin sections.

I have found that the only way that I could understand any of the metamorphic history was to try to relate the metamorphism to the fabrics which formed syn-kinematically during the different tectonic episodes.

Therefore I shall discuss first the metamorphism as related to the fabrics in the various rock types, including Scourie-dykes, early basic rocks, metasediments and acid gneisses. The Scourie-dykes are specially significant, as in many it is relatively easy to date the fabric and as they show only metamorphic assemblages of Laxfordian age. They show very different metamorphic assemblages from the other rocks of South Uist, however, and I believe that it was because Dearnley and Dunning placed too much literal emphasis on the Scourie-dyke assemblages that they reached conclusions which do not seem to be borne out by a more general survey.

In the second part of the chapter I shall attempt to summarise the Laxfordian metamorphic history and shall also discuss the significance of the term "granulite facies" in relation to the rocks of South Uist.

3.2 Problems in fabric analysis

When the rock fabric is determined by the preferred orientation of individual mineral grains, with no later mimetic recrystallisation, then it ought to be very easy to assign certain metamorphic minerals to metamorphic growth during a specific tectonic phase. There is the problem, however, that certain minerals, horneblende for example, may not have fully recrystallised to produce the new fabric, but may have been rotated to their new fabric position by the latest strain increments. There is also the problem that many of the South Uist gneisses, especially the acid gneisses, show various stages of late "annealing", from mimetic recrystallisation over the tectonite fabric to the production of completely recrystallised and homogenous rocks. This post-kinematic recrystallisation is accompanied by an increase in grain size; and sometimes the recrystallisation of the rocks into what appears to be a different metamorphic

facies from that in which the fabric originally developed.

The main problem with this post-kinematic recrystallisation is that it is always very difficult to assign it to any particular period of time. In a few places, as at Ardivachar or Garry-a-siar, one can trace rocks which have a fabric of a certain age into rocks without this fabric but with the same mineralogy, and in these areas one can make a reasonable estimate as to the age of the mimetic minerals. Generally, this is not the case.

3.3 The metamorphic history of the individual rock types

3.3A The metamorphism of the Scourie-dykes

Six dyke assemblages have been chosen to begin this discussion, in the hope that these will not only be useful for determining the metamorphic history but also for discussing the sequence and significance of early Laxfordian fabrics (see section 4.4). In these six dyke assemblages I shall try to take the fabric development from the early stages of dyke deformation and metamorphism to the condition seen in the more highly deformed zones. There is obviously not one simple sequence as the dykes concerned have suffered several deformation phases and it is often difficult to date the individual mineral fabrics.

One needs to start with the original mineral composition of the dyke. Orthopyroxene has been found in a small number of relatively undeformed dykes which still appear to carry an original igneous texture. Clinopyroxene can occasionally be found as large poikilitic laths, probably in its original igneous state. Nowhere, however, can both undeformed and unmetamorphosed dykes be found and any thoughts on the original nature of the dykes can be no better than intelligent guesses. It would appear that the dykes were originally tholeiitic. (see also Dearnley 1963), and

that they originally contained clinopyroxene and orthopyroxene in unknown proportions.

1) This first example is taken from the centre of the large Scourie-dyke mass north of Loch Eynort (782292). A most striking feature about this rock, both in the field and in thin section, is the texture, in which mafic minerals form relatively large clusters separated by the feldspar and quartz. The centre of most clusters contains a large lath of clinopyroxene, altering to hornblende at the edges, and surrounded by a mass of smaller idioblastic grains of dark green hornblende and clinopyroxene. This mass is intergrown with and rimmed by small sub-idioblastic garnets and iron ore, the iron ore occasionally forming an almost complete rim around the mafic cluster.

The mineralogy and the mineral fabric of this rock can be considered to be due to the relatively static metamorphism of an originally ophitic igneous texture. Garnet presumably formed at the edge of the mafic assemblages, by the interaction of pyroxene and feldspar, and the iron ore rim may have resulted from exsolution from some of the mafic minerals and from recrystallisation of the pre-existing iron ore grains. The large laths of clinopyroxene probably represent unaltered grains of original igneous pyroxene.

2) The second example is taken from the centre of the basic mass at Carnan (824448). Once again the mafic minerals form clusters, but in this case these clusters have been drawn out to give a very weak linear fabric. This local patch of rock is however different from that mentioned above in that garnet is rare and the mafic clusters are made up of small

equant grains of clinopyroxene and dark green hornblende with some orthopyroxene. The orthopyroxene forms small grains generally showing some stage of alteration to hornblende. It could be original igneous orthopyroxene or a metamorphic orthopyroxene. One cannot tell.

3) From the same rock mass as the above this specimen shows a similar relict igneous texture, but here the mafic clusters consist largely of dark green hornblende, clinopyroxene and garnet, the garnet locally forming quite large poikiloblastic masses. The garnet presumably formed by the reaction:-

Orthopyroxene + plagioclase = almandine + clinopyroxene + quartz
(Ringwood and Green 1966)

Much of the basic mass has this mineralogy and the localised anomalous occurences of orthopyroxene, as in specimen 2, could be due to local chemical differences in the rock or to variations in the reaction rate between orthopyroxene and feldspar.

The mafic clusters, as in specimen 2, above, show a certain tectonic fabric, relatable to the f_2 deformation phase. One can date the breakdown of the old igneous assemblages and the recrystallisation of the small grains of pyroxene, hornblende, garnet and plagioclase as being closely related to this phase. There is a later stage of recrystallisation shown by this specimen, in that there are also some large grains of green hornblende growing over the earlier hornblendes, and some of the plagioclase shows late alteration to sericite, epidote, etc.. The age of this later alteration is completely unknown. All one can conclude is that there was recrystallisation at some period following the f_2 phase, when the rock was slightly cooler.

4) This next specimen is taken from a much more highly deformed dyke from north of Culla Bay, Benbecula (757546), where in the field the dyke shows a strong schistosity of f_1 age. In thin section this schistosity can be seen to be caused by the preferred orientation of biotite, and a lineation to be caused by the intersection of a number of the biotite grains and by the orientation of the hornblende. The intervening plagioclases and quartz form small irregular grains and there are often elongated irregular blobs of sphene lying along the schistosity plane. There is no clinopyroxene, no orthopyroxene and no garnet.

5) This specimen, from a deformed dyke at Garry-a-siar, Benbecula, shows a well-developed linear fabric of f_2 age, which in hand specimen appears to be produced by the orientation of relatively small hornblende grains. In thin section, however, there appears to be no such uniformity of hornblende orientation. There is an obvious mineral banding, shown by concentrations of both large and small idioblastic hornblendes and of garnets. There is no obvious lineation, and it would appear that the hornblendes show an almost random orientation upon this mineral banding. It would therefore appear that in this case, the linear fabric is not made up by the orientation of individual mineral grains, but by the linear orientation of small groups of hornblende. If this interpretation is correct, then it would appear that there are two types of fabric within this rock: a "lithological fabric" caused by the shape and orientation of very small groups of minerals, and a mineral fabric caused by the preferred orientation of individual mineral grains. These two fabrics need not be co-eval, neither need they show the same fabric class (i.e. planar, planar-linear, etc.). This significant point will be brought up

again in section 4.4. It would indicate that the individual mineral grains need not necessarily give a fabric related to the finite strain within the rock, but could give fabrics related more to the strain increments which occurred during the crystallisation. It will be argued later (section 4.4) that the lithological fabrics suffer strain by rotation and hence their fabric shows the state of finite strain in the rock. Individual minerals appear to deform not only by simple rotation but also by recrystallisation.

This now leads one on to another problem, to decide how much hornblende or, in other cases, how much pyroxene, has deformed by body rotation and how much by recrystallisation. In this case it may be that almost all the individual mineral grains deformed by recrystallisation, while only the small groups of minerals, which give the linear fabric in hand specimen, are indicative of the finite strain. There is no pyroxene in this rock.

6. This next specimen, also from Garry-a-siar, shows the effects of further recrystallisation on the dyke fabric. In hand specimen, the lineation is coarse, but still obvious, and is produced by the orientation of medium to large sized hornblende grains. In thin section, the hornblende appears as relatively large dark green grains with sutured edges. There is very little obvious microscopic fabric and it would appear that there has been mimetic recrystallisation over the tectonite fabric and the original groups of small mineral grains recrystallised to give large irregular hornblendes.

In the above tectonised dykes (specimens 4, 5 and 6), pyroxene is absent. This is by no means the case in every dyke specimen;

Dearnley has described examples of tectonised clinopyroxene and garnet bearing dykes, and it appears that in the centres of the less deformed dykes, as at Ardivachar and Garry-a-siar, small pale green grains of clinopyroxene can form up to 10% of the rock.

Sometimes there is a difficulty in distinguishing the age of the pyroxene; the grains do not usually show a preferred orientation and it is difficult to date them relative to the deformation phase. Some pyroxene could be igneous, but it is probable that much of it was coeval with the tectonised hornblende and therefore metamorphic. The amount of pyroxene relative to hornblende appears to decrease with the intensity of the deformation and deformation fabric, and also with the amount of later mimetic recrystallisation. In rocks which have suffered mimetic recrystallisation, any pyroxene often shows evidence of alteration to hornblende, and one has no idea of its origin, either igneous or metamorphic.

The above discussion deals with mineral assemblages found associated with the early deformation fabrics. From an examination of similar assemblages associated with fabrics of all the different fold phases, one can gain some ideas of the metamorphic history of the dykes. The lists of mineral assemblages associated with the different fabrics are summarised in Table 3.1.

Where the dykes have suffered low amounts of deformation, there is a partial preservation of the original texture. a replacement of the original sub-ophitic pyroxene by granoblastic aggregates of secondary clino-pyroxene and hornblende and the recrystallisation of the original plagioclase laths into granoblastic grains. With more deformation the amount of pyroxene decreases along with the relict original texture and the rocks become (pyroxene) amphibolites. This situation can be

With f_1 fabric	Planar fabric, with hornblende - plagioclase - biotite.
With f_2 fabric in areas of low def.	Linear fabric with hornblende, plagioclase, (clino-pyroxene) and occasionally garnet.
With f_2 fabric in areas of high def.	Planar/linear fabric with hornblende - clino-pyroxene - (garnet) - plagioclase or hornblende - (garnet) - plagioclase.
f_2 metamorphism with no tectonic fabric.	Relict igneous texture with hornblende - plagioclase - clinopyroxene - garnet and occasionally hornblende - plagioclase - clinopyroxene - orthopyroxene.
post-f_2 syn- tect. metamorphism on an f_2 fabric.	Linear fabric, with hornblende - plagioclase.. (garnet rare)
post-f_2 syn-tect. metamorphism on no f_2 tectonic fabric.	Planar or planar/linear fabric with hornblende clinopyroxene - plagioclase. or hornblende - plagioclase.

Table 3.1

The fabrics and mineral assemblages of the Scourie-dykes.

found no matter which deformation phase was the first to deform the dykes, the only difference being that garnet usually formed along with clinopyroxene and hornblende during the early phases, but was absent during the f_3 and f_4 events.

Pyroxene therefore occurs as a stable mineral in Scourie-dyke mineral fabrics of all the Laxfordian fold phases. It is a more important constituent of the f_2 fabrics than those of later age and has only once been found in fabrics of f_4 age. In this one instance, at Beinn Bheag Dheas, the f_4 fabric was the first to deform an original igneous texture. Hornblende is also present in all fabrics from f_1 to f_4, and during any mimetic recrystallisation, it is hornblende which appears as the stable mineral and the pyroxene shows various stages of alteration.

The syn-tectonic f_2 metamorphism gave a garnet-bearing assemblage in certain amphibolites. Clinopyroxene is the stable pyroxene and orthopyroxene occurs only rarely. Green and Ringwood (1966) have shown experimentally that under a relatively high pressure, orthopyroxene and plagioclase will recrystallise as clinopyroxene and almandine garnet. The presence of garnet and clinopyroxene in the South Uist assemblages may therefore indicate a rather high pressure metamorphism.

It may be noted that the assemblages listed in Table 3.1 do not indicate a granulite facies metamorphism as defined by Fyfe, Turner and Verhoogen (1958), but indicate more an amphibolite facies metamorphism. This problem will be discussed in much more detail in section 3.4

From Scourie-dykes alone, there is no evidence for a variation in metamorphic conditions across the western gneisses of South Uist during the syn-tectonic f_2 metamorphism. This apparent uniformity is unfortunately different from the metamorphic variations given by other rocks (section 3.3B,C). It could be due to:

a) the possibility that the other rocks are the more sensitive indicators. It is reasonably evident, however, that the dykes are much more sensitive to local variation in P_{H_2O} than the other rocks. They ought therefore to be highly sensitive to changes in metamorphic intensity which involve changes in water fugacity.

b) the relative infrequency of dykes in the most critical areas and therefore to a sampling error.

c) the possibility that there are no variations in f_2 metamorphic conditions across South Uist. The variations shown by the other rock types may only be mimicking earlier (Scourian ?) variations (section 3.4, and Chapter VII).

3.3B Metamorphism of the early basic rocks

The mineral assemblages shown by these rocks are given in Table 3.2. Some large bodies, such as the ultrabasic masses, do not require much discussion; they just shown early assemblages at the centres and later assemblages at the mar ins. As there is rarely a fabric within such rocks, many of the minerals cannot be dated. The early assemblages, therefore, could be of any age prior to the f_3 and f_4 deformation phases, as the production of the green hornblendite at the edges of these bodies often appears to be catalysed by deformation related to the f_3 or f_4 phases.

As for the early amphibolites, there is a notable variation in

	f_2 metamorphism	f_3 and f_4 metamorphism
Ultra-basic rocks	ortho-pyroxene - clino-pyroxene - (brown hornblende).	green hornblende
Early-amphibolites	(clino-pyroxene) - hornblende - plagioclase - biotite. hornblende - plagioclase - biotite - garnet.	hornblende - biotite - plagioclase.
Gabbro-anorthosite	plagioclase(An_{60-70}) clino-pyroxene - hornblende - garnet.	plagioclase(An_{30-40}) - hornblende.

Table 3.2

Mineral assemblages within the pre-Scourie-dyke basic and ultra-basic rocks.

mineralogy and apparent metamorphic state of these bodies across South Uist, and for simplicity one can divide the island into four sub-areas (Fig 3.1a) and discuss the main characters of each in turn.

(1) In this area the amount of f_3 and f_4 deformation is lower than it is to the east and south, and the amphibolites usually contain earlier mineral assemblages. An f_2 mineral banding is strong, shown by variations in relative amounts of garnet, plagioclase and hornblende, and large post-f_2 garnets often grow across this foliation. In some of the basic rocks, sporadic quartzo-feldspathic veins define f_2 folds or an f_3 planar fabric, indicating possible earlier (Scourian) migmatitic events.

(11) Here and to the east, the f_3 or f_4 recrystallisation is much stronger, and the early fabrics and mineralogy can only be found at the centres of the large basic bodies. Generally the rocks are totally amphibolite, or garnet amphibolite, but locally (clino-) pyroxene occurs in certain layers of banded gneiss.

(111) Within this area the early amphibolites show a completely different mineralogy and fabric and generally show agmatite structures of more than one generation. Many contain biotite and have a streaky biotite/hornblende fabric produced not just by a preferred mineral orientation but also by the shape and orientation of drawn-out hornblende/biotite clots. Much of this fabric is of f_2 age, but a similar fabric is locally developed by intense f_4 strains, and within many basic rocks there was a well-developed biotite-hornblende-(pyroxene) mineral banding in existence prior to the f_2 phase (section 7.2).

It appears that there have been several phases of agmatisation. Much of this is pre-Laxfordian in age. Basic rocks of Scourie-dyke age show relatively few agmatisation effects developed prior

to the f_4 phase, and some of the intra-basic quartzo-feldspathic material bears an f_2 fabric, or is folded and re-orientated by the f_2 structures. There was a certain, but small, amount of re-mobilisation of the quartzo-feldspathic material followingthe f_2 phase, as the agmatites locally break up f_2 folds, and the f_4 migmatisation again remobilised some of the veins.

(IV) The basic rocks of this area are very similar to those described above (cf Sub-area III), except that the early agmatisation is more widespread and the amphibolites are locally rich in pyroxenes. The most pyroxene occurs in the gneisses close to Loch Teanga (817383). Basic gneisses here contain diopside, hypersthene ± hornblende and plagioclase, or diopside, garnet, plagioclase ± hornblende. Diopside is found in some of the adjacent country rock. Later f_4 metamorphism recrystallised and locally re-mobilised the gneisses, producing hornblende and biotite at the expense of pyroxene, and therefore much of the area now consists of hornblende or biotite bearing rocks.

From this discussion, several points can be made:

(i) to the west, the metamorphic history as shown by these early amphibolites is essentially similar to that shown by the Scourie-dykes. There is relatively little evidence for an early Laxfordian granulite facies metamorphism, though it is possible that the f_2 metamorphism was of a high pressure type, indicated by the growth of garnet during and just after the f_2 phase.

(ii) the early Laxfordian (pre-f_4 and probably pre-f_3) metamorphism round Loch Skiport was of a different nature from that of other parts of South Uist. Here two pyroxenes appear as apparently stabilised

mineral phases, within a previously agmatised group of rocks.

(iii) The degree of f_4 metamorphism increases to the east.

3.3C The metamorphism of the metasediments

Table 3.3 gives the distribution of metamorphic minerals which grew during the Laxfordian metamorphism in the metasedimentary belts of South Uist. Variations in both intensity and type of metamorphism are amply shown by the metasediments, and are summarised below:

(i) Variations in the state of the f_2 metamorphism.

Around Loch Skiport (Fig. 3.1b) the metasediments show an early metamorphism which produced a high grade mineral assemblage. West of Loch Teanga (at 809383) there occur alternating bands of pyroxene bearing or hornblende-biotite bearing metasediments, and typical assemblages include:

garnet-biotite-quartz ± hornblende

garnet-biotite-quartz-diopside-hypersthene ± hornblende

biotite-anthophyllite-quartz ± hornblende.

Following the f_2 deformation period there was a period of static and sometimes pegmatitic mineral growth, with the crystallisation of large grains of pyroxene, biotite, garnet and anthophyllite. The bands or pods of alternating pyroxene and hornblende rock vary in width from a few centimetres to a metre, and all gradations can be found from metasediments in which the mafic constituents are purely pyroxene, to those in which they are purely hornblende or biotite.

Elsewhere on the Uists, the f_2 metamorphism gave garnet-biotite-hornblende bearing assemblages. In parts of the Loch Bee group the f_2 schistosity was defined by syn-kinematic growth of biotite and staurolite, and in the metasediments at the quarry, close to the market stance,

		F_2	Post-F_2	F_3	F_4 and post-F_4
A	Sillimanite	X	X		
	Staurolite	X	(X)		
	Anthophyllite	X	X		
	Biotite	X	X	X	X
	Garnet	?	X	X?	-
	Hornblende	X	X	X	X
B	Clinopyroxene	(X)	X		
	Orthopyroxene	(X)	X		
	Hornblende	X	X	X	X
	Anthophyllite	X	X		
	Biotite	X	X	X	X
	Garnet	X	X	?	

TABLE 3.3

Minerals of successive phases in the metasediment belts of A, the western regions of South Uist, B, the eastern regions.

Fig. 3.1

Variation of Laxfordian metamorphism across South Uist

a) Laxfordian metamorphic zones as defined by the metamorphism of the early amphibolites (see section 3.3B)

b) The area of South Uist where pyroxene occurs in the metasedimentary rocks.

c) Syn-kinematic f_4 metamorphic zones as defined by the metamorphism of the acid gneiss (section 3.3D).

d) Area of late (post-kinematic) retrogression

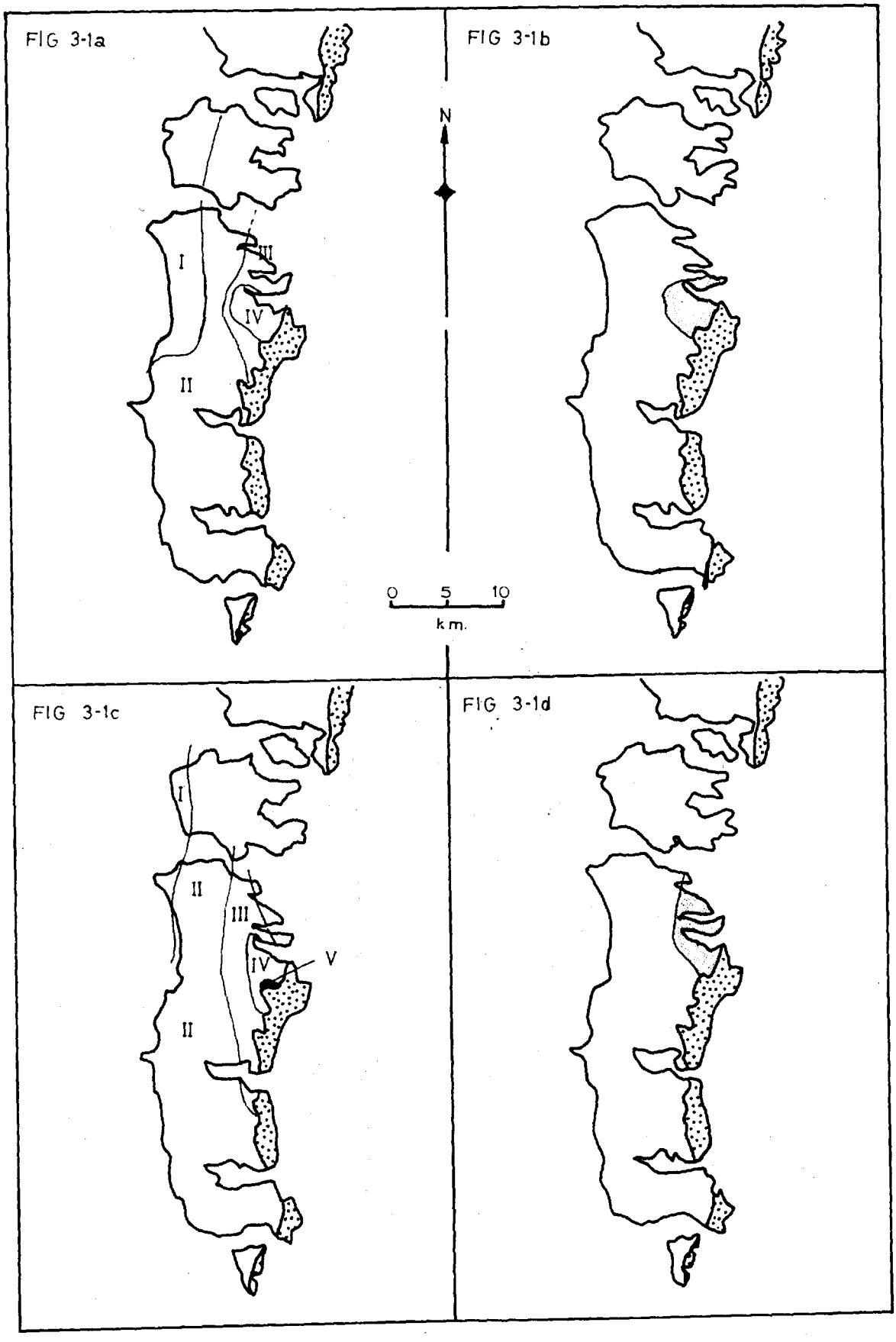


FIG 3-1a
FIG 3-1b
FIG 3-1c
FIG 3-1d
N
I
II
III
IV
V
0 5 10
km.

Benbecula, biotite and silliminite define the f_2 fabric. In these areas there is no trace of pyroxene.

(ii) Variations in amount of late metamorphism.

The groups of metasediment on the east side of South Uist (apart from some patches of the Skiport group) have lost much of their early fabric as a result of late migmatisation and metamorphism. Granite masses were both formed in situ and intruded, and some areas were homogenised into a quartzo-feldspathic mass. In such areas, many of the metasediment belts found it hard to survive. Several belts can be traced along their length into a migmatitic granite gneiss. Granites and pegmatites occupied strips wit in the belts and the remaining metasediments tended to change their mineralogy; biotite, quartz and feldspar being the dominant minerals to crystallise. In the east, therefore, the amount of obvious metasediment was drastically reduced by t the Laxfordian migmatisation.

3.3D The metamorphism of the quartzo-feldspathic gneisses

These are largely quartz-oligoclase-hornblende rocks with variable amounts of biotite and garnet.

(i) Early mineral assemblages

At Ardivachar Point, Dearnley and Dunning (1968, p.363) state that they found acid gneiss showing a relict granulite facies mineral assemblage. In this area, however, I only found an assemblage of plagioclase-hornblende-biotite-garnet-quartz. and this assemblage may be due to the fact the rocks were related to the metasediments in origin, and not necessarily to a granulite facies metamorphism.

At a few localities in the Skiport area, however, (as mentioned

previously), clinopyroxene may be found in acid gneiss close to agmatitic early amphibolites, but elsewhere no trace of pyroxene has been found in the acid gneiss. If pyroxene did once occur in patches of the acid gneiss, at say Ardivachar or Garry-a-siar, then it could produce competence variations within the gneiss and would be most useful in accounting for varying amounts of early Laxfordian strain (see Chapter V). Unfortunately most of the acid gneiss appears to have been made up of quartz-feldspar-hornblende-biotite assemblages, and though of course one cannot be absolutely certain, it would appear that there was no mineralogical difference between the areas of low and high Laxfordian deformation.

(ii) The effects of the f_4 metamorphism.

Most of the mineral assemblages shown by the acid gneiss were formed during the f_4 migmatisation and by late retrogression. Associated with this late recrystallisation there was a marked increase in grain size, and a homogenisation and simplification of the rock fabrics. The gneisses become granitic in texture and often in mineralogy, and there is often evidence for partial melting of the gneiss during and just following the f_4 deformation phase. From field and petrographic studies I have not noticed, however, any change in bulk composition of the rocks. Potash feldspar-bearing granites, which are relatively rare, could have been potash bearing granites before the f_4 effects, and there is no evidence for mass migration of ions as claimed for some other migmatitic areas.

In Fig.3.1c, I have tried to zone the amounts of late metamorphism. This map must, of course, be highly subjective, but it should show the approximate variations in this late recrystallisation. The areas can

be briefly described as follows:-

Area 1.

Within this area, the gneisses show very few effects of late recrystallisation or migmatisation. The early fabrics are often very well preserved and the only evidence of a late metamorphic phase is given by the presence of large discordant pegmatites.

Area 2.

Within this area there has been a certain amount of f_4 syn-tectonic metamorphism. The f_4 folds usually have an f_4 fabric and there is evidence of later recrystallisation or even pegmatite formed in the more deformed regions of these folds.

Area 3.

This area comprises much of South Uist. The rocks show a certain amount of syn-tectonic f_4 metamorphism but many of the fabrics have been obliterated by late static recrystallisation.

Area 4.

Here, there has been syn-tectonic migmatisation, to the extent that the f_4 deformation phase produced disharmonic or "wild" folds (see Chapter VI). Patches of gneiss have been transformed into granite or pegmatite.

Area 5.

Within this small patch almost the whole gneiss has been transformed into a granitic mass, the gneissosity is destroyed and the basic rocks or basic gneissose foliae remain as hornblende-biotite rich schlieren.

(iii) Late retrogression (post-migmatisation)

Fig. 3.1d shows the distribution of rocks which show a lower-grade

mineralogy than those elsewhere. This low-grade metamorphism is indicated by the presence of biotite and chlorite as the mafic constituents and of plagioclases with a lower anorthite content. The plagioclase is often sericitised, and the rock becomes a chlorite-epidote-sericite gneiss. Iron ore in the gneiss is oxidised to haematite/limonite and the rock may be tinted a creamy brown colour as a result. This late alteration, I believe, reflects areas of slower cooling.

3.4 Synthesis of the Laxfordian metamorphic history

In attempting a summary of the Laxfordian metamorphic history, one is faced with several major problems. Such a summary usually entails placing the metamorphic mineral assemblages in specific "metamorphic facies", in order that one can then deduce the metamorphic grade and hence, whether the rocks have been upgraded or retrogressed. Two phases of Laxfordian deformation, however, were accompanied by metamorphism which does not fall easily into the usual definitions of metamorphic facies or metamorphic grades. Of these two phases, the first was the f_2 deformation phase, with its problematical granulite facies metamorphism. The second was the last phase, with its accompanying migmatisation. The evidence concerning the other two deformation phases is apparently much more simple: where f_1 and f_3 fabrics can be found, they both show mineral assemblages characteristic of the amphibolite facies.

Dearnley (1962) and Dearnley and Dunning (1968) have claimed that the early Laxfordian metamorphic phases formed in the granulite facies. I would claim that the answer is by no means so simple, and in this next section, I would like to give a few arguments for and against Dearnley's use of the term 'granulite facies' for the South Uist rocks.

(i) One has first to define the term 'granulite facies'. In the literature granulites are defined as being "high grade metamorphites devoid of micas. They are characterised by kyanite or sillimanite + garnet instead of hornblende or other amphiboles" (Winkler, 1965). The granulite facies, as introduced by Eskola (1939) was thought to originate through a further rise in temperature after the field of the highest-temperature subfacies of the almandine amphibolite facies has been traversed, and according to Turner and Verhoogen, the granulite facies field comprises temperatures between 700 - 800°C. Under the metamorphic conditions envisaged by Fyfe, Turner and Verhoogen (1958) the mineral assemblages from meta-basic rocks are of hornblende + plagioclase + diopside in the amphibolite facies and of hypersthene + diopside + plagioclase in the granulite facies. Recently, however, de Waard (1965) and Green and Ringwood (1966) proposed a subdivision of the granulite facies into a low pressure form, characterised by orthopyroxene and plagioclase, and a high pressure form characterised by the higher density clinopyroxene-almandine-quartz assemblages in meta-basic rocks containing excess silica.

If one examines the metamorphic assemblages of South Uist in the light of the above definitions, then certain rocks could fall into the high pressure sub-facies of the granulite facies (or, more strictly, the hornblende-granulite facies). One has the problem, however, that as pyroxene is stable or metastable in metamorphic mineral assemblages from f_2 to f_4, then one could consider each phase being locally accompanied by hornblende granulite facies metamorphism or high grade amphibolite facies metamorphism.

There is a problem though of the limiting temperatures of this

hornblende granulite facies. When one examines the graph (Yoder and Tilley, 1962) for the hornblende/pyroxene dehydration reaction, it is obvious that pyroxene is stable over a very wide range of temperatures, depending upon the ratio of the partial pressure of water, P_{H_2O}, to the total pressure, P_T. Should one therefore consider rocks as being in the granulite facies, if they crystallised pyroxene at 600°C when P_{H_2O} was negligible, or are they only in the granulite facies when crystallising pyroxene at over 1000°C with $P_{H_2O} = P_T$? I feel that such high temperatures as 1000°C were not realised in the South Uist rocks and that for the pyroxene to crystallise, the P_{H_2O} must have been fairly low. I also feel that until more is known about the original P_{H_2O} of the rocks, then the temperature and pressure conditions cannot be estimated with any degree of accuracy. It could be argued, for instance, that the increase of hornblende in metamorphosed dykes with both deformation and time need not indicate an important lowering of temperature conditions, but might indicate largely an increase in the water fugacity.

It is evident from the previous section that in South Uist, both pyroxene and hornblende crystallised at the same time, both in the Scourie-dykes and in the Loch Skiport metasediments and that this variation must, to some extent, have been due to local variations in P_{H_2O} (c.f. Buddington, 1963). As the presence of pyroxene within the least deformed dykes throughout South Uist appears anomalous with respect to the mineral assemblages shown by all the other rocks, it could be argued that the early metamorphism within these dykes took place in a closed system, and the dykes therefore crystallised pyroxene under lower P_{H_2O} than that of the enclosing rocks. This would imply that if the dykes were more deformed and the system were allowed to become more open or if

more time were available for recrystallisation, then the P_{H_2O} would rise, pyroxene would become unstable and so hornblende or biotite would crystallise as the stable mafic minerals.

It appears, therefore, that though the dyke mineral assemblages indicate that pyroxene was locally stable, the distribution of pyroxene does not reflect local temperature variations but, more likely, variations in amount of available water.

(ii) Dearnley and Dunning (1968) also examined individual minerals to see if they supported their argument for a variation in metamorphic facies with time. They noted a subtle colour change from an olive green hornblende in the centre of an Ardivachar dyke, to a pale green hornblende at the margins, where the dyke was mimetically recrystallised. Associated with this colour change there was also a change in refractive indices, from $\gamma = 1.675 - 1.685$ to $\gamma = 1.660\pm$. From this they concluded that there was a change in some metamorphic condition between the crystallisation times of the two types of hornblende, i.e. between the f_2 deformation phase and some unknown period following this. An examination of dykes from the rest of South Uist substantiates the above. There is a similar subtle colour change in hornblendes, from olive green to green to blue green. This change was associated with a late (post-f_4) retrogression, presumably when the rocks were finally cooling down.

It has been noted previously that in the centres of the large ultrabasic masses, in addition to clino- and ortho-pyroxene, one finds a dark brown hornblende. If these ultrabasic mineral assemblages are of f_2 age, then it would indicate that brown hornblende could exist in rocks of one composition while olive green hornblende existed in rocks of another. This would indicate, therefore, that the colour difference

is partly due to different amounts of certain chemical elements originally inherent in the rocks and I would suggest that until more is known about the detailed composition of these hornblendes, they cannot be used to indicate the conditions of metamorphism.

A rough examination of the anorthite content of the different plagioclase feldspars (using the refractive index or the twinning) possibly could also be used to indicate some metamorphic variations. Feldspars, however, are very difficult to date relative to tectonic fabrics as they often show equant textures which may be indicative of static annealing or recrystallisation. However, feldspars from Scourie-dykes with an f_2 hornblende mineral fabric often have a higher anorthite content than those with later mimetic fabrics (Fig 3.2). Conversely, plagioclases apparently of f_2 age from early basic masses show lower anorthite percentages, while plagioclases from the anorthosite have an anorthite content of up to 70%. The variation in anorthite content therefore reflects a variation in original composition of the rock.

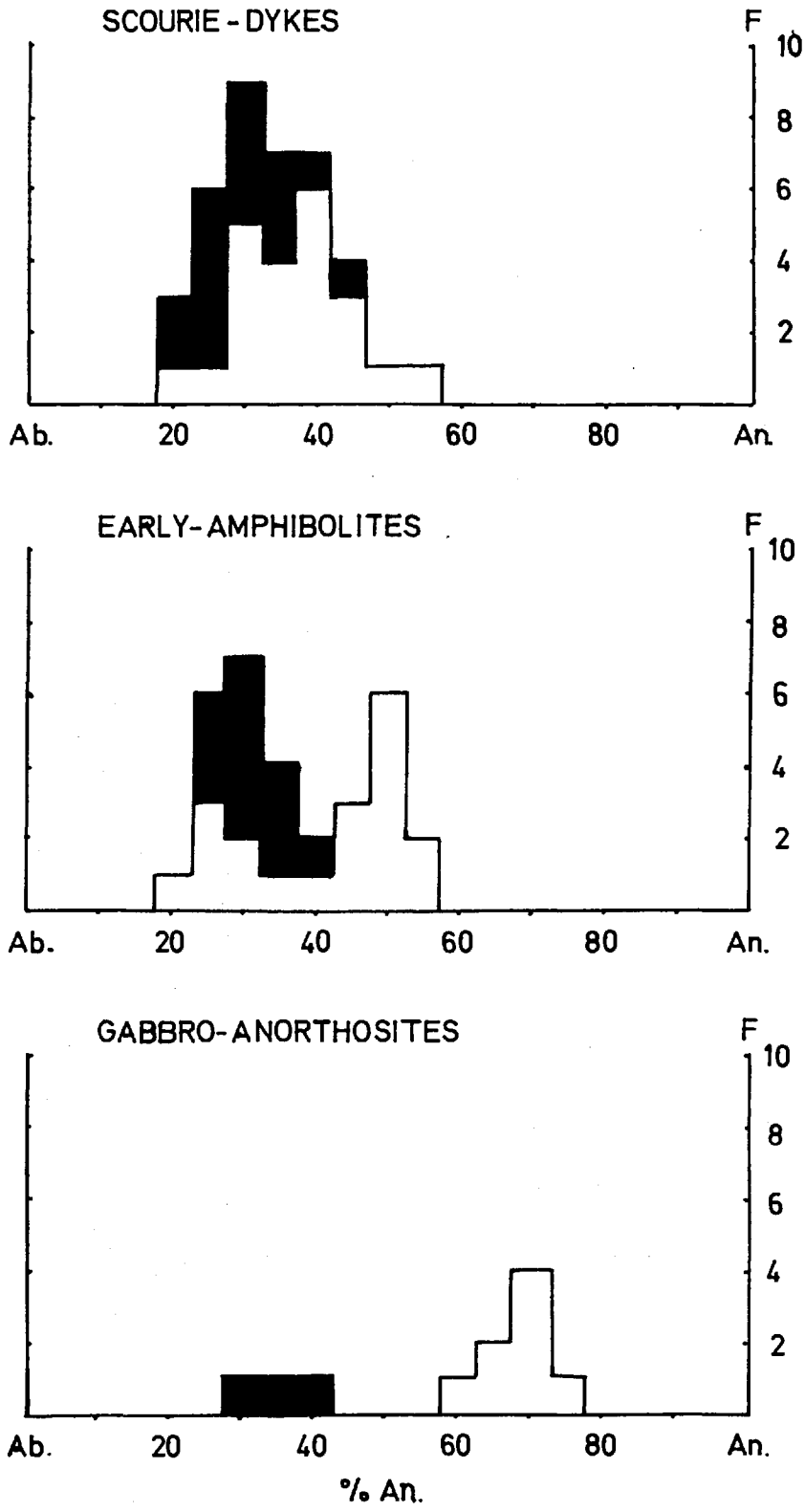
One must conclude that though some of the individual minerals show some variation with date of crystallisation, these variations often involve either non-tectonic textures of which one does not know the exact age, or changes which are associated with the late post-migmatite recrystallisation (section 3.3D(iii)) and probably the slow cooling and retrogression of the rock. The variations in minerals which can be tied down to a particular deformation phase are probably largely due to variations in original rock chemistry.

(iii) Since one cannot always use the mineral assemblages of the Scourie-dykes, nor the variations in composition of the individual minerals, to indicate metamorphic variations in time and space, one has

Fig. 3.2

The composition of plagioclase feldspars (% anorthite) within Scourie-dykes, early amphibolites and gabbro-anorthosites as measured from extinction angles and refractive indices. The clear histograms give the frequency (F) of plagioclases with specific anorthite values of f_1 and f_2 age, the black histograms give those of f_3 and f_4 age.

FIG. 3-2


SCOURIE - DYKES
F
10
8
6
4
2
Ab.
20
40
60
80
An.
EARLY-AMPHIBOLITES
F
10
8
6
4
2
Ab.
20
40
60
80
An.
GABBRO-ANORTHOSITES
F
10
8
6
4
2
Ab.
20
40
60
80
An.
% An.

to fall back on the general assemblages shown by other rocks. These demonstrate a change in mineral assemblage from west to east. In the west, sediments and early amphibolites give hydrous minerals indicative of the amphibolite facies, while to the east anhydrous minerals sometimes occur. During the f_2 phase there was therefore a regional variation in some metamorphic variable across the island.

One cannot be certain, however, that the anhydrous nature of much of the eastern rocks is wholly of f_2 age. The localised variations from anhydrous to hydrous rock do not appear to be related to the different amount of f_2 deformation and it is possible that they may have been in existence before the onset of the Laxfordian. It will be argued in section 7.2 that the early basic rocks of the Skiport area may be similar in origin to a group of pre-Scourie-dyke intrustions on Barra and also possibly of the same age as the large igneous bodies east of the South Uist thrust zone, or in South Harris. A feature that all of these have in common is that a great deal of the meta-igneous rock and also much of the surrounding country rock, bears pyroxene.

(iv) A final point concerns the early metamorphism and its subsequent "retrogression" (Dearnley and Dunning, 1968) to migmatite conditions. Some changes in metamorphic parameters can be seen from the change in mineral assemblages from the f_2 to f_3 phases, e.g. from garnet amphibolite to normal amphibolite. As the presence of garnet in the tholeiitic Scourie-dykes during the f_2 phase probably indicates a high pressure metamorphism, then the alteration of garnet to plagioclase and hornblende during the f_3 and f_4 phases, could indicate a lower pressure or higher temperature metamorphism.

The fact that pyroxene occurs more commonly in Scourie-dykes with an

f_2 fabric than in those with f_3 or f_4 fabrics could, and in certain cases probably does, indicate lower temperature conditions associated with this later deformation. It also indicates a general rise in the amount of water and it is debatable as to whether "retrogressive conditions during the f_4 deformation phase would give rise to the migmatisation.

One therefore has the problem, that during the f_2 period, pyroxene locally was stable in rocks of low P_{H_2O}, while there was relatively little migmatisation. During the f_4 phase, there was migmatisation, yet pyroxene is rare in these rocks. Also, the area of most intense late recrystallisation and migmatisation sits directly over the area which following the f_2 phase was most nearly anhydrous. There are several possible solutions:-

a) Partial melting could have been hindered during the f_2 phase by relatively low P_{H_2O} conditions. If there was then an episode of water migration through the crust, then possibly lower temperatures during the f_4 deformation might cause more migmatisation. Winkler (1965, p.120) has shown, however, that only very small amounts of water are required (approx. 5 wt.%) for low temperature melting (700°C) to take place and as far as I could tell, even within the Skiport area, many rocks contained as many hydrous minerals during the f_2 phase as during the f_4 phase. One has also to explain where the water came from. Did it come from below, from above, from the side? If the water did migrate, then it could not have come from the local area, as that was supposed to be too anhydrous

b) A second possibility is that the f_2 conditions of temperature and pressure were too low to produce migmatisation or granitisation. The

pyroxene bearing assemblages could be due to localised closed system metamorphic effects on rocks which were anhydrous prior to the f_2 deformation. One would expect a granulite facies metamorphism to be associated with a certain amount of partial melting and agmatisation, and I have already shown that the best developed agmatites in South Uist are not only intimately associated with these granulites but also can be dated as being pre-Laxfordian. I would therefore suggest that much of the anhydrous nature of the country rock is due to the same metamorphic effects as those which caused the partial melting of the country rock and the formation of the agmatites, and that the local granulite facies mineral assemblages owe their origin to a pre-Laxfordian event. The f_2 metamorphism in the east need then only be intense enough to recrystallise pyroxene in anhydrous rocks, i.e. it need not necessarily be high enough to cause granitisation or partial melting.

c) A third possibility is that the f_4 metamorphism of the eastern area took place under much higher temperature conditions than that of the f_2, possibly even under temperature conditions which people might assign to the "granulite facies". Within this area one only very rarely finds good mineral assemblages which were syn-kinematic with the f_4 phase and generally finds coarse grained mimetic fabrics which could have formed when the migmatite area was cooling down. Very little is known, therefore, about the conditions under which the main migmatisation and granitisation took place.

d) A slightly different possibility to the above is that the factor of time might have played an important part in determining the amount of late Laxfordian metamorphism. Very little is known about the absolute length of time at which temperatures were high. One can, however, recognise a

great deal of post-kinematic f_4 redrystallisation and it could be that it was the long period of time spent at a high temperature, in the east, which produced the excessive migmatisation effects.

Of these four possible explanations, I find that the third is possible, but rather extreme, while the first (a) invokes external processes of which we have no knowledge. I therefore prefer a combination of the second and fourth (b and d), i.e. the anhydrous nature of the eastern rocks could be due to a local pre-Laxfordian event, while the granitisation and migmatisation could be due to prolonged high temperature conditions.

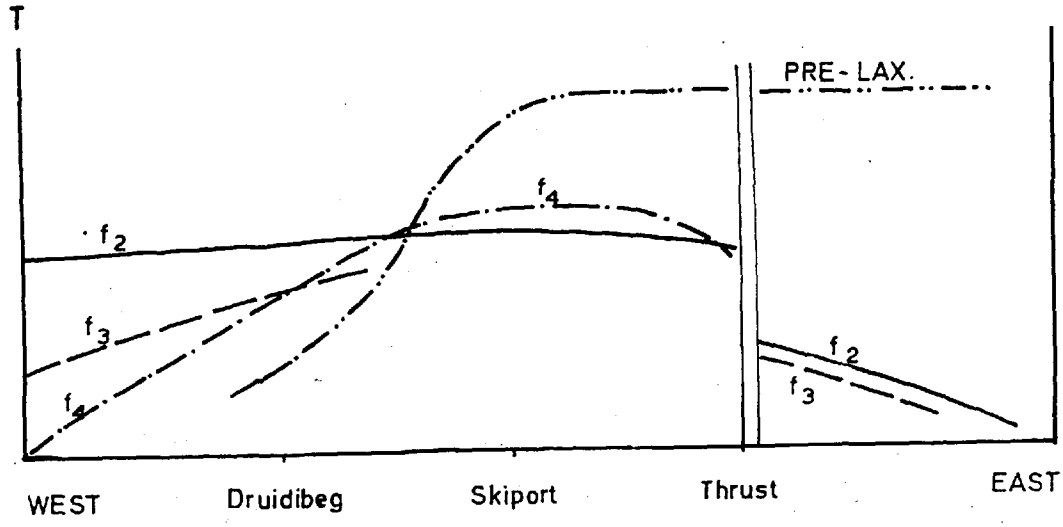
My conclusions about the general variations in relative metamorphic temperature from west to east and south to north across South Uist, and for the different deformation phases, is given in Fig. 3.3. The early pre-Laxfordian temperature variation is, to a large extent, hypothetical, but it would solve a few problems such as those surrounding the pyroxene bearing rocks and the agmatites in the east. I shall discuss it further in the section on pre-Laxfordian metamorphic history (Chapter VII).

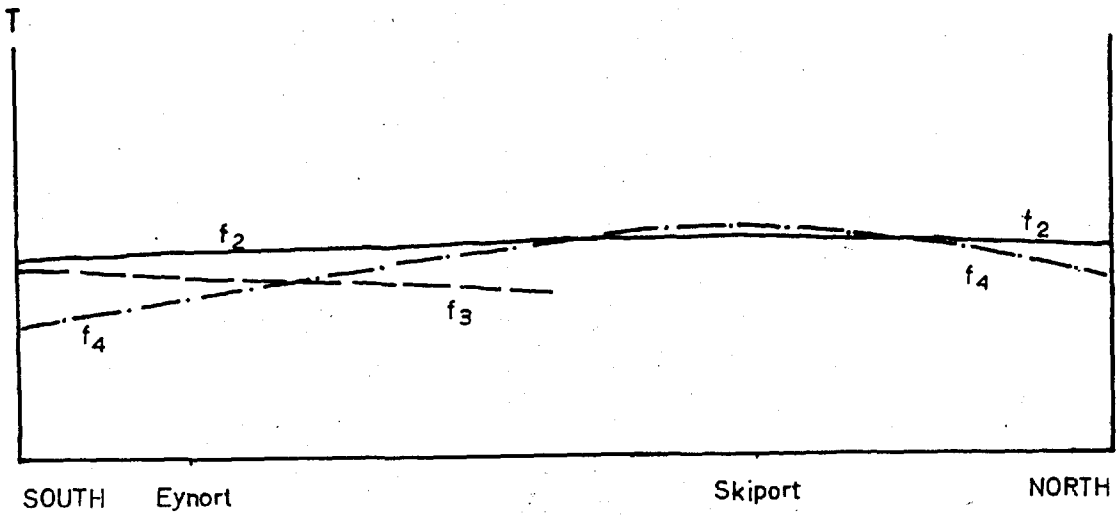
Very little is known about the metamorphism associated with the f_1 deformation phase, except that f_1 fabrics usually consist of biotite and hornblende. One knows much more about the metamorphism associated with the f_2 phase. It seems to have been in the high amphibolite facies (or hornblende-granulite facies, if one wishes to use that term) and pyroxene was locally a stable mineral. This early metamorphism also seems to have been of a high pressure form, as garnet is stable prior to the f_3 phase. It is shown (Fig. 3.3) to be fairly uniform throughout South Uist but for reasons which will become apparent in a later section I have shown it to diminish to the east, beyond the thrust zone.

Fig. 3.3

Variations in relative metamorphic temperatures for the different metamorphic phases from west to east and south to north, across South Uist.

FIG. 3-3


T
PRE-LAX.
f2
f3
f4
f4
f2
f3
WEST
Druidibeg
Skiport
Thrust
EAST


T
f2
f2
f4
f3
f4
SOUTH
Eynort
Skiport
NORTH

The f_3 metamorphism was generally confined to the higher deformation zones, which were themselves partly delimited by the earlier metamorphic state of the rock (see section 6.2). The metamorphism was generally in the amphibolite facies, and both pyroxene and garnet appear to have been unstable. Around East Geirinish, even in the tighter fold zones, there appears to have been relatively little f_3 metamorphism and so on Fig 3.3 I have shown the amount of f_3 metamorphism as decreasing to the east.

The f_4 metamorphism was extremely variable. On the west and east coasts the amount of recrystallisation was low, but, as I have shown, it increases in intensity inland to reach a peak in the Skiport - Hecla region. The northern part of this region also shows the most retrogressive metamorphism, containing minerals such as epidote, chlorite and sericite, and is possibly the area which was slowest to cool.

To summarise the above, one therefore has a pattern of relatively high pressure, uniform metamorphism accompanying the early deformation phases, followed by a decrease in pressure, but possibly a localised increase in temperature culminating in the f_4 deformation. All this appears to have been superimposed on a still earlier metamorphic gradient, which increased from west to east. The influence of these metamorphic events on the f_2, the f_3 and the f_4 deformation history will largely be the topic of the following chapters.

CHAPTER IV

METHODS OF STRAIN ASSESSMENT

Basement rocks are notorious for their deficiency in material from which one can measure quantitatively the amount of strain; they lack the coids, the pisolites and the conglomerates which enable one to determine the strain in deformed cover rocks. To a certain extent, however, qualitative results may be obtained from fabric analysis or from analysis of the changes in the discordant relationships between dykes and gneiss, and semi-quantitative results may be obtained from measurements of the amount of shortening along the folded layers, or from a study of the orientations of folded or boudinaged veins. These methods have been tried in the Uists and in this chapter I would like to comment on some of their applications or on some of their failings.

4.1 Defining directly the shape of the strain ellipsoid using 'retrogressed' garnets.

During the period of metamorphism associated with the f_3 and f_4 fold phases, garnet, in basic rocks, tended to alter to an aggregate of hornblende and feldspar, and whereas the garnet crystals tended to act as competent solid bodies the aggregates had a lower competence and became deformed, taking on an ellipsoid shape. The shape of this mass would, in general, appear to be related to the strain increments which it suffered following its retrogression from garnet. If deformation accompanied retrogression, however, then many of the bodies will have had competent solid cores of garnet with outer rims of less competent hornblende/feldspar.

This phenomenon of variation in competence within a single garnet/

aggregate mass may account for some of the anomalous results given by this method, such as those obtained from a shear zone in a basic mass from the Parc district of Lewis. The zone was several metres wide, but wholly contained within a large garnetiferous amphibolite. Within the shear zone the garnets were retrogressed and each hornblende/feldspar mass took on an ellipsoidal shape, the long axis of which curved into the zone of maximum shear as in a typical case of simple shear (Ramsay and Graham, 1970). In the zone of maximum shear the retrogressed masses broke up into long feldspathic streaks, the dimensions of which were immeasurable, though the strain ellipsoid could be measured from garnets in the less deformed zones. The most interesting point about these garnets is that they did not give a plane strain ellipsoid as one could expect in a case of simple shear, but gave a constrictional strain ellipsoid of $k = \infty$. B.K. Tan (personal communication) has shown that a constrictional strain ellipsoid is to be expected if the originally spherical mass had varying competence (increasing inwards) during deformation. The strain path formed by progressive deformation of such objects would be highly variable, especially if the competence of the mass was changing actually during deformation.

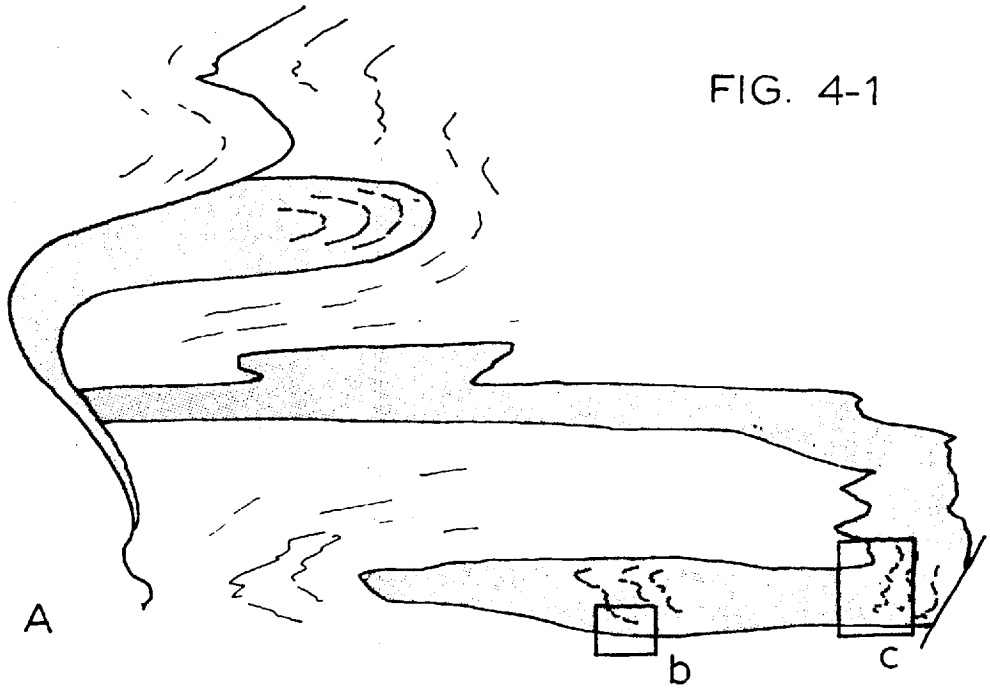
It is a common feature of these retrogressed garnets in the Hebrides that garnet cores can still be observed in the centres of the deformed aggregates and as the amount of garnet often tends to decrease with increasing intensity of deformation, it is likely that the retrogression actually accompanied the strain. It is clear therefore that in most cases the retrogressed garnets cannot be used for quantitative measurements.

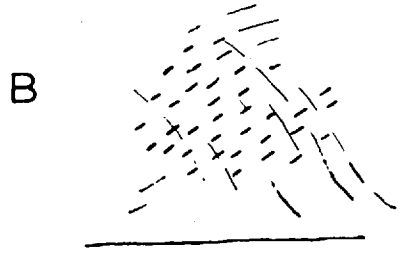
The shape of these altered garnets can be used, however, to show

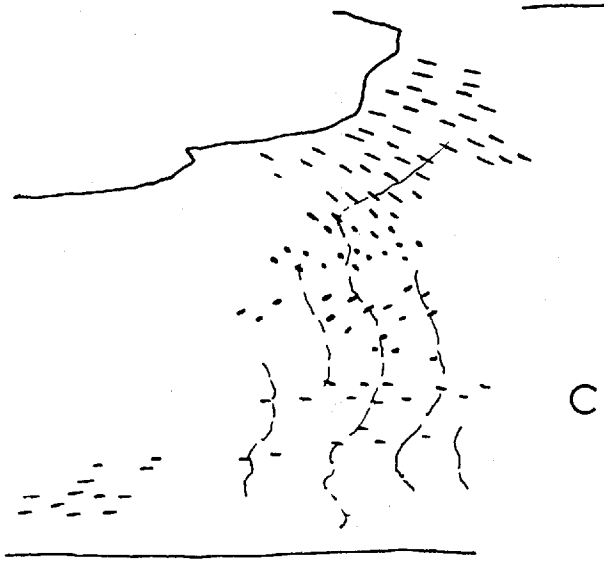
Fig. 4.1

A folded Scourie-dyke (shaded) from East Geirinish. The heavy lines in b and c show the orientation and approximate length of the "retrogressed garnet" masses from two parts of the fold.

FIG. 4-1







up the orientation of the strain axes and to show qualitatively a certain amount of the heterogeneity of the deformation within many amphibolites. An example of this is given in Fig. 4.1, where the altered garnets indicate some of the complexities of deformation both at the hinge and on the limbs of an f_3 fold.

4.2 Defining the surface of no finite longitudinal strain within the strain ellipsoid

This method was first suggested by Flinn (1962) and was then developed by Talbot (unpublished PhD thesis, 1966). It involves the stereographic projection of the orientations of both folded and boudinaged veins, and should enable one to find the position, shape and orientation of the surface which separates the extensional field of the strain ellipsoid from the compressional field. The shape of this surface defines exactly the principal strain ratios, and from the position and orientation, orientations of the principal axes of strain can be deduced.

Because of the unpublished nature of much of Talbot's work, I do not wish to add much comment on this method except to say that it has not been practicable to use it in South Uist for a number of reasons -

a) there is too much small scale folding causing local inhomogeneities in the amount of strain.

b) often the veins did not have a random orientation prior to the deformation and, with flattening strains, a plot gives only a π pole great circle distribution. The distance across the surface of no finite longitudinal strain can be accurately measured only along this one great circle and this will by no means define the full shape of the surface of the form of the strain ellipsoid.

c) The shape of the no finite longitudianl strain surface has to be measured with a high degree of accuracy. A small error of about 1° in measuring across the cone may give highly erroneous strain values. It id generally accepted that foliation readings can only be taken to an accuracy of $\pm 2^{o}$ to 5°. For accurate strain measurements a very large number of originally randomly orientated veins are required and unfortunately in South Uist in areas of apparently relatively homogeneous deformation the number of originally randomly orientated veins is usually too small.

d) If I may permit myself a general criticism of the method as a whole, the attempts that I have made to define the strain ellipsoid using deformed veins have shown up some of the shortcomings of the technique. At Ardivachar, following the early deformation phases, garnet/feldspar veins, with apparently random orientation, segregated out within the Scourie-dyke amphibolites. These veins were then either folded or boudinaged by the f_3 deformation. The surface of no finite longitudinal strain could not be defined from these veins, as the compressive strain or extension close to this surface was not enough to cause folding or boudinage respectively; and if the surface of no finite longitudinal strain is defined by the limit of folded veins as suggested by Talbot, then impossible strain values result (negative values of b and very low values of a were obtained). The surface defining the limit of folded veins is largely controlled by factors such as the viscosity contrast as well as the strain ratio, as will be dealt with in the next section.

4.3 Using folded or boudinaged layers to calculate the amount of strain

4.3A Using folded layers

4.3A1 Fold classification

In this thesis, fold geometry has been analysed in two ways: by measuring the thickness variations round folds and by comparing the curvatures of the layer-bounding surfaces. In considering thickness variations, the orthogonal thickness paramener t' has been used, where $t' = t_a/t_o$, t_a being the orthogonal thickness at dip $\alpha°$ and t_o being the thickness at the hinge, and these values have been plotted on the t' against α graph as given by Ramsay (1967, p.413). For the curvature comparisons, the dip isogon parameter ϕ_α has been used, where ϕ_α is the angle between the dip isogon (for "dip isogon" see Ramsay, 1967, pp. 363 - 372, and Elliot, 1965) and the tangent to the layer at dip $\alpha°$. In this case the values of ϕ_α are plotted against α, and each fold should give a characteristic curve on this graph (Huddleston, 1969).

Both methods have their advantages and disadvantages. The t' parameter was found to be the most useful when analysing individual layers, when the hinge point can easily be seen. This parameter is not as sensitive to minor irregularities in the original thickness as the ϕ_α parameter, and is also the most useful for isoclinal or very tight folds. Generally, however, it is far more convenient to use the Huddleston ϕ_α parameter, as this is less time consuming and no calculations are required, but irregularities in original thickness or in the amount of superimposed strain give highly irregular and sometimes incomprehensible results. Where the hinge cannot be easily defined, the Huddleston parameter is far more accurate.

From such fold analyses the folds can be classified into the

following types (from Ramsay, pp. 363 - 372):

Class 1. The curvature on the inner fold always exceeds that on the outer arc, i.e. the folds have convergent isogons and have an orthogonal thickness ratio of greater than cos α. This class can be sub-divided into:

Class 1a. Folds with strongly converging isogons.

Class 1b. "Parallel folds", where the orthogonal thickness is constant

Class 1c. Folds with weakly converging isogons.

Class 2. The curvature on the inner and outer arcs is equal, i.e. the isogons are parallel. These have often been called "similar folds".

Class 3. The curvature on the inner fold arc is always less than that on the outer arc, i.e. the folds have divergent isogons.

4.3A2 Strain measurement from folded single layers

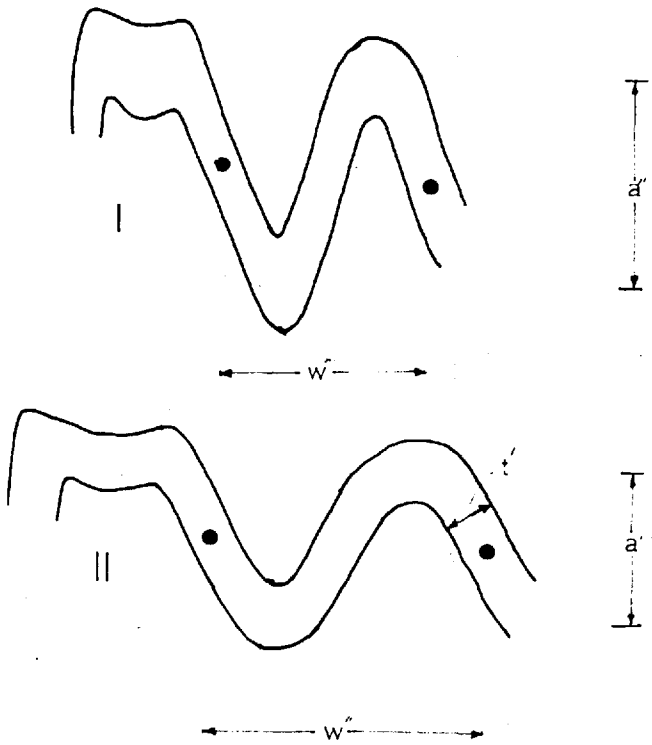
It has been shown by Ramberg (1959,1960,1963) and by Biot (1957, 1961) that when a layer of high viscosity embedded in a media of low viscosity is compressed, an instability is set up, producing folds of various wavelengths. There is however always a dominant fold wavelength which is generated more rapidly than the others and this dominant wavelength is characteristic for layers of certain viscosity contrasts and thicknesses. For any folded layers, therefore, if one can identify this dominant wavelength then one can estimate the viscosity ratio between the layer and the matrix.

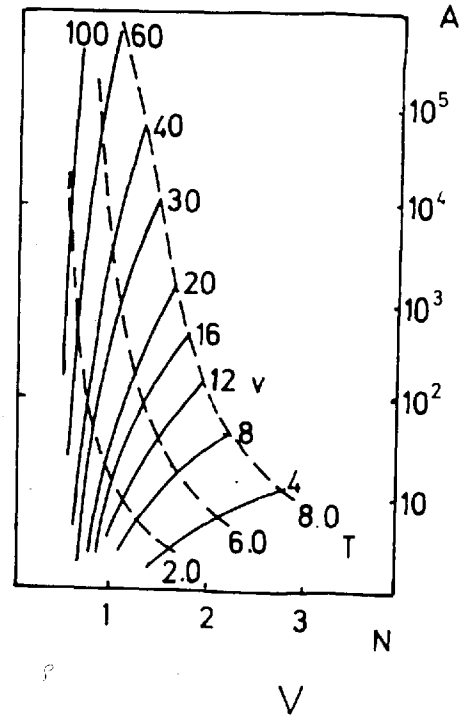
There is one problem, however, that one is not dealing with simple buckling but with a layer shortening in which buckling may play

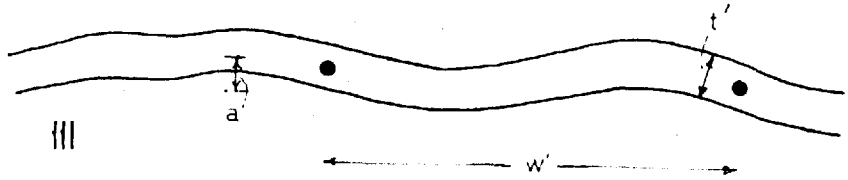
Fig. 4.2

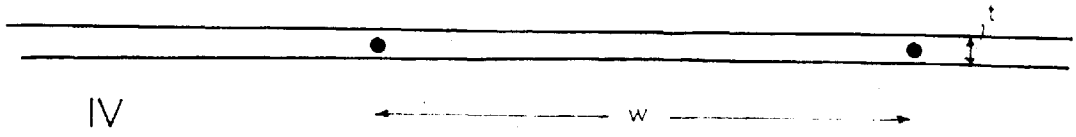
The method for determining the total amount of strain along a folded layer (see text). Fig. 4.2V gives the amplification of the dominant wavelength as a function of the non-dimensional dominant wavenumber ($= \frac{2\pi t'}{w}$) for various values of viscosity ratio (v) and of uniform shortening (T), after Sherwin and Chapple (1968).

FIG. 4-2.


I
a''
w''
II
t'
a'
w''


A
100
160
40
30
20
16
12
v
8
4
8.0
6.0
T
2.0
10^5
10^4
10^3
10^2
10
1
2
3
N
V


t'
a'
III
w'


t
IV
w

only a minor part. It has been shown by Biot and by Sherwin and Chapple, that layers with a low viscosity contrast will suffer a high amount of strain, up to 40% shortening, prior to buckle amplification, and thus any dominant wavelength/thickness relationship will change. A solution to this problem has been give by Sherwin and Chapple (1968) and later modified by Huddleston (1969). For the detailed theory and mathematics I must refer the reader to the above accounts and I shall here only briefly describe the method using as an example a fold from north-west Benbecula, shown in Fig.3.2i.

In this diagram, "I" shows the fold drawn in a profile section, from a photograph looking down the fold plunge. From isogon measurements or from t'/a measurements one can determine the fold geometry (see previous section), which in this case is of the class 1c of the particular type which would form if a flattening strain of 28% (i.e. $1 + e_1 / 1 + e_2 = 2 : 1$) were added to a pure parallel buckle fold. This flattening strain can be removed graphically to give the buckle fold (class 1b) as shown in Fig. 3.2ii. This fold has still a low wavelength/amplitude relationship, caused by buckle, so the effects of buckle can be removed to give the fold as in Fig. 3.2iii. Huddleston (1969) has shown that prior to "buckle" the amplitude of the fold rises only slightly, and much of the strain is taken up by homogeneous layer shortening. After a limb dip of about 10° to 50° is reached (an amplification of between 10 and 50). then most of the strain is taken up by a straightforward buckle without at first very much homogeneous strain. If homogeneous strain does take place at the same time as buckle, then it can be identified from isogon measurement etc. by the fact that the limbs would show higher amounts of flattening strain than the hinge zone. To return to the above method,

one can therefore graphically unfold the buckle fold of Fig. 3.2ii to a fold with limb dips of about 10° to 15° as in Fig. 3.2iii, and it is the amount of shortening represented by this operation which is considered as being due to buckle alone. If one then estimates the wavelength:thickness relationship of the layer, then both the viscosity ratio between the cempetent/incompetent media, and the original layer shortening prior to buckle, can both be read off the graph provided by Sherwin and Chapple (Fig. 3.2v). In this example, the strain due to simple buckle is of the order of 2.8 : 1, the post-buckle flattening strain in the order of 2 : 1 and the strain due to the original layer shortening is 1.5 : 1. The total strain as given by this folded layer is therefore of $1 + e_1 / 1 + e_2 = 8.4 : 1$.

Using this method, very high strain values have been obtained from some folded layers. Folded dykes at north-west Benbecula give ratios of 35 : 1, made up of a shortening due to buckle of 4 : 1, a shortening due to flattening of 3 : 1 and an original layer shortening of 3 : 1. With original layer shortenings as large as this, they cannot be ignored; they make a difference of up to and over three times the total amount of strain.

This leads us on to the accuracy of these results. Possible errors include graphical and mechanical errors: -

a) The fold measurement not exactly in the profile plane.

b) Errors in constructing isogons, or in t'/a measurements.

c) Graphical errors in unflattening or unbuckling the folds.

Some of the above can be reduced by care or by selecting the best examples, but isogon measurements and t'/a measurements are both very innccurate for high values of strain, over 5 : 1.

Errors may also occur in estimating the wavelength/thickness ratio. The wavelength should be measured from the inflection points on the limbs, but as shown in Fig. 3.2ii the inflection point is often difficult to pin-point. There is the possibility that the wavelength is not characteristic for that particular viscosity and thickness. Several wavelength/thickness ratios are required to find the mean. There is also the possibility that the layer may have had an undulatory form originally and hence not only would folds with a non-characteristic wavelength amplify but also other folds could develop not only by buckle but by homogeneous flattening. In this situation the above theory and method would be invalid. Though this is unlikely to be the case in folded sedimentary layers, the chances are that it is much more common in gneissose regions where one is dealing with folded dykes and quartz veins and where some of the layers may have been previously folded.

One can conclude therefore that this method cannot be particularly accurate when one is dealing with high strains where, through multiplication, any inaccuracies become very much enlarged. It also confines strain measurement to two dimensions, normal to the fold hinge. For certain of the early fold phases in South Uist, fold hinges are believed to be parallel to the principal strain axes in three dimensions, but this is not the general case, as will be shown later. One does not know, therefore, which strain ratios one is dealing with. Since many of the earlier folds in South Uist maintain a constant plunge, this method does show the variation in two "fixed" dimensions, and as such it has been found very useful. It also appears to be the only method of field quantitative strain measurement applicable to most basement terrains.

4.3A3 Strain measurement from multilayer folds

It has been shown by Ramberg (1961, 1963, 1964) that in a rock consisting of layers of alternately viscous and less viscous material, each layer will tend to produce folds with its own domaiant wavelength, but because of the proximity of the other layers there will be interference and the wavelengths will be modified. Biot (1964a) has shown that the dominant wavelength is larger than that of a single layer by a factor of $n^{\frac{1}{3}}$ (n = number of layers with equal thickness and equal viscosity). Biot (1964b) has also examined the case where the layers are enclosed by two straight rigid boundaries, and found that here the wavelength was controlled more by the thickness and the spacing of the layers and less by the viscosities. Ramberg found that the force necessary to produce buckle increased not only with the lower viscosity contrasts but also with a lower spacing between the layers. One can therefore conclude that a great many features combine to determine the wavelength of a multi-layer system, and, as yet, no simple solution has been found for the problem.

In natural fold multilayers it is therefore very difficult to make any accurate estimate of the relative viscosities or of the amount of shortening prior to buckle folding. One can only measure the amount of flattening superimposed on the folds, and the amount of shortening due to buckle. In many examples, one is even denied the privilege of measuring the amount of shortening due to buckle, as when one graphically removes the flattening on many of the folds, one often finds that the amount of buckle was very small. Most of the folding would appear to have been taken up by homogeneous strain superimposed on only gentle warps. When dealing with multilayers one has therefore no idea of the total amount

of strain!.

When faced with such facts, it is very disappointing to find that the majority of the Hebridean folds are multi- layers, even many of the Scourie-dykes or the quartz veins have been folded in association with layered gneiss, and one cannot always presume that this banded gneiss played no part in the production of these folds. I have therefore had to content myself with a simple geometric analysis of the folds, as described previously, and from such an analysis I have only been able to determine the relative viscosities (not the ratios) and also the amount of flattening suffered by the folds.

As far as viscosities are concerned, these have been estimated from the geometry of the folds, from the way certain layers form class 1c folds when enclosed in one medium, but class 3 folds in another. Though at first this sounds very simple, it is often ,in practice, difficult to find much consistency and in some cases it is even difficult to decide which is the competent layer. A layer will often form a type 1c fold in, say, an anticlinal area, but a type 3 fold in the adjacent syncline. This may indicate that a whole group of layers is acting as the competent unit and the different types of folds are due to variations in the amounts of internal strain set up during buckling.

As far as the flattening is concerned, this has been used to show certain variations in amount of strain. Over 270 individual competent layers have been measured. Most of these concern the late f_4 folds and indicate a general increase in the amount of flattening from west to east. This increase corresponds with a variation in f_4 strain as estimated using other methods (see Chapter VI). The earlier folds show variable amounts of flattening. Generally, the amount of flattening suffered by

the f_2 folds is greater than that suffered by the f_3 folds, but this is to be expected of earlier structures. The values of the flattening strain have been plotted in histogram form for each fold phase and for several different areas. They will be given later, in the section on high Laxfordian deformation. A more detailed description of the more interesting folds will also be given later.

3.3B Using boudinaged layers

Alpine geologists (Goguel, 1962, Badoux, 1963) have shown how strain can be measured by determining the linear changes in the rock using boudinaged belemnites. In a similar way it could be thought possible to determine the amount of strain in gneissose terrains, using boudinaged amphibolites or quartz veins. For the following reasons, however, it has been found impracticable to use these structures in South Uist:

1) Prior to boudinage, there was presumably ductile extension along the layer. One has no idea of the amount of this.
2) There is the problem that one does not know exactly what causes boudinage, whether it is the amount of strain, the ductility of the rock or the rate of strain. In the quarry at Howmore, South Uist, for example, there is a rapid transition from basic layers which have suffered extensive boudinage to layers which are still continuous. I have not found it possible to decide whether this change is due to a change in the amount of extension or to a change in the relative viscosities of the acid and basic gneiss.
3) Most of the boudinage occured early in the fold phase sequence, and thus the boudinaged rocks have suffered a later compressive strain. The effect of this later deformation varies for different boudins. If gneiss had been sucked into the boudin necks during separation,

then the boudins cannot be reunited and often they are "rotated" relative to the gneiss (see Ramsey, 1962) and sometimes folded. At one locality at Garry-a-siar, however, the bouding have been shunted together and partially pushed over one another.

4.4 Using rock fabrics

As a result of strain the linear and planar elements of rocks will undergo some re-arrangement. Rocks which were originally isotropic will develop an anisotropy related to the form of strain which they have undergone. With strain ellipsoids of type $k = 0$, poles to planes are re-orientated and become concentrated round the Z axis of the ellipsoid, while linear elements concentrate along the X - Y plane. With prolate ellipsoids ($k = \infty$), linear elements become concentrated round the X axis while poles to planes lie on the Z - Y plane. Flinn (1962 and 1965) has suggested that ellipsoids of $k > 1$ will therefore produce a linear fabric ($l > s$), while ellipsoids of $k < 1$ will produce a planar tectonic fabric ($s > l$).

Lithological fabrics (i.e. fabrics made up from fold hinges, quartz rods, etc.) must be related to the finite strain but mineral fabrics are also related to the increments of strain under which they formed. Planar fabrics produced during deformation have usually been found (e.g. Cloos, 1947) to lie perpendicular to the Z direction of the strain ellipsoid. Linear fabrics which often accompany cleavage where the strain has a k value close to or greater than one usually lie along the maximum extension direction (X) in the X - Y plane. In simple, single phase deformation environments, therefore, the fabric is due to re-orientation of minerals with some recrystallisation and is related directly to the strain which is superimposed on the rock. In multi-phase deformation

it is more difficult to decide whether the mineral fabric represents the total finite strain or whether it is controlled by the stresses which existed while the rock was last recrystallised.

The type of mineral fabric therefore depends on how individual minerals deform in a particular circumstance (see also Flinn, 1965), whether they are "rotated" by the new strain increments or whether they are deformed by recrystallisation. It may well be that all minerals, including most biotite, hornblende and possibly even pyroxene, deformed largely by recrystallisation when under high-grade metamorphic conditions.

I have mentioned previously (section 3.3A) that certain of the fabrics shown by the dykes at Garry-a-siar are not formed by individual grains but by the orientations of groups of small grains, such as those which would form from a deformed igneous texture. As there is no evidence for large scale recrystallisation prior to the f_2 period then these groups of small grains have not resulted from the breakdown of a large recrystallised hornblende, but from other small mineral grains. If this is the case, then this fabric ought to indicate the state of finite strain in the rock. Other fabrics which should give some indication of the finite strain are those given by the drawn-out relict ophitic textures (section 3.3A) and those given by the large relict feldspar phenocrysts often found in Scourie-dykes at Ardivachar and Garry-a-siar. These relict feldspar phenocrysts now consist of a great many individual feldspar grains, which presumably recrystallised at some time during the f_2 deformation phase. The phenocryst mass as a whole ought to show the finite strain.

One must conclude that in a fabric analysis one needs to distinguish between fabrics produced by different elements. In South Uist, for instance, the gneisses will show lithological fabrics resulting from

numerous Scourian and Laxfordian deformation phases, the Scourie-dykes will show lithological fabrics resulting from a sequence of Laxfordian strains, while a petrofabric analysis of the individual mineral grains in both may only show fabrics related to the last strain increments. In South Uist, I have therefore confined myself to the large scale fabrics, those generated by dyke/foliation intersections, quartz and hornblende rods, etc., and the small scale lithological fabrics, the deformed relict ophitic texture, the deformed phenocrysts, etc. Individual mineral fabrics have not been used to demonstrate the state of strain.

In areas of buckle folding, the internal strains set up by the buckling will cause local strain heterogeneities (see section 5.6E) and thus an analysis has to be confined to a very small area. Large scale lithological fabrics are useless in this case and small scale lithological fabrics have often been almost obliterated by high deformation or by excessive recrystallisation. Over much of South Uist, therefore, a detailed lithological fabric analysis is impossible. A discussion of the strain history derived from fabric analysis is therefore confined to the areas of low deformation, especially those at Ardivachar and Garry-a-siar, and is given in the next chapter, section 5.6.

4.5 Using the deformation of discordant relationships such as those between dykes and gneissose foliation

In the past, the extent of obliteration of discordant relationships between dykes and gneiss has been used to indicate the amount of deformation in basement areas. Sutton and Watson (1951) used the decrease in angular discordance between Scourie-dykes and foliation to show an increase in Laxfordian deformation between Scourie and Laxford

Bridge. Ramsay (1963) showed that with an increase in amount of Caledonian deformation, in the area between Sleat and Glen Arnisdale, Scotland, the angular discordance between dykes and gneissose foliation decreased and was almost totally obliterated. Similarly, it was partly from the degree of discordance of Hebridean Scourie-dykes that Dearnley (1962) and Dearnley and Dunning (1968) proposed that there was a "Central Block" of relatively undeformed Scourian separating two re-worked Laxfordian areas.

In this section, I propose only to summarise some of the effects of dyke deformation; a more detailed analysis is unnecessary for this thesis and will be published elsewhere. The main points are:

(i) The amount of strain necessary to obliterate discordant dyke relationships is dependant largely upon the orientations of the dykes and foliation relative to the principal strain axes.

(ii) When a dyke is buckle folded, the dihedral angle between dyke and foliation suffers internal strains due to the buckling, i.e. it suffers flexural shear, tangential longitudinal strain or a combination of the two. Flexural shear will have similar effects on the dihedral angle on the inner and outer arcs, while tangential longitudinal strain will give very different results (see Ramsay, 1967, pp 492 - 504). It is thus possible to make a rough estimate of the amount of flexural shear relative to tangential longitudinal strain. In the Uists, I have found that it is the former which has the most effect on the edge of t the dyke (while internally tangential longitudinal strain would appear to be the dominant mechanism, see section 5.6F).

(iii) When the significance of a fold shape is in doubt, one can sometimes use the variation in dihedral angle round the fold and also

the complex form traced out by the deformed foliation to indicate whether the fold formed by buckle instead of heterogeneous simple shear (see also Ramsay, 1967, Figs. 9.18 and 9.30). From such analyses one can sometimes also rule out the possibility that the fold shape formed by homogeneous flattening, or as an irregular ptygmatic intrusion.

(iv) When a homogeneous strain is added to the buckle fold, the dihedral angle is modified by different amounts around the fold. Except for the case of a superimposed strain of type $k = \infty$ when X is parallel to the fold axis, this modification can be easily recognised in the final form of the dihedral angle round some flexural slip folds. With a strain as above, however, there is relatively little modification. Thus many of the folds at Garry-a-siar and Ardivachar can be shown to have been modified by a strain other than that above. The significance of this point will become apparent later (in section 5.6).

The amount of homogeneous strain cannot be measured without first knowing the original dihedral angle and the orientation of the axes of the flattening strain.

(v) When a lineation lies on a foliation plane which is discordant to the dyke and the dyke is then folded, the lineation will trace out a complex locus whether deformed by flexural slip or tangential longitudinal strain. The form of this complex locus will depend on the orientation of the lineation and foliation relative to the fold axis in the dyke and also to the mechanism of internal buckling strain. Many of the complex forms taken up by the folded lineation at Ardivachar and Garry-a-siar (section 5.10) probably have such an origin.

CHAPTER V

THE AREAS OF LOW DEFORMATION

5.1 General

On the north-west coast of South Uist at Ardivachar Point, and on the west coast of Benbecula between Garry-a-siar and Balivanich, areas of low Laxfordian deformation are exposed in the coastal strips. Though these areas only show clear rock in a narrow inter-tidal zone 40 - 50 m. wide the outcrop is continuous, an important factor when dealing with areas of low deformation.

Mapping was carried out on a very large scale. The general data were plotted on to enlarged aerial photographs on a scale of 1:4220 (15" to 1 mile). But as this failed to show all the significant variations, much of the data had to be plotted on maps roughly constructed (by pace and compass and field sketches) in the field. Maps 7 and 8 are the result. There are bound to be localised inaccuracies in scale and positioning on these maps.

The aims of this detailed survey were:

(1) To see whether there was any similarity in structural sequence between the areas of high deformation and low deformation, and to see which Laxfordian phases were poorly developed or absent.

(2) To gain an insight into the early Laxfordian and Scourian structural history, which is often marked by later deformation elsewhere.

(3) To see if there was any ready significance in, or explanation for, the local or regional variations in the amount of Laxfordian strain.

5.2 Previous work

To a certain extent this study was a follow up to work done by Dearnley and Dunning (Q.J.G.S. 1968). Dearnley and Dunning had noticed that within the general zones of low deformation present on these west coasts, there were also local variations in the amount of deformation. They mapped five Scourie-dykes in detail to illustrate the increase in complexity of the Laxfordian structures. The dyke which showed least deformation (their dyke "A" at Garry-a-siar) showed very clear discordant relationships with the earlier gneissosity; groups of thin branching apophyses appeared to be undeformed. In areas of Laxfordian deformation the foliation was more nearly discordant with the dykes and both the dykes and apophyses showed effects of folding. Locally at Garry-a-siar Dearnley and Dunning noted that variations in amounts of Laxfordian deformation could take place over a very short distance, within several metres.

5.3 General structure of the areas of low deformation

The main points of field evidence which led me to my structural interpretation are given below:

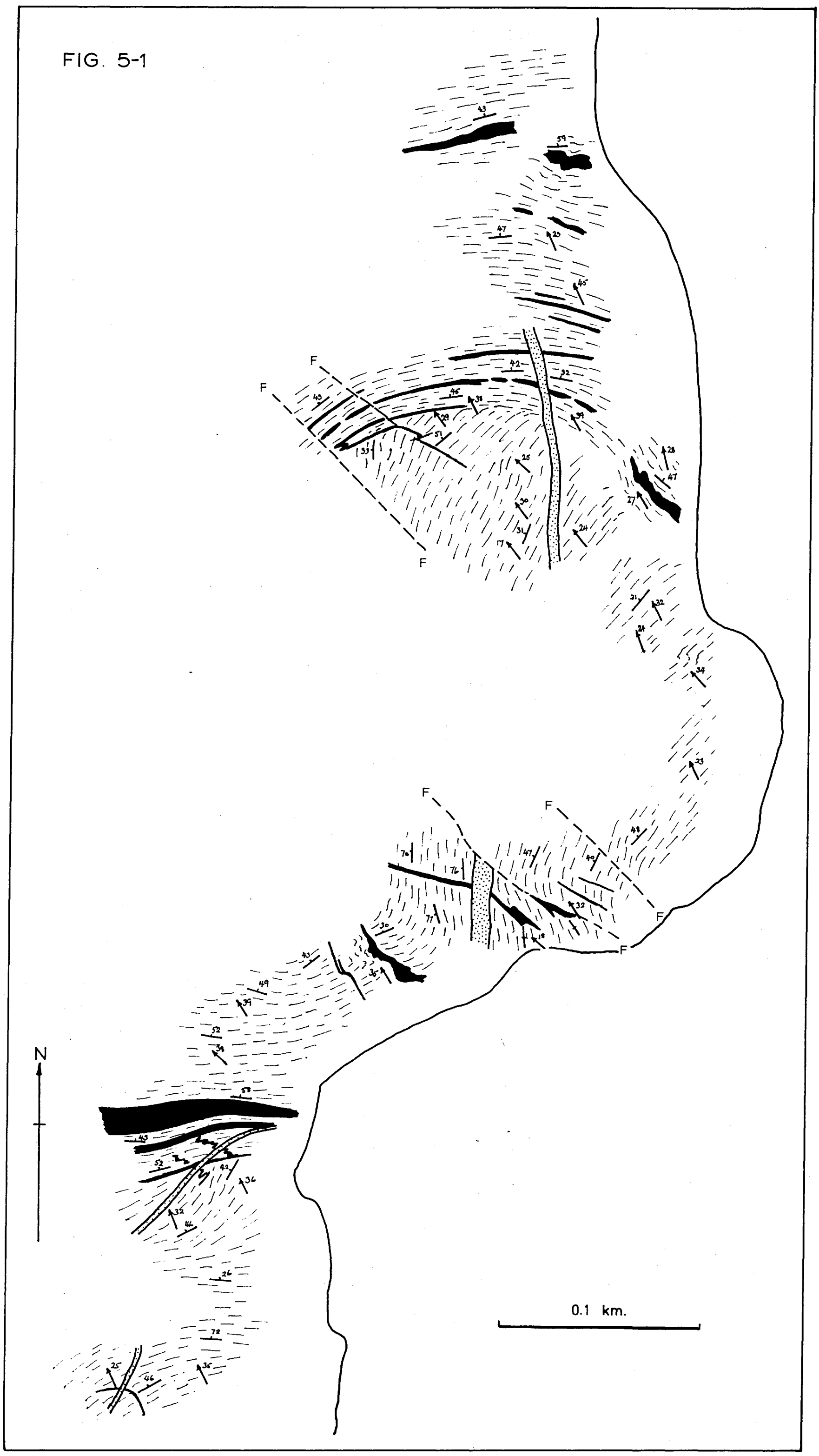
1) The dykes cut an already existing gneissose foliation, mineral banding and complex fold sequence, details of which will be mentioned elsewhere (section 7.4)

2) There is a strong linear fabric which over much of the area is relatively constant in plunge value and orientation (Maps 7 and 8). This linear fabric is, to me, one of the most striking features of the whole area of low deformation, and a feature characteristic of all the low deformation regions on the Uists (apart from Barra). In many

Fig. 5.1

Detailed map of the area between Eilean Crossan and Geo Ban, north of Culla Bay. Late Laxfordian granites and pegmatites are shown stippled, Scourie-dykes are black. Note the sudden transition from the area where the dykes are undeformed, in the centre of the map, to the area with highly deformed dykes in the north. There is also a zone of high f_2 deformation around the thick dykes in the south-east.

FIG. 5-1



parts there are large areas almost devoid of any planar fabric, and it is often impossible to measure a "foliation" orientation. The fabric may be defined by fold hinges, elongate pods or rods of quartzo-feldspathic gneiss or amphibolite, or by the orientation of individual mineral grains. Except in areas of obvious late re-folding this lineation has a constant direction.

3) A fabric of some kind can be found in all the Scourie-dykes, and dykes with a mostly mimetic igneous fabric show some evidence of a tectonite fabric at the edges. The strength of the fabric appears to be related to:

a) the amount of deformation and the structural position of the dyke.

b) the thickness of the dyke. Large dykes often have cores of a recrystallised igneous fabric.

c) the mineral composition of the dyke; dykes with feldspar phenocrysts or dykes which are rich in hornblende or biotite show a better fabric than those consisting mainly of pyroxene and garnet (see Chapter III).

The dyke fabrics vary from perfectly linear to weak planar (from $L \gg S$ to $S \gg L$ using Flinn's 1965 terminology). Perfectly planar fabrics rarely occur in these areas of low deformation.

4) The linear fabric within the dykes is parallel to that of the country rocks and also closely parallel to that of the π -pole girdles of dyke orientations and foliation orientations (Maps 7 and 8). Much of the linear fabric outside the dykes is demonstrated by Scourian structures, while that inside the dykes must be Laxfordian. One has therefore to explain why lineations which formed at several different periods should be co-axial.

5) Open folds affect many of the dykes, the axial planes striking between north-east - south-west and north - south and dipping to the west. These folds are co-axial with the linear fabric.

6) The above mentioned folds locally may fold a set of earlier structures. At G7 (Map 8) isoclinal folds affecting dyke apophyses are refolded. At G4 an earlier set of boudins is folded. These early structures which have a north-west - south-east orientation have been termed f_1, the north-east - south-west trending folds f_2.

7) At both Ardivachar and Garry-a-siar these structures (f_1 and f_2) have been re-orientated round large open antiformal structures. These large open structures have been termed f_3 and they appear to be co-axial with the earlier folds.

8) The linear fabric has been locally folded by a late set of structures which strike approximately east - west and dip to the south. These structures, though they are only locally developed, vary from open warps to almost isoclinal folds. They have been termed f_4.

9) These late folds are followed only by a group of pegmatites and by faulting.

10) The deformation is extremely heterogeneous. Dykes which exhibit isoclinal folds sit next to dykes which have suffered no folding. In many localities the f_4 structures are the first to fold the dykes, though there is always evidence of earlier strains in the form of dyke fabrics.

The sequence of fold phases mentioned above is very similar to that of the more highly deformed regions, and, probably, direct correlations can be made. All the deformation phases as recognised elsewhere in South

Uist are recorded in these low deformation areas.

5.4 The f_1 phase of deformation

5.4a General

There are only three areas where the f_1 deformation phase can be unequivocably established. At G7 (Map 8) dykes and apophyses are folded by a pre-f_2 phase of deformation. At G4 there is a hint of a pre-existent fabric folded by f_2, and the dykes have been boudinaged prior to the f_2 phase. Similarly at G8 dykes appear to have suffered pinch and swell structures and obtained a planar fabric before being affected by the f_2 phase.

There are however other possibly areas of f_1 deformation. In the low deformation zone north of Culla Bay (Fig. 5.1), the dykes often bear a planar fabric parallel to their strike, which has a north-north-west orientation. This foliation is at a high angle to the intense deformation zones which cut across the area with an east - west to north-east - south-west orientation, and may therefore be earlier, of f_1 age.

At G2 the foliation appears to have been brought into concordance with the dyke prior to the f_2 folding (Map 8), and at Ardivachar A8 there is a well-developed planar fabric within the dykes and an interbanding between dykes and foliation in a zone with an orientation similar to that of other f_1 structures (Map 7).

The f_1 deformation zones and f_1 fabric strike north-west - south-east and have a steep to vertical dip. The form of the structures depended on the dyke orientation relative to the f_1 strain axes. The f_1 fabric is usually almost parallel to the dykes, and pinch and swell or boudinage is more frequent than folding, indicating that in the zones of f_1 deformation

Fig. 5.2

A group of branching dykes deformed by the f_1 deformation phase, G8, Garry-a-siar. Within the f_1 deformation zone the rocks have become banded and all trace of any discordant nature has been obliterated.

Within this and the following diagrams in this chapter, Scourie dykes are shown as clear areas, pegmatites are stippled.

FIG. 5-2.

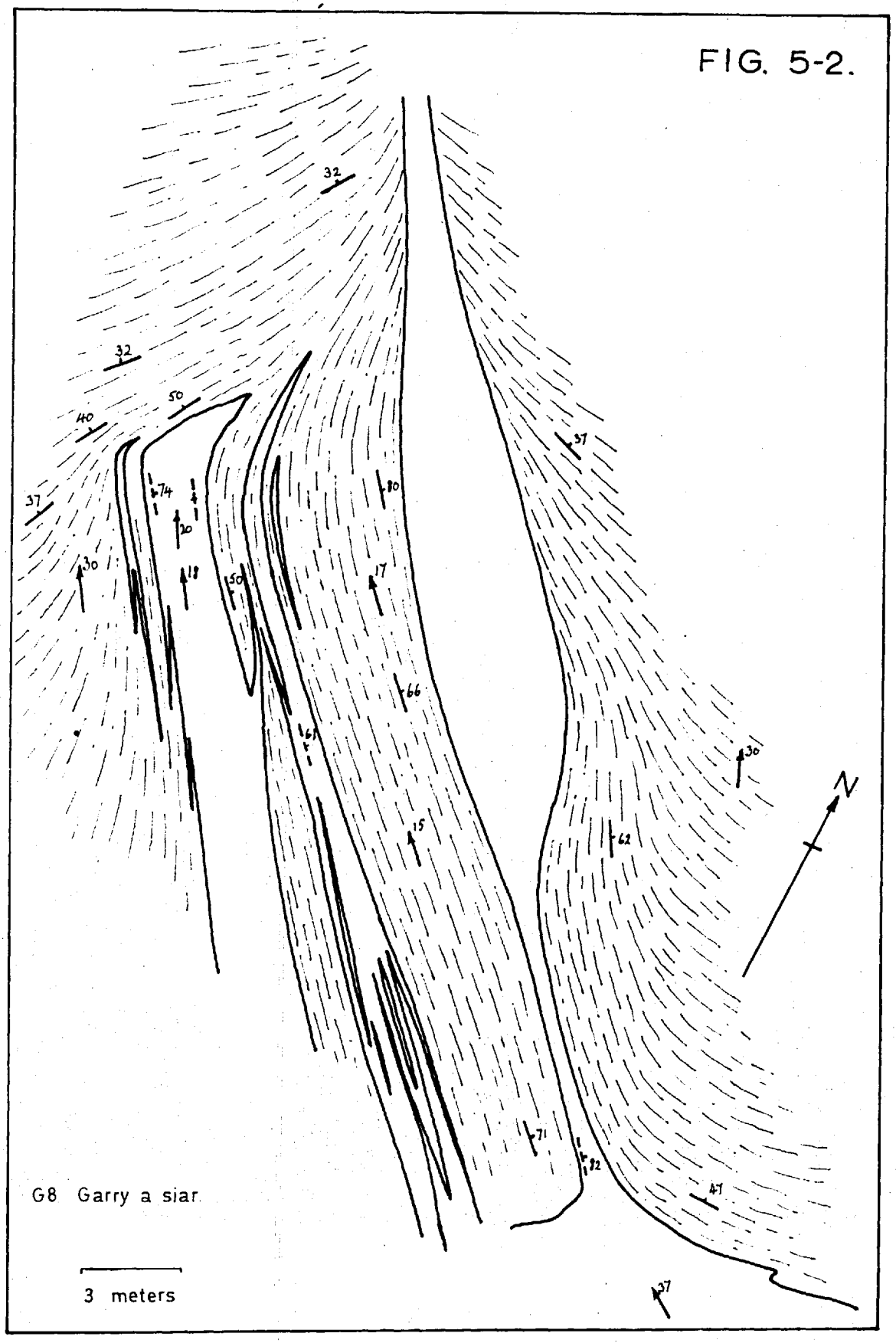
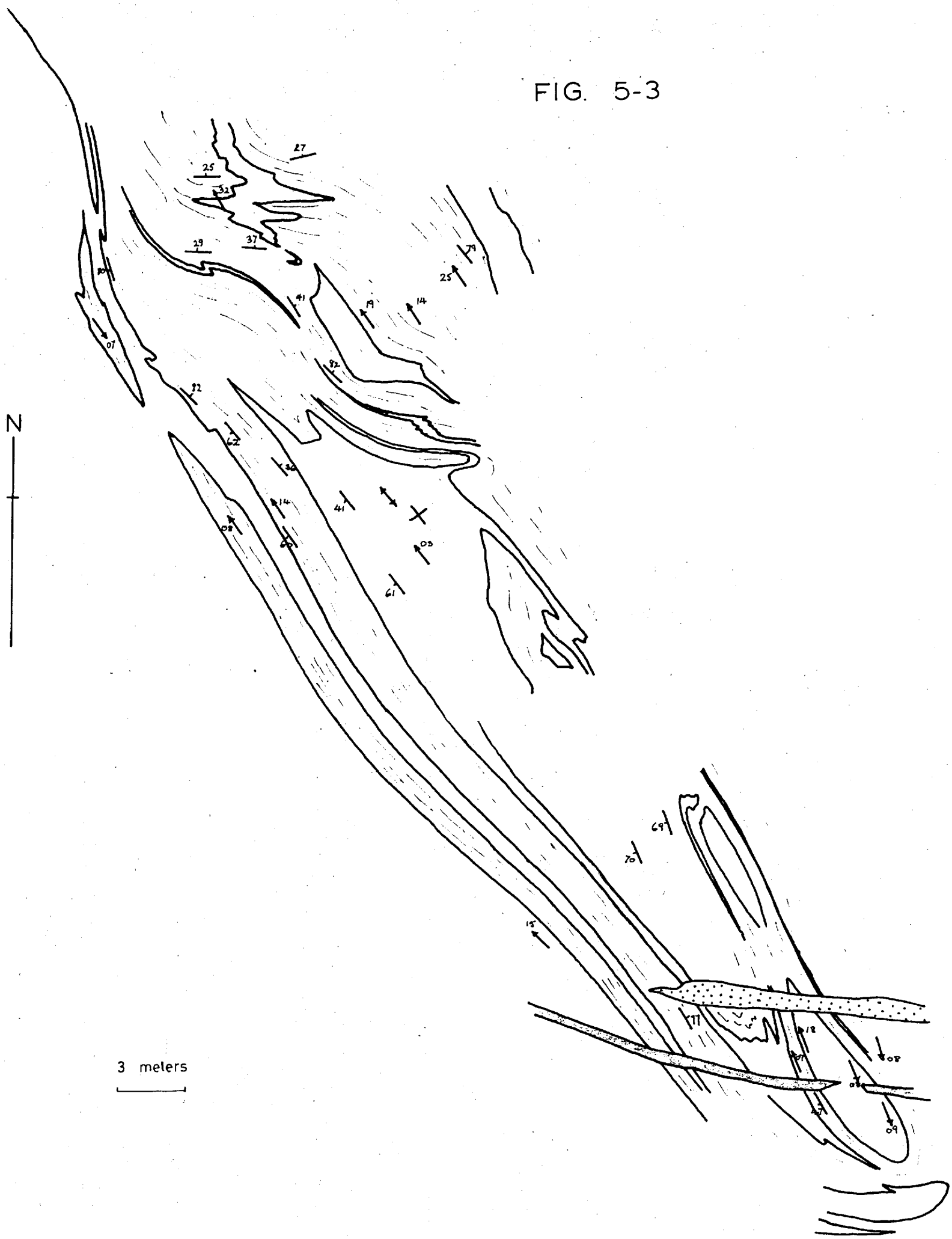

32
32
50
40
37
74
20
30
18
50
80
17
66
61
15
37
30
62
71
82
47
37
N
G8 Garry a siar
3 meters

FIG. 5-3



Deformed dykes at G7 - Garry-a-siar

the f_1 X - Y plane and the dykes had a similar orientation. However the gneissose foliation now has a north-east - south-west strike, normal to f_1 X - Y plane and presumably lay within the contractional field of the f_1 strain ellipsoid; yet there are no effects of f_1 folding in the gneisses.

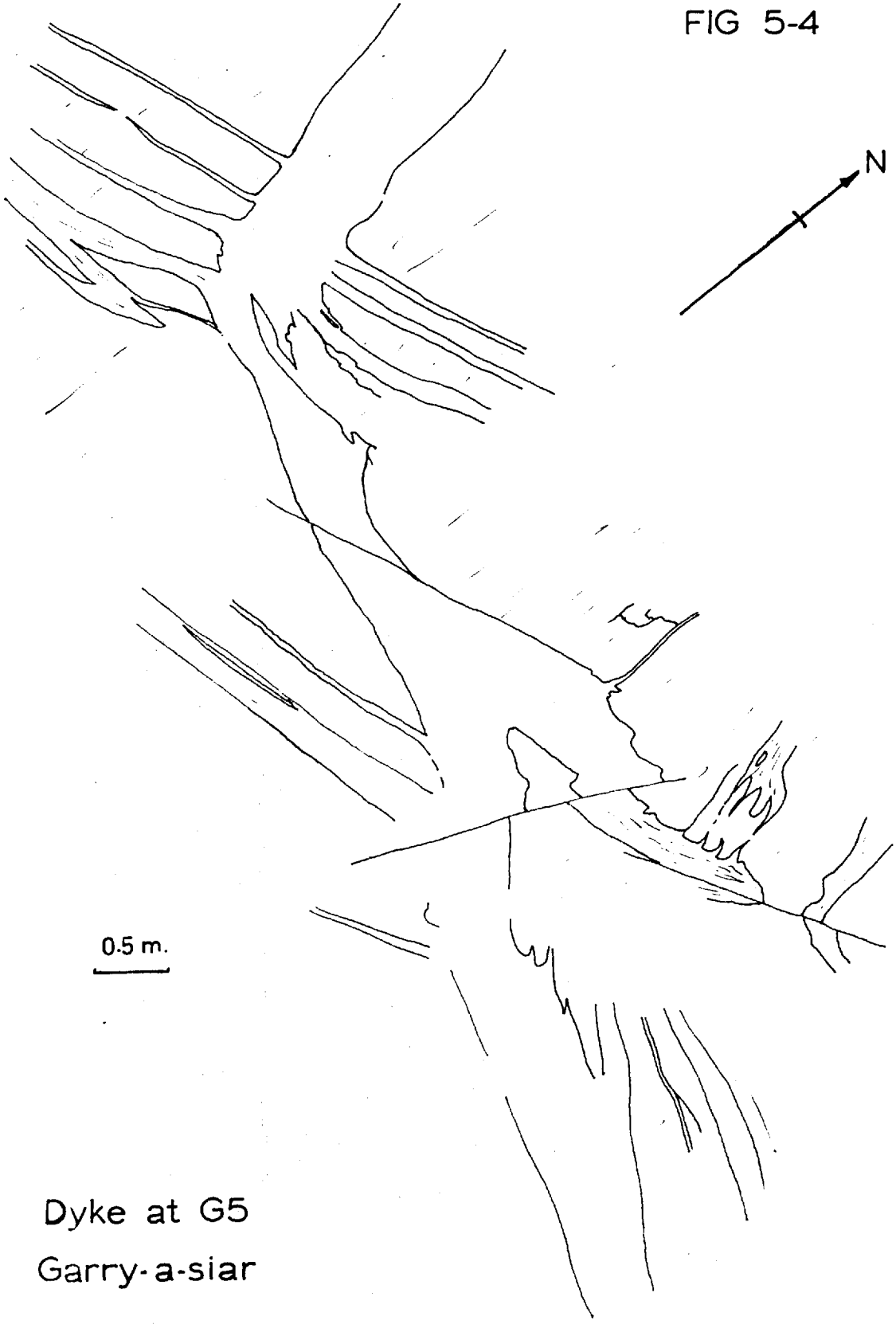
5.4b The f_1 deformation zones

The f_1 deformation will be examined in three areas, at G4, G7 and G8, Garry-a-siar, as these show certain characteristics of the f_1 deformation and of heterogeneous deformation zones in general.

At G8 (Fig. 5.2) the deformation zone is narrow and well-defined and is concentrated largely within a group of Scourie- dykes. The dykes and gneisses are brought into concordance and both have a north-west - south-east strike. From the way the foliation curves into concordance with the dykes on both sides of the deformation zone, one can infer that the deformation was due to heterogeneous flattening rather than to heterogeneous simple shear, and within the deformation zone a strain ratio for $\frac{1 + e_1}{1 + e_2}$ of at least 8 : 1 is indicated (from the degree of obliteration of the discordant structures). The amount of deformation decreases to the north-west. where in the acid gneiss there is no obvious sign of f_1 strain.

This heterogeneous character of the f_1 strain could be interpreted as being due to a higher strain rate due to a lower competency within the basic rocks. However, the core of the large basic mass in the south-east consists of a relatively undeformed rock with mimetic original fabric. Another of the basic dykes shows what appears to be a pinch and swell structure which could not have formed tectonically if the dyke had a lower competence than the acid gneiss.

FIG 5-4



Dyke at G5

Garry-a-siar

At G7 (Fig. 5.3) the intense f_1 deformation occurs at the edge of a thick Scourie-dyke and it is probably apophyses of this basic dyke which show the f_1 folds. These folds are isoclinal and show variable amounts of superimposed flattening. The dykes always show folds of class 3, but the adjacent gneiss bears folds of class 1c with a strain ratio due to flattening alone of 2 : 1 in the north and 3.3 : 1 in the south. Total strain values, taking into account the shortening due to buckle folding and the variable limb length prior to buckle, are of the order of 7 : 1 to 10.5 : 1.

As with the zone at G8, this deformation zone shows an apparent variation in strain not only across the strike but also in a north-west - south-east direction along the strike. The rate of change in amount of strain appears most rapid across the strike, and the whole zone is less than 20 metres wide. Only 30 metres to the north of G7, at G5, the dyke with its array of branching apophyses shows no evidence of any folding (Fig. 5.4).

At G4 (Fig. 5.5) the dykes had suffered boudinage prior to the f_2 folding and some isoclinal folds which affect apophyses of the north-east dyke may be of this age. From the shape of the cusps in these folded dykes, it would appear that the acid gneiss had a viscosity similar to, or higher than the amphibolite dykes. These thin apophyses, however, lie in relatively homogenous gneiss in which it is difficult to find any other obvious layer which because of its higher competence would initiate a buckle fold. The fact that the main dyke had boudinaged and contains patches with relict igneous fabric would indicate that the basic rock had a higher competence than the acid gneiss. One is therefore faced with the possibility of varying competence differences within the same rocks with

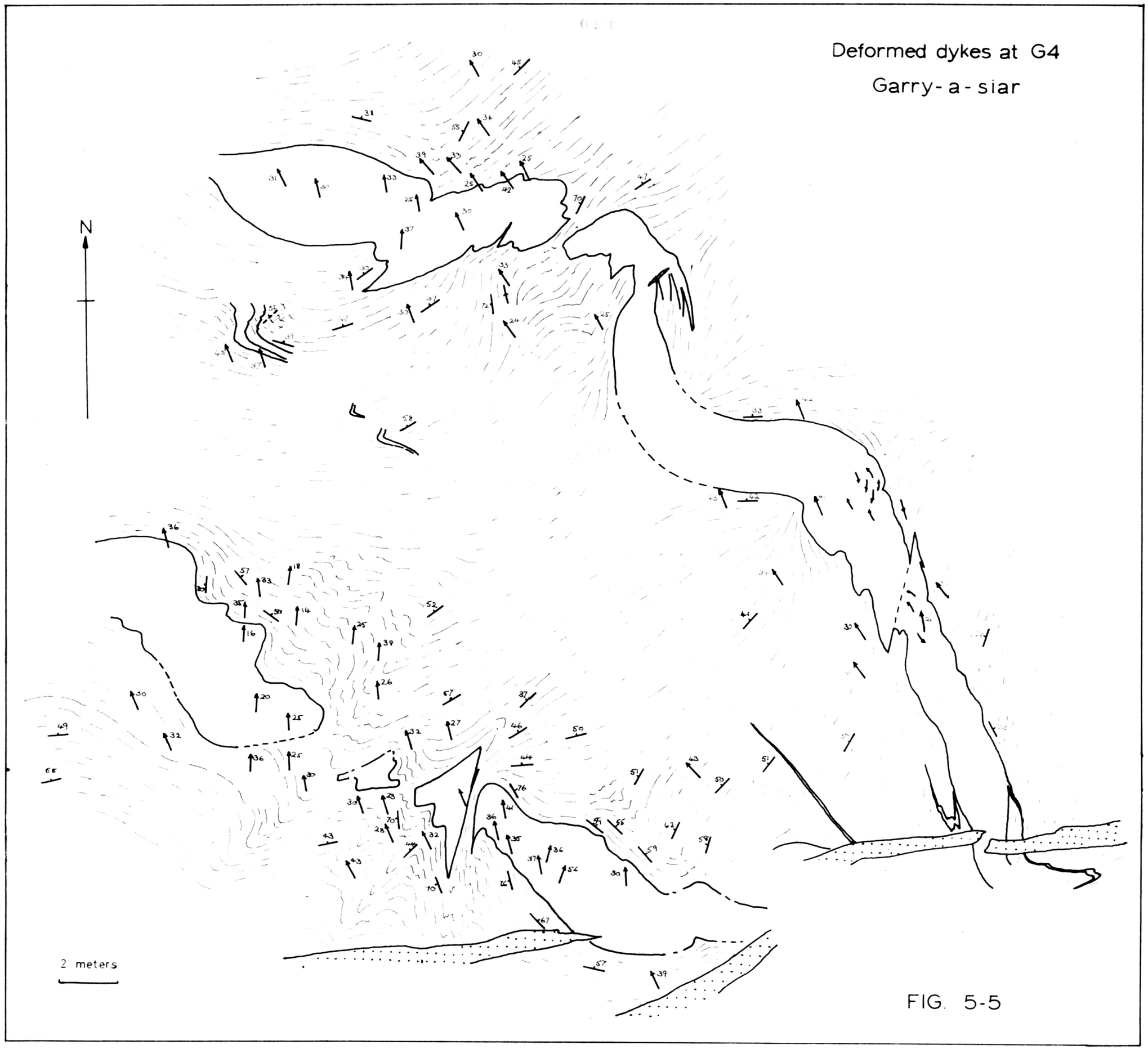


FIG. 5-5

time and deformation. I shall return to this important point later.

5.4c The f_1 fabric

Within the deformation zones the gneisses sometimes attain a more strongly banded nature as a result of the f_1 strain. The dykes and foliation were brought into parallelism at G8 and A8 and a planar fabric, demonstrated by biotite, hornblende and by feldspar phenocryst orientations, was developed.

5.4d F_1 deformation outside the deformation zones

Outside the deformation zones as shown in Maps 7 and 8, there is very little eveidence for any f_1 structure, though this does not necessarily mean that there was no f_1 strain. Both dykes and foliation could have suffered a certain amount of re-orientation, and , as I shall discuss later, (section 5.6), the f_1 strain may have had an important effect on many of the f_2 fabrics.

5.5 The F_2 deformation

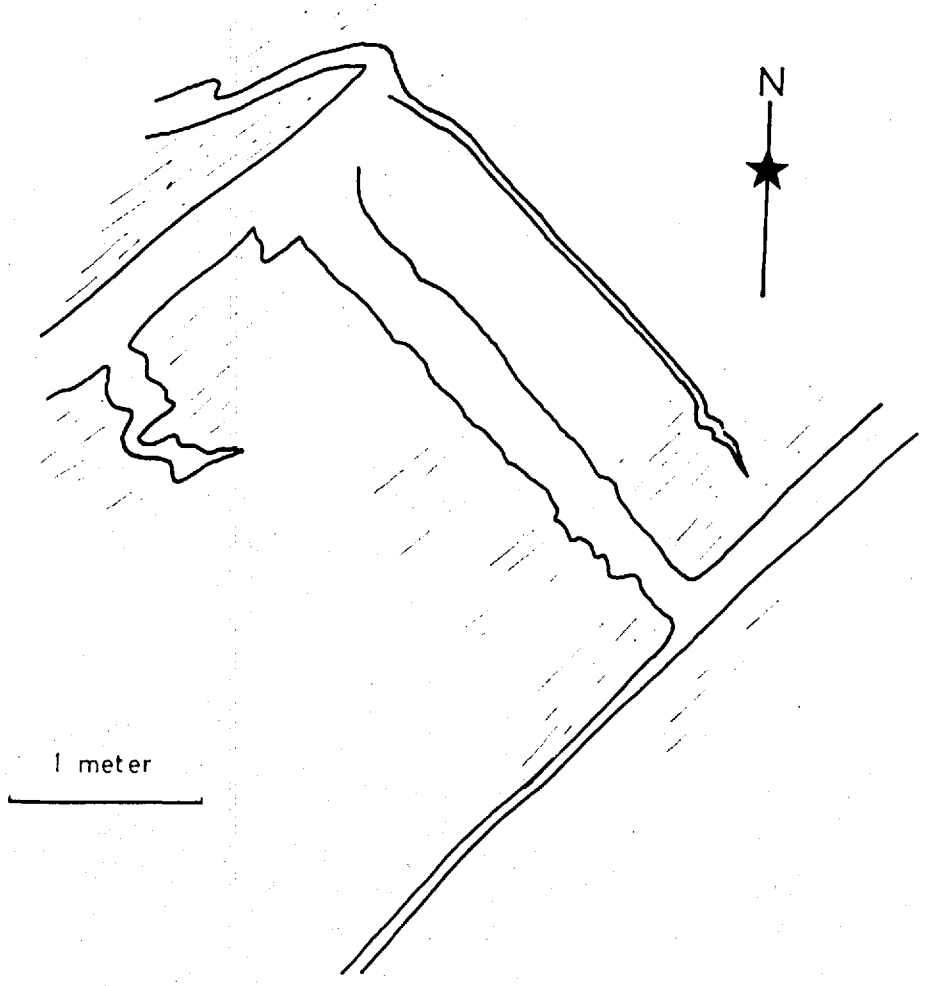
5.5A General

F_2 folds are best developed at north-west Garry-a-siar, east Ardivachar and locally at west Ardivachar. The axial plane now shows a variable orientation ranging from north - south to north-east - south-west and in parts, north of Culla Bay, to east - west; this variation is probably due to later (f_3) re-orientation. Many of the dykes, prior to the f_2 deformation, had a north-west - south-east strike and lay largely within the contractional field of the strain ellipsoid and were thus duly folded. However, the dyke at A10 presumably had an orientation originally close to north-east - south-west, as it was only boudinaged by the f_2 deformation. Other dykes however (e.g. at G12) still have a

Fig. 5.6

A branching dyke with folded apophyses from north of A14, Ardivachar.

FIG. 5-6.



west-north-west - east-south-east orientation, at a high angle to the f_2 compressive strains, and yet remain apparently unfolded.

As can be seen in Maps 7 and 8, it is the dykes which are usually folded while the gneissose foliation now often lies parallel to the f_2 axial plane, and probably originally lay close to the f_2 X - Y plane. The resulting pattern of folded dykes with relatively unaffected foliation can be exemplified by that of north-east Ardivachar between A1 and A4, where there are folds of different tightness affecting the dykes.

At A4 (Map 7) the dyke traces out a structure which appears to be a reclined fold which plunges at 35° to 40° to the north-west and closes to the east. In the rocks in the core of this fold, close to the shingle beach, however, the gneissose foliation traces out an older Scourian reclined isoclinal fold with the same plunge and strike as the Laxfordian fold, but with a closure to the west.

Often dykes are folded while their apophyses are not, or vice-versa, depending on their orientation relative to the f_2 strain axes. This is demonstrated by structures in apophyses at west Ardivachar at A15 and north of A14 (Fig. 5.6)

5.5B The effect of f_2 deformation on the discordant structures

The subject of the deformation of discordant dyke/gneiss relationships is one which could take up a great deal of time and space, but as the more detailed theory and the examples are of little relevance to the structural story of this thesis, they have been largely skimped over here and will be published separately elsewhere. It is, however, worthwhile here mentioning some of the more interesting examples of such deformation, taken from the folded dykes at Ardivachar and Garry-a-siar.

Fig. 5.7

Deformation of discordant relationships between dyke and gneiss on the northern part of the large dyke at G7, Garry-a-seer.

FIG. 5-7.

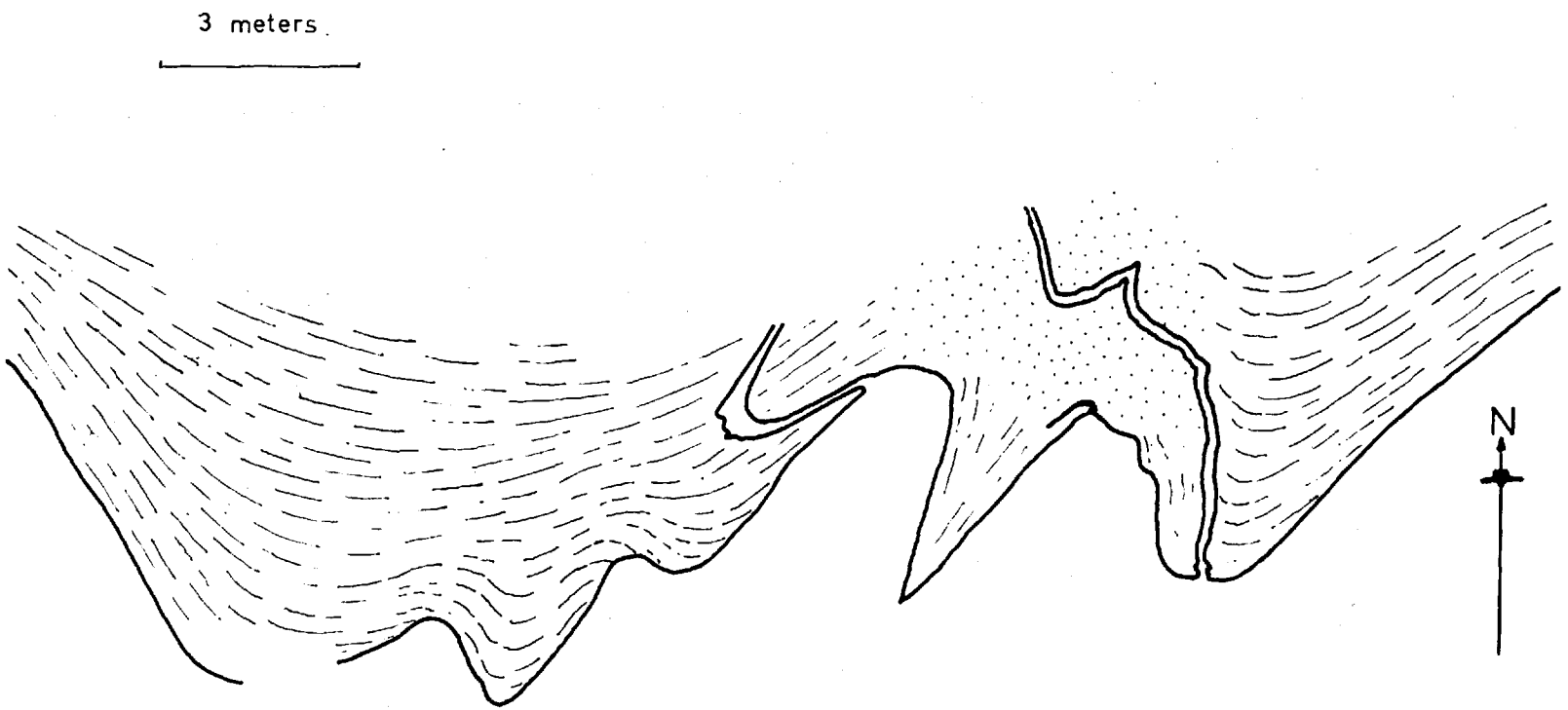
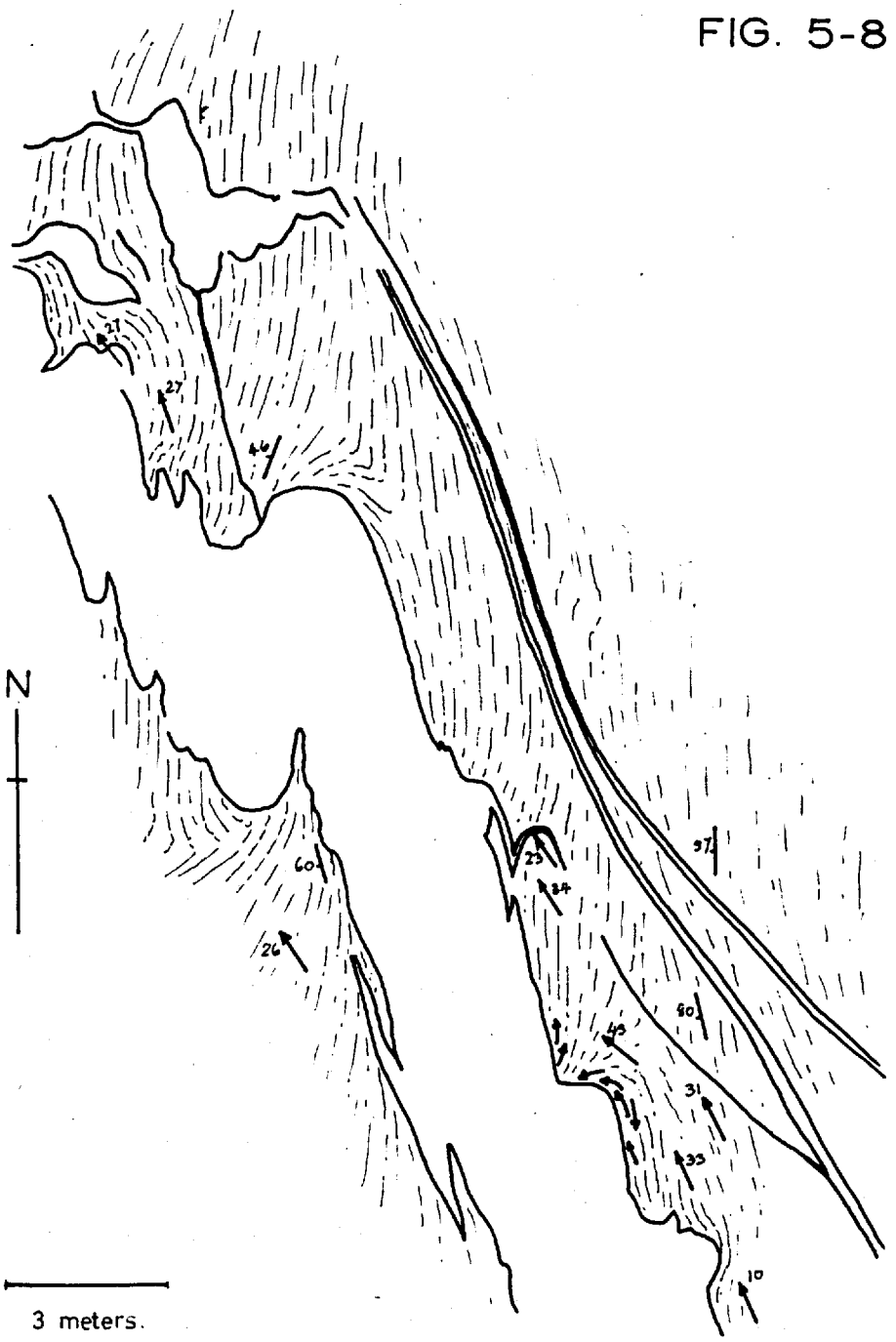


FIG. 5-8



Deformed dyke at G6 Garry a siar

A very good example is given by the large dyke at G7 (Fig. 5.7 and also Map 8). The dihedral angle varies from almost nil on the north-west - south-east limb, to an angle of about 45-50° close to the hinge, and then back again to a low angle on the north-east - south-west limb. It gives the pattern of a cross-cutting relationship which has suffered first folding, followed by a component of homogeneous strain. It may also be noted (Fig. 5.7) that the intersection angles of the apophyses to the dyke vary over different positions of the f_2 fold.

To the north, at G6 (Fig. 5.8), similar phenomena can be seen, but here the original discordant angle was more variable and Scourian folds were involved in the deformation. As is to be expected, the interlimb angle of the Scourian fold changes when traced round the f_2 fold. In the south-east parts of the dyke, dyke, gneiss and lineation are folded by an f_4 fold.

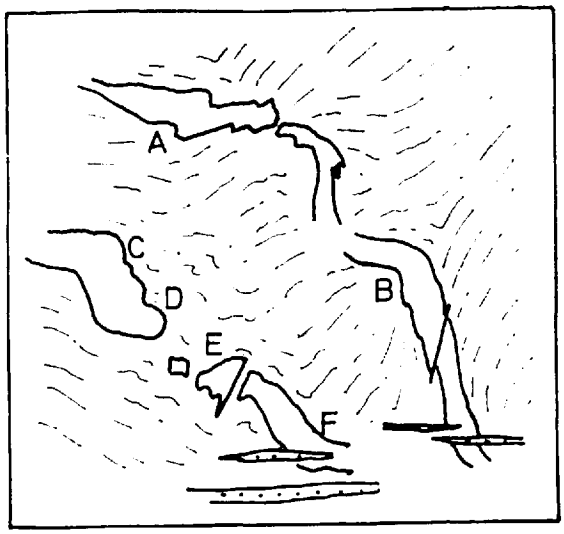
At G4 the structures are made highly complex by the presence of large Scourian folds, and by the fact that there was a high component of f_1 Laxfordian deformation close to the dykes. However, one can still plot the variation in dihedral angle round some of the f_2 folds (Fig. 5.9) and can make some estimate of the original angle before f_2 deformation. Though these original angles (Fig. 5.10) cannot be particularly accurate, there does seem to be a systematic variation in original angle along both of the main dykes, which appears to trace out a fold, presumably of Scourian age. Apart from a few areas in the south, and the regions close to the dykes, this Scourian fold has been almost totally obliterated by the Laxfordian deformation.

An examination of the deformation of discordant dykes should discuss the examples in three-dimensions, but these will be dealt with on another

Fig. 5.9

Plots of the dihedral angles (δ) between dykes and gneiss round f_2 folds at G4, Garry-a-siar.

FIG. 5-9



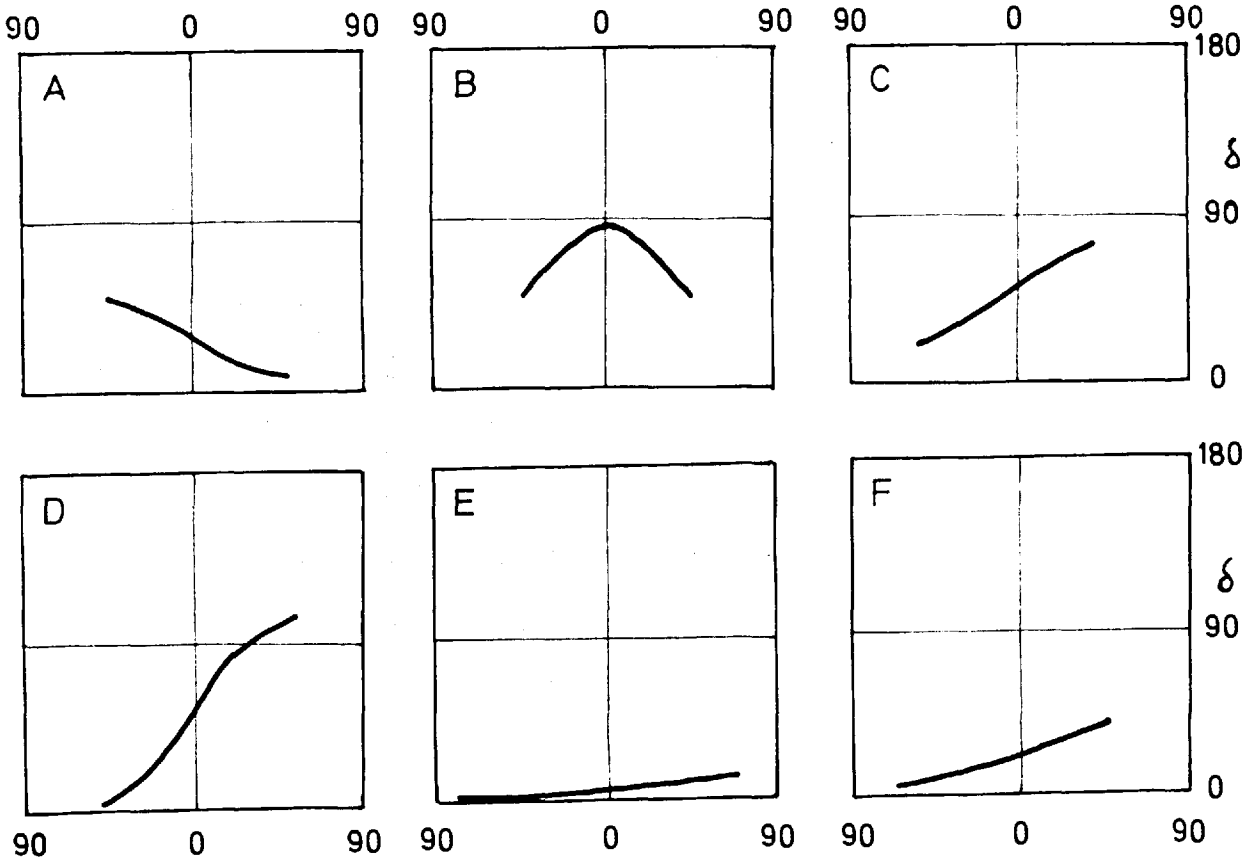


Fig. 5.10

Diagram to show the variation in original dihedral angle between dykes and gneiss along two dykes at G4, Garry-a-siar. This variation in original discordant angle appears to trace out a pre-Scourie-dyke fold which elsewhere has been almost obliterated by the Laxfordian deformation.

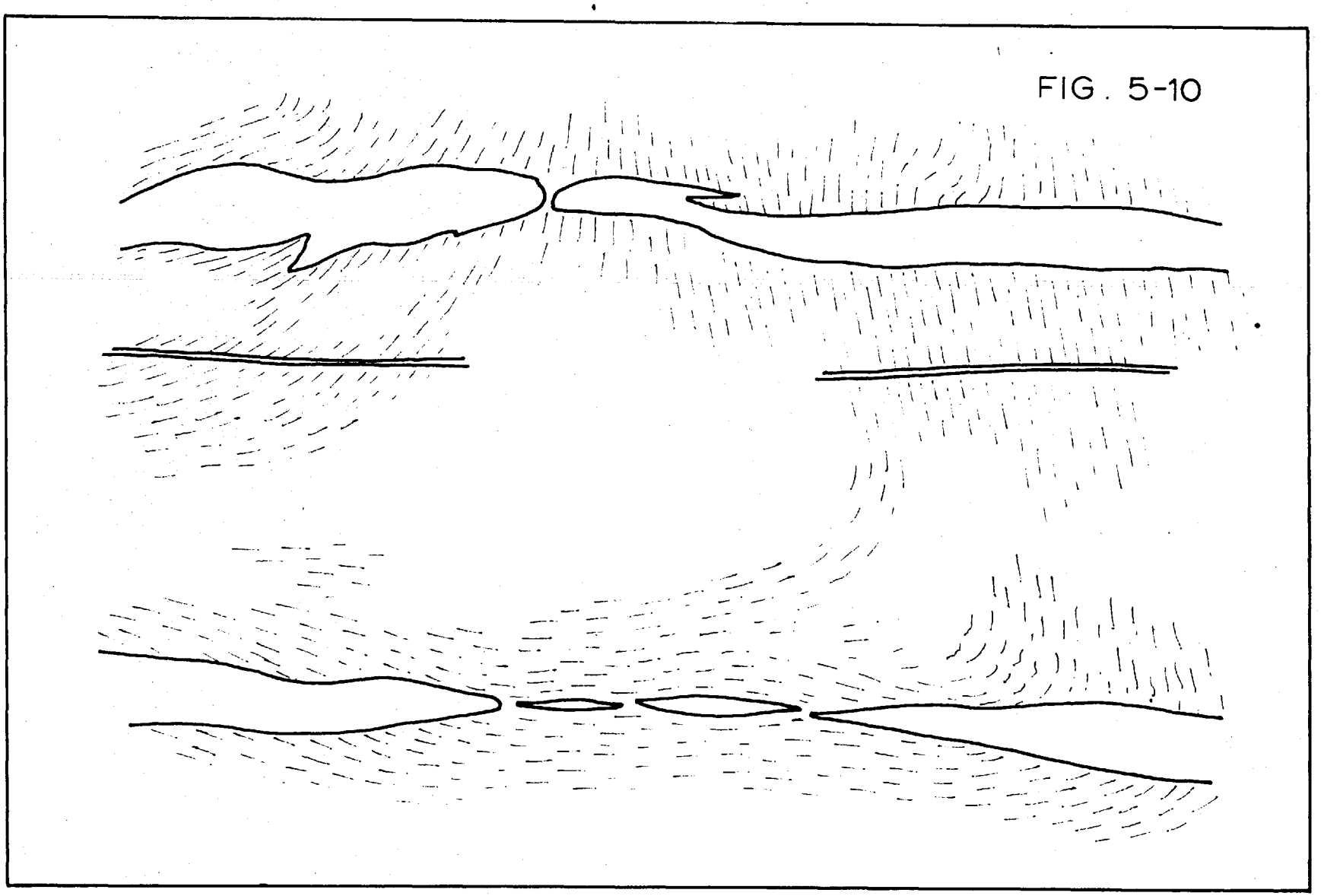


FIG. 5-10

occasion.

If I may therefore put forward some of the results before mentioning the theory and the examples, then one can summarise the main effects of the f_2 deformation as follows:-

a) The value of the dihedral angle varies round the folds in a manner which would indicate buckle folding, rather than similar folding.

b) The folding often appears to have been accompanied by a higher component of flexural shear than of tangential longitudianl strain, as far as the contact between dyke and gneiss is concerned. Quite often the dihedral angle does not have a peak close to the fold hinge, as one would expect from tangential longitudinal strain deformation, and there does not appear to be much difference in the dihedral angle plots on the inner and outer arcs of the folded dykes.

c) Many of the folds show the effects of a modification by a homogenous strain. In some cases the dihedral angle reaches a peak to one side of the fold hinge on the outer arc of the fold (c.f. Fig. 5.7). This is characteristic of a homogeneous strain superimposed on flexural slip folds, where the homogeneous strain is not of the type k= ∞ with X parallel to the fold hinge.

5.5C The f_2 fabric

The type of f_2 fabric would appear to be related to the amount of f_2 deformation. Where the f_2 strain was low, then there is only a linear fabric in the dyke or gneiss. F_2 planar fabrics are fairly common outside the low deformation areas of Ardivachar and Garry-a-siar but occur only rarely within them, and then in areas of localised higher deformation. Thus a planar fabric is developed in the dyke at A1 Ardivachar, in the axial region of a fairly tight fold and locally at G4 at Garry-a-siar

where the dykes and gneisses are tightly folded and an axial plane fabric is produced in the gneiss.

As mentioned previously, all the dykes show some mineral lineation which has a constant orientation and is co-axial with other linear rock fabrics of various ages. All the lineations, intersection lineations and linear fabrics as listed below, appear co-axial.

1) Scourian structures

   a) Scourian fold hinges of three fold generations.

   b) The long axis of boudins (note: not necessarily the extension direction of boudins), of pods of hornblendic material and of rods of quartzo-feldspathic material.

   c) The intersection of axial plane fabrics of Scourian age with the gneissose foliation.

   d) Scourian mineral lineations of hornblende, quartz and feldspar.

2) Post-Scourian, pre-Laxfordian structures.

   a) The intersections of groups of dykes in areas of low deformation. Thus orientations of the dyke apophyses of the dyke at G5 (Fig. 5.4) when plotted, have an intersection co-axial with the other structures.

   b) The intersection of dykes and gneissosity.

3) Laxfordian structures.

   a) Laxfordian fold hinges.

   b) The poles to π-pole girdles of folded dykes or foliation.

   c) The intersection of Laxfordian planar fabrics (of f_1 or f_2 age) with the gneissose foliation.

   d) Laxfordian mineral lineations within or outside the dykes, as demonstrated by the preferred orientations of feldspars, feldspar phenocrysts, hornblendes, etc.

e) The long axis (not usually the extension direction) of f_1 boudins.

There are a few exceptions to the co-axiality:-

i) The linear fabrics may be folded by the f_4 structures as will be mentioned later.(Section 5.10).

ii) The lineation may vary in orientation round certain structural inhomogeneities in the rock. Within many of the large Scourie-dykes, the change from pyroxene bearing rock with a mimetic original igneous texture to amphibolite with a tectonic linear fabric was not uniform. There is therefore a marked strain heterogeneity between the non-tectonised basic rock and the tectonised amphibolites (assuming that the presence of fabric is an indicator of a certain amount of strain). The fabric in the dykes wraps itself round these non-tectonised basic bodies (Fig. 5.11) as would be expected. In the south-east corner of the northern dyke at G4 the dyke shows possible internal boudinage structures, the boudin necks now being filled with quartzo-feldspathic material (Fig. 5.11). The lineation in these dykes wraps itself round these more competant bodies.

iii) The lineation may vary in its orientation around some of the f_2 folds. This is shown by the orientations of the linear fabric in the dyke at G7 and in both dyke and country rock at G4 (Map 8, Fig. 5.5). The possible explanations for this variation are probably best left until I have discussed the origins of the co-axiality in the next section.

5.6 The sequence of early finite strains

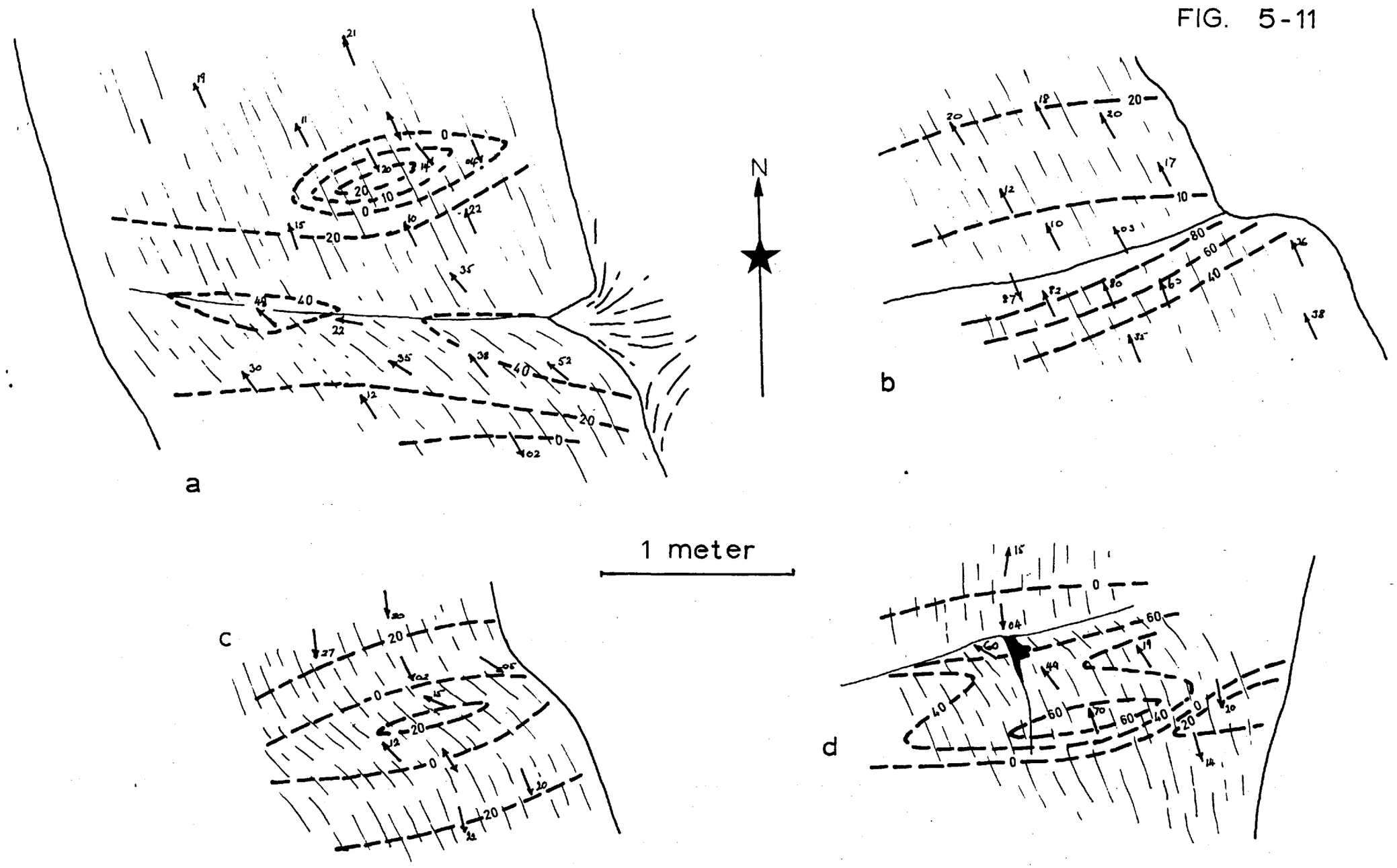
5.6A The f_1 strain

Within the f_1 deformation zones, where the strain was more intense,

Fig. 5.11

Internal deformation within the Scourie-dykes at G4, Garry-a-siar. The heavy lines are plunge isogons (see Elliot, 1965) i.e. they join points of equal plunge. The less heavy lines indicate the trend of the plunge orientations and in the case of 5.11d, the trend of the schistosity.

FIG. 5-11



a banded nature was imposed on the rock and the f_1 mineral fabric, where still visible, was planar ($s \geqslant L$). This would indicate, therefore, that the f_1 strain was of the type $1 \geqslant k > 0$.

5.6B The f_2 strain

As I have shown, when the f_2 deformation is intense, a planar lithological and mineral fabric is developed, indicating again that the strain was of the type $1 \geqslant k > 0$.

However, as one goes into areas of less intense deformation, there is far less planarity to the fabric and over much of Ardivaehar and Garry-a-siar only a linear fabric can be seen. If fabric can be used to indicate the state of strain, then these rocks must bear a finite strain of type $k = \infty$ or $\infty > k > 1$

5.6C Sequence of strains

Using the above and other data previously mentioned, one can set up the sequence of types of finite strain which the rocks have passed through, making the assumption (which appears to be fairly valid) that the low deformation areas have been through the same tectonic sequence as the areas of higher deformation, but at each stage have suffered less actual strain.

1. The strain associated with the first deformation phase, f_1, appears to have been of the type $1 \geqslant k > 0$, where the X - Y plane lay with a west-north-west - east-south-east to north-west - south-east orientation, and the Z direction was almost horizontal.
2. Areas of low deformation appear to have a finite strain of the type $\infty > k > 1$, with an X axis (at Ardivachar Point and Garry-a-siar) plunging at 30° to 40° to the north-west.
3. When the f_2 deformation was higher, the finite strain was of the

type $k = 1$ pr $1 > k > 0$. The X - Y plane had a strike of north-east - south-west with a variable dip, usually between 30° and 50° to the west, and an X direction (from the mineral lineation) plunging at 30° to 40° to the north-west. The f_2 X - Y plane therefore appears to have lain normal to the f_1 X - Y plane.

To achieve this strain history there are two possibilities:

i) The early deformation took place under certain stress conditions which directly produced a $k = \infty$ strain following the f_1 and prior to the main f_2 deformation. I have however noted (section 5.5B) that the discordant relationships of the foliation to dykes round f_2 folds indicate that such a prolate strain was not superimposed on the rocks.

ii) The states of strain, $k = \infty$ for low f_2 deformation and $1 > k > 0$ for high deformation, were brought about by the superimposition of the f_2 strain increments on the f_1 strain fabric. I consider this to be the most likely possibility and will now examine the possible conditions in more detail.

5.6D Superimposition of the two strains

The characteristics of progressive deformation of liquid or plastic materials have been summarised by Ramsay (1967, pp 322 to 332). If there is an f_1 finite strain within the rock and increments of f_2 strain are superimposed on this strained state, a new finite strain results. The successive changes in strain can be plotted on a Flinn plot (Fig. 5.12) to give a strain path for the f_2 finite strain deformation. Each point along the path represents an f_2 finite strain after a certain number of small f_2 strain increments have been added.

Strain can be represented by matrix notation, the general strain increment matrix is given by:- (Ramsay 1967, p.327)

$$\begin{matrix} 1 + e_x & \frac{\gamma_{xy}}{2} - \omega_2 & \frac{\gamma_{xz}}{2} + \omega_2 \\ \frac{\gamma_{xy}}{2} + \omega_3 & 1 + e_y & \frac{\gamma_{y^2}}{2} - \omega_1 \\ \frac{\gamma_{xz}}{2} - \omega_2 & \frac{\gamma_{yy^2}}{2} - \omega_1 & 1 + e_z \end{matrix}$$

and the finite strain is determined by successive multiplication of such strain increment matrices with the previous finite strain matrix.

One can simplify the above matrix mulitplication considerably by examining the cases where the f_2 principal extensions, X_2, Y_2, Z_2 co-incided exactly with the co-ordinate directions x, y, z set up by the f_1 principal extensions X_1, Y_1, Z_1. One therefore considers the six positions:

$$\begin{matrix} x\ X_2 & x\ X_2 & x\ Y_2 & x\ Y_2 & x\ Z_2 & x\ Z_2 \\ y\ Y_2 & y\ Z_2 & y\ X_2 & y\ Z_2 & y\ X_2 & y\ Y_2 \\ z\ Z_2 & z\ Y_2 & z\ Z_2 & z\ X_2 & z\ Y_2 & z\ X_2 \end{matrix}$$

The orientations of the principal strains now remain constant and there is no rotation. The increment strain matrix simplifies to:

$$\begin{matrix} 1 + e_x & 0 & 0 \\ 0 & 1 + e_y & 0 \\ 0 & 0 & 1 + e_z \end{matrix}$$

and strains can be 'added' by

a simple matrix multiplication.

One can partially justify this simplification by the facts that in the areas of low deformation in South Uist the f_1 and f_2 X - Y planes are normal to each other, and that the orientation of the strong linear fabric is constant in both high and low f_2 deformation areas and appears to have suffered no appreciable rotation.

On Fig. 5.12 I have plotted the theoretical strain paths for (i) f_1 and f_2 k values both equal to one, (ii) f_1 and f_2 k values of 0, (iii) the general cases of f_1 and f_2 having different k values in the range of $1 > k > 0$. In this general case the resultant finite strains may tend towards $k = 0$ or $k = \infty$ with orientations of $\begin{matrix} x & Y_2 \\ y & X_2 \\ z & Z_2 \end{matrix}$ $\begin{matrix} x & X_2 \\ y & Z_2 \\ z & Y_2 \end{matrix}$ respectively, while $\begin{matrix} x & X_2 \\ y & Y_2 \\ z & Z_2 \end{matrix}$ gives a more normal path in the $1 > k > 0$ field. With the orientation $\begin{matrix} x & Z_2 \\ y & Y_2 \\ z & X_2 \end{matrix}$ the path moves to the $k = 0$ field, then reverses, crosses the $k = 1$ line to a low value of the $k = \infty$ strain field. The path then moves back to the $k = 0$ field but at slightly higher strain values, and finally moves off into the $1 > k > 0$ field. A high f_2 strain is required before the finite strain leaves the low deformation field, and there are five changes in the orientation of the principal strain axes during progressive deformation with this arrangement of f_1 and f_2 strain. These five changes in X, Y and Z, the f_2 finite strain axes, relative to the x, y and z co-ordinates are shown in Fig. 5.13.

With the f_2 strair in the orientation $\begin{matrix} x & Z_2 \\ y & X_2 \\ z & Y_2 \end{matrix}$ the strain path first proceeds to the $k = 0$ field and then reverses into the $k = \infty$ and then the $\infty > k > 1$ fields.

The path, though, which appears to be the most relevqnt to South

Fig. 5.12

Flinn plots giving the strain paths for finite strain deformation for

a) f_1 and f_2 k values both equal to one.

b) f_1 and f_2 k values both equal to zero.

c) f_1 and f_2 having different k values in the range $1 > k > 0$.

Orientations:

A.	x	X_2	B.	x	X_2	C.	x	Y_2
	y	Y_2		y	Z_2		y	X_2
	z	Z_2		z	Y_2		z	Z_2
D.	x	Z_2	E.	x	Y_2	F.	x	Z_2
	y	Y_2		y	Z_2		y	X_2
	z	X_2		z	X_2		z	Y_2

solid circle - f_1 strain

open circle - f_2 strain increments.

FIG. 5-12

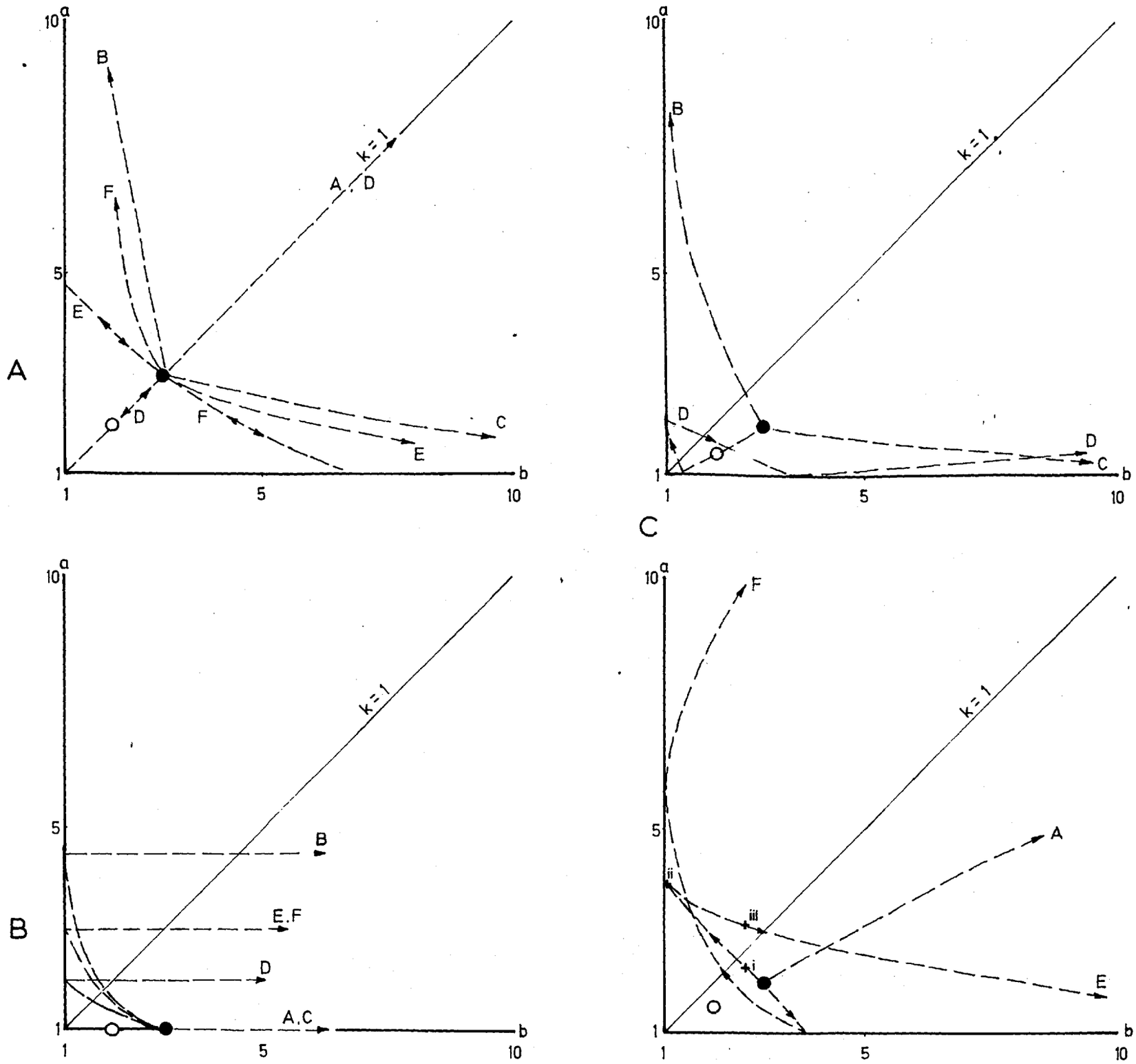


Fig. 5.13

Change in orientation of X,Y and Z, the finite strain axes, in relation to the x,y and z co-ordinates for

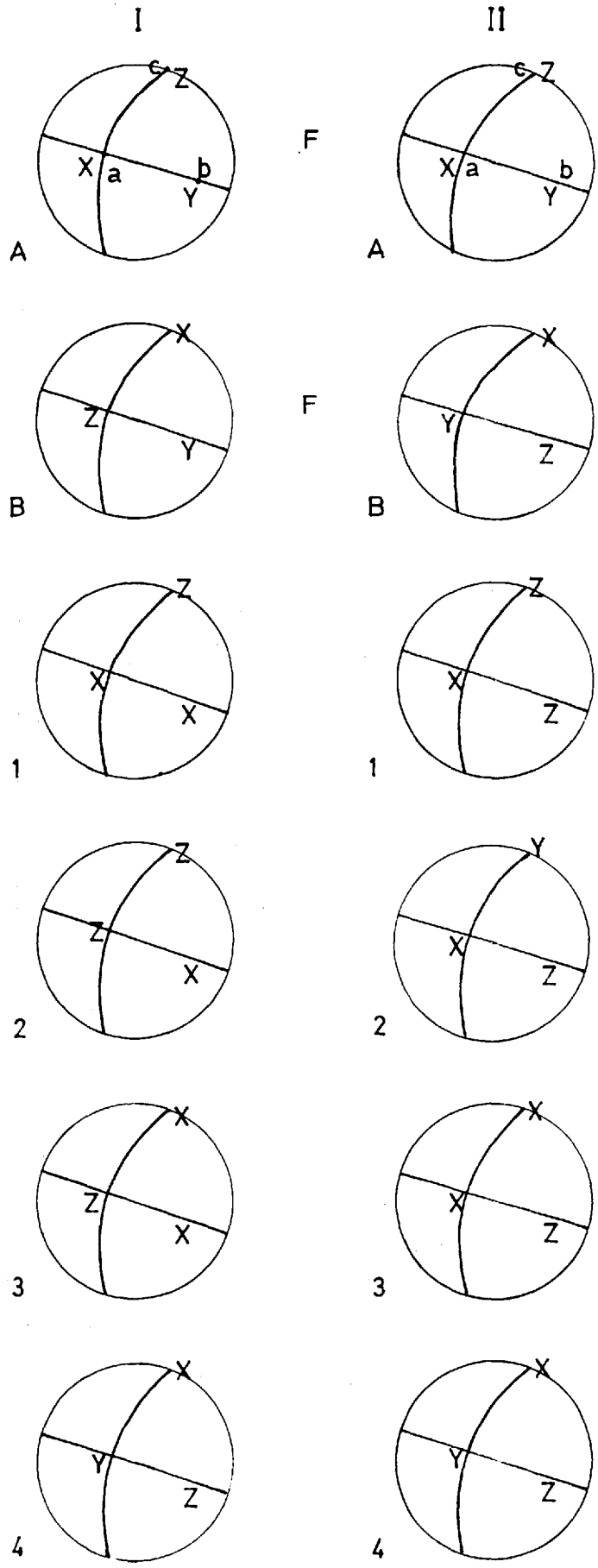
I. the orientation $\begin{matrix} x & Z_2 \\ y & Y_2 \\ z & X_2 \end{matrix}$ II. the orientation $\begin{matrix} x & Y_2 \\ y & Z_2 \\ z & X_2 \end{matrix}$

A. the f_1 finite strain axes

B. the f_2 incremental strain axes

1 to 4. the f_2 finite strain axes after various f_2 strain increments.

FIG. 5-13



Uist is that given by the orientation $\begin{matrix} x & Y_2 \\ y & Z_2 \\ z & X_2 \end{matrix}$. Here, the strain path passes to the $k = \infty$ field, and then returns to the $k = 0$ and the $1 > k > 0$ fields. Fig. 5.13 shows the sequence of X, Y and Z positions during this progressive deformation.

The above discussion deals not only with the case when the f_1 strain was low. When the f_1 strain was high, the sequence is exactly the same, except that the resulting finite strains are correspondingly much higher. It must be noted that in the case in which we are most interested, $\begin{matrix} x & Y_2 \\ y & Z_2 \\ z & X_2 \end{matrix}$, only very low increments of f_2 strain are required to send the finite strain into the $k = \infty$ field.

On Fig. 5.14 I have compared the positions of the finite strains for the different orientations along the various strain paths for a series of increments of f_2 strain using here the logarithmic ratios, plotting $\log_e \frac{1 + e_1}{1 + e_2} = \log_e a$ against $\log_e \frac{1 + e_2}{1 + e_3} = \log_e b$.

Thus the following points can be made from this theoretical discussion:-

1) The k value of the finite strain never remains constant, varying from $k = 0$ to $k = \infty$, but eventually after sufficient f_2 strain increments, all paths take on the gradiaent $a = b^k$ (where k is the function for the power curves for the incremental strains, in this case (Fig.5.14) = o.66). The log. deformation paths follow straight lines, and are only deflected at t the $k = 0$ and $k = \infty$ field boundaries.

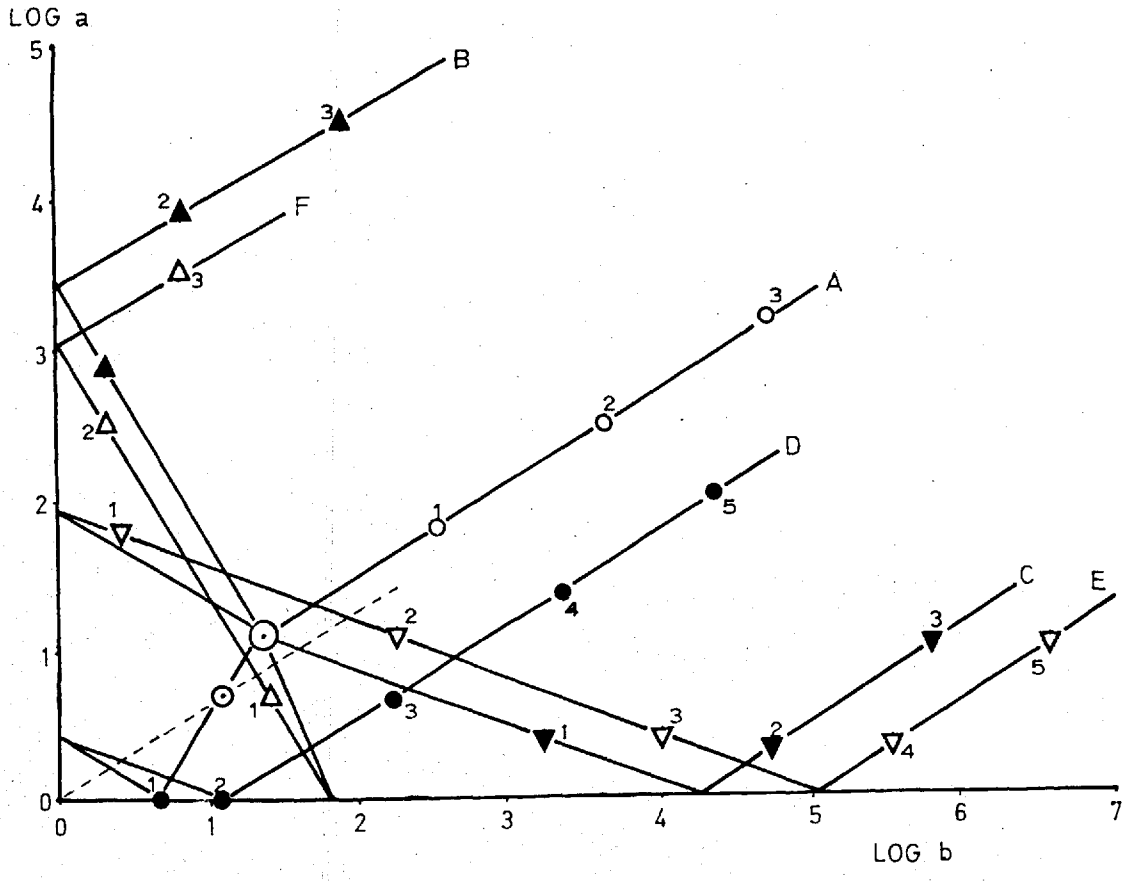
If the deformation had a rotational component, the strain paths would not follow straight lines but would curve.

2) Depending on the orientation of the strain increments, some finite

Fig. 5.14

The deformation paths of Fig. 5.12c plotted using the logarithmic ratios of the three principal strains.

FIG. 5-14



strains develop with very high strain ratios, others remain very low. Thus with the orientation $\begin{matrix} x & Z \\ y & Y \\ z & X \end{matrix}$ a very high f_2 strain is required before the finite strain leaves the low deformation area.

3) Linear strains are developed by the orientations $\begin{matrix} x & X \\ y & Z \\ z & Y \end{matrix}$ and $\begin{matrix} x & Z \\ y & X, \\ z & Y \end{matrix}$ planar strains by $\begin{matrix} x & Y \\ y & X \\ z & Z \end{matrix}$ and $\begin{matrix} x & Y \\ y & Z \\ z & X \end{matrix}$.

4) The strain paths given by $\begin{matrix} x & Y \\ y & Z, \\ z & X \end{matrix}$ $\begin{matrix} x & Y \\ y & X \\ z & Z \end{matrix}$ pass through the $k = \infty$, $k = 0$ fields respectively, to the opposite strain fields. In practice, one generally only sees the finite strain in the rock, not the previous increments, and so any effects of, say, the $k = \infty$ strain should have been dispensed with when the finite strain reaches the $k = 1$ field. If folding takes place, however, then rotational and shear components are brought in, and in parts of the fold, a $k = \infty$ strain might be re-developed.

5.6E The effect of buckle folding on the strain paths

When the rocks are folded, predictions of the finite strain are made much more difficult, as on the limbs the finite strains suffer both body rotation and the superimposition of strain increments due to the internal strains of buckling. Even at the fold hinge, where there is no rotation, there will probably be some component of tangential longitudinal strain. It ought therefore to be possible to describe almost any spread of points on a Flinn plot by incorporating buckle folding and progressive deformation.

This can be exemplified by taking the path produced with the orientation $\begin{matrix} x & Y_2 \\ y & Z_2 \\ z & X_2 \end{matrix}$ and considering the effects of buckle folding at different positions along the path. The position along the path will

depend on the number of increments of f_2 strain which have been added prior to buckle amplification, which will, in turn, depend upon such factors as the relative viscosities of the rocks (see section 4.3A2). In rocks of variable viscosities, therefore, folds will develop at many different positions along the path.

The folding can take place by flexural shear or by tangential longitudinal strain. With flexural slip deformation, the finite strain will suffer a high internal deformation on the limbs. Ramberg and Ghosh (1968) have shown that the shape and orientation of the finite strains will depend on the previous finite strain ratio and thus convergently or divergently fanning axial plane fabrics can form. The strain paths which develop with further f_2 homogeneous strain increments on such flexural slip folds will be extremely variable and will not be considered here. To simplify the case we can consider the deformation taking place purely by tangential longitudinal strain, and also consider that there was a very high rate of amplification.

(i) If the buckle folding develops when the f_2 strain increments are low, at point (i) on the strain path in Fig. 5.12, and the fold axis is parallel to x (i.e. Y_2), then the effect of the rotation on the limbs is such that new f_2 increments will be added with the orientation $\begin{matrix} X_1 & Y_2 \\ Y_1 & X_2 \\ Z_1 & Z_2 \end{matrix}$

The strain path on the limbs will therefore be completely different from that at the hinge (see Fig. 5.15a); planar atrain ellipsoids will develop on the limbs while only linear ellipsoids will develop at the hinge. The spread of points from the whole fold zone, on a Flinn plot, will be as in Fig. 5.15b; a large spread from the $\infty > k > 1$ to the $1 > k > 0$ fields.

If the folding develops about the z (i.e. X_2) axis, then the defor-

Fig. 5,15

The effect of buckle folding on the strain path produced with the orientation

x	Y_2
y	Z_2
z	X_2

a. the strain paths on the limbs and at the hinge when buckle folding develops at point (i) on the path given in Fig. 5.12c, when the fold axis is parallel to x (i.e. Y_2).

b. the spread of finite strains throughout the above fold after successive strain increments.

c. the strain paths on the limbs and at the hinge when the fold axis is parallel to z (i.e. X_2).

d. the spread of finite strains throughout the above fold after successive strain increments.

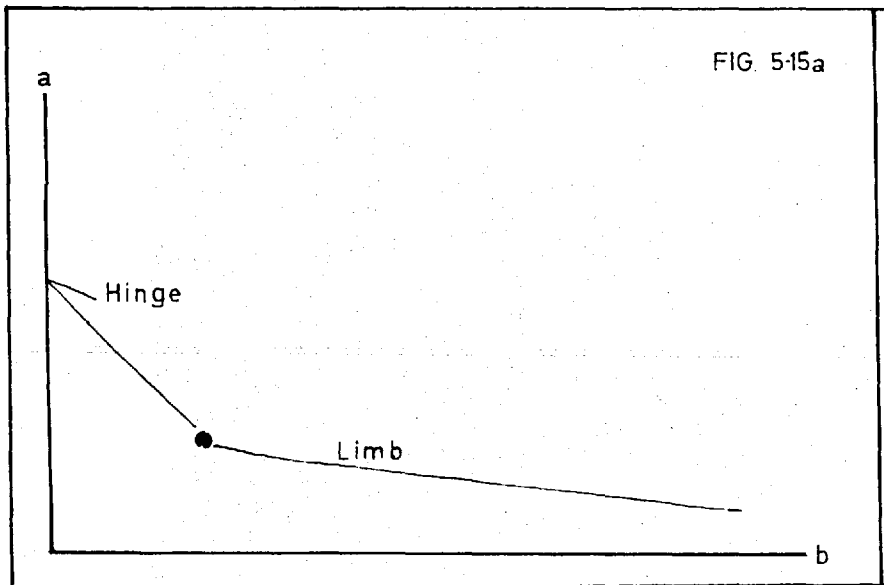


FIG. 5·15a

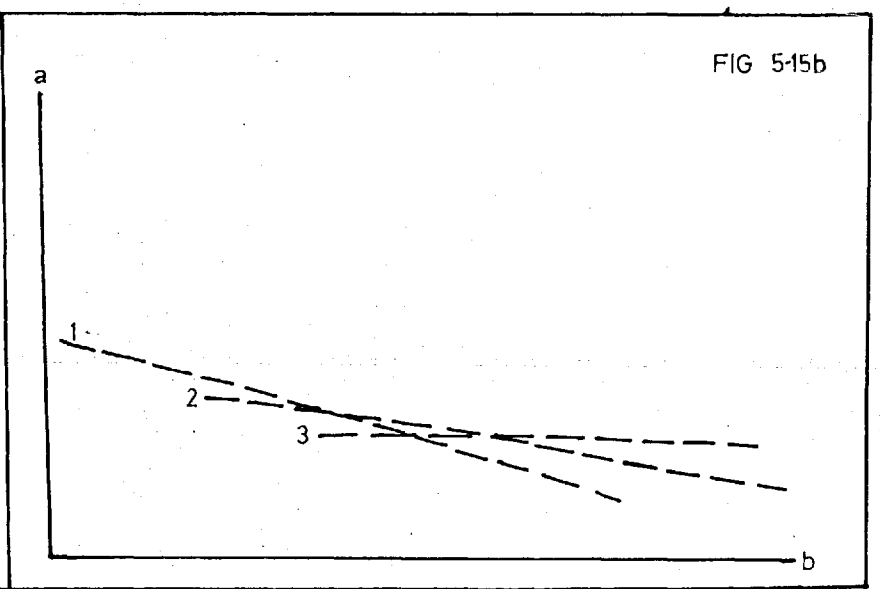


FIG 5·15b

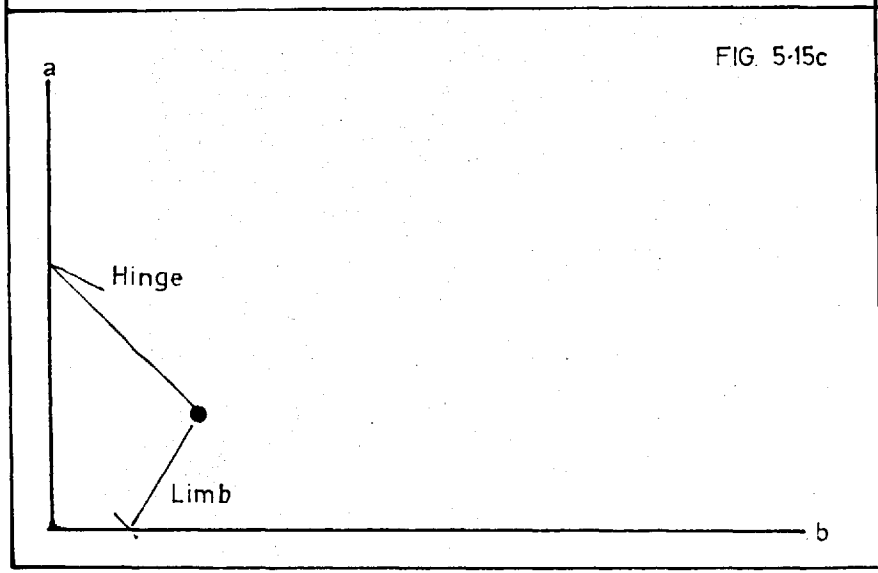


FIG. 5·15c

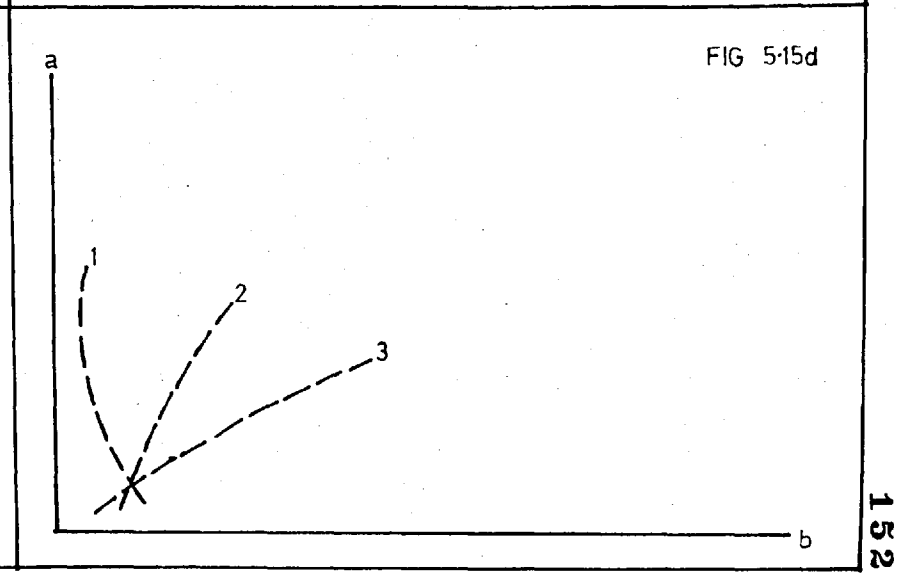


FIG 5·15d

mation path at the hinge will contine to follow the path $\begin{matrix} x & Y_2 \\ y & Z_2 \\ z & X_2 \end{matrix}$, while the strain on the limbs will be re-orientated into $\begin{matrix} X_1 & Z_2 \\ Y_1 & Y_2 \\ Z_1 & X_2 \end{matrix}$. These strain paths are given in Fig. 5.15c. On the limbs the finite strains will be negligible, while a linear strain will develop in the hinge zone. The spread of points on the Flinn plot will be very different from that given by the former orientation, and as shown in Fig. 5.15d will spread from low strains, close to zero, to well into the constrictional field.

If the fold axis was intermediary between y and z (i.e. any point on the f_2 X - Y plane, then all three finite strain axes will suffer rotation. Thus the strain axes will trace out complex loci round the fold, and with further homogeneous deformation following buckling, the new increments will cause further rotation of the finite strain. In such cases, there should therefore be a noticeable variation in orientation of all three strain axes, and a strong variation in fabric orientation should be expected. The spread of points on the Flinn plot will be a complex intermediary between the two mentioned previously.

One can conclude, therefore, that if buckle folding took place after only a few increments of f_2 strain had been added, then any constancy in fabric and strain would be destroyed, and generally there should be a marked variation in strain axes.

(ii) If the buckle fold amplifies at the stage when the finite strain had a k value of close to ∞ (point (ii)), then as all planar elements should intersect at x (Y_2) and should intersect the new X_2-Y_2 plane at x (Y_2), only one possible fold axis orientation can result. The rotation due to buckle folding will make no difference to the

finite strains and the path will continue as before. The buckle folding will therefore have little effect on either orientations of the strain axes or on the final finite strain.

(iii) If the buckle folding develops after moderately high amounts of f_2 strain (point (iii)), then again rotation will play an important part in determining the strain path. If the fold axis is parallel to x (Y_2), then on the limbs a new orientation of $\begin{matrix} X_1 & Y_2 \\ Y_1 & Z_2 \\ Z_1 & X_2 \end{matrix}$ will develop. Thus with with further f_2 strain increments an oblate strain will develop at the hinge while a linear strain will form on the limbs. If the fold axis is not parallel to a principal strain axis, then again complex strain patterns will develop.

Application to the Uists

1) The strain sequences in the low deformation areas of South Uist and Benbecula generally appear to be closest to the path given by the orientation $\begin{matrix} x & Y_2 \\ y & Z_2 \\ z & X_2 \end{matrix}$. There does not appear to be much change in orientation of the X f_2 finite strain axis during progressive deformation; the strong linear fabric maintains a fairly constant direction for all amounts of f_2 strain. It could be argued, though, that this constancy of lineation is due to a strengthening of the fabric, or a re-oreintation, during the f_3 phase. One also has no evidence whether or not the Y or Z f_2 finite strain axes remained constant, though it would appear from the low deformation areas that the f_1 and f_2 X - Y planes were originally almost perpendicular.

It could be argued also that the Uist strain sequence could have developed during the single f_2 deformation phase where there was a rotation of the incremental strain axes during deformation. This must

of course remain an alternative possibility, but it does require some extra mechanism of which evidence has not been found. One can however recognise two phases of folding and, to me, the simplest mechanism which will cause the Uist type path is the one given above.

One can consider therefore that the f_1 phase of deformation gave generally a low strain as in Fig. 5.12, though with locally higher strain values. With very small increments of f_2 deformation, the finite strain would be of the linear type, with the X axis plunging at 30° to 40° to the north-west. This, I believe, is the situation which one can now see over much of Ardivachar and Garry-a-siar. Folding probably only developed locally, depending on factors such as the viscosity contrast.

With further f_2 deformation, the X direction would maintain its orientation but a new X - Y plane would develop, striking north-east - south-west. The finite strain would now be in the $\infty > k > 1$ or $k = 1$ field. This state is probably reached only in the more deformed parts of Ardivachar and Garry-a-siar, but is the more general state outside the areas of low deformation.

2) At Garry-a-siar there is a well marked spread of the lineation round many of the f_2 folds, though the folds are almost co-axial with the general lineation direction. Such fanning could best be explained by considering the f_2 folds to have begun to amplify after relatively few f_2 strain increments, before all the fabric became co-axial and when the intersection of the f_2 incremental X - Y plane on the dyke did not coincide with the X strain axis. From the orientations of the fabric, much of the internal deformation within the dyke at G7 (Fig. 5.3) would appear to be due to tangential lonitudinal strain.

5.7 Variations in amounts of f_2 deformation

Throughout Garry-a-siar and Ardivachar, there are no sharp boundaries between areas of high and areas of low f_2 strain; these boundaries are often rather vague and diffuse. One can, however, see a general pattern of areas where most basic dykes are deformed adjacent to areas where there is very little apparent deformation. Actual strain ratios are difficult to obtain from the areas of low f_2 strain, but from localised patches of higher deformation a few strain ratios may be obtained from folded dykes. Strain ratios of 5.5 : 1 have been obtained from a group of folded dykes at G15, in the southern part of Garry-a-siar (see Table 5.1), and a ratio of 8.4 : 1 from a folded dyke in a high deformation zone north of Culla Bay.

There are also anomalies, as at G5, where no portion of the dyke appears to be folded or boudinaged, and yet it lies in an area of moderate f_2 deformation (see Map 8). I would suggest therefore that there are two problems concerning the variable f_2 deformation, in that both the large scale variations and also the local anomalies need to be explained. They need not be due to the same cause. First, the relationship of these low deformation areas to the f_3 and f_4 structures needs to be known and it may therefore be better to leave a discussion of these problems till later.

5.8 The viscosity ratios between dykes and gneisses

As has been explained previously (section 4.3A2), viscosity ratios can be obtained from some folded layers, and Table 5.1 gives some of the viscosity ratios obtained from f_2 folds along the west coast of Benbecula. One should note that the viscosity ratios for dykes and gneiss are rather

	AMOUNT OF POST-BUCKLE FLATTENING	PRE-BUCKLE HOMOG.STRAIN	SHORTENING DUE TO BUCKLE	TOTAL STRAIN	VISCOSITY RATIO
1	1.5 : 1	1.5 : 1	1,3 : 1	2.95: 1	65+
2	1.5 : 1	1.5 : 1	4 : 1	9 : 1	65+
3	1.4 : 1	1.5 : 1	3 : 1	6.75: 1	40+
4	3 : 1	1.5 : 1	3 : 1	13.5: 1	40+
5	2 : 1	1.5 : 1	2.8 : 1	8.4 : 1	25
6	9 : 1	2.25 : 1	1.8 : 1	36 : 1	16
7	4 : 1	3 : 1	3.1 : 1	37 : 1	11
8	2 : 1	2 : 1	1.75 : 1	7 : 1	0.05

Table 5.1

Variation in amount of total strain and viscosity ratio between Scourie-dykes and acid gneiss for eight groups of folds from the west coast of Benbecula.

variable, from 70 : 1 for thin dykes at G15 at the southern part of Garry-a-siar to between 10 and 15 : 1 for folded dykes on the north-west coast of Benbecula, to less than 1 for many dykes, e.g. at G7.

To a certain extent, an idea of the viscosity ratio of the large dykes to country rock can be gained from very rough estimates of the wave-length - thickness relationship made in the field. It can easily be seen from Maps 7 and 8 that many of the folds are rather chubby structures, with a low wavelength - thickness ratio, indicating low viscosity ratios. The dyke at G7, for instance, gives a wavelength - thickness ratio of less than 5 : 1.

An interesting point which arises from this is that if one accepts the above estimates of viscosity ratios, and if one accepts Sherwin and Chapple's (1968) and Huddleston's (1969) ideas on fold development, then for some of these buckle folds to be developed, quite a high amount of homogeneous strain must have been superimposed on the rocks prior to buckle fold amplification. In the area of north-west Garry-a-siar this preliminary homogeneous strain must have been in the order of at least 2 : 1 and probably higher.

Many of the dykes show structures which indicate that they were less viscous than the country rock; they often show cuspate edges with the rounded part of this cusp facing into the dyke rock. From fold isogon measurements it would appear that many of the folded dykes are of class 3, indicating that they were the less competent unit, and from an f_1 fold at G7 a dyke/gneiss viscosity ratio of 1 : 20 has been obtained. Other dykes show what would appear to be good buckle folds, indicating that they were the more competant unit, and yet at the edge there are often cuspate structures indicating that locally at least the opposite

was true. Dykes which have competent cores can therefore give off less competent apophyses (section 5.4A) or have less competent edges.

This change in competence is usually paralleled by a mineralogical change from garnet-(pyroxene)-amphibolite in the conpetent parts of the dyke to biotite amphibolite in the less competent parts. Most of the dykes which have very low ($\ll 1$) viscosity ratios with the country rock are biotite bearing amphibolites.

One feature concerning this change in competence with mineral composition has already been mentioned in section 5.5C, when I indicated that the tectonite fabrics often wrapped round the more competent garnet/pyroxene bearing cores to the dykes.

Anomalies, such as that of the dyke at G5, which shows no evidence of folding although surrounded by areas of relatively high f_1 and f_2 deformation, may be due not just to rapid variations in the amount of strain, but to low viscosity ratios between dyke and gneiss. The dyke at G5 carries a hornblende and biotite rich mineral assemblage, and a low competence could, therefore, account for its relatively unmodified branching arrangement.

Many of the dyke apophyses at G4 and G7, Garry-a-siar, and north of Culla Bay, show a whole assemblege of folds, almost ptygmatic in places, and yet from fold analyses they appear as the less competent units. As I have shown, these apophyses are often the only structures which could have initiated buckle folding. If one therefore accepts that these folds were due to buckle folding, then there must have been a change in viscosity ratio during deformation, and this change is most likely due to a change in mineral content during metamorphism. During deformation, the relict igneous fabric presumably broke down to a tectonite fabric

made up of more hydrous minerals, hence changing the viscosity of the rock. In the centres of the dykes, where deformation was less, or where there was less local migration of water, etc., then pyroxene, garnet and hornblende recrystallised and the rock remained competent.

It would appear that variations in temperature,, water fugacity, mineral content, etc., can change the mechanical properties of rocks, and if such variables patchily affected the acid gneiss as well as the basic rocks, then a host of extremely variable competence relationships and viscosity ratios could have been set up. If this were true then it is quite remarkable that there is any sense or homogeneity in any basement deformation.

5.9 The f_3 deformation phase

The striking feature of the f_3 deformation phase is that it once again exhibits large areal variations in the amount of strain, as well as local heterogenities.

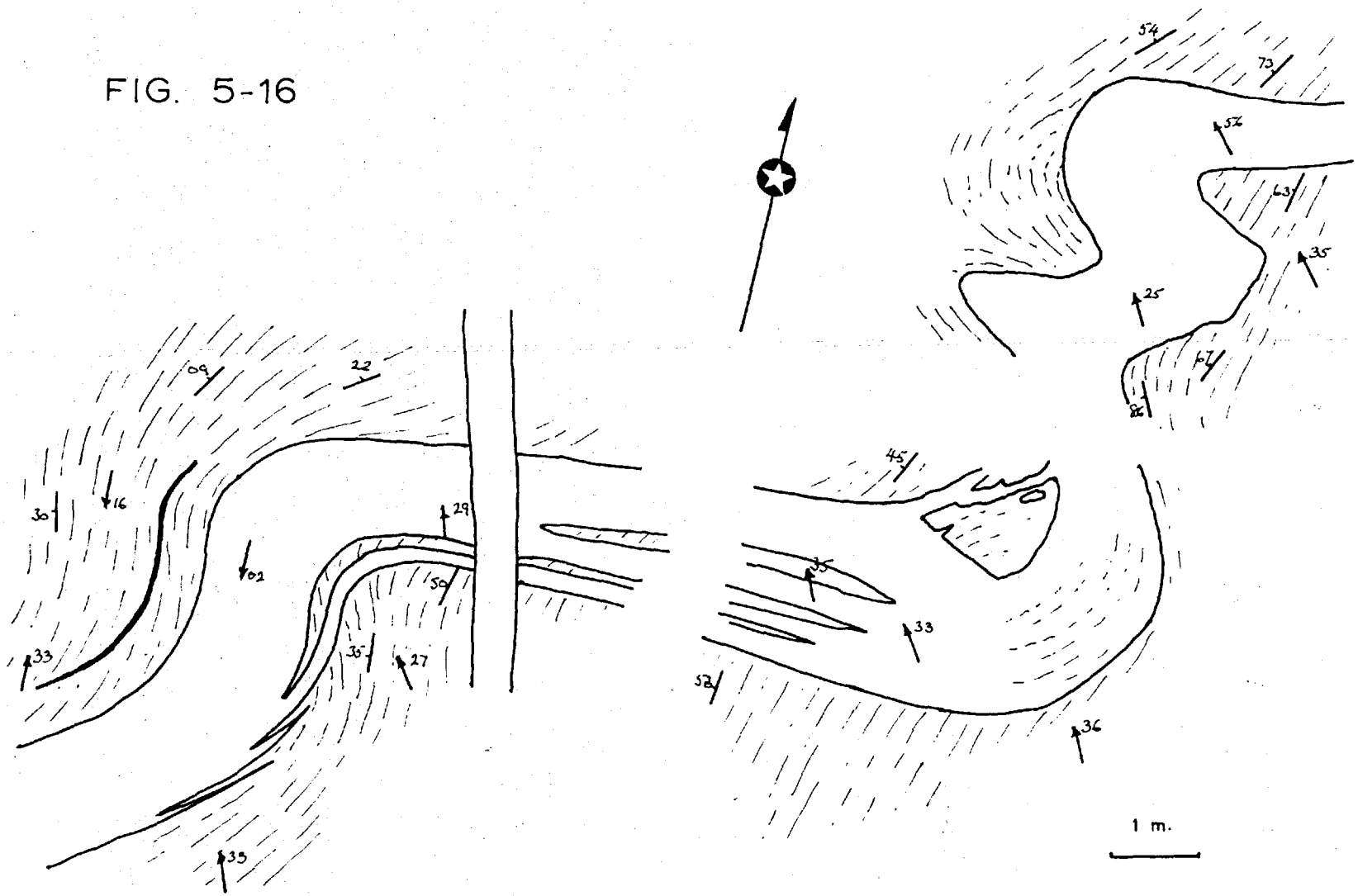
Maps 7 and 8 show the orientation of the foliation and f_2 axial traces in the areas of low deformation. It can be seen that both foliation and axial planes curve round large open structures which would appear to be of f_3 age. Apart from these large open structures, there are folds on a smaller scale belonging to this generation. At G13, Garry-a-siar, the dykes are folded in a synformal (but almost reclined) fold, which lies in the core of an earlier fold complex. This fold which modifies the dykes may be of f_3 age. It is apparently co-axial with the f_2 structures.

At Ardivachar, there is a similar set of east - west to south-east - north-west trending structures which are very much localised and hetero-

Fig. 5.16

A deformed dyke at A9, Ardivachar. Note the variation in discordant angle between dyke and gneiss in the north-east part of the map, where the dyke must have been intruded in an irregular form. A strain of 6 : 1, however, has been obtained from folded garnet/feldspar veins within the Scourie-dyke from this region. In the west, the discordant angle varies much more systematically round the f_3 fold, but the structure is complicated by f_4 warping.

FIG. 5-16


54
73
57
63
35
25
47
86
45
09
22
30
16
02
29
50
35
27
33
35
33
52
36
35
1 m.

geneous and again almost co-axial with the f_2 structures. The dyke at A9 is quite tightly folded by these structures. Garnetiferous, quartzo-feldspathic layers which formed previously within the amphibolite, have been tightly folded and give a two-dimensional strain ratio of 6 : 1. The dihedral angle between dyke and foliation varies round these folds but in an unsystematic fashion (Fig. 5.16) which is probably due to strong variations in the pre-existing dihedral angles. To the south these basic dykes are folded round large open warps (interlimb angle of 90° to 110°), which nicely show the effects of flexural shear on the discordant relationships (Fig. 5.16). At A10, boudins, presumably of f_2 age, are squashed together and have suffered a certain amount of dextral shear instead of normal buckle folding.

South of the small bay, near A 12, dextral shear is a more important mechanism than simple buckling within the f_3 deformation zones, and in places these appear as typical "shear zones". One such heterogeneous deformation zone is illustrated in Fig. 5.17, another in Dearnley and Dunning, 1968, p.358. In both, pegmatite is intruded along the shorter and more deformed limb of the fold. If one examines the change in dihedral angle round the folded dykes within these zones, one finds that the dihedral angle decreases relatively regularly from the northern, or east-west limb, to the crest, and is very low on the north - south or north-west - south-east limb (Fig. 5.17). This form of variation in the dihedral angle rules out a mechanism of only heterogeneous simple shear for the production of the folds. Though there was probably a strong simple shear component present here, most of the deformation would seem to be caused by heterogeneous flattening.

I would like therefore to stress this point again. Many of these

Fig. 5-17

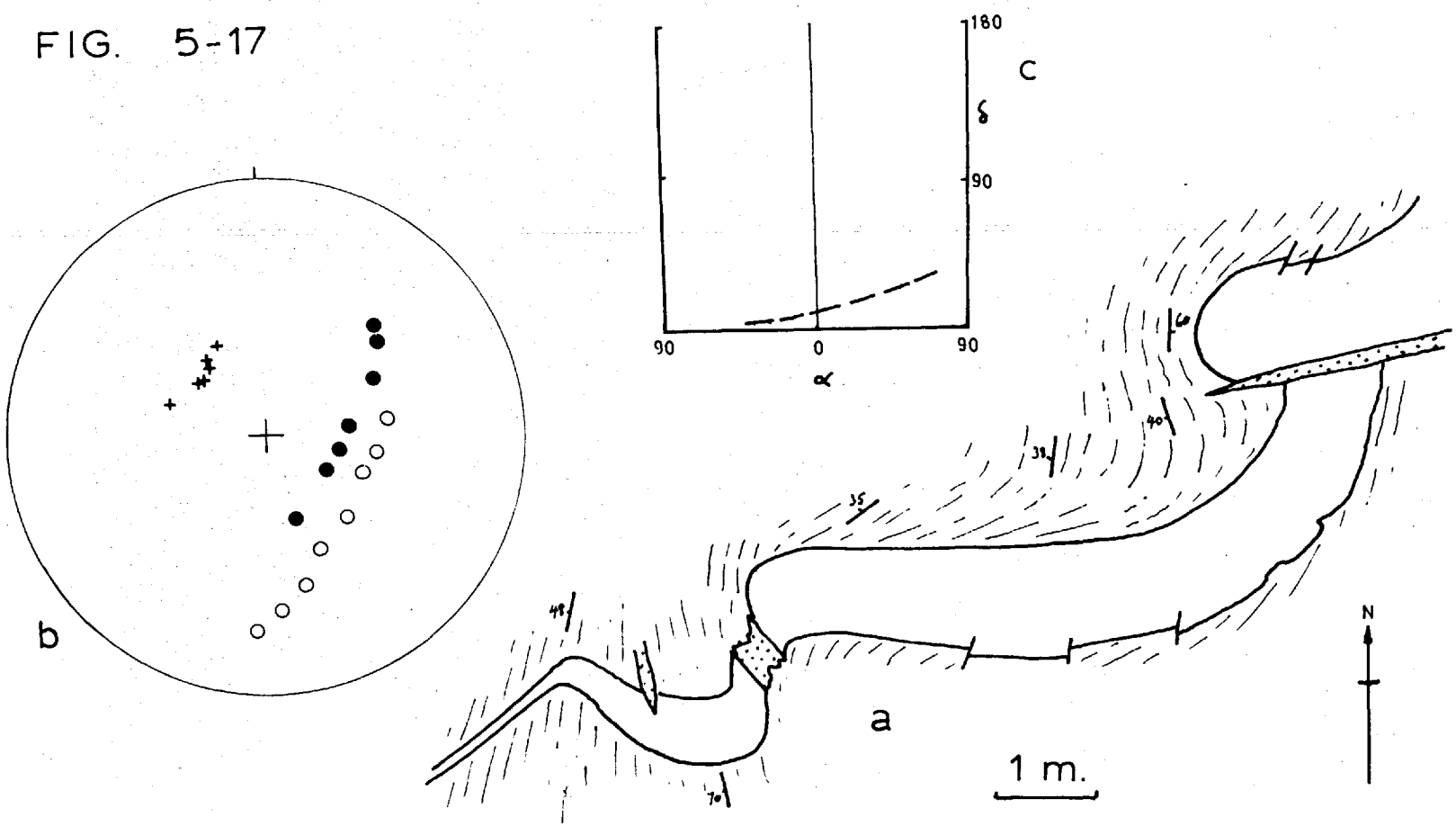
F_3 deformation close to A12 Ardivachar.

a) Map of the deformed dyke. Pegmatite is shown stippled

b) Stereographic plot of poles to the gneissose foliation (solid circles) and the dyke (open circles) showing that they are folded about slightly different axes. The poles to the foliation do not trace out a great circle, but trace out a more complex surface. The spread of lineations in the gneiss is shown by the group of small crosses.

c) Plot of the dihedral angle (δ) between dyke and gneissose foliation.

FIG. 5-17



zones of more intense deformation, which are very similar to structures elsewhere called "shear zones", are not produced by a mechanism of simple shear, but largely by heterogeneous flattening strains.

5.10 The f_4 deformation phase

As with the previous deformation phases, the striking feature about this phase is the heterogeneity of strain; the deformation is confined to a few narrow belts of folding. The phase is characterised by folds striking east - west (from north-west - south-east to south-west - north-east), with a variable dip, from 10° to 90° to the south. They locally form open fold interference patterns with the earlier deformation phases (for example between A9 and A10 and near G9), and generally fold the earlier linear fabric.

The deformed lineation usually traces out the form of a complex surface, (Figs 5.18), a form between a cone and a cylinder. This occurs even in areas of fairly gentle warping, for example at west Ardivachar, where a component of flattening following the buckle is not to be expected. The complex lineation form is probably due to the fact that the linear fabrics lie on a foliation surface which is oblique to the dykes, and has therefore suffered a component of oblique flexural slip or tangential longitudinal strain, as well as rotation round the fold axis.

At east Ardivachar, in North Bay, the f_4 folds are much tighter, almost isoclinal, and have low dips to the south. It is the foliation itself which is folded here, and the lineation traces out the form of a modified small circle, in this case probably due to the intersection of a localised homogeneous strain superimposed on the buckle fold.

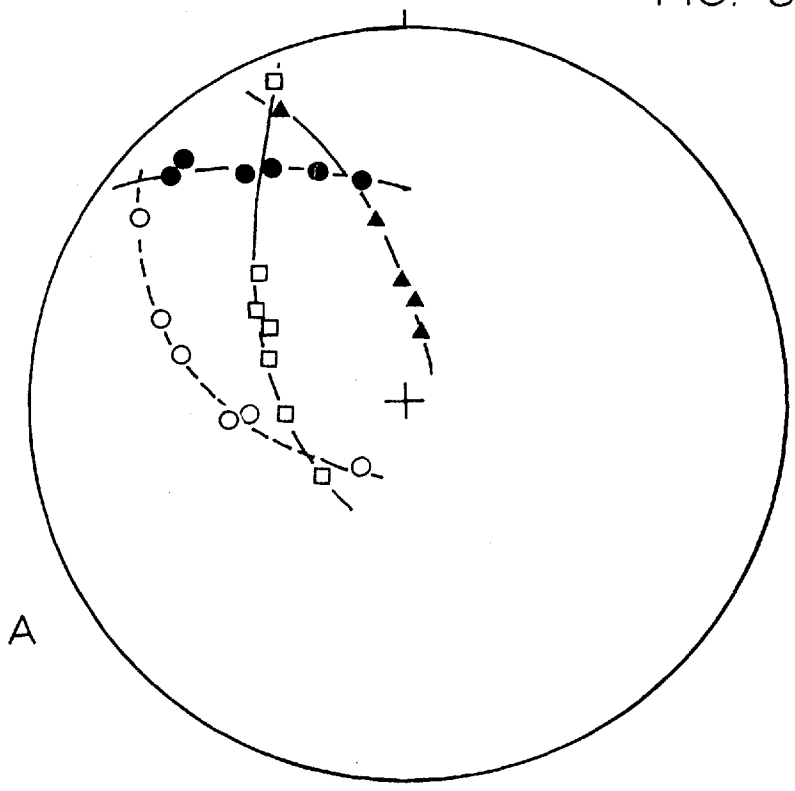
Fig. 5.18A

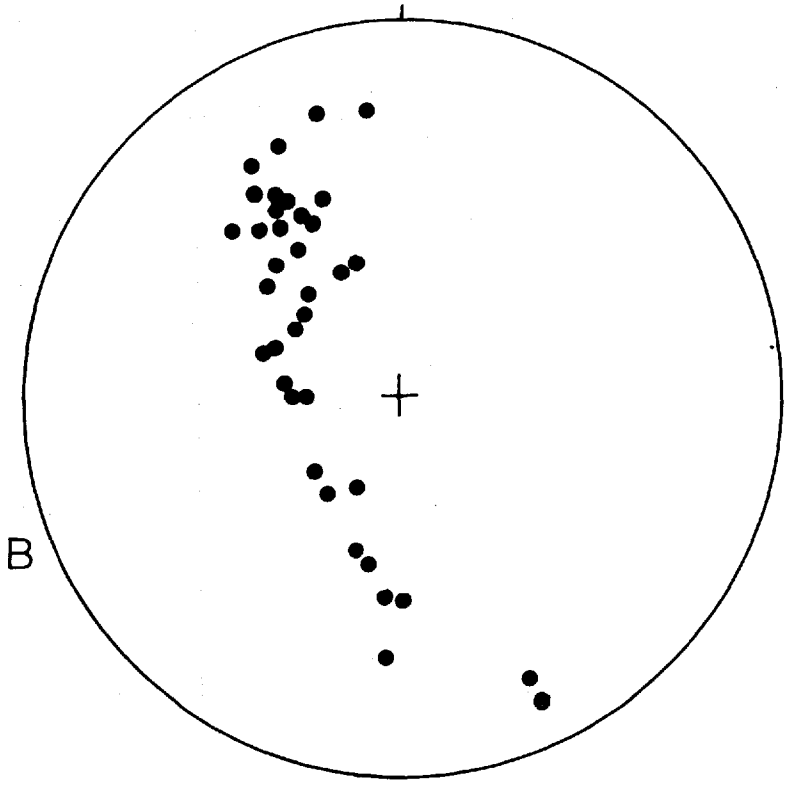
A plot of four groups of deformed lineations from close to A9, Ardivachar. The loci of the deformed lineations trace out the forms of complex surfaces, part-way between cones and cylinders.

Fig. 5.18B

A plot of the deformed lineations from North Bay, east Ardivachar.

FIG. 5-18





At north-east Ardivachar within one of these fold zones, part of the gneissose foliation strikes north-west - south-east and is folded by low dipping f_4 structures striking east - west. The result is that the foliation appears to be only slightly warped, but the lineation appears to be quite intensely deformed on the foliation surface.

5.11 The structures at the edge of areas of low deformation

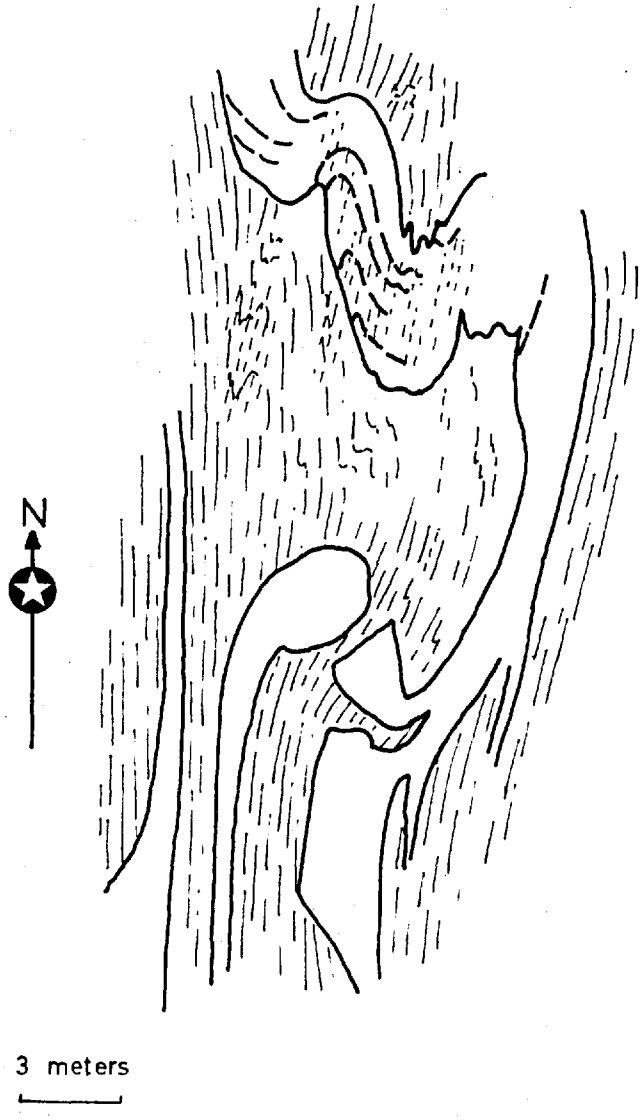
In this section, some of the structures adjacent to the low deformation zones will be briefly examined. The actual edge of these areas is usually obscured by beach sand or peat bog except in north-west Benbecula where the transition is fully exposed on the coastal strip (Fig. 5.1). In this region the edge of the low deformation belt is marked by a sudden rise in the amount of f_2 strain (Fig. 5.1). A dyke can be traced northwards from an area where it is strongly discordant, to an area where it suddenly bends round into full concordance with the gneissose foliation. This edge is marked also by the development of a strong planar fabric in both dykes and gneiss. This rapid increase in the amount of f_2 deformation appears to have caused a certain amount of slip at the edge of the low deformation block, which is now marked by a plane of shearing and quartz veining. The whole transitional zone is only a $\frac{1}{2}$ metre wide.

Such sudden transitions have been recorded from the f_1 phase at Garry-a-siar but are unusual in the f_2 group of structures. There is however a rapid local variation in the amount of f_2 strain about 50 metres to the south of the above mentioned locality (Fig. 5.1) where a transition from a relatively weak f_2 strain to f_2 strains in the order of between 8 and 9 : 1 (as measured from fold profiles in dyke apophyses).

Fig. 5.19

Deformed dyke at Peinavalla (762520), south of Garry-a-siar. The dyke is folded by f_2 structures but still shows discordant relationships. Note the variation in dihedral angle between the dykes and quartzo-feldspathic veins (heavy dashed lines). The Laxfordian f_2 fabric is represented by the thin dashed lines.

FIG. 5-19



takes place again within a half metre.

In North Bay, Ardivachar, the actual transition is obscured, but must take place in less than three metres; from relatively unstrained rocks, to rocks showing almost isoclinal f_2 folds. South of Garry-a-siar, the transition is again obscured. In the rocks of Peinavalla (762523) the dykes still show discordant relationships but also demonstrate well-developed f_2 structures (Fig. 5.19).

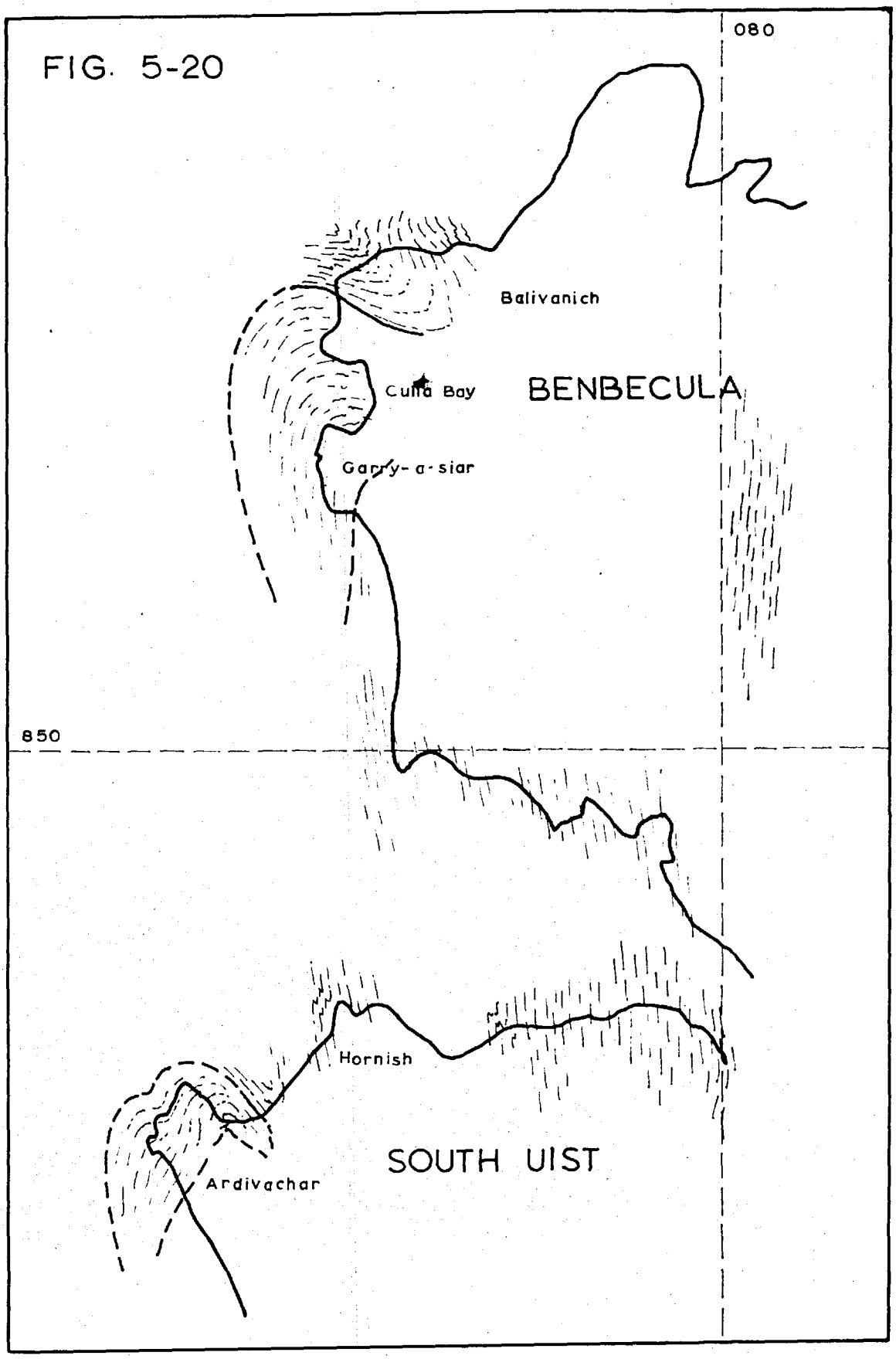
Outside the low deformation zones the full succession of Laxfordian structures can be recognised. It is difficult to trace each of the individual fold sets across the transition, but from corresponding fold sequences and fold orientations one can make fairly certain correlations. As I have shown above, the f_2 structures can be traced across quite easily, and at North Bay, Ardivachar, the f_4 set of structures can be traced across the transition into the high deformation zone. Since one can correlate the f_2 and f_4 phases one can be relatively certain about correlating the intervening f_1 and f_3 fold phases. In fact, correlations between structures in the areas of high and low deformation are often easier than similar correlations between structures in different regions of high deformation.

On the north-west coast of Benbecula, the most important deformation phase as far as the present structure is concerned is the f_3 phase which forms a complex fairly tight synformal region. Fig. 5.20 shows the relationship of the areas of low deformation to these f_3 folds, both at Ardivachar and on Benbecula. The areas of low deformation form open rounded f_3 antiforms and are separated by synformal regions of higher f_3 deformation. It may be that these areas of low deformation were folded en masse, and instead of being homogeneously affected by the f_3

Fig. 5-20

Map of the northern part of South Uist and Benbecula showing the form of the f_3 structures in the regions adjacent to the areas of low deformation. The areas of low deformation are shown bounded by the heavier lines.

FIG. 5-20


080
Balivanich
Culla Bay
BENBECULA
Garry-a-siar
850
Hornish
SOUTH UIST
Ardivachar

deformation, themselves influenced the shape and position of the f_3 structures. Prior to the f_3 deformation they appear to have had a lensoid shape, the long axis of which was orientated in the f_2 X - Y plane. This lens then acted as a competent plate, controlling the shape of the f_3 fold structure.

5.12 Low deformation areas elsewhere in the Uists

Variations in f_2 deformation are not confined to the coast sections described in this chapter. Similar variations are to be found in several other areas, where the f_3 or f_4 strain is locally low enough to allow one to observe the earlier strain history. Such variations have thus been found on the island of Wiay, off the east coast of Benbecula, where branching and only slightly folded dykes show strong discordance with thegneissose foliation (Fig. 5.21a), and also patchily within the area south of Loch Eynort. Fig. 5.21b shows a sketch of such a dyke on the eastern slopes of Trinival.

It was mentioned in the previous section that the low deformation zones at Ardivachar and Garry-a-siar are folded en masse by the f_3 structures. They were platy or lensoid structures and so responed to the f_3 strain by producing large scale fold shapes. A similar mass of slightly deformed rock at Reineval (755255), however, originally appears to have had a more circular section. This structure has responded to the f_3 deformation by remianing relatively stationary, its weak foliation remaining with a north-east - south-west strike, while the strongly planar gneisses outside rotated into a north-west - south-east orientation. In this respect, therefore, it shows a structural pattern similar to many so-called "rotated" garnets in schistose rocks (Ramsay, 1962), or to the large igneous bodies of South Harris (Graham

Fig. 5.21

a. Sketch map of an almost undeformed dyke from Wiay, Benbecula.

b. Map of an apparently undeformed branching dyke, east of Trinival (785263).

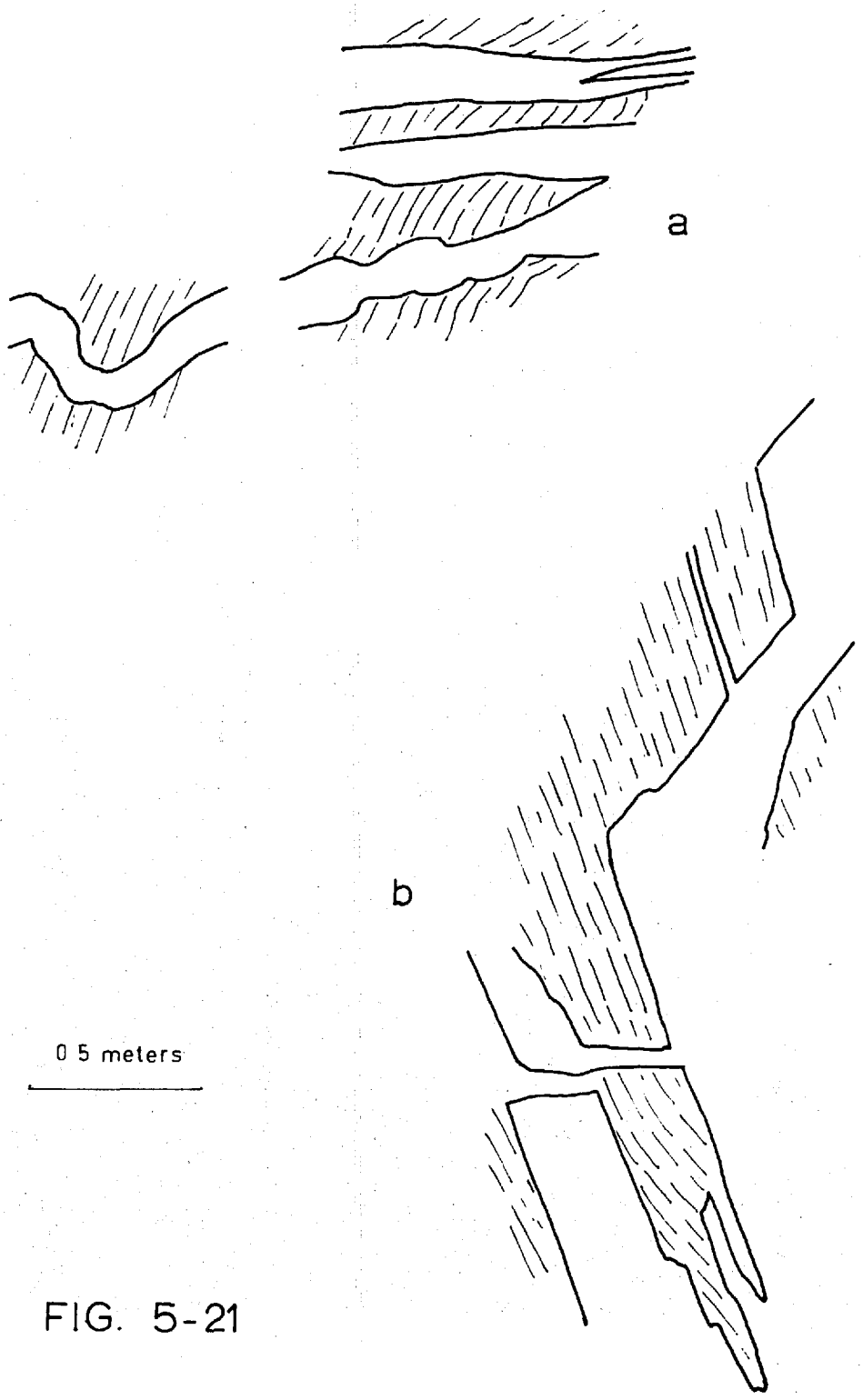


FIG. 5-21

1969).

5.13 The origin of the low deformation areas

Before discussing the origin of areas of low deformation, one has first to define what one means by this term. If one takes the Ardivachar and Garry-a-siar examples as type cases, then areas of low deformation will be defined as areas where in each phase of Laxfordian deformation the amount of strain has been relatively low. I have shown however that these areas were in existence as corporate units prior to the f_3 deformation and they tended to modify the shape of the f_3 folds. Their origin, therefore, must lie in the f_1 and f_2 phases.

I have listed below the most relevant points in this discussion.

1) All the areas have suffered a certain amount of f_1 and f_2 strain.
2) Both these phases have variable amounts of strain. The variations may be extremely rapid, or they may be more gentle
3) The rapid variations in strain tended to occur when and where certain planar units originally had an orientation closely parallel to the X - Y strain plane. Thus the f_1 deformation phase is characterised by very rapid changes in amounts of deformation, the high deformation concentrated along dykes orientated in the f_1 X - Y plane. The f_2 strain variations are usually more gentle, possibly because there were very few planar competence changes in the form of dykes parallel to the f_2 X - Y plane. In the cases where this was so, however, as in north-west Benbecula, then the amount of f_2 strain shows very rapid variations.
4) There are also variations of strain within the high deformation zones.

There is no particularly good evidence that the areas of low

deformation lie in specific low strain positions on very large f_2 folds.

5) A point made in chapter III is that there is no apparent difference in mineralogy between the areas of low and high deformation.

Heterogeneous deformation is a common feature of all basement geology. By taking examples from many different areas, one could produce a whole spectrum of deformation patterns, from that shown by the Scourie-Laxford area (Sutton and Watson, 1951, 1962), where a few strips of high deformation sit in the middle of a general low deformation region, to that shown by South Uist, where there are a few areas of low deformation sitting in a general high deformation region. I would suggest that these lie at different points along a general deformation sequence.

I would suggest that in basement rocks in early stages of a tectonic phase, deformation was concentrated in two areas:

(i) where there was a lower viscosity, enabling a higher strain rate to be attained.

(ii) where there were discontinuities, especially parallel to the X - Y plane (or the plane of maximum shearing strain) of the strain ellipsoid

Within normal cover rocks, heterogeneities are usually confined to planar sedimentary layers or to single massive igneous bodies. These rocks respond to these heterogeneities by producing buckle folds in the layers, or producing zones of high deformation around the igneous bodies (see also Chapter VIII). In basement rocks such heterogeneities would be much more irregular. In the Uists, for example, heterogeneities in the amount of strain could be caused by a local preponderance of competent dykes at a high angle to the X - Y strain plane (or to the

$\sigma_1 - \sigma_2$ stress plane?). These bodies would cause local strain shadows in the adjacent gneiss, but because of the irregular nature of the dykes, the original bifurcation and the original thickness variations, these strain inhomogeneities would be very patchy. If there were regions where dykes were closer to the X - Y plane, then higher strains would be possible, and adjustments in the stress and strain trajectories in the rock would concentrate the strain here. In many ways, therefore, these planar concentrations may be similar to the gleitbretter or micro-lithon structure which is found on a much smaller scale, where hig deformation is concentrated along the limbs of small folds, i.e. the planar discontinuities closely parallel to the X - Y strain plane. I would suggest, therefore, that as a result of the early strain heterogeneities, this basement area could be considered as a large scale example of an augen gneiss, with pods or bands of low deformation separated by zones of slightly higher deformation.

As I have shown, we are dealing with areas which have suffered syn-tectonic metamorphism. The areas which suffered the most deformation would therefore be susceptible to more metamorphic changes, to mineral changes and to changes in the relative viscosities of the rocks, enabling therefore more deformation. It would appear that we are dealing with a "strain softening" process, in that areas which first produce higher strains continue to do so. To a certain extent one can see this in the way that the more deformed parts of the dykes change their mineralogy and become less competent. I would suggest that strain softening took place to a much wider extent within the gneisses of South Uist , thus enhancing the zones of heterogeneous deformation.

CHAPTER VI

THE AREAS OF HIGH DEFORMATION

The so-called areas of high deformation occupy most of the Uists, and as such have been described in a fair amount of detail in Chapter II. Many points concerning the deformation history of these areas have also been discussed in the previous chapter, especially those concerning the earlier Laxfordian phases. In this chapter, therefore, I shall concentrate on the later Laxfordian fold episodes and a discussion of the earlier deformation phases will be largely confined to a more detailed description of two small areas, in north-west Benbecula and at Sgeir Dremisdale, South Uist. These areas were chosen partly because they were beside the sea and partly because they amply showed the relationship of the early structures to the f_3 and f_4 fold phases.

6.1 The early Laxfordian deformation phases

6.1A North-West Benbecula

In this section, I intend first to examine the structure close to the low deformation region, taking specifically the north-west Benbecula coast (Fig. 6.1) as here one can recognise most of the Laxfordian deformation phases and one has a continuous exposure from high to low deformation areas.

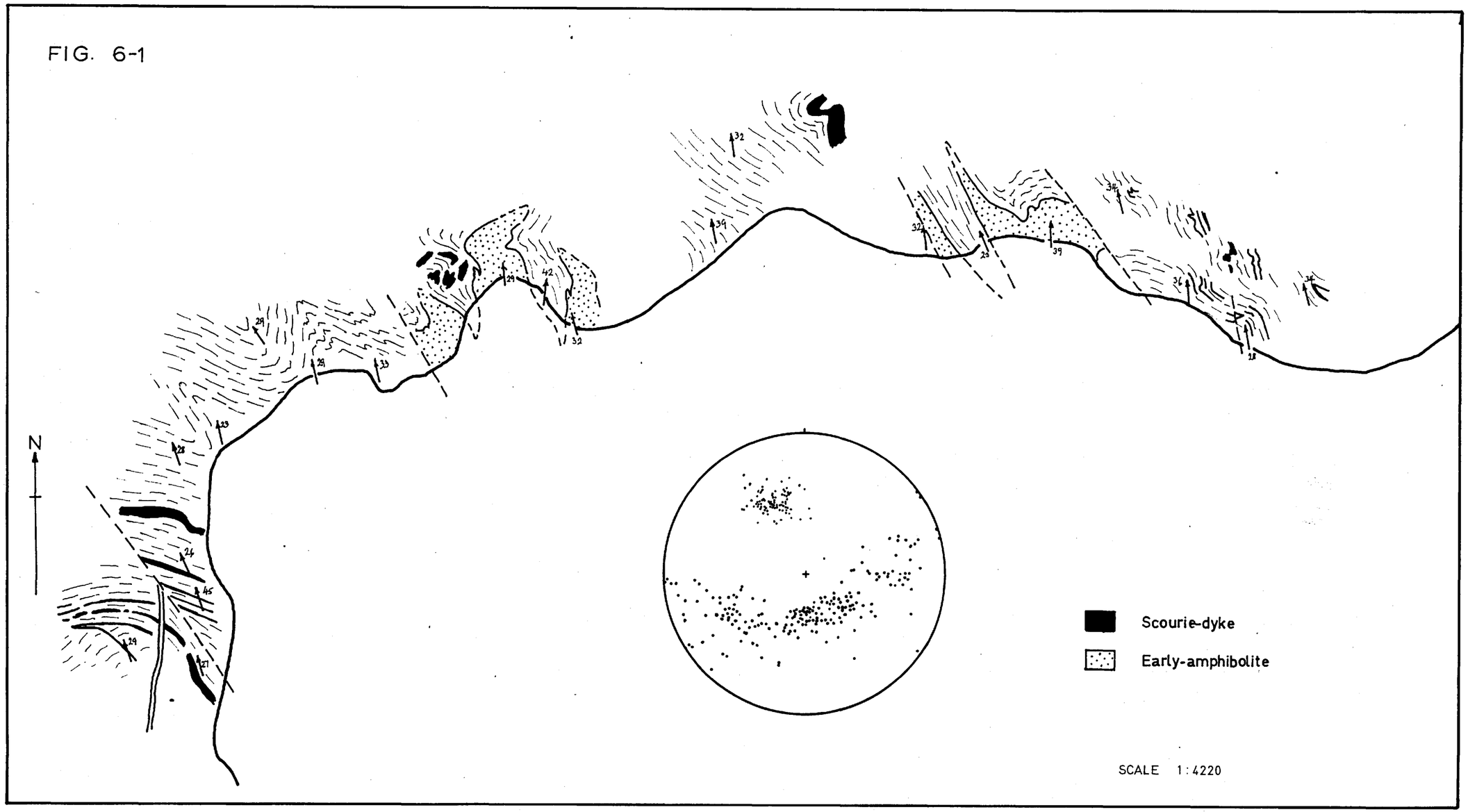
Within many of the Scourie-dykes in this region, the f_2 folds fold an earlier f_1 fabric, and there are also a few fold structures presumably of f_1 age. One cannot however make any estimate of the intensity of the f_1 strain, nor of how widespread was the f_1 deformation.

The f_2 strain is generally high and the edge of this region of high

Fig. 6.1

Map of Northwest Benbecula and stereographic plot of the fold and lineation axes and foliation poles.

FIG. 6-1



deformation is marked by banded gneisses made up of Scourie-dykes and acid gneiss. A well-developed planar fabric can be found in many of the dykes and is often emphasised by quartzo-feldspathic veining and incipient migmatisation.

Within the zone as a whole, very large strains can be recognised. The folds show a high component of flattentng strain, generally in the order of $1 + e_2/1 + e_1$ of between 0.4 and 0.2 (i.e. $1 + e_1/1 + e_2$ of between 2.5 : 1 and 5 : 1), as shown in Fig. 6.4, and locally f_2 flattening strains of over 10 : 1 occur. These values are taken from folded quartzo-feldspathic and amphibolite layers, from an early basic rock which is folded into a succession of anticlines and synclines, and thus recurs regularly along the coast (Fig. 6.1). From an analysis of the folded Scourie-dykes, an idea of the total f_2 strain can be obtained. The dykes are often ptygmatic and the limbs are often almost sheared out leaving the hinge floating in the acid gneiss. This can be seen on all scales in large dykes several metres thick (see Fig. 6.2) and in dykes only a few centimetres wide. Flattening components of up to 10 : 1 have been found from some folded dykes and very high total strain values of $1 + e_1/1 + e_2 = 36 : 1$ have been obtained.

6.1B Sgeir Dremisdale

The f_2 structures have also been examined in detail in a small section of rock exposed on the coast at Sgeir Dremisdale, South Uist (749370), where again all four Laxfordian phases can be seen. Once again, though, the f_1 phase can only be recognised from a folded foliation within the dykes. No f_1 fold structures can be identified with any certainty.

Fig. 6.2

Detailed map of part of the North-West Benbecula coast showing dykes deformed by f_2 structures, trending north-east - south-west and f_3 structures, trending east - west. The Scourie-dyke mass in the centre shows an f_1 foliation folded by f_2 structures. The stipple indicates areas of dominantly linear fabric.

FIG. 6-2

Fig. 6.3

Diagram showing the probable form of the f_2 structures in the area covered by Fig. 6.2. During the f_3 deformation phase, the f_2 fold hinges suffered relatively little body rotation and most of the f_3 strain was taken up by the gneiss.

FIG 6-3

The f_2 deformation produced isoclinal folds and boudinaged the dykes on the limbs of these structures. The boudinage is of the chocolate tablet kind, i.e. there was extension in two perpendicular directions and the strain was presumably of the type $1 > k > 0$. Some of the dykes show internal boudinage, the garnet amphibolite layers being more competent than the normal amphibolites. Mineral banding and incipient migmatisation developed on the limbs, along what was probably an f_2 planar fabric, and in areas where the f_3 deformation has only rotated the boudinaged blocks and not caused much apparent deformation, then planar fabrics can still be recognised. Most of the dyke fabric, however. is linear, as a result of the superimposition of the f_3 strain on the earlier fabric, and in most instances the original nature of the f_2 fabric can only be surmised.

It is difficult to estimate the amount of f_2 strain because of the high f_3 deformation. The f_2 folds are often close to similar, of classes 1c or 3, approaching class 2.

6.1C South Uist as a whole

Apart from the regions mentioned in an earlier section (5.12), the f_2 strain in South Uist as a whole is variable, but generally high. Folded quartzo-feldspathic veins in early amphibolites give post-buckl flattening strains of over 8 : 1 and from some more detailed fold measurements on some of these veins, total strain values of 23.5 : 1 have been obtained.

The amount of flattening for 44 folds is given in Fig. 6.4. Variations in amounts of flattening within these f_2 folds appears to be due to local viscosity contrasts and to the relative position in the

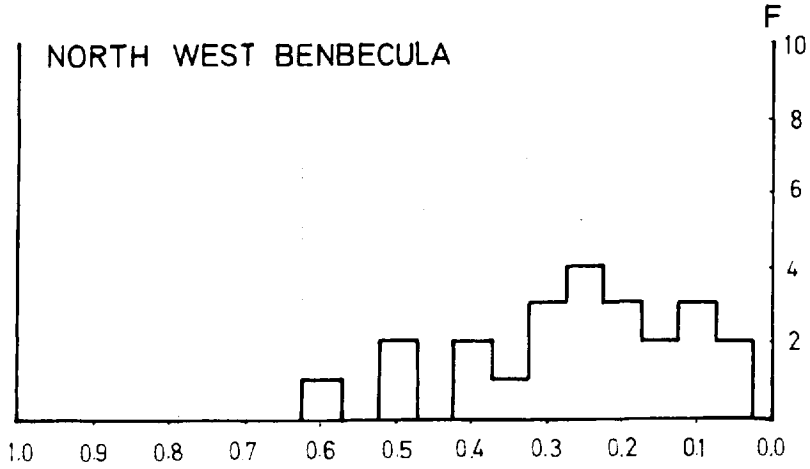
Fig. 6.4

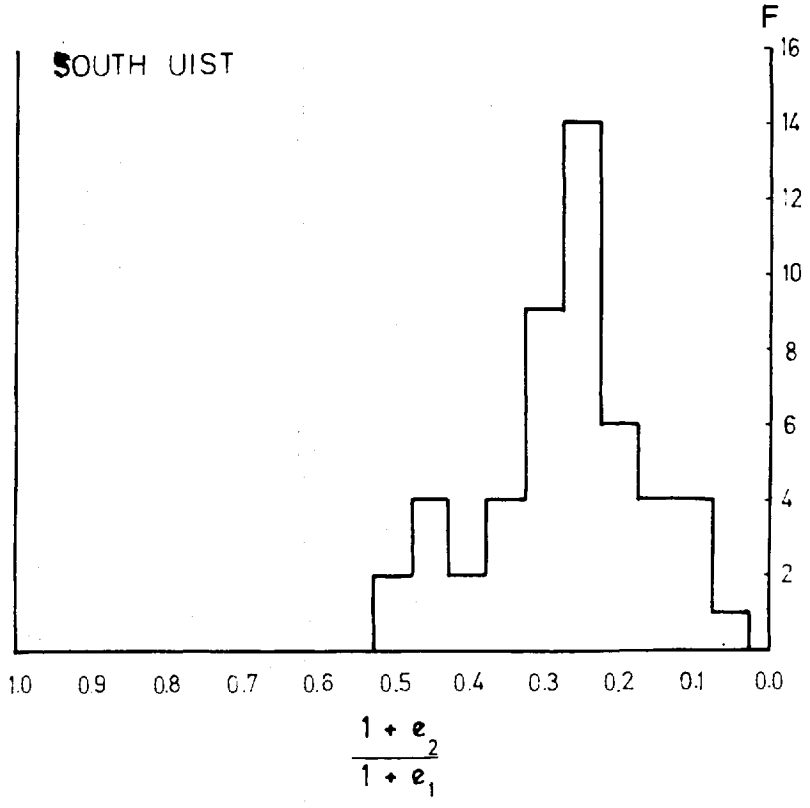
Histograms showing the amount of post-buckle flattening

on 23 f_2 folds from North-West Benbecula

44 f_2 folds from South Uist.

FIG. 6-4


NORTH WEST BENBECULA
F
10
8
6
4
2
1.0
0.9
0.8
0.7
0.6
0.5
0.4
0.3
0.2
0.1
0.0


SOUTH UIST
F
16
14
12
10
8
6
4
2
1.0
0.9
0.8
0.7
0.6
0.5
0.4
0.3
0.2
0.1
0.0
$\frac{1 + e_2}{1 + e_1}$

multilayered fold system.

From fold analyses, one can obtain an idea of the relative viscosities of the different layers. I have found that, for the early phase of folding, the following competence sequence holds:

pure quartz or pure feldspar > garnet-amphibolite(pyroxene) > garnet-amphibolite > amphibolite (without garnet) biotite amphibolite > biotitic acid gneiss. Hornblendic acid gneiss is sometimes more competent than normal amphibolite, sometimes less.

Pure quartz, pure feldspar and pure pyroxene rocks appear therefore to be the most competent. The presence of hornblende decreases the competence somewhat, as does any mixed mineral assemblage, and biotite usually occurs in the least competent assemblages.

6.1D Conclusions

In addition to the conclusions given in the previous chapter, one can also add the following:

i) Over much of South Uist, the amount of f_2 strain is usually high, though variable. There is generally, however, too much f_3 deformation to make any quantative comparisons.

ii) The strain generally appears to be of the type $1 > k > 0$. The lineation was constant in orientation, as in the areas of low deformation, and the f_2 finite strain probably lies on the strain path given in section 5.6F.

iii) As I have shown in Chapter II, the f_2 plunge varies from north-west to south-east (Map 4) and this variation is presumably due to heterogeneities in the f_3 deformation. One can therefore only speculate

about the original strike and dip of the f_2 structures. However, if one makes the assumption that the f_2 plunge was re-orientated by equal amounts on either side of the plunge culmination, then the f_2 X - Y plane would originally be almost horizontal and the f_2 folds would be flat-lying structures.

6.2 The f_3 phase of deformation

In this section I would like to examine first the effects of the f_3 deformation in the two areas previously discussed in the section on the f_2 phase, namely in north-west Benbecula and at Sgeir Dremisdale. They appear to show strain variations typical of the f_3 phase in South Uist as a whole.

6.2A North West Benbecula

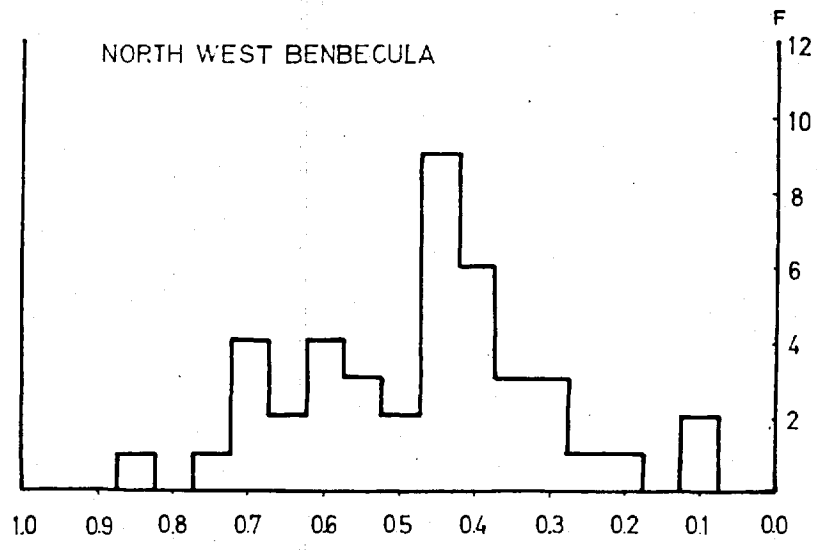
The structure of north-west Benbecula is shown in Fig. 6.1. It is made up of several tight f_3 fold zones, which, from their profiles, indicate a synformal structure to the north. This synformal zone lies on the northern flanks of the broad f_3 antiform formed by and within the Benbeculan region of low deformation. As I have suggested in section 5.11, this region of low f_1 and f_2 deformation appears to have controlled the f_3 fold form and wavelength. The general structure is composed of a large rounded antiformal domain with more highly deformed pinched-in synforms on either side.

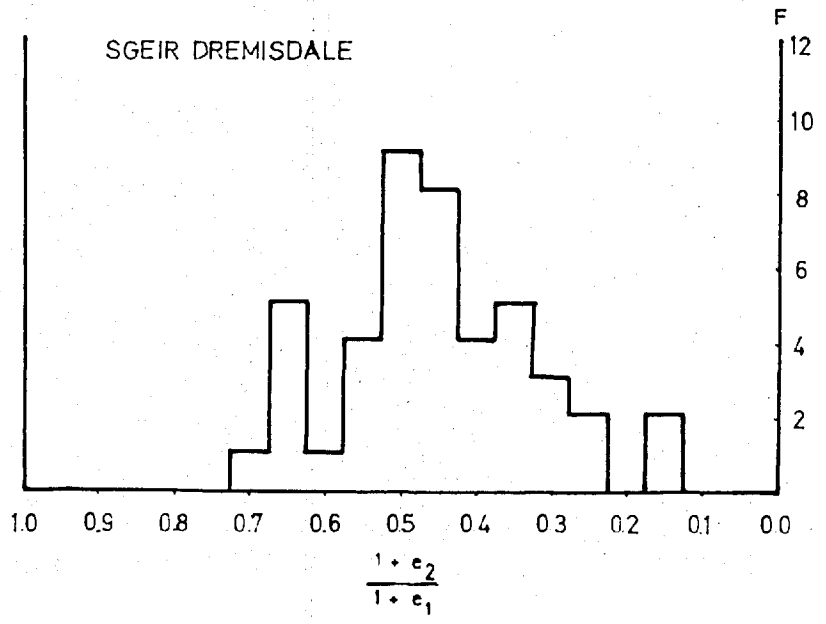
On a smaller scale, interesting structures are produced by the behaviour of the Scourie-dykes. The f_2 boudins and fold hinges often remain orientated in the f_2 X - Y direction, while the acid gneiss is deflected in such a way as to suggest that it was rotated around them

Fig. 6.5

Histogram showing the amount of post-buckle flattening on 43 f_3 folds from Northwest Benbecula and 44 f_3 folds from Sgeir Dremisdale, South Uist.

FIG. 6-5


NORTH WEST BENBECULA
F
12
10
8
6
4
2
1.0
0.9
0.8
0.7
0.6
0.5
0.4
0.3
0.2
0.1
0.0


SGEIR DREMISDALE
F
12
10
8
6
4
2
1.0
0.9
0.8
0.7
0.6
0.5
0.4
0.3
0.2
0.1
0.0
$\frac{1 + e_2}{1 + e_1}$

(Fig. 6.3 and 6.4).

The amount of f_3 strain is high, though it by no means compares with the f_2 strain values from this region. The folds are usually fairly tight and are made up of alternating layers of classes 1c and 3. The amount of flattening from 43 folded layers is given in Fig. 6.5. Onecan see that the amounts of flattening strain are varied, usually between 1.5 : 1 and 3 : 1, with a peak at 2.25 : 1. From folded Scourie-dykes, the total amount of f_3 strain can be estimated and this has values of between 7 : 1 and 13.5 : 1.

6.2B Sgeir Dremisdale

The structure of part of Sgeir Dremisdale is shown in Fig. 6.6. This area amply shows the effects of the Scourie - dykes and the f_2 structures on the form of the f_3 folds. The area exhibits a series of boudinaged f_2 fold limbs. When these boudinaged blocks were fairly closely spaced, then the f_3 deformation was highest close to the boudin neck infolds, but decreased away from these. The boudins appear to have been shunted together by the f_3 deformation and shortened largely by homogeneous strain or by gentle warping. They appear to have protected the adjacent gneiss from the highest f_3 strains, and close to these basic blocks the gneiss has a more planar appearance and early folds and early fold interference patterns can be found. When the boudins were not so close (see western part of Fig. 6.6), the f_3 structures became much more complex. The gneiss appears to have suffered more fold deformation and fold interference patterns abound, but the boudin blocks tended to act as individual units and played a relatively small part in developing the large scale fold pattern. Locally, however, they caused complex structures and variations in the amount of strain developed. Rounded

Fig. 6.6

Detailed map of part of the coast, south-west of Sgeir Dremisdale, showing the form of the folded and boudinaged Scourie-dykes. The stipple indicates an area of dominantly linear fabric. Late pegmatite is shown shaded.

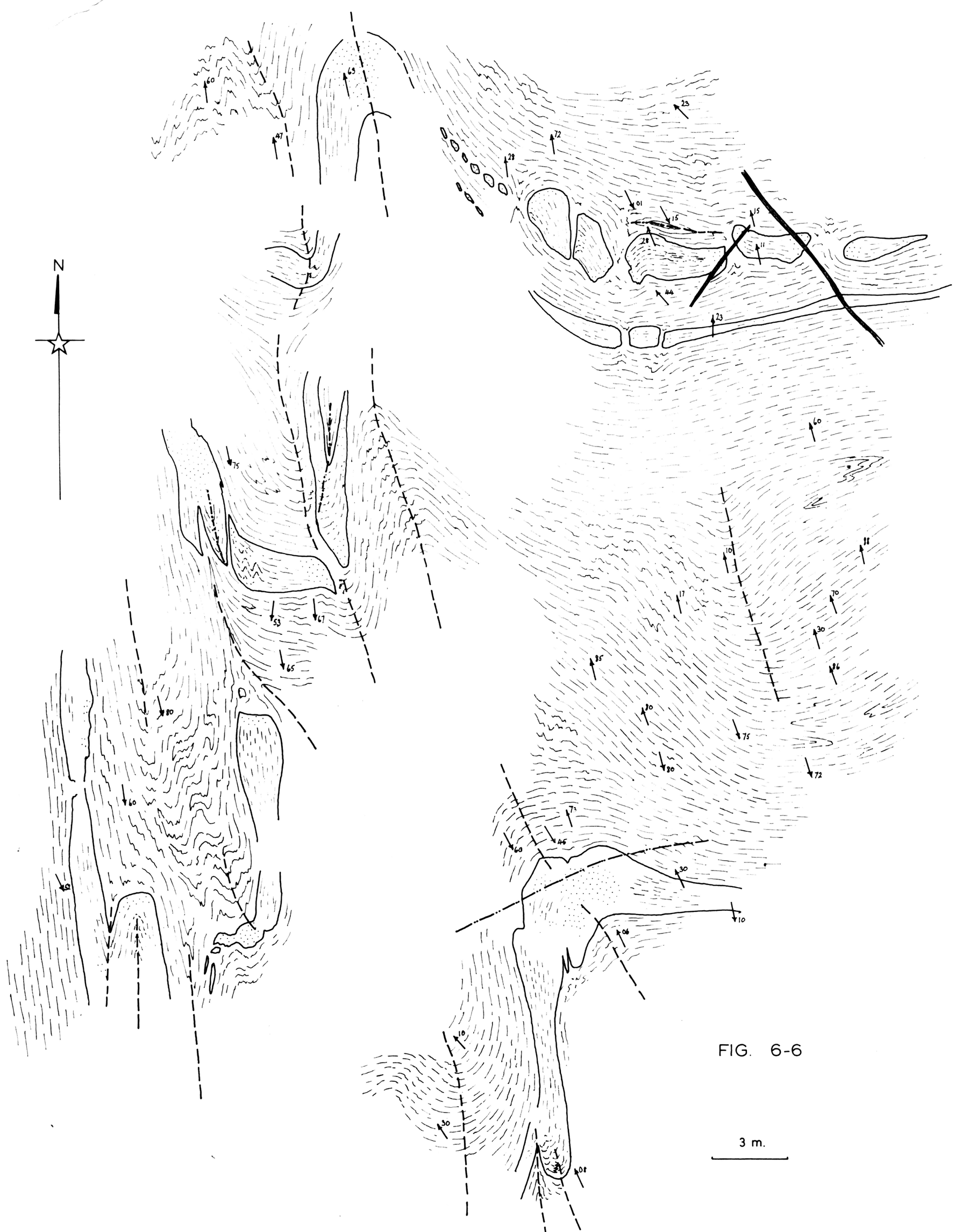


FIG. 6-6

folds alternate with pinched-in structures, depending on the closeness of the competent bodies. One can easily see these structural differences by comparing the different regions of Fig. 6.6 and by comparing these with the f_3 structures of Fig. 6.7, which shows an area where the gneiss consisted originally of only a banded.gneiss.

The country rock gneisses of parts of Sgeir Dremisdale contain abundant folded quartzo-feldspathic veins, from which an idea of the amount of strain can be obtained. Table 6.1 gives two-dimensional strain ratios and also the viscosity ratios obtained from 15 of such sets of veins. They are also given in a $\log_e$ form (i.e. natural strain) in Fig. 6.8. The f_3 strain is generally between 9 : 1 and 14 : 1.

6.2C South Uist as a whole

Most of the large f_3 structures if South Uist have been described in Chapter II and in this section I only wish to discuss briefly two points about these.

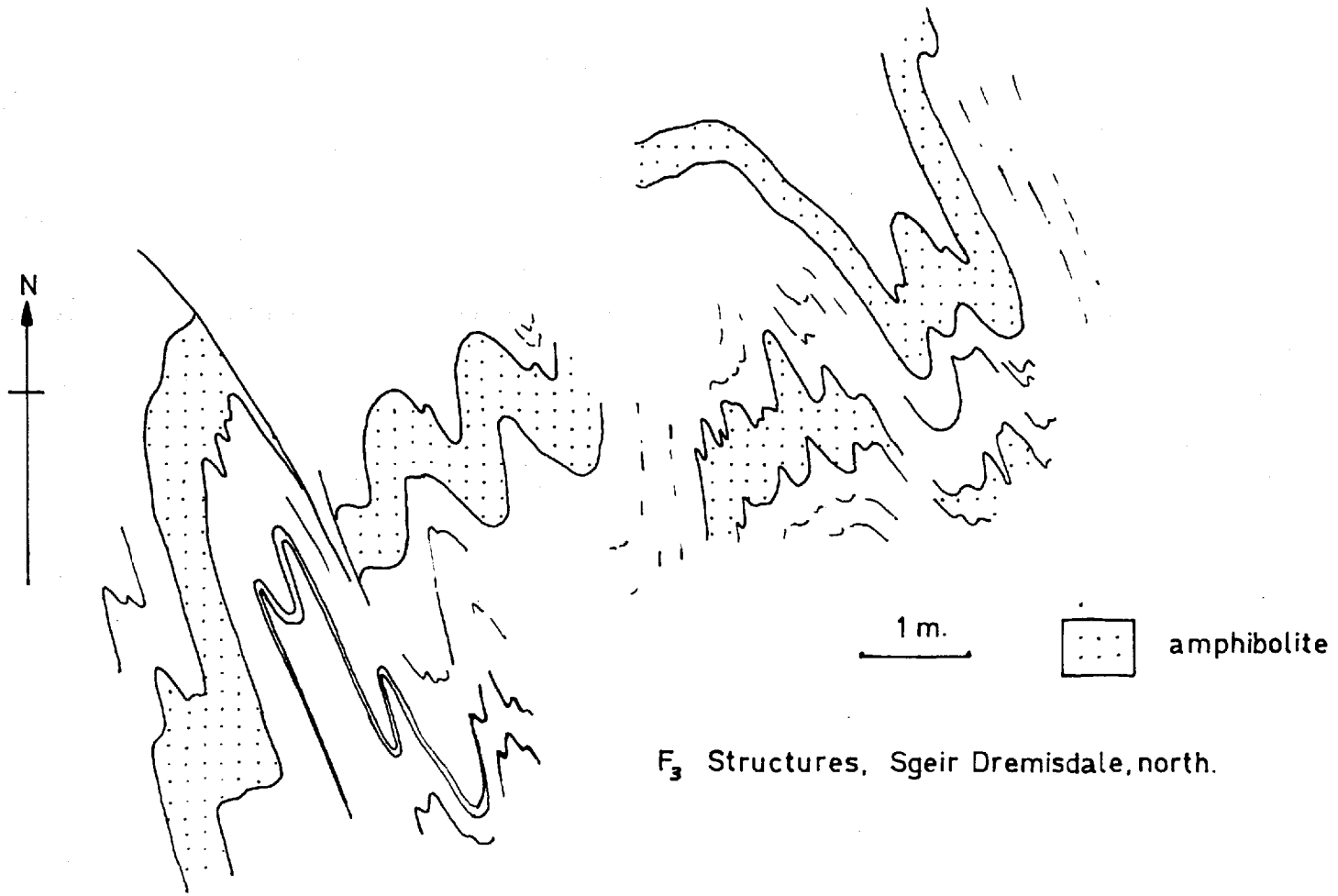
(i) The heterogeneity of the f_3 deformation. In the north, the f_3 folds are usually large open structures, and the rocks generally maintain their earlier fabrics. In the south-west, however, the structures are much tighter, often isoclinal, and strong new penetrative fabrics develop.

(ii) The fanning of the f_3 axial planes.

(i) The f_3 strain heterogeneity

The f_3 strain heterogeneity appears to be much easier to explain than that of the f_2 phase, and generally appears to be related to the internal deformation associated with the large-order structures. The form of these large-order structures is that of a large broad open

FIG. 6-7.



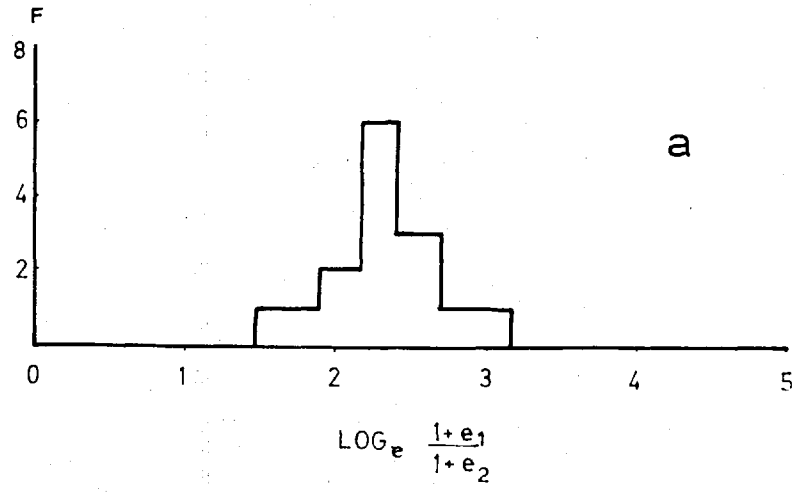
F_3 Structures, Sgeir Dremisdale, north.

Fig. 6.8

a) Histogram of natural strain ratios obtained from folded quartzo-feldspathic veins from Sgeir Dremisdale, South Uist.

b) Histogram of viscosity ratios obtained from folded quartzo-feldspathic veins from Sgeir Dremisdale.

FIG. 6-8


F
8
6
4
2
0
1
2
3
4
5
a
$\mathrm{LOG}_e \frac{1+e_1}{1+e_2}$

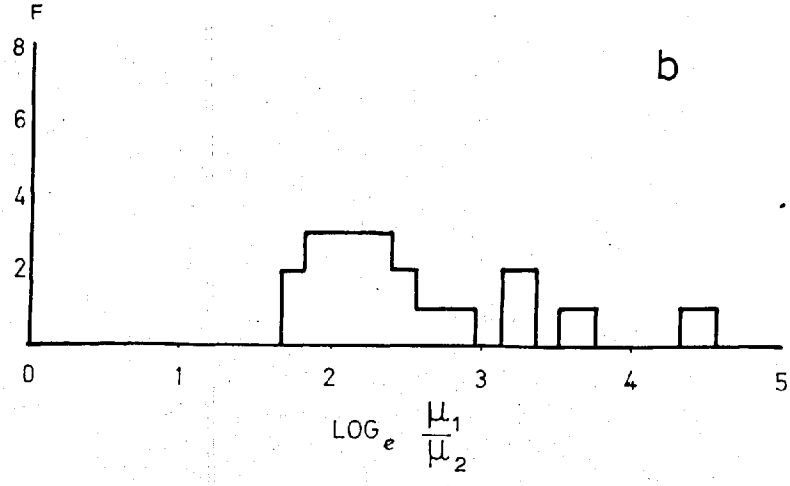

F
8
6
4
2
0
1
2
3
4
5
b
$\mathrm{LOG}_e \frac{\mu_1}{\mu_2}$

Table 6.1

Total f_3 shortening measured from folded veins, Sgeir Dremisdale, South Uist.

VEIN	AMOUNT OF FLATTENING	2 t/w SHERWIN & CHAPPLE	VISCOSITY RATIO	PRE-BUCKLE SHORTENING	SHORTENING DUE TO BUCKLE	TOTAL STRAIN	NATURAL STRAIN (LOG$_e$ STRAIN RATIO)
quartz/ feldspar	2:1	1.26	12	3:1	1.65:1	9.9:1	2.3
"	2.2:1	1.9	7	5:1	1.8:1	19.8:1	3.0
"	3:1	1.4	100	3:1	1.25:1	1.25:1	2.4
"	2.1	1.2	14	2.5	2:1	10:1	2.3
"	2:1	0.75	25	1.8:1	1.4:1	5.05:1	1.62
"	3:1	0.6	40	1.5:1	1.5:1	6.65:1	1.9
"	1.5:1	1.5	10	3:1	3:1	13.5:1	2.6
"	2:1	0.78	25	1.75:1	2.8:1	10.2:1	2.3
"	2:1	0.97	16	2:1	2.2:1	8.8:1	2.2
"	1.5:1	1.68	8	3.9:1	2.3 :1	13.5:1	2.6
"	1.5:1	2.3	6	6:1	1.6:1	14.5:1	2.7
"	1.7:1	0.425	80	1.5:1	3.2:1	8.25:1	2.1
"	none	1.71	8	4:1	2.7:1	10.8:1	2.4
"	none	2.3	6	6:1	2.1:1	12.6:1	2.5
Sc.dyke	2:1	1.55	9	3:75	none	7.5:1	2.01

antiformal domain in the north and a tight, upright pinched-in synformal region south-west of Loch Eynort. This synformal region can be recognised from the inward dips of the foliation on either side, and by the S and Z profiles of the f_3 folds (Maps 1 and 3). It is however difficult to be certain about the position of the major axial trace (if there is one major axial trace), though it would seem to stem from the Bornish region to the area north of Lochboisdale. These structures are shown on the series of sections I-J, K-L, M-N, Map 3.

This form of structure, characterised by tight cuspate synforms separated by broad lobate antiformal regions is a feature of many of the f_3 folds (cf. the previous section, 5.11, on the f_3 structures in areas of low deformation), and is a feature of all the f_3 large order structures in the Hebrides from Barra Head to the Butt of Lewis. It is the theme of a publication (Coward, Francis, Graham and Watson, 1970, in press), which compares the form of these structures with those produced experimentally (Ramsay, 1967, p.383; Cobbold, 1969) at the interface of two materials of different competence. It is also suggested that such structures are characteristic of other basement regions and that large basement domes of central Africa, or the mantled-gneiss domes of Scandinavia, may be of similar origin. In the Hebrides these structures could be produced by the buckling of either a thick layer of competent material which lies to the east of South Uist, or by the buckling of the interface between an eastern competent rock at low structural levels and the less competent rocks to the west.

Much of the evidence for this regional competence difference has been given in Chapter III, which mentions the presence of pyroxene bearing rocks, and of rocks rich in igneous bodies in the eastern part of South

Uist. Similar lithologies have been reported from Eastern Barra (Francis, 1969) from the core of the corresponding antiformal structure to the south, and further evidence has been given by a gravity survey of the Uists and the Minch (Annual reports of the I.G.S., 1964-68) which reveals a major positive gravity anomaly situated just to the east of the islands.

(ii) The fanning of f_3 axial planes

Within the tight synformal region, the f_3 fold axial planes are usually upright parallel structures, but to the north they tend to fan round the large antiformal domain (section 2.6E). I would suggest that the variation in orientation of these axial planes is due to the internal deformation produced by the large order fold structures. Before discussing this, however, some of the current theories on the origin of parasitic folds and on the orientation of axial planes within folds of different orders will be mentioned.

Ramberg (1963, 1964) has shown that parasitic folds can become asymmetric by the later amplification of the larger wavelength structures. The internal strains produced by the development of the large wavelength fold will further deform the small-scale folds, and change the orientation if their axial planes, and one can compute the variations in orientation of small-scale fold axial planes due to tangential longitudinal strains or flexural slip strains (Fig. 6.9).

If the large order folding was largely formed by flexural slip strains, assuming originally symmetric small scale folds, then the final angular divergence of the small scale fold axial planes will be small, and for low values of dip for the limbs of the large scale fold, the axial planes will remain parallel. The folds will become markedly asym-

Fig. 6.9

The variation in orientation of small scale fold axial planes when the rocks are subjected to folding of a higher order,

where a is the dip of the high order axial plane

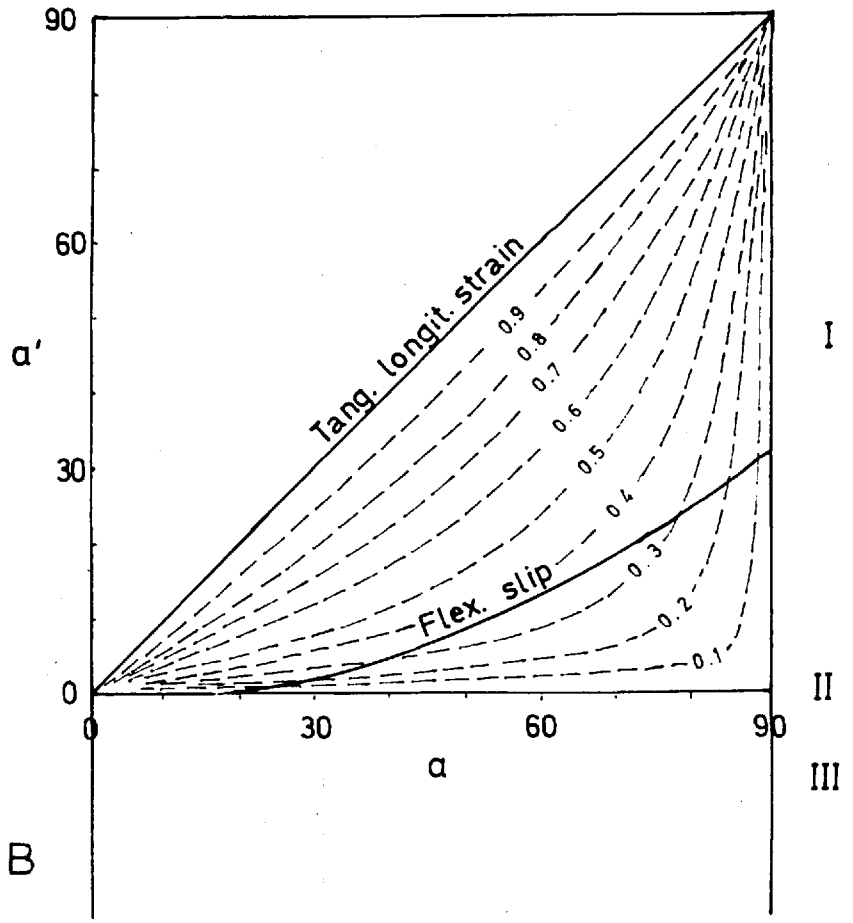
a' the angle between the low order axial plane and the high order axial plane.

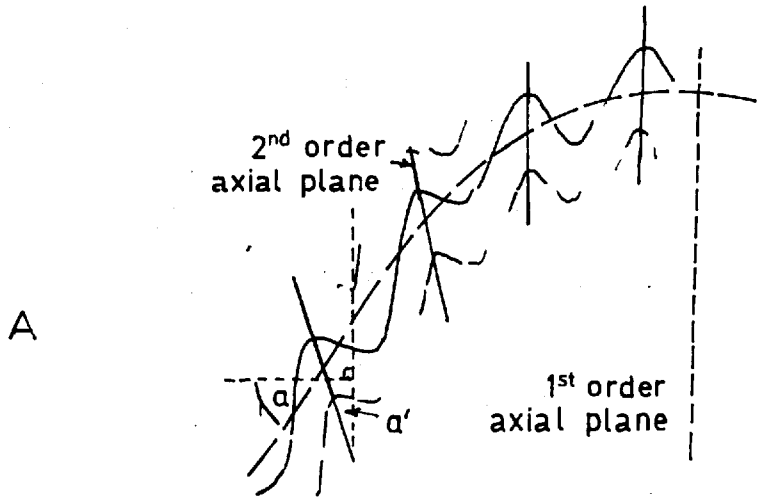
(see Fig. 6.9A)

The plots of a/a' are given in Fig. 6.9B. Three classes of high order folds occur, and these are equivalent to Ramsay's (1967) classes for individual fold layers.

The plots of a/a' for purely tangential longitudinal strain and purely flexural slip are shown. The thin lines show the plots of a/a' for purely tangential longitudinal strain plus components of homogenous flattening.

FIG 6-9


90
60
30
0
α'
Tang. longit. strain
0.9
0.8
0.7
0.6
0.5
0.4
0.3
0.2
0.1
Flex. slip
I
II
III
0
30
60
90
α
B


2nd order
axial plane
A
α
α'
1st order
axial plane

metrical, and the effect on the fold can be considered as a case of simple shear at a variable angle to the axial plane. The effect of this on the fold profile is the same as that of a fold suffering an oblique homogeneous pure shear and a rotation (Huddleston, 1969). Along the 'limbs' of the large order fold, the small scale folds should suffer an increasing amount of flexural shear with increase in dip, and small parasitic folds on the limbs therefore should show much higher amounts of flattening than those at the crest of the fold.

If the large order folding involved tangential longitudinal strain then the final angular divergences of the axial planes will be very large (Fig. 6.9), and the folds will remain more or less symmetrical throughout. The amount of deformation will depend on the tightness of the large order fold and the distance from the neutral surface. It will be greatest at the crest (or trough) on the inner arc of the large order fold structure.

The above discussion takes only the simple case where the small folds have formed before the amplification of the large order structure. For this to be valid, the "layer" forming the large order fold should have a low rate of amplification (generally equated with a low viscosity contrast).

Applying the above theory to the Uists, the f_3 folds appear to have been modified by a large component of tangential longitudinal strain. On the limbs of the large order structures, some folds, such as the Spin anti-form, show a certain amount of asymmetry, but this could be related more to the angle of intersection of the f_3 strains on the original foliation, rather than to a large order flexural shear modification. It must be mentioned here that the variations in amount of f_3 strain also agree with this hypothesis of large scale tangential longitudinal strain deformation, in that very high strains are developed in the inner arc, i.e. the cuspate

synform, while very low strains develop on the outer arc, the broad antiformal domain.

6.2D The variations in f_2 and f_3 plunge

These variations are shown in Map 4, and have been mentioned previously in Chapter II, section 2.6D. As I then showed, the f_3 axial plane, as well as the dip of the non-vertical foliation, changed systematically with the change in lineation plunge. This shows quite conclusively that the variation in lineation plunge is not due to variations in shape of the initial surface prior to the f_3 deformation. The variations, therefore, may be due to:

(i) a later phase of folding

(ii) heterogeneous strain during the f_3 phase of folding

(i) The later phase of folding.

The individual major f_3 folds are periclinal structures, trending north-west - south-east, plunging to the north-west and south-east. The line joining all the f_3 periclinal culminations, however, trends north-north-east - south-south-west, and if this change in plunge is due to a post-f_3, pre-f_4 fold phase, then the folds would also have to trend north-north-east - south-south-west. The trend of the vertical foliation and the trend of the vertical axial planes remains constant no matter what the plunge, and for the lineation to be folded upon this vertical foliation, the post-f_3 fold would have to be of the "similar" type (see Ramsay, 1967, pp 426 and 473).

(ii) Heterogeneous strain during the f_3 phase.

Ramsay (1967, pp 434-436) has shown how inhomogeneous strain must lead to differential flow and therefore to the production of folds in

layered rocks. The trend of the axial plane culminations in South Uist follows exactly that of the axial planes of the f_2 structures and the trend of much of the foliation prior to the f_3 phase. Heterogeneous f_3 deformation at depth, due to a higher strain developing in one of these structural units, would cause a "similar fold" to develop, which would follow the f_2 trend.

It would appear, therefore, that this f_2 structural trend was of fundamental importance in determining the later strain inhomogeneities. Most of the main Lewisian features follow this trend; the f_2 strain inhomogeneities, the f_3 enveloping surface, the f_3 axial culmination, gravity anomalies and the main thrust zone. It must be noted that most of the important structural features in North-West Scotland, including the Moine Thrust, have this orientation!

6.3 The f_4 phase of deformation

Anyone who examines the geology of South Uist (see Map 1) will be struck by the abundance of f_4 structures along the east side of the island. Immediately south of Loch Skiport, it is this f_4 phase which dominates the whole structural pattern. Yet on the west side of the island there are no major f_4 structures, and small scale structures are often few and far between. There would therefore appear to be far less f_4 deformation in the west than in the east, and in this section I should like to examine the other evidence for and the implications of such a deformation gradient.

6.3A F_4 axial plane trajectories

It is generally considered that a fold axial plane lies fairly close to the X - Y principal plane of that fold phase, and any curvature of the fold axial plane can often be considered as due to a later "rotation" effect

or to an initial curvature of the strain trajectory. From a plot of the f_4 axial plane orientations, one should obtain some idea of the variation in orientation of the f_4 strain axes. In South Uist, a pattern emerges (Map 5) in which the axial planes systematically tend to converge towards the east, especially in the Skiport area. This plot, I believe, reflects the variation in orientation of the f_4 strain trajectories in South Uist.

6.3 B Variations in fold geometry and in the amount of flattening on the folded layers

There is a change in style of the f_4 folds from west to east, partly associated with the eastward increase in deformation and also partly with the changes in syn-tectonic metamorphic conditions (see Chapter III). A rather subjective map has been made (Fig. 6.10) which splits South Uist up into several zones based on the field appearance of the folds, but backed up by a number of fold analyses. Localities from which fold analyses were made are shown on this diagram.

Zone 1

Within this zone there are relatively few small scale folds and large scale folds occur only as gentle warps. Folds of this generation have already been described from Ardivachar and Garry-a-siar (section 5.10) and it was noted that the f_4 deformation was highly variable and localised. Within this zone there was no syn-tectonic metamorphism or even major recrystallisation.

Zone 2

Small scale f_4 folds may be locally abundant within this zone (notably on Rueval (781401) or at Cnoca Breac (734342), though large order folds generally occur only as open warps. Within these areas,

Fig. 6.10

Map of South Uist to show the structural zones based on the f_4 fold geometry. Localities, from which analyses were made, are shown by the solid triangles.

FIG. 6-10

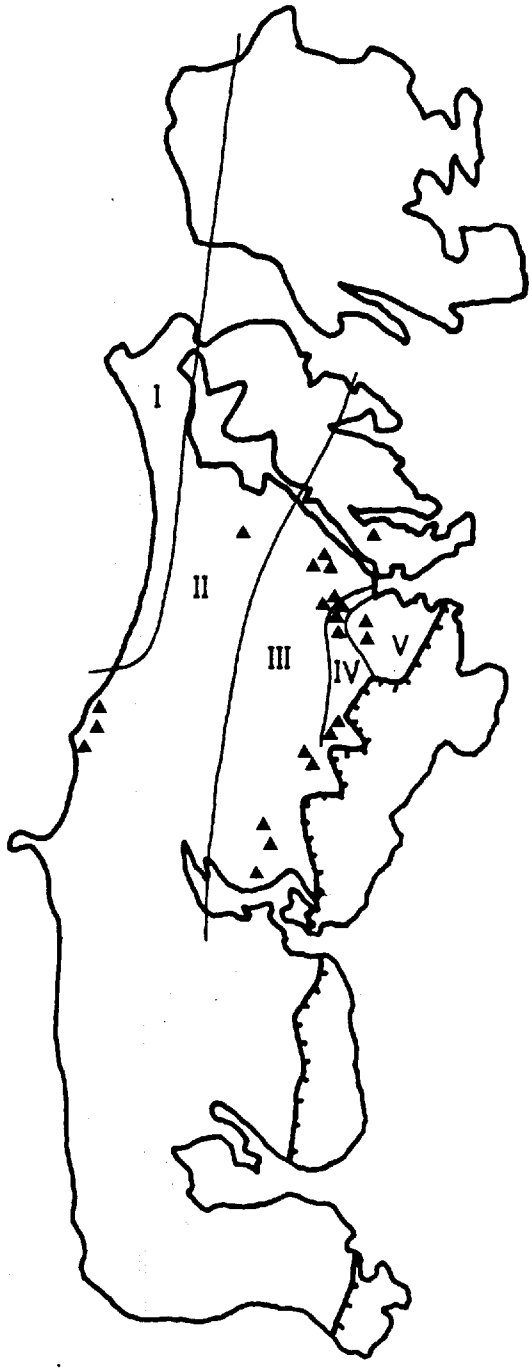
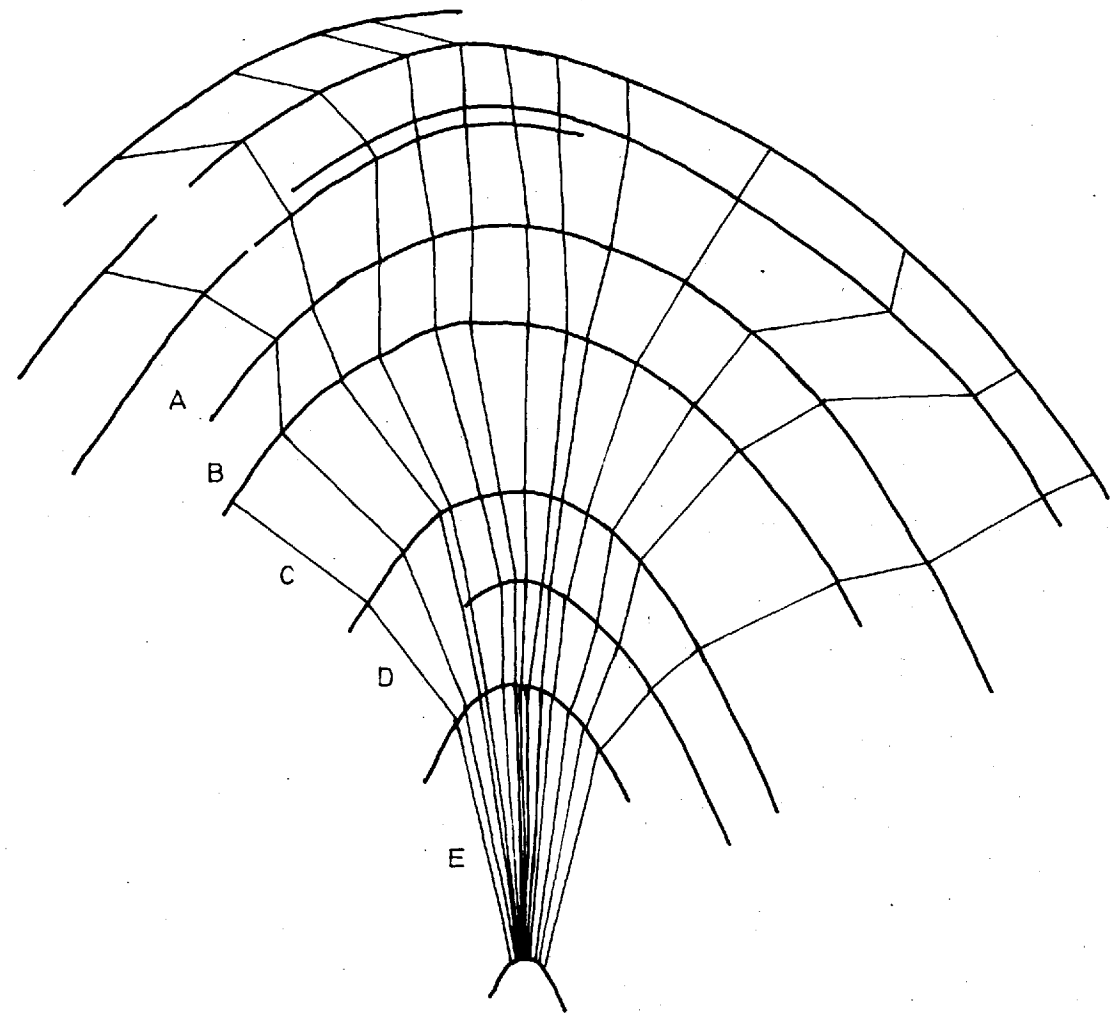


Fig. 6.11

An f_4 fold profile showing the effects of tangential longitudinal strain on the geometry of a multilayer fold system. The layers on the outer part of the fold form folds of classes Ia and Ib; the ϕ/α plots are made highly irregular by the initial thickness variations. The layers on the inner part of the multilayer fold form folds of class Ic with 25% flattening.

FIG. 6-11



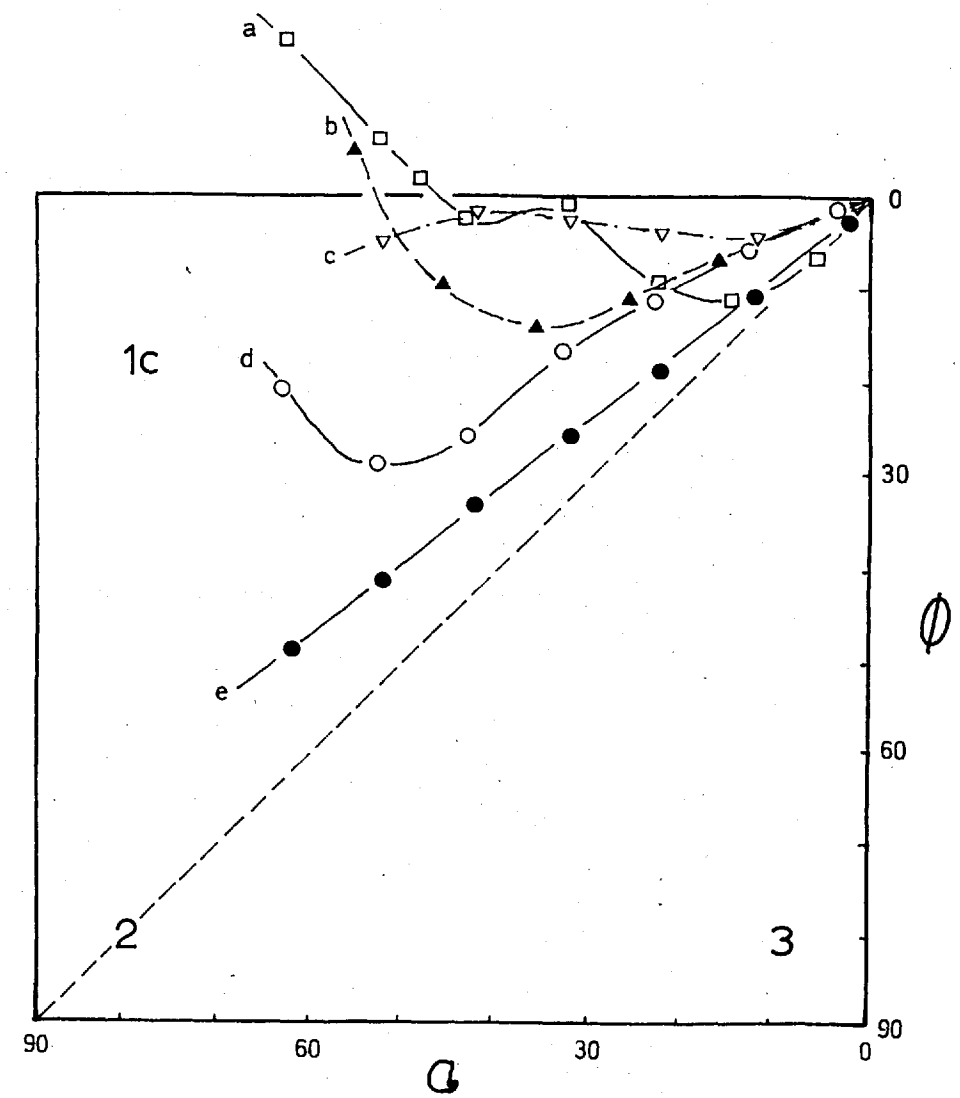


Fig. 6.12

Profile through an f_4 fold from Ben Tarbert.

FIG. 6-12

variations in the original layer thickness are very important in determining the geometry of the folded layers. The t'/a plots and especially the isogon ϕ/a plots are made highly irregular by such variations and often it is impossible to make any estimate of the amount of flattening deformation suffered by these folds.

The internal deformation, especially the tangential longitudinal strain, is very important, especially in multilayer fold systems. Fig. 6.11 gives a good example of this effect. The geometry of each folded layer depends on its position in the fold structure, and thus varies from type 1a on the outer arc to type 1c with up to 25% flattening on the inner arc. The outer arc of this fold shows irregular plots due to the exaggeration of what are probably initial thickness variations. One can conclude, therefore, that though certain layers show the 1c class of fold, with up to 25% flattening, the whole multilayer has a geometry close to 1b, with a low amount of homogeneous flattening.

Another feature of the f_4 folds in this zone, is the tendency for box-fold development (Fig. 6.12). This phenomena may partially explain a phenomena shown in many of the ϕ/a or t'/a plots (Fig. 6.13) where the plot does not lie on a simple $t' = (\cos^2 a + \frac{\lambda 2}{\lambda 1} \sin^2 a)$ curve. I found that certain t'/a plots show a tendency to curve up, indicating lower deformation on the limbs than on the hinge. and layers with this property often alternated with layers where t'/a plots or ϕ/a plots curved down, indicating the opposite. Heterogeneous flattening of a fold ought to produce one of these features, but not both in the same fold system and the phenomena cannot be explained by an original thickness variation. Huddleston (1969) has shown how that if folds suffer homogeneous strain and buckle at the same time, then t'/a or

Ø /a plot should curve down; but this would not explain the reverse case One possible explanation for the folds showing higher deformation at the crest than on the limbs may be that the layers measured were suffering from the effects of tangential longitudinal strain and therefore suffering a high component of extra strain at the crest of the inner arc. Such structures, however, have been found in positions other than at the inner arcs of multilayer folds, and so this cannot be the complete explanation. Another possible explanation is that the folds are tending to become multi-hinged. When buckling had ceased at the fold crest, and this area was deforming by homogeneous flattening, the limbs may still have suffered deformation due to buckle rotation about a new 'hinge' part way down the limb and thus show the effects of less actual flattening strain.

The folds within this zone also show a certain amount of heterogeneous deformation, in that there is often excess deformation on certain fold limbs, leading to the development of "megalithon structures". In many regions these structures are essentially large scale examples of the gleitbretter or microlithon structures which lead to the development of strain-slip cleavage in fine-grained low-grade metamorphic rocks. It has been found (Hills, 1963) that microlithon structures are developed in multi-layer complexes and that the width of the lithon structure is controlled by the thickness of the competent layers. Hinge zones are displaced relative to each other by a zone of shearing in the fold limbs. In the Uists I found that it is usually not adjacent hinges which were displaced (Fig. 6.13) and it is possible that several layers within the multi-layer complex may have combined to define the width of the lithon structure.

I have examined these structures in detail from a few examples (e.g. Figs. 6.13 and 6.14), both where the shear appears to cut through an originally folded multi-layer and also where the amount of original buckle folding is slight. In the former case (Fig. 6.13), the zone of shear runs parallel to the fold axial plane. As would be expected, a certain amount of simple shear is indicated, in that the hinge points of the folds are deflected between alternat, layers.

There is often a fabric, demonstrated by biotite or hornblende orientation, developed within and parallel to these shears, even when the adjacent fold axial planes have a different orientation. This would indicate that these structures developed closely parallel to the local X - Y plane. It would appear, therefore, that these structures are the results of localised adjustments of strain within multi-layer complexes, which because of their structure and orientation have a certain amount of resistance to homogeneous deformation.

Gnnerally within zone 2 the amount of flattening strain, superimposed an the folds, is fairly low. Most folds appear to be of fold classes 1a, 1b and 1c with variable flattening strain ratios, alternating with folds of class 3. The amount of flattening is variable, from zero to 3 : 1, as shown in Fig. 6.15a, which gives the amounts of flattening from 58 measured fold layers. Much of this variation is due to internal strains set up during folding, but there are local variations in amount of deformation.

Zone 3

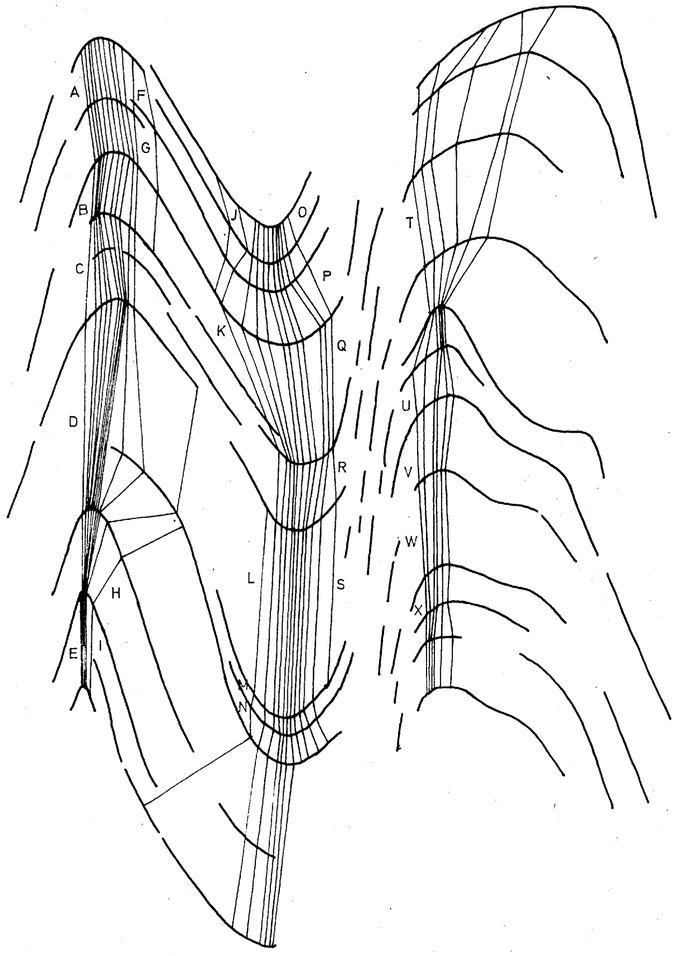
Structures within this zone show flattening strains similar to those of zone two, and in fact have been incorporated in Fig. 6.15a, but within this zone there was a strong tendency for the production of

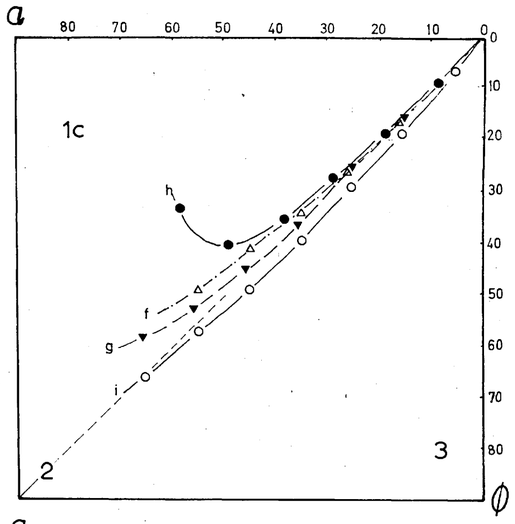
Fig. 6.13

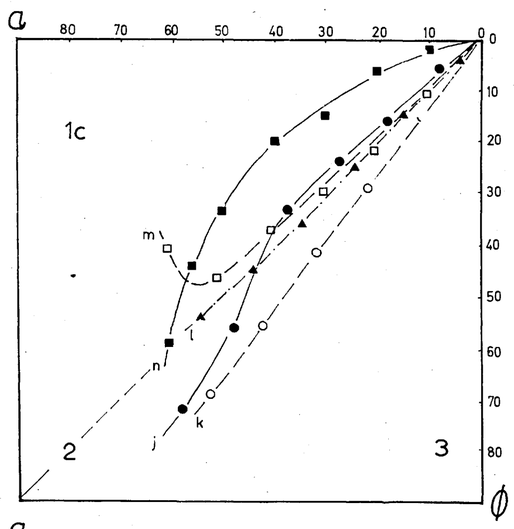
An example of an f_4 fold showing :-

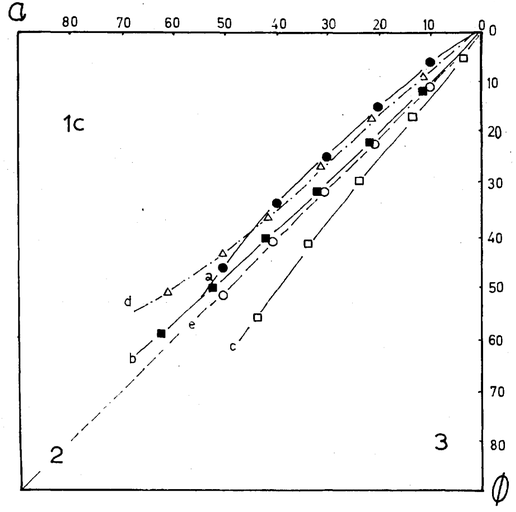
i) a zone of shear (i.e. a zone of high deformation) on one limb.

ii) plots of ϕ against a which do not lie on simple curves, but sometimes curve up, sometimes down (see text).


A
B
C
D
E
F
G
H
I
J
K
L
M
N
O
P
Q
R
S
T
U
V
W
X


a
1c
2
3
ϕ
h
f
g
i


a
1c
2
3
ϕ
m
l
n
k
j


a
1c
2
3
ϕ
d
a
e
b
c

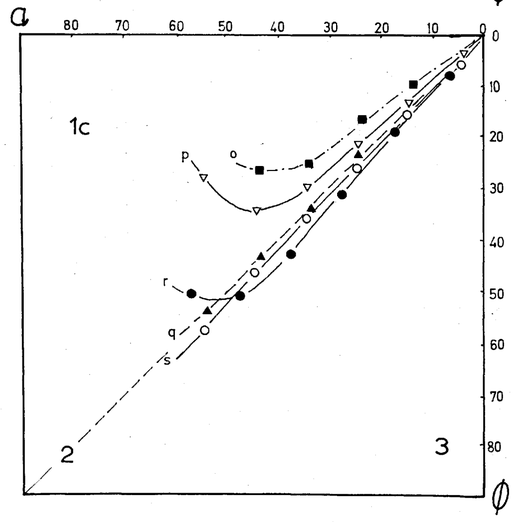

a
1c
2
3
ϕ
p
o
r
q
s

FIG. 6-13

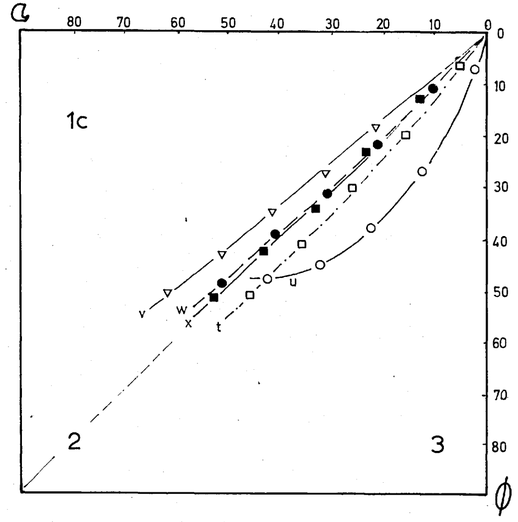
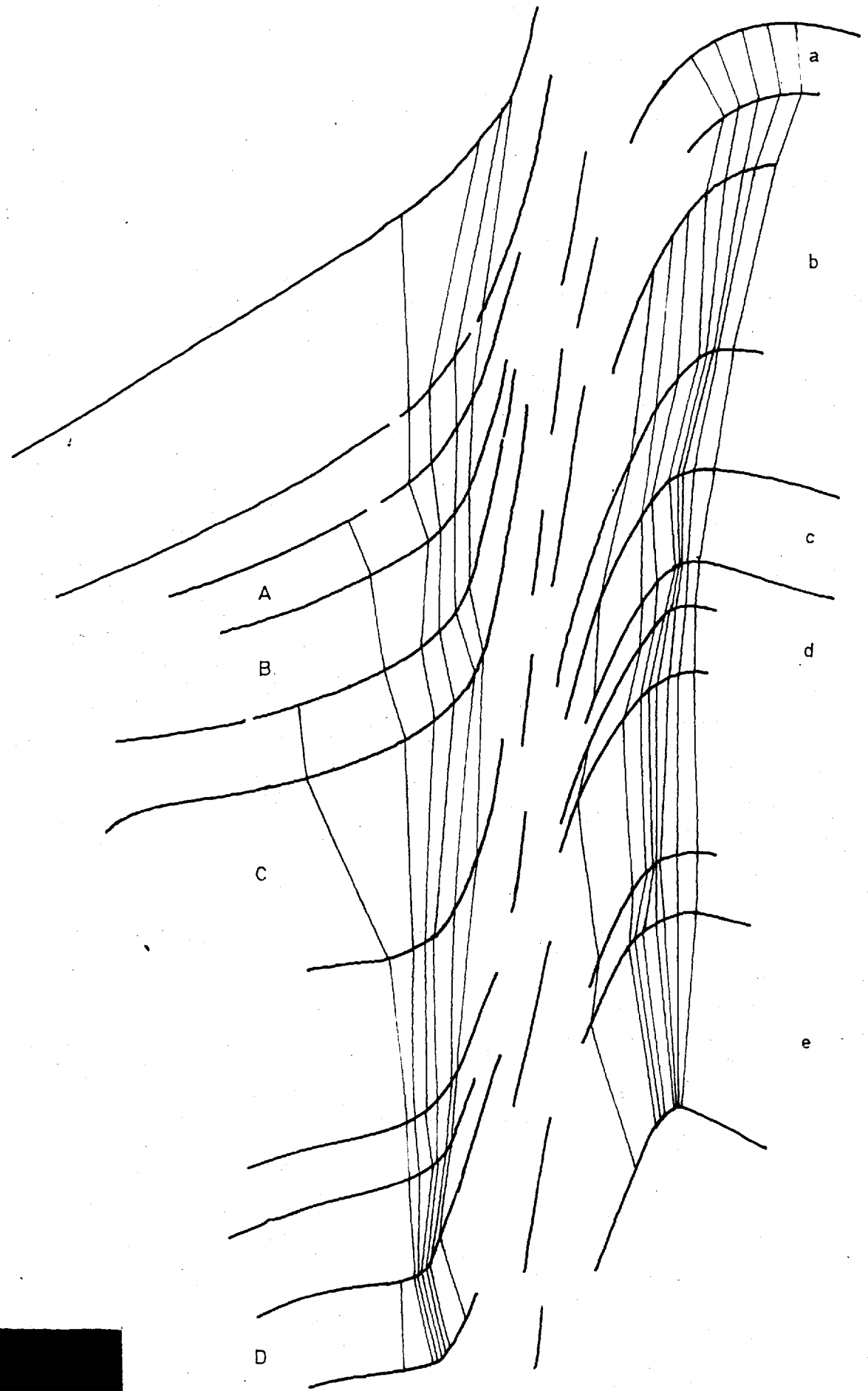
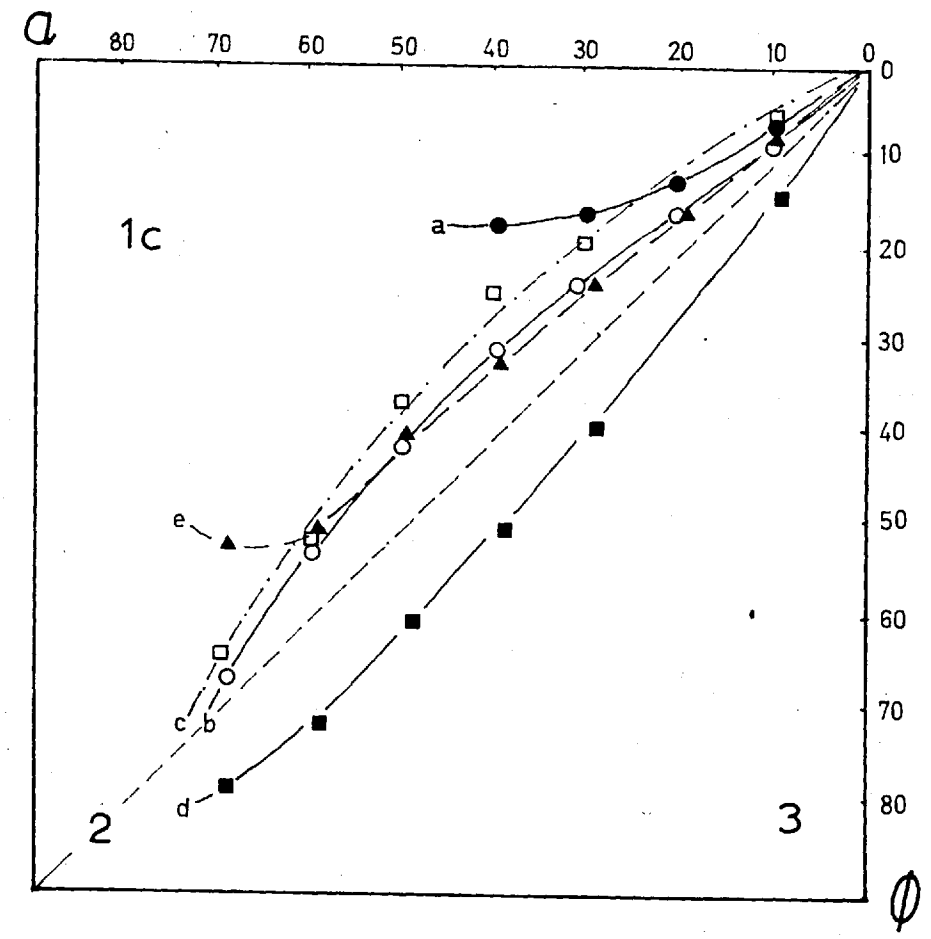

a
1c
2
3
ϕ
v
w
x
t
u

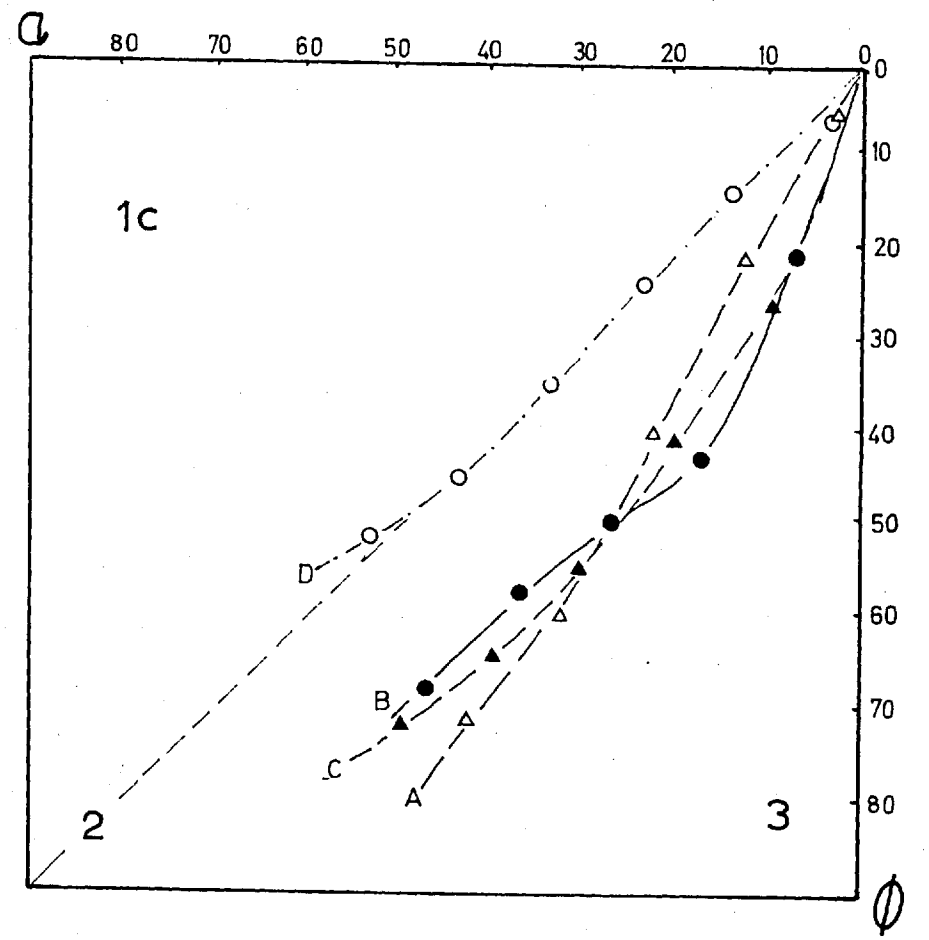
Fig. 6.14

An f_4 zone of shear in layered quartzo-feldspathic rocks. Note that this is not a zone of solely simple shear; the layers show both converging and diverging isogons.

FIG. 6-14


a
b
c
d
e
A
B
C
D


a
80 70 60 50 40 30 20 10 0
0 10 20 30 40 50 60 70 80
1c
a
b
c
d
e
2
3
ϕ


a
80 70 60 50 40 30 20 10 0
0 10 20 30 40 50 60 70 80
1c
A
B
C
D
2
3
ϕ

of disharmonic structures. In many localities the rocks cease to act as simple multilayer systems and localised complications occur, probably the result of variations in competence due to the f_4 metamorphism. Well-marked viscosity differences occur within the gneiss itself, depending on the response of the particular rocks to the f_4 granitisation. The nearly massive quartz-feldspar-biotite/hornblende gneisses appearing closest to granite in composition tended to act as the least competent material. These rocks often recrystallised as granite or pegmatite, and often show mobilisation structures, in that they sometimes appear to intrude the adjacent gneisses. Disharmonic structures often result, and the general pattern is often similar to that of "wild-folding" in other migmatitic regimes. Certain more competent gneissose layers or amphibolite bands often trace out fold shapes which appear to "float" in a sea of granitic or granodioritic rock.

It is obvious that strain variations within this region will have little relevance to those to the west, and will largely depend on local competence variations. There are, however, a few areas of high deformation, such as one south of the Skiport road, where the folds are caught in the inner part of a large order f_4 structure and therefore show a higher degree of flattening deformation, often leading to the near-similar appearance of the folds.

Zone 4

This zone shows similar f_4 folds to zone 3, but the folds are usually tighter and show a higher amount of flattening strain. Flattening strains of from 1.5 : 1 to over 8 : 1 have been recorded in this area, though the average is about 3 : 1. Fig. 6.15b gives the amount of flattening for 52 such folded layers.

Fig. 6.15

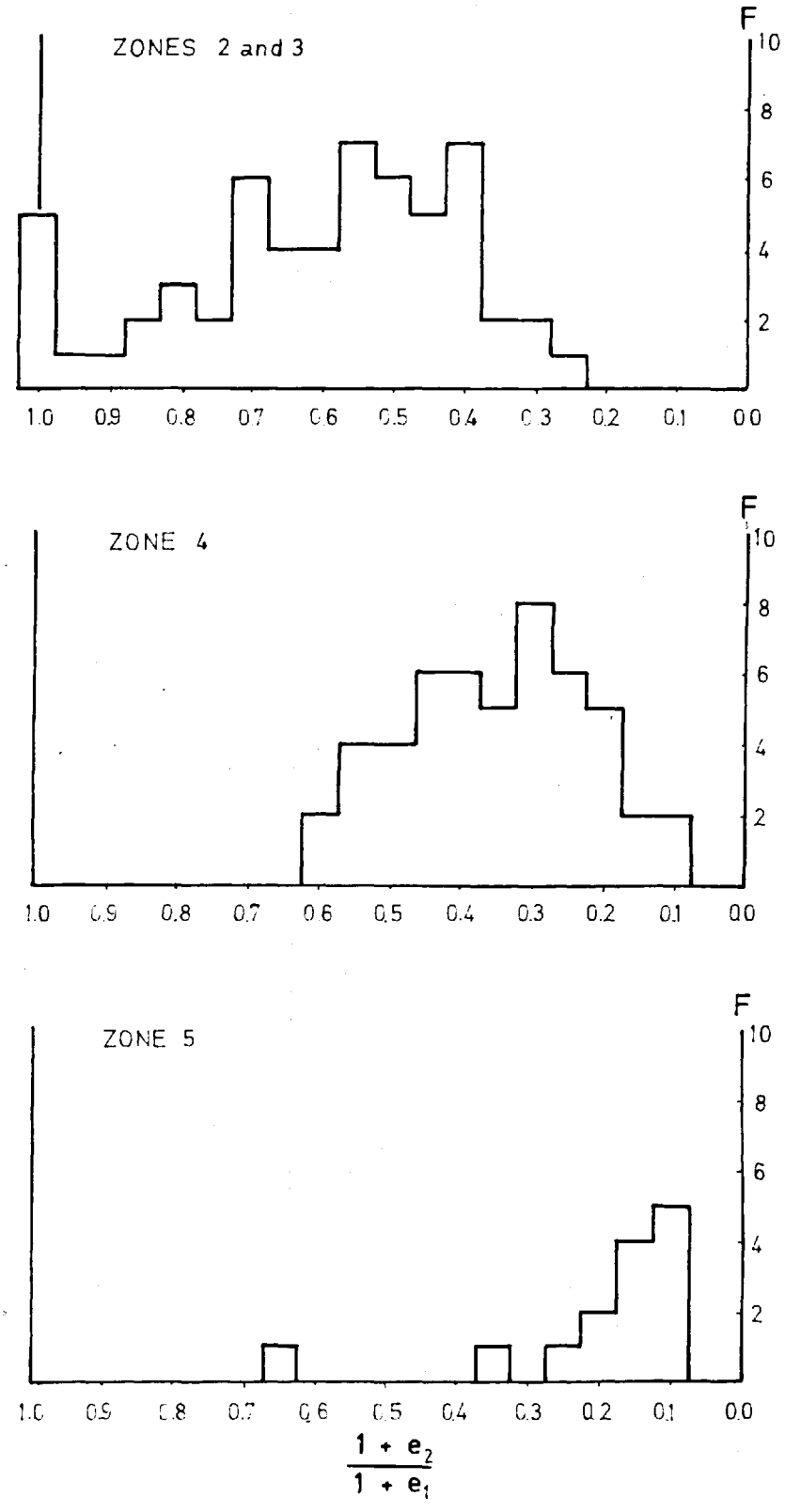
Histograms showing the amounts of post-buckle flattening

on 58 folds from zones 2 and 3

50 folds from zone 4

18 folds from zone 5

FIG. 6-15


ZONES 2 and 3
F
10
8
6
4
2
1.0 0.9 0.8 0.7 0.6 0.5 0.4 0.3 0.2 0.1 0.0
ZONE 4
F
10
8
6
4
2
1.0 0.9 0.8 0.7 0.6 0.5 0.4 0.3 0.2 0.1 0.0
ZONE 5
F
10
8
6
4
2
1.0 0.9 0.8 0.7 0.6 0.5 0.4 0.3 0.2 0.1 0.0
$\frac{1 + e_2}{1 + e_1}$

Zone 5

Within this region the f_4 deformation becomes much more intense, producing tight to isoclinal folds. The fold geometry is of classes 1c or 3, both approaching that of class 2, and in many areas of normal quartzo-feldspathic gneiss most of the folds are almost of class 2. The higher the deformation then the less sensitive are the methods of strain measurement and so heterogeneities due to original thickness variations or tangential longitudinal strain cease to measurable. Fig. 6.15c gives the amount of flattening from 14 folded layers within this zone. As one can see, many of the strains are up to 10 : 1.

Comments on the f_4 strain variation

If one compares this section with section 3.3D (also Fig. 3.1c) in Chapter III, one will notice the obvious relationship between the area of high f_4' deformation and the area of high syn-kinematic and post-kinematic f_4 metamorphism. The problem is - which came first? Which was the cause and which the result? I believe that there is no clear-cut answer to this problem and there may be several interacting explanations for the localised high f_4 effects. Two comments, however, come to mind.

(i) The f_4 strain trajectories do not converge on to a simple f_4 synform, and so the heterogeneous deformation cannot be put down so easily as being due to the buckling of an interface between rocks of different competencies, as can the heterogeneities within the f_3 phase. However, if one looks further afield, for example to Barra, Francis (1969) has noticed that the f_4 axial planes are there either vertical or dip to the north. In South Uist the are vertical or southwards dipping. It may be, then, that the f_4 strain trajectories converge downwards as well as

eastwards. This would indicate that in South Uist, deformation increases with depth, into the hottest parts of the f_4 migmatitic area.

It may be worthwhile adding that many people consider that gravity plays an important part in determining the structure of a migmatitic area. Migmatitic regions are supposed to rise, while non-migmatitic areas sink. The strain trajectories for the f_4 strucutres, however, do not tie up with any such theories about the migmatitic parts of South Uist rising and the intervening parts going down. In fact, they could indicate the opposite!

(ii) The region of most intense f_4 metamorphism co-incides almost exactly with the region of the most anhydrous f_2 metamorphism in South Uist. It is also an area rich in early intrusives, and may have had a high metamorphic grade in the Scourian. If this is co-incidence then it would be rather remarkable. It would appear that there is something about the basement to this region which on at least two occasions has either possessed a higher geothermal gradient than other regions, or has been more susceptible to deformation, facilitating higher heat flow. One is now obviously in the realms of pure speculation, and so I would now like to leave this problem for any reader to ponder on.

CHAPTER VII

THE PRE-LAXFORDIAN HISTORY

The previous five chapters have all dealt with the dominant geological structures of South Uist which are of Laxfordian age. This chapter will therefore briefly fill in the rest of the story and discuss the earlier, the pre-Laxfordian, events.

7.1 The Scourie-dykes

As mentioned previously (Chapters I and III), the Scourie-dykes form a roup of relatively homogeneous tholeiitic basic bodies now usually represented by (olinopyroxene) ambhibolites. It is difficult to be certain how far the hydrous nature of some of the dykes reflects amphibolite facies metamprphism and how far it refelcts the original composition, i.e. some of the dykes could have been biotite/hornblende rich originally. Most of the dykes, however, appear to have originated as typical tholeiitic basalts or dolerites (Dearnley, 1963, table IX, nos. 3, 5 and 6).

Within the larger dykes there is a certain amount of compositional banding, marked by varying amounts of leuco- or mafic minerals. The large mass at Carnan (823445) gives a very good example of such banding, which presumably formed by igneous cumulative layering. The large dykes show a decrease in grain size towards the edge. In some cases this could represent a tectonic grain-size reduction, but at Carnan it appears to be original and might indicate a slight chilling of the rock. Some of the dykes (e.g. at G7) contain feldspar phenocrysts and most of the large dykes exhibit a relict doleritic or gabbroic tecture (see also Chapter III).

The intrusive form of the dykes was varied. Some occur as large

intrusive sheets (Beinn Bheag Dheas), others as saucer-shaped bodies. On Rueval (782393), the Scourie-dyke mass forms a plug from which dykes radiate in all directions (Map 1). Branching forms are very common; G5, Garry-a-siar, (Fig. 5.4) is a typical example, and that shown in Fig. 5.21b from east of Trinival is another. Both of these example are typical of the undeformed branching dykes, in that the branches appear to follow some pre-existing joint or fracture pattern in the rock.

Many of the Scourie-dykes are very thin, as shown on Fig. 7.1b. They can be contrasted with Scourie-dykes from other parts of the Hebrides, from Great Barnera, for example (Watson 1968), where the dykes are thick and locally make up to 40% of the rock. It appears therefore that there are variations in number and thickness of the dykes throughout the Hebrides, phenomena which should be expected from such a dyke swarm.

Fig. 7.1a gives the orientations of the dykes from a section of coast at Garry-a-siar, where the amount of Laxfordian deformation is low and where the amount of tectonic re-orientation is minimal. The majority have a north-west - south-east strike. The dip of the dykes is generally steep and most of the "sills" in South Uist have been formed by deformation and by the obliteration of discordant structures.

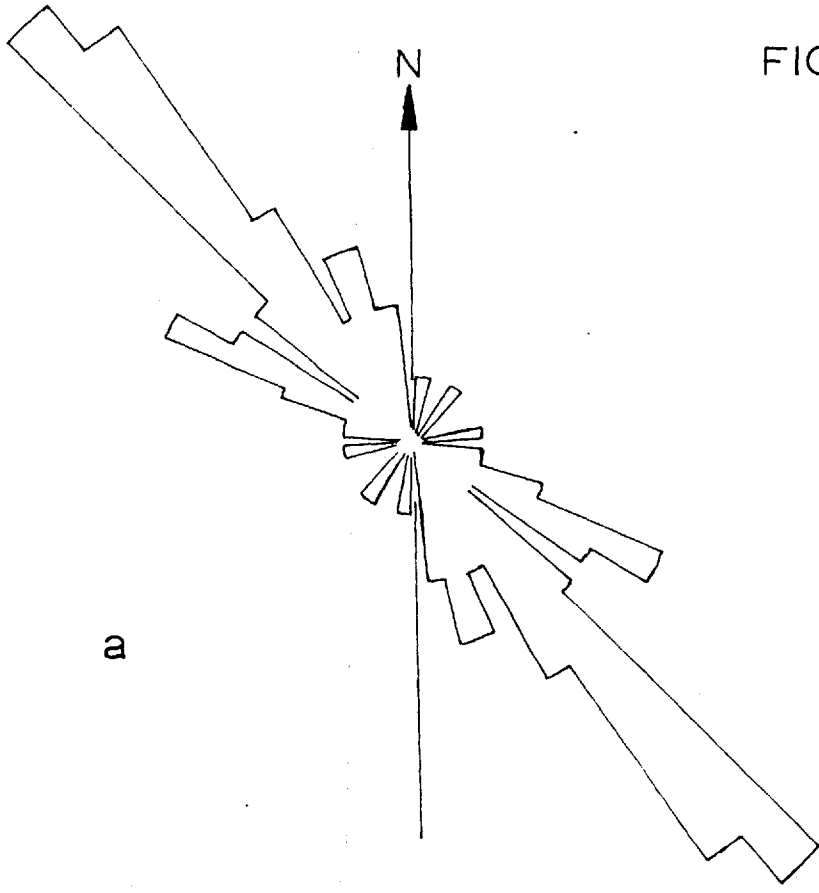
To conclude, the Scourie-dyke swarm appears very similar to many of the more recent Tertiary swarms. The dykes vary in thickness and intensity throughout the Hebrides, though they maintain a dominant orientation (see also Sutton and Watson, 1951, and Watson, 1968); variations in orientation may be partly controlled by earlier fractures or partings in the rock. The predominance of thin dykes in South Uist may indicate a relatively fluid magma, but there is no evidence for intrusion into a hot country rock as has been claimed for some dykes

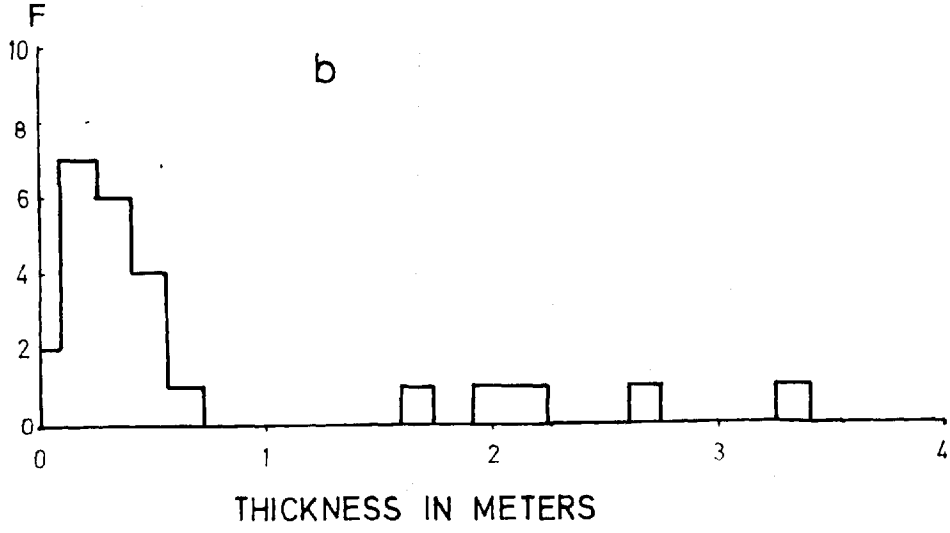
Fig. 7.1

a. Rose diagram showing the orientation of 45 apparently undeformed dykes from Garry-a-siar.

b. Histogram showing the frequency (F) of dykes of various thicknesses, from the northern part of Garry-a-siar.

FIG. 7-1





in the Scourie district of the mainland (O'Hara, 1961).

7.2 Pre-Scourie-dyke intrusives

It has been noted, in Chapter I, that in South Uist there are basic rocks of an earlier age than the Scourie-dykes. Certain of these basic rocks may be similar in age to the acid gneisses or metasediments, but there is a distinctive group of rocks, ranging in composition from pyroxenites to leuco-gabbros to granodiorites, which may belong to a separate intrusive suite. The rocks of this early suite form lenticular or sheet-like bodies usually about 50+ metres in thickness.

They occur in a relatively undeformed state in the Carnan - East Geirinish area where the original igneous structures are often still visible. The mass at East Geirinish (838393), for example, shows igneous layering; the base consists of about five metres of pyroxenite and is followed by a thick cumulative layered series of banded meta-gabbro. The upper part of the mass consists largely of diorite or granodiorite.

Where Laxfordian deformation is more intense, the masses lose their igneous structures and those with igneous layering become transformed into basic gneiss. Such banded bodies occur on Arnaval (783251), at Allt Volagir (802292) and at Carnan (817446, see also Map 1). On Arnaval, the centre of the mass consists of pyroxenite, rimmed by tectonically banded gneiss and locally by garnetiferous diorite. At Carnan the pyroxenite occurs only in the core of an f_2 synform; the rest of the fold structure is taken up by banded gneisses.

There are several points to be considered concerning the age and the origin of these bodies, as summarised below:

(i) Where the bodies are least deformed, it appears that the f_2

deformation phase was the first to affect the rocks.

In regions of higher Laxfordian deformation, then there is no evidence as to the amount, if any, of a pre-Laxfordian deformation. At Allt Volagir, Scourie-dykes cross-cut the banding of an early amphibolite, but this banding is presumably only a transposed igneous layering and so this discordant relationship only indicates that the rocks pre-date the Scourie-dyke intrusion phase.

(ii) The rocks have a distinctive mineralogy; the less deformed and metamorphosed rocks contain assemblages such as -

clino-pyroxene - ortho-pyroxene ($\pm$ hornblende)

clino-pyroxene - prtho-pyroxene - biotite - plagioclase - quartz ($\pm$ hornblende)

biotite - plagioclase - quartz $\pm$ pyroxene $\pm$ hornblende

When the rocks have been extensively deformed and metamorphosed, they also include the assemblage -

garnet - clino-pyroxene - biotite - plagioclase - hornblende - quartz.

The rocks are locally extremely garnetiferous and are rich in biotite. They would therefore appear to be iron- and magnesium-rich rocks (possibly also potash rich), and the closest analogy is with lamprophyric rocks associated with the Caledonian Orogeny.

(iii) Very similar rocks have been found on the east coast of Barra (Francis, 1969) where again a full suite of intrusions ranging from ultrabasic rocks to diorites are relatively undeformed and yet are cross-cut by intrusions of Scourie-dyke age. These early igneous rocks have given apparent radiometric ages of 2600 m.ys (Francis, personal communication).

(iv) They are often rich in quartzo-feldspathic material and often veined and patched with potash feldspar bearing pegmatite, much of which

has an f_2 fabric and must therefore be syn- or pre-f_2 in age. I have found no Scourie-dykes which bear these pegmatites even when adjacent to these early basic bodies (856399). This could indicate that much of this granite and pegmatite was of pre-Scourie-dyke age.

(v) Certain of these bodies appear to be agmatised, in that they consist of numerous individual blocks separated by quartzo-feldspathic material. The large masses of agmatitised basic rock in the Skiport area (Chapter III) appear to be of this type; they are rich in biotite and carry biotite - hornblende - (clino and orthopyroxene) - plagioclase - quartz assemblages.

The origin of these agmatites is itself of interest. Some masses are only agmatised at the margins, others, up to a half kilometre across, are agmatised throughout. If the agmatites are tectonic in origin, the quartzo-feldspathic material being due to migmatisation, then these large masses of rock must have suffered simultaneous extension in three perpendicular directions. Personally, I find this mechanism hard to imagine. Analogous structures such as net-veined bodies are however very common in igneous environments and form where there has been a mixed intrusion, or where the igneous rocks have intruded unconsolidated acid magmas. It may be possible that the Skiport agmatites formed as a result of the intrusion of basic magma into hot (?) country rock, thus causing localised melting and mixed intrusion.

It appears likely, therefore, that these rocks are analogous to the late orogenic lamprophyric suites of intrusions which can be found in many other orogenic belts. They would appear to be of late Scourian age

(2600 m. ys) and locally may have intruded while the gneiss was still hot. In these areas they formed chiefly as agmatitic masses but elsewhere formed as composite distinct igneous bodies.

To conclude, there are two groups of igneous intrusives separating the Scourian deformation and metamorphism from that of Laxfordian age, as shown on Table 7.1. There remains the problem, however, of the amount of deformation or metamorphism (if any) separating the two periods of intrusion. There is no evidence of any such deformation in the Eastern part of South Uist (Table 7.1) but on the west coast at Garry-a-siar at G13 there are thin biotite rich dioritic dykes which have suffered at least one phase of Scourian activity and yet are slightly discordant. If these dykes are of the same age as the igneous masses to the east, it would appear that there was a period of deformation separating the two main igneous phases, this period of deformation occuring largely in the west (Table 7.1).

7.3 Basic rocks of inderterminate origin

Apart from the basic rocks discussed in the previous sections, there are other groups of an earlier age. Dearnley and Dunning (1968) recognised a group of these early migmatised and boudinaged rocks at Ardivachar and at Garry-a-siar. Some early basic rocks could be of intrusive origin, but others seem to be integral parts of the gneiss complex and could be of sedimentary or volcanic origin.

At Stoneybridge point (734338), there occurs a meta-gabbro-anorthosite, with layers of pure plagioclase (up to 70% anorthite), separated by garnet amphibolite, ($\pm$ clinopyroxene). A similar rock occurs on Orosay, North Boisdale and scattered throughout the island there are a few other garnet amphibolites which, from field appearance, could be of

the same age. The age relationship of these garnet amphibolites and anorthosites to the other early basics is purely speculative. They may be of a similar age to the other anorthosites and gabbro-anorthosites of the Outer Hebrides, in South Harris (Dearnley 1963) and Ness (Watson, 1969).

7.4 Scourian fold history

Scourian structures can be recognised in several localities throughout South Uist, where they are sometimes cut by Scourie-dykes or refolded by f_1 or f_2 folds, but as such observations are so sporadic, it has not been possible to consider them constructively. At Ardivachar and Garry-a-siar a much clearer sequence of Scourian structures can be seen as noted below.

At least three pre-Laxfordian fold phases can be recognised: f_{S1}, f_{S2} and f_{S3}. The first of these, f_{S1}, can only be recognised from the fold interference patterns which it makes with the main phase structures, f_{S2}. These interference structures are of the closed type, forming "eyes" or "mushrooms". Much of the migmatitic veining in the early amphibolites may be of f_{S1} age. The main phase of deformation, f_{S2}, produced isoclinal folds, often on a small scale. It re-folded the early migmatite veins and boudinaged the basic blocks giving the rock an agmatitic appearance.(Plates 29B and 30 in Dearnley and Dunning, 1968). The main phase structures now lie with a north-east - south-west trend, parallel to the main phase Laxfordian structures. The late deformation phase, f_{S3}, had a more east - west trend. It generally produced small-scale structures as well as large-scale folds which locally re-orientated the gneissose foliation into a north-west - south-east trend (A16 Ardivachar). Either closed or open interference patterns were made

with the main phase folds, depending on the relative orientations of the two sets of structures. All the Scourian structures are now co-axial; see Chapter V.

7.5 Scourian metamorphic history

It is difficult to give a Scourian age to any mineral assemblage, but several periods of migmatite veining, granite and pegmatite growth can be recognised. Close to A11 Ardivachar the country rock consists of a Scourian granite with an f_{S2} fabric. Many pegmatite and migmatite veins also formed prior to the f_{S2} deformation and other granites and pegmatites formed following this phase. These later veins cross-cut the f_{S2} structures and the early pegmatites and they often form migmatitic augen growths within the older granite gneiss. The pegmatite close to A14 Ardivachar is an example of such a mixed growth. Much of the rock is a deformed granite and pegmatite with an f_{S2} foliation (c.f. plate 29a in Dearnley and Dunning, 1968), but growing across this fabric there are feldspathic augen of post-f_{S2} age which have themselves been deformed by late Scourian and Laxfordian deformation phases. This pegmatite has given an apparent age of 2300 m.ys (Dearnley and Dunning, 1968), an age which falls between the Scourian (2600 my) and the Laxfordian (1800 - 1500 my.) periods of the mainland (Watson, 1965). It may be that this pegmatite gives a Scourian age which has been modified by Laxfordian events. The other alternative is that it gives an unmodified age and the deformation phases at Ardivachar and Garry-a-siar may represent the last stages of the Scourian period (c.f. section 7.2 and Table 7.1). The pegmatite cannot be taken to represent a post-Katarchaean and pre-Scourian event (as suggested by Dearnley and Dunnnng, 1968) until the base of the "Scourian" has been more firmly established (c.f. Dr. N. J.

Snelling in the discussion on Dearnley and Dunning, 1968).

In Chapter III it was mentioned that the anhydrous nature of certain localised rocks in the Skiport area might be due to a pre-Laxfordian event, and could possibly be due to the same event which agmatised the basic rocks of the early igneous suite mentioned in section 7.2. It has also been suggested, in section 7.2, above, that this agmatisation occured at the time of intrusion. If conditions of intrusion were such as to cause complex agmatite structures to develop, then it is conceivable that temperatures were reached high enough to cause local dehydration of some hornblende and biotite. The anhydrous rocks could be considered as being due to high grade contact metamorphism superimposed on the late stages of the Scourian regional metamorphism.

In this Skiport region there is an association between the distribution of anhydrous ferromagnesian minerals and rocks of undisputed sedimentary origin, and both are closely associated with the early basic intrusives. It may be that these early basic rocks preserved both the anhydrous nature and the sedimentary nature of the rock from much of the Laxfordian deformation and metamorphism. Alternatively, the anhydrous nature of the rock may only reflect one feature of a contact metamorphism which gave the metasediments a high chance of preservation (Coward et al, 1970).

7.6 Synthesis of the Scourian history

A summary of the pre-Laxfordian history is given on Table 7.1 which shows also a possible correlation with the pre-Laxfordian history of Barra as set up by Francis (1969). It is a long and complex history with many phases of igneous activity, deformation and metamorphism. The main

WEST BENBECULA	EASTERN UISTS	SOUTH-EAST BARRA
Slight Laxfordian deformation	Laxfordian deform.	Slight Laxfordian def.
Scourie-dyke intrus.	Scourie-dyke intrus.	Scourie-dyke intrus.
f_{S2} and f_{S3} deform. and metamorphism, (age of 2300m.y.?).	No apparent deform.	No apparent deform.
Intusion of diorite and granodiorite	Intrusion of ultra-basic rocks, layered metagabbros and lamprophyric rocks	Intrusion of granite and pegmatite, Intrus. of a suite of dioritic rocks (2600 m.y.)
f_{S1} def. and metamorphism	Main Scourian def. and metamorphism. Gneissification of the rocks	Main Scourian def. and metamorphism.

Table 7.1

A possible correlation between the Lewisian histories of west Benbecula, eastern Uists (Carnon - Skiport region) and south-east Barra (after Francis, 1969).

tectonic period concluded with a phase of igneous activity; lamprophyric, dioritic or ultrabsic dykes were intruded at a time when the structurally lower levels of South Uist, now seen in the east (see Chapter VI) may still have been hot. If the correlations on Table 7.1 are correct, then this igneous phase was followed by phases of localised deformation and metamorphism, now seen only in the west. The Scourie-dykes appear to have been intruded into "cool" non-orogenic environments, though in the structurally lower levels (c.f. Francis, 1969, O'Hara, 1961) this need not have been the case.

Unitl more is known about the detailed structural and metamorphic history of the mainland, it is unwise to make any correlations with the "Scourian" and the "Inverian" as set up in the Lochinver area by Evans, 1965, Bowes, 1968.

7.7 The origin of the gneissose foliation

At Ardivachar and at Garry-a-siar, all the fold phases, including the first recognisable phase, f_{S1}, fold a pre-existing gneissose foliation. The same situation is found everywhere in South Uist where Scourian folds can be recognised. This original foliation has been modified, transposed and strengthened by all the Scourian and Laxfordian deformation phases, but locally homogeneised and almost obliterated by Scourian and Laxfordian migmatisation. The gneissose foliation therefore appears to have had its beginnings in the original pre-tectonic rock but to have suffered extensive alteration since then.

7.8 The origin of the gneiss

Within the normal, acid gneisses, as mentioned in Chapter I, meta-sedimentary relics are still preserved and occur in all forms, from vague,

poorly defined areas, to narrow, well-defined bands traceable for several kilometres along their length (Map 1). Similar metasedimentary relics have been found elsewhere in the Hebrides and have been discussed at some length in a separate paper (Coward, Francis, Graham, Myers and Watson, 1970).

The conclusions of this paper are that the metasedimentary strips may represent either

cover rocks to a pre-existing basement gneiss

or only part of a supra-crustal group of rocks which included all the Western Gneiss, in which for various reasons the metasediments have survived through both the Scourian and Laxfordian periods.

Though the first of these possibilities cannot be completely ruled out, t ere is no evidence for a basement/cover relationship in the "Western Gneisses" of South Uist. As Dearnley and Dunning (1968) point out, the nature and appearance of the compositional banding in the gneiss suggest a supracrustal series consisting of greywacke and volcanic material which had undergone polymetamorphic mafic and quartzo-feldspathic segregation. There may be evidence of an earlier basement in parts of the "Eastern Gneisses" in Barra (Francis, 1969) where the country rock is different from that to the west, being much more anhydrous and dense (Coward et al, 1970a, 1970b). There is definitely no evidence of a basement cover relationship existing in the central part of South Uist, as was suggested by Bowes, 1968.

CHAPTER VIII

THE EASTERN GNEISS

(Incluling the "Usinish Mylonites" and "Mashed Gneisses")

8.1 General

The "Eastern Gneisses" of South Uist is the name given to the group of rocks which lies in and above the thrust zone (see Chapter I). In many areas they are highly crushed and there it is almost impossible to get anything constructuve in the way of a structural or metamorphic history. A study of the Eastern Gneiss, therefore, has to be confined to the relatively uncrushed regions (see Chapter IX), such as the area between Ornish and Maol Martaig, and the eastern slopes of Beinn Mhor.

The main components of the Eastern Gniess form an almost uniform suite, stretching from Ornish Island to the south of Lochboisdale (Francis, 1969); they have the following characteristics which differentiate them from the gneisses below the thrust belt:

(i) The rocks show relatively little mineralogical banding, and in many areas they appear homogeneous. Small-scale folds are uncommon, but a strong mineral fabric or cleavage is developed locally.

(ii) Prior to crushing, the rock wasequigranular except in areas where there was a well-developed cleavage.

(iii) The rock is quartz-poor to quartz-free.

(iv) It often appears to be iron-rich. The ferro-magnesian minerals sometimes show exsolution of magnetite/ilmenite. Iron ore rims many of the grains and is locally concentrated into quite large slivers.

(v) In many areas the rock bears granulite facies assemblages con-

sisting of pyroxene, garnet and plagioclase feldspar.

8.2 Lithology

Most of the Eastern Gneiss is made up of a leucocratic garnet-plagioclase (usually andesine)-diopside-brown hornblende gneiss. Locally, as on the summit of Hecla, hypersthene as well as diopside may occur. In the Lindale area there are a few patches of leuco-gneiss containing up to 10% of quartz.

Interbanded with the leuco-gneisses are layers richer in mafic constituents, i.e.:

pyroxene - (hornblende) - garnet - plagioclase gneiss

pyroxene - (hornblende) - garnet - gneiss

pyroxene - (hornblende) - gneiss

These mafic or ultramafic gneisses occasionally occur as large pod-shaped bodies, but more frequently as bands from a few millimetres to several metres thick. They are best developed on the low hills close to the Minch, especially on the Eastern slopes of Beinn Mhor.

At a few localities (815284, 827295, 826283) there are some distinctive finer grained, equigranular basic rocks, containing hypersthene, diopside, plagioclase with a small amount of quartz and garnet. In the field they are different texturally from the normal mafic Eastern Gneisses, and in all the above mentioned localities they appear to be discordant to some lithological banding. They may therefore be members of a later suite of dykes, possibly of Scourie-dyke age. Dearnley (1962) mentions the existence of such dykes in the Ornish area.

Whether or not the discordant relationship referred to above is original may be open to some doubt. All the Eastern Gneisses have

suffered some crushing; along any major lithological boundary there are thin stringers of altered or mylonitised rock and the above-mentioned basic rocks are no exception. Their discordant relationship to the lithological banding may therefore be a later tectonic effect. However, because of their different textures (finer-grained but often with ellipsoidal concentrations of mafic and leuco-minerals, which may mark a deformed igneous texture), and their different mineralogy (abundant orthopyroxene), I find it easy to consider that these basic rocks are the thinly distributed members of a dyke suite which was intruded some time after the Eastern Gneiss series had formed. Dearnley (1962) has correlated similar basic rocks from the Ornish area with basic rocks of the Scourie-dyke suite in the Western Gneisses. If Dearnley is correct, and the discordant relationships are original, then much of the lithological banding in the Eastern Gneiss would presumably be of Scourian age, while most of the later deformation which affects these basic rocks as well as the Eastern Gneiss ss. would be post-Scourie-dyke and probably Laxfordian. The above, however, is all the evidence that there is for the age of these rocks. It is all based on correlations of dyke intrusions and structural phases across the thrust zone.

8.3 Structural history of the Eastern Gneiss

Over parts of the Eastern Gneiss, a good schistosity is visible, demonstrated by the preferred orientation of pyroxenes, hornblendes and feldspars, and due not to syn-tectonic recrystallisation but to the cataclastic rotation of the mineral grains. Where the fabric is more intense, then syn-tectonic minerals also occur, but these are usually of hornblende or chlorite.

This fabric can be seen to be later than the mineral banding, but

to be re-orientated by the later deformation. The fabric therefore makes a good time marker around which the other phases can be placed, and so in this section I intend to describe the structures of this phase first, and demonstrate the other deformation phases by their effects on this cleavage.

8.3A Main phase deformation

As mentioned above, the relatively widespread cleavage is due to a re-orientation of all the rock-forming minerals. In the west, there was relatively little syn-kinematic mineral growth associated with this phase, and the areas with a regular equant-grained fabric appear to have been areas where the strain was less intense. In some areas, the earlier banding and main phase cleavage are parallel, but along the north-east side of the Ornish Peninsula and in the low ground between Rudha Hellisdale (835307) and Rudha Bolum, they are generally oblique. This could be due to variations in the amount of deformation, as in some areas the banding may have been transposed into the main phase foliation, but I feel that it is more likely to be due to variations in initial orientation of the banding. Syn-tectonic recrystallisation may have occured locally in the west, but it appears to be largely confined to the east where hornblende, chlorite, and other low-grade minerals define the fabric, and where the amount of this recrystallisation appears to be related to the intensity of the deformation; the stronger the fabric, the more low-grade minerals there are.

Small-scale folds are rare, but at some localities on the east coast (notably 827291 and 826286) layers of mafic and ultramafic gneiss trace out folds between 10 and 20 metres in wavelength and 30 to 40 metres in amplitude. The main phase cleavage lies parallel to the axial

planes of these structures. East and south-east of Bolum small-order folds of this generation occur, with a similar axial plane fabric. Because of the nature of the rock, however, the amount of crushing and the telescoping of the whole succession, I have found it hopeless to try to map out any major structuresof this generation.

In the areas south-east of Beinn Mhor and south-east of Hecla, the main phase foliation has a relatively constant orientation, with a steep dip and a strike of about N.20° (Map 6). In areas of later folding, this direction marks the trend of the enveloping surface of the large structures. This p se also appears to be the first to deform the basic rocks which Dearnley correlated with the Scourie-dyke suite. Because of this circumstantial evidence, I have tentatively correlated this phase with the Laxfordian f_2 phase (see Chapter II) and shall thus consider the later phases as f_3 and f_4 (table 8.1).

8.3B Structures earlier than the main phase fabrics

Most of the mineral and lithological banding of the Eastern Gneiss existed prior to the f_2 deformation. Many of the bands show what could be early tectonic folds or boudinage, but could just as easily be original structures if the rock was formed from an original layered igneous body.

Some of the mafic gneisses contain feldspathic streaks and patches reminiscent of migmatised rock, and there are localised pegmatitic patches of plagioclase and potash feldspar material, all of which are deformed by the main phase structures. It may be that the rocks suffered some pre-f_2 (Scourian?) metamorphism and migmatisation, but as to its amount or extent, I would not like to comment at this stage.

Table 8.1

Suggested correlation between the Eastern and Western Gneisses

EASTERN GNEISSES		WESTERN GNEISSES
ORNISH AREA	LIADALE ARAE	
	late pegmatites	late pegmatites
		late migmatisation
"f_4" folding and shears	Glen Liadale shear zones	f_4 folding, east-west trending axial planes
f_3 folding, north-south-trending axial planes, with dip to east	f_3 folding, large antiformal structure, Beinn nan Coaroach shear zone	f_3 folding
f_2 main phase deformation, cataclastic fabric, local low-grade metamorphism	f_2 main phase deformation, cataclastic fabric, low-grade metamoprhism in the east	f_2 main phase deformation
		f_1 deformation
Intrusion of basic dykes	Intrustion of basic dykes	Scourie-dyke intrusion
Early deformation?		
Formation of the Eastern Gneiss		Intrusion of pre-Scourie-dyke igneous bodies

8.3C The f_3 phase of deformation

Over much of the Eastern Gneiss outcrop, this phase re-orientates the f_2 cleavage and the mineral banding to produce quite large, but fairly open, inclined fold structures (see Map 6, section E). The axial planes trend between north - south and north-north-east - south--south-west, with a moderate dip to the east. It must be noted that this orientation is oblique to the f_3 trend in the western gneisses, indicating either a non-correlation or a change in f_3 orientation to the east.

Few small scale structures have been found in the south, but between Maol Martaig and Ornish (notable at 847346 and 853376) folds occur with a wavelength down to ⅔ to 1 metre. I have been unable to identify with certainty any planar fabrics of this generation, though some areas have a strong mineral lineation which is presumably due to the intersection of the f_3 strain on the main phase cleavage. This mineaal fabric is demonstrated by the preferred orientations of hornblende and chlorite.

8.3D Beinn nan Caorach "Shear Zone"

On the southern ridge of Beinn Mhor, at 815290, the gneisses are involved in a zone of intensive deformation. Several local phases of folding occur. The main phase (f_2) fabric is isoclinally folded, and a new planar fabric, demonstrated by preferred orientations of plagioclase, hornblende, chlorite and epidote is imposed on the rock. This fabric itself is deformed by a series of phases of crinkle, chevron and kink-band folds, and all these are deformed by structures associated with the thrusting.

The orientation of the zone as a whole, and of the new planar

fabric, is approximately north - south. It is of a different orientation from shear zones in Glen Liadale to the east, and because of its complex fold history, may be of an earlier age.

8.3F The f_4 phase

On Map 6, it can be seen that east of Maol Martaig the f_2 cleavage is contorted by a series of structures trending east - west to east-north-east - west-south-west. Folds with a wavelength of down to one or two metres occur in this region. Both the steep and gently-dipping limbs of the f_3 structures are folded. North of Corrodale Bay, there is a spread of lineation plunges, probably due to open folds of this generation.

East of Maol Martaig, in "shear zones" (zones of more intense deformation, not necessarily produced solely by simple shear) and probably associated with this phase of deformation, there is a new planar fabric, striking at between N.90° and N.120°.

8.3F Glen Liadale Shear Zones

Between Glen Hellisdale and the head of Glen Cormascot (818297) there are several zones of intense late deformation striking at about N.100° (Map 6). They impose a new planar fabric upon the rock in which plagioclase, hornblende, chlorite and epidote show a preferred orientation and also a compositional banding. This deformation locally transposes all the previous fabrics and lithological banding in the Eastern Gneiss, as well as developing its own schistosity. From the orientation, it may be possibly to correlate it with the f_4 shears and folds at Maol Martaig.

8.4 The structural history of the other bodies above the Outer Isles Thrust Zone

As was mentioned previously (section 1.5), the Eastern Gneiss ss. was not the only rock group to occur above the main thrust zone. The Eastern Gneisses are bounded to the west by a group of phyllonites followed eastwards by a group of highly sheared acid gneisses. According to Jehu and Craig (1925), these rocks gained their essential characteristics during the late thrust period, but as I explain below, I consider them to be of an earlier age, related to the main structures in the Eastern Gneiss mass itself.

8.4A (I) The mashed gneisses

The rocks to which Jehu and Craig applied this term lie on the most easterly promontories of South Uist, above the "Eastern Gneiss" and the "Usinish Mylonotes". They are generally acid gneisses, apparently of much the same composition as the quartzo-feldspathic gneisses to the west of the Outer Isles Thrust Zone. Occasionally, bands of more basic material can be found and it often appears that one is dealing with rocks of essentially the same composition as the " Western Gneiss".

Structurally and mineralogically, however , these rocks are vastly different from those to the west. They are intensely crushed and have suffered countless phases of fracturing, shearing and cataclasis. In the field, the structures appear as a meshwork of shear and mylonite zones. In thin section, the rocks appear as an almost homogeneous mass of quartz, epidote, chlorite, sericite and other alteration minerals. I have been unable to make any structural sense out of the rocks; the shears appear to be too randomly orientated and though sequences of shears and minor faults can be worked out, these sequences apply only for localised

areas. Over the whole of the Usinish Peninsula I have found very few (three) areas where any pre-cataclastic foliation could be identified and the age of this foliation is of course unknown. The cataclastic deformation may be wholly of thrust plane age or it may have begun earlier. I prefer the latter alternative, but there is no sure way of being certain as to which is the case.

8.4B The Usinish mylonites (phyllonites)

The rocks to which Jehu and Craig applied this term lie sandwiched between the "Eastern Gneiss" and the "Mashed Gneiss" and crop out on several peninsulas or islands off the east coast. If one defines the term "mylonite" as a rock which has formed strictly from cataclastic deformation, with no associated recrystallisation, then these rocks are not true "mylonites". They are rocks in which a strong planar fabric has developed, this planar fabric being produced by the preferred orientation of the micaceous and chloritic minerals, and platy grains of quartz. They are better described, therefore, as "phyllonites" and it is as phyllonites that I shall subsequently refer to them. Their composition varies. The lower phyllonites are often rich in chlorite while the upper phyllonites are richer in quartz and feldspar.

Several phases of deformation can be recognised at each of the main outcrops, and table 8.2 gives my structural correlation. The earliest fabric can be most easily recognised in the phyllonites of Stuley Island and at a few localities in Usinish Bay. It consists of an early "phyllonitisation" resulting in an interbanding of fine-grained quartz and feldspar layers or mafic layers with a matrix of fine grained normal quartz/feldspar rock. These layers are tightly folded and a second phyllonite fabric or schistosity is developed parallel to the

Table 8.2

Structural correlation between the different outcrops of "Usinish Mylonites"

ORNISH	RUDHA ROSSEL	USINISH BAY	BOLUM	STULEY ISLAND
late mylonites	late mylonites		late mylonites	late faults
				quartz-feldspar injection
Tilting on north-south trending axes	north-south tilting	north-south tilting	north-south trending warps	north-south warps
		north-south trending crenulation folds		
	east-west trending folds	east-west trending kinks and small folds	east-west warps	east-west warps
Second phyonite fabric	phyllonite fabric	second fabric produced axial-planar to tight folds	phyllonite fabric	second phyllonite fabric, produced axial planar to tight folds
first phyllonite fabric	?	first phyllonite fabric	?	first phyllonite fabric

axial planes of these folds; a strong linear fabric developing at the intersection of these two planar fabrics. In some other localities, as at Bolum, the decond deformation appears to have been parallel to the first; no folds occur and the linear fabric is not so well-developed.

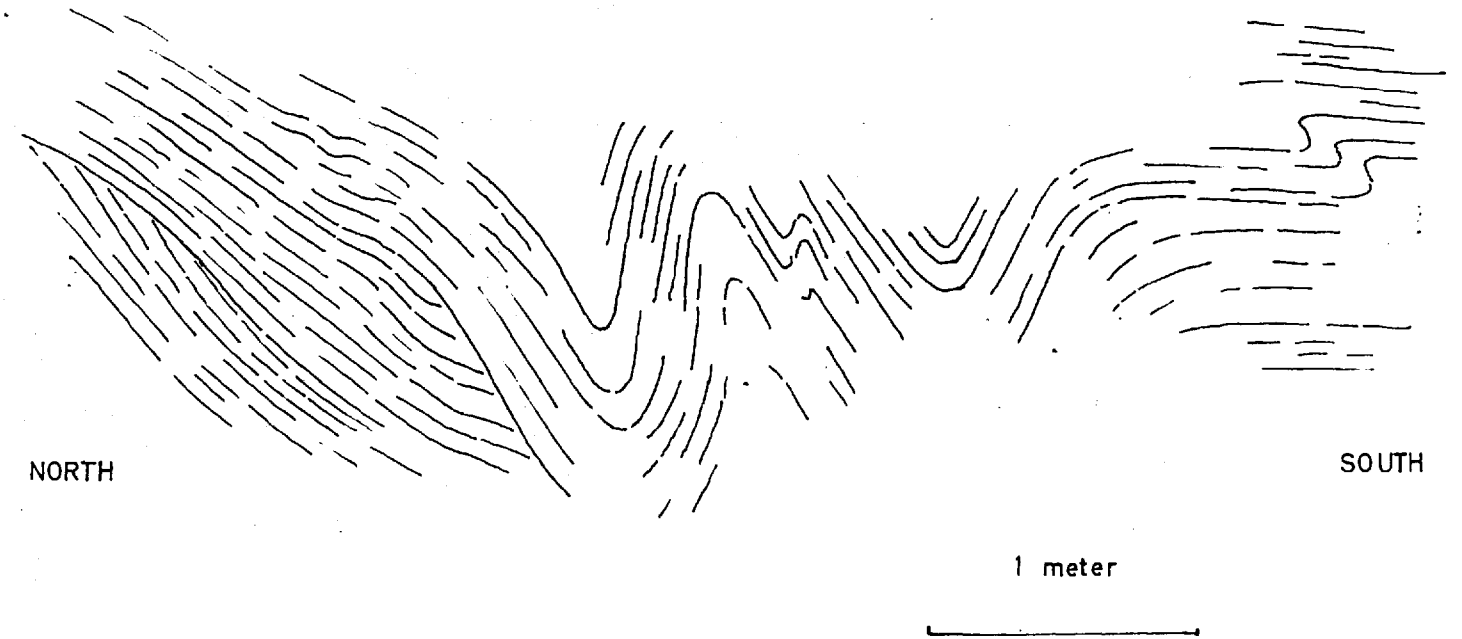
This complex phyllonite fabric has been folded and the lineation re-orientated by a succession of fold phases. At Rudha Rossel (857366) these late folds are locally tight structures of east - west trend, which are themselves then warped by late open folds. At Usinish Bay, the late structures consist of east - west trending kink-bands and crinkle folds, again warped by late structures. At Bolum, (828283) and Stuley Island (830233) the late structures are represented only by open warps producing dome and basin structures.

Somewhat late in this fold sequence there was a phase of quartz and quartz-feldspar injection. These intrusions appear as spidery irregular masses rather than as planar dyke-like bodies. They have suffered cataclasis and locally mylonitisation but generally their irregular shape is not due to deformation. They are useful in that they show that there was a late phase of mylonitisation separate from the early phyllonite producing phases. Locally these late mylonites are associated with thrust faults and several low dipping shears or faults can be recognised in most localities (see Kursten 1957). Generally, the last phase of deformation has been that producing steeply dipping faults which bring the "Usinish Mylonites" into contact with the Eastern Gneiss at Stuley, or the Mashed Gneiss at Usinish Bay. These faults and thrusts are probably partly responsible for the variation in thickness and width of outcrop of the "Usinish Mylonites" at different localities along the east coast.

Fig. 8.1

Section through 'f_4' structures within the Usinish Mylonites at Rudha Rossel (857365).

FIG. 8-1



Because of these late faults the true contact relationships between the phyllonites and the other rocks are not always seen. However, at Ornish, at Rudha Rossel and at Bolum the phyllonites appear to grade into the rocks above and below, though these gradations may be condensed by late mylonitisation. At Bolum there is an interesting section showing a full gradation between the phyllonites and Eastern Gneiss, where it can be seen that the phyllonite fabric appears to grade into the main phase fabric of the Eastern Gneiss. It would therefore appear that the phyllonites are the contact facies between the Eastern Gneiss and the Hashed Gneiss and are probably the result of deformation at the edge of the large igneous body which forms the Eastern Gneiss.

There are two hypotheses for the age of these phyllonitic rocks. Jehu and Craig (1925) suggested that they may be the result of intense deformation along a thrust plane. If this suggestion is correct then it would indicate that much of the fabric and large scale folds in the Eastern Gneiss as well as in the phyllonites were part of a complex deformation sequence which culminated in the formation of the Outer Isles Thrust. The other hypothesis is that the phyllonites formed much earlier than the thrust and were related to the main phase deformation in the Eastern Gneisses, i.e. were of Laxfordian age. Low grade metamorphism associated with intense deformation at the edge of a massive competent body would thus be the cause of the phyllonite formation, and these rocks could then be considered as being completely unrelated to the thrust deformation.

8.5 General structure of the Eastern Gneiss

The general structure of the Eastern Gneiss can be seen from the series of sections on Map 6. The structure consists of a large open f_3

antiform, whose westerly limb is steeply dipping, often overturned, and whose eastern limb has a more gentle dip to the east. The present trend of this antiform is north-north-east - south-south-west, but it is obvious from Map 6 that to a large extent this trend has been produced because the southern part of the structure has been thrust further to the west by more late crush zones. There has also been a certain amount of re-orientation by the f_4 folds, especially in the north. The plunge of th the structure is varied, due to f_4 re-orientation, but probably was originally low.

The eastern and upper edge of the Eastern Gneiss mass is bounded by a zone of phyllonites, the "Usinish Mylonites" of Jehu and Craig, who thought that they were the products of a thrust plane, higher than the Outer Isles Thrust. They show, however, a structural sequence very similar to that of the Eastern Gneisses and the phyllonite fabrics appear to be better developed equivalents of the Eastern Gneiss fabrics (section 8.4B). It therefore appears most likely that these rocks are the result of intense f_2 deformation at the edge of the Eastern Gneiss mass.

The structural history of the Eastern Gneiss can be summarised as follows:

1) The relatively massive and homogeneous body suffered a main phase deformation, resulting in a low dipping (or even horizontal?) cleavage being superimposed on the rocks, At the upper boundary of the mass, this cleavage was much more intense, resulting in the formation of phyllonites.

2) The mass was folded into a large antiformal structure. Generally no new planar fabric was formed, though a strong linear fabric can occasionally be found, resulting from the superimposition of the f_3

strain on the early cleavage. The Beinn nan Coarach shear zone may be of this age, possibly formed by a localised semi-brittle or plastic failure of the rock, resulting originally in a zone of simple shear. The phyllonites on the outer edge of the Eastern Gneiss mass were further deformed, possibly partly by flexural slip deformation on the flanks of the large f_3 antiform.

3) The northern region was then deformed by a group of east - west trending folds and shears. These affected both the massive Eastern Gneiss and the phyllonites. In the south, conditions were different and the f_4 deformation resulted in the development of a group of zones of intense deformation, again possibly originally forming as zones of simple shear.

This structural history is also given in Table 8.1, along with the most probable correlation with structures elsewhere in the Hebrides. I would suggest that all the structures referred to can be correlated with those of Laxfordian age to the west of the Outer Isles Thrust.

8.6 Metamorphism of the Eastern Gneiss

The main problem with the Eastern Gneiss metamorphic history concerns the age of the granulite facies metamorphism. Where the f_2 fabric is most intense, then the granulite facies minerals are retrogressed, and low amphibolite - greenschist facies assemblages result. When the f_2 fabric is almost non-existent, then equant grained granulite facies mineral assemblages occur. Certain granulite facies minerals appear to show a preferred orientation parallel to the f_2 cleavage, but on closer examination this mineral orientation can be seen to be due largely to a cataclastic deformation and not to a syn-tectonic recrystallisation. One can therefore conclude that prior to the f_2 deformation the Eastern

Gneiss was made up of a relatively equigranular, equant grained rock, with locally a certain amount of lithological banding. If the structural sequence from f_2 to f_4 is of Laxfordian age, then the granulite facies minerals may well be Scourian.

There is also a problem as to the age of the assemblages from the rocks which have been correlated with the Scourie-dyke suite. Possible hypotheses include:

a) The mineral assemblages in the dykes may be of the same age as those in the Eastern Gneiss s.s. This would indicate either

i) Granulite facies static recrystallisation following the period of Scourie-dyke intrusion

ii) The possibility that the dykes are not of Scourie-dyke age, but earlier.

b) The mineral assemblages may be wholly igneous and the fine- but equant- grained texture could be a deformed igneous fabric.

The Eastern Gneisses have an apparently non-igneous fabric, but this could be due to igneous crystallisation under unusual conditions. The amount of pre-Laxfordian deformation is unknown. The structures and quartzo-feldspathic material mentioned previously could have formed during the igneous activity and the Eastern Gneiss mass as a whole may be fairly young, say of Scourie-dyke age. I feel, however, that the equant grained mineralogy is due to a period of static metamorphic activity, giving garnet/clinopyroxene bearing assemblages and during which time iron was exsolved from much of the pyroxene lattice, to rim many of the crystals.

This static recrystallisation may appear as a granulite facies metamorphism producing pyroxene and plagioclase, etc. or as a high grade

amphibolite facies metamorphism with hornblende as the only mafic constituent. For reasons similar to those described in section 3.4, these alternations of pyroxene and hornblende bearing rocks are probably due to variations in water fugacity at the time of crystallisation. The brown colour of the hornblende and the stability of the pyroxene when present would indicate a general high grade of metamorphism, probably under or close to "granulite facies" conditions.

The presence of garnet and the occurence of garnet-clinopyroxene rocks may indicate a high pressure metamorphism - the clinopyroxene-garnet-granulite facies of de Waard (1965). However, quartz is absent from the Eastern Gneiss and Yoder and Tilley (1962) have shown that clinopyroxene -almandine/pyrope - plagioclase assemblages are stable in silica deficient rocks at normal granulite facies conditions. According to Green and Ringwood (1966), the reactions

anorthite + enstatite = aluminous diopside + quartz

anorthite + enstatite = almandine/pyrope + clinopyroxene

occur at a lower pressure for undersaturated rocks, than for those with excess quartz. Also according to Green and Ringwood (1966), a high $\frac{Fe}{Mg + Fe}$% would lower the pressure at which garnet will form.

Therefore the clinopyroxene, the clinopyroxene-garnet and the clinopyroxene-garnet-andesine assemblages of the Eastern Gneiss do not necessarily indicate a near eclogite or high pressure granulite facies, but are more likely to have formed under medium metamorphic pressures due to the low silica and high iron content of the gneisses.

Where any of the tectonite fabrics are well developed, the rock is made up of plagioclase (low anorthite content) plus green hornblende with some chlorite and other low grade minerals. The grade of meta-

morphism would thus appear to be in the low amphibolite - greenschist facies. On the east coast of South Uist, such low grade metamorphism has also altered areas of rock outside the higher deformation areas. If the correlation with the Western Gneisses is correct, then it would indicate a lowering of the f_2 Laxfordian metamorphic grade across the Outer Isles Thrust from west to east. This lowering of grade has already been indicated on Fig. 3.3 in section 3.4.

Pegmatites containing quartz, potash feldspar, plagioclase ($\pm$ biotite and magnetite) cut the Eastern Gneiss fabrics. They occur as east - west striking steeply dipping sheets ($\frac{1}{3}$ to 1 metre thick) and form local discontinuities along which later crush movements concentrated. Bands of crushed pegmatites are therefore found in relatively uncrushed Eastern Gneiss.

8.7 Origin of the Eastern Gneiss

From the homogeneous appearance of much of the Eastern Gneiss and from the mineralogical composition, and the structure, the rock would appear to be of igneous origin. Much of the lithological banding is probably a deformed original igneous layering, and the whole sequence from leuco-gneiss to ultramafic gneiss may be considered as part of a layered basic complex. There is of course too much deformation to estimate any "way-up" of the gneiss, but the more feldspathic and less banded rocks occupy the area in the outer and therefore structurally higher part of the f_3 antiform, while the banded gneisses lie in the core of the main fold. The mass was very large; taking into account the tectonic deformation, it must have originally been over a kilometre in thickness, and in length and breadth over 20 kilometres by at least

4 or 5 kilometres.

From a structural correlation across the thrust zone, the age of the rock would appear to be pre-Laxfordian, and if Dearnley (1962) is correct in correlating the dykes across the thrust zone, it is also probably of pre-Scourie-dyke age. Onepossible hypothesis is that the Eastern Gneisses were co-eval with the large basic masses of pre-Scourie-dyke age (section 7.2) in the Western Gneiss.

CHAPTER IX

THE OUTER ISLES CRUSH ZONE

9.1 Introduction

The Outer Isles Crush Zone extends from Sandray (south of Barra) to Tolsta Head, Lewis, a distance of 200 kilometres. 70 kilometres of this, however, lie beneath the sea, under the Minch, or in the Sounds of Barra, Eriskay and Harris, and many more kilometres are obscured by the thick bogs of Lewis. The main thrust zone tends to occupy the most inaccessible regions of the Outer Isles, both in the Uists and in the Parc area of Lewis, and generally forms a chain of hills down the east side of the islands. On South Uist it can be traced southwards from the small islands north of Ornish (855380), round the hills of Hecla, Corrodale and Beinn Mhor, where the irregular outcrop is due to the topography, then to the south over the hills Stulaval, Truiribheinn, Ben Kenneth, Roneval and on to Eriskay.

This statement, however, is very much a generalisation. The thrust zone is a whole region of crushed rock, in which some belts were more highly crushed than others, and often acted as fault planes. One of the most notable of these intensely crushed zones is that separating the Western Gneiss from the Eastern Gneiss, as recognised by Jehu and Craig (1925). It is not, however, the only belt of intensely crushed rock, nor is it probably the most important, as will be discussed later. The general distribution of crushed rock and thrust belts is shown on Map 6. The large faults and thrust zones can be traced for many kilometres along their strike and there are innumerable smaller crush zones, which because of their small scale have been left off this map. Between

Fig. 9.1

Thrust zones and fault blocks in the central region of South Uist. Thrust zones :-

A. The Eynort Crush Zone

B. The Meall Mhor Crush Zone

C. The Outer Isles Crush Zone

D. The Corrodale Crush Zone

E. The Liadale Crush Zone

Fault blocks :-

I The Foreland Block

II The Eynort Fault Block

III The Meall Mhor Block

IV The Ornish Fault Block

V The Hellisdale Block

VI The Liadale Block

U The Usinish Mylonites

M The Mashed Gneisses.

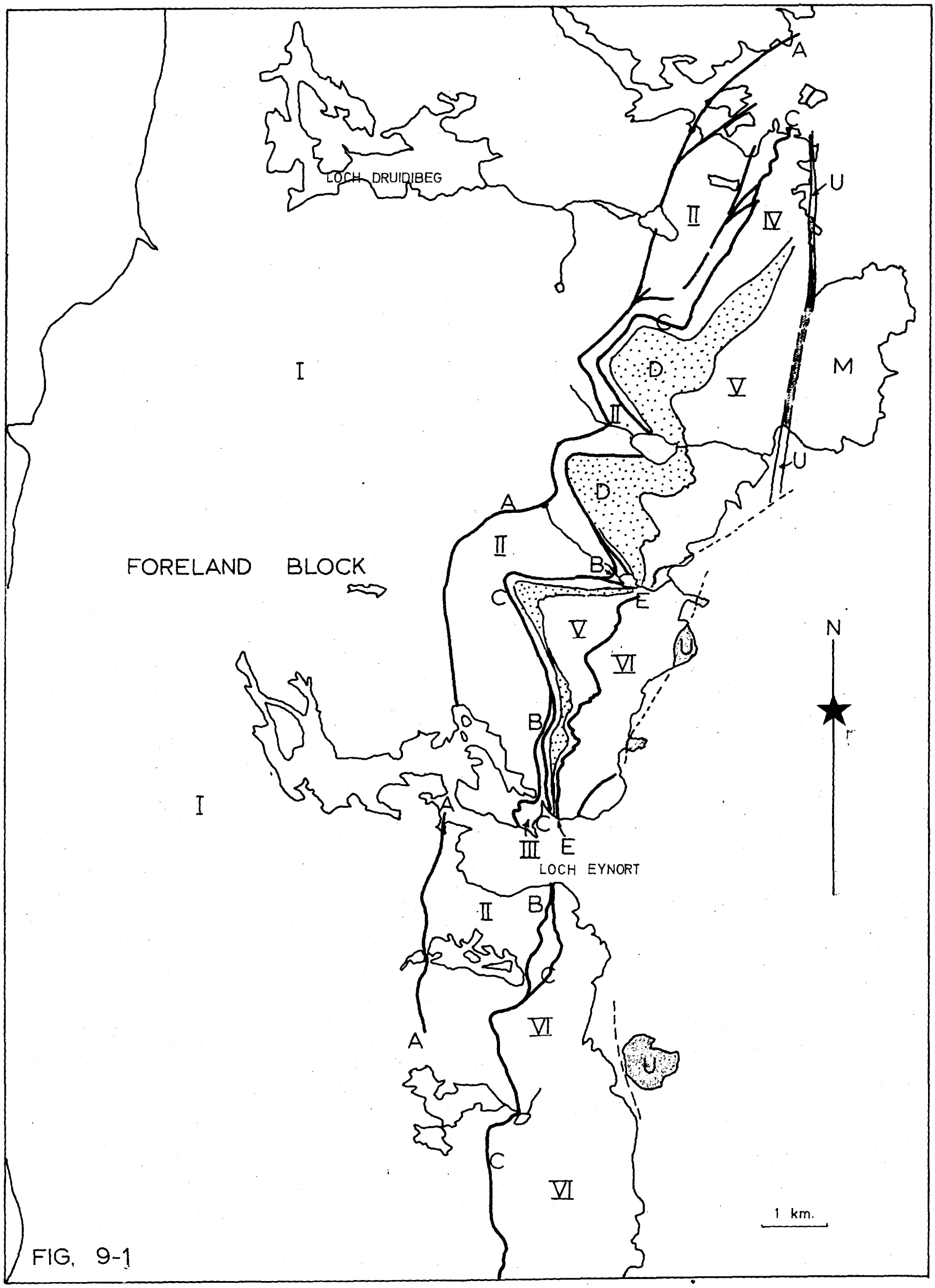

LOCH DRUIDIBEG
FORELAND BLOCK
LOCH EYNORT
I
II
III
IV
V
VI
A
B
C
D
E
U
M
N
1 km.

FIG. 9-1

the crush and fault belts, the rocks may be relatively undeformed and such masses can be considered as discrete fault blocks. Thus South Uist can be split up into a sequence of thrust zones and fault blocks, as shown in Fig. 9.1.

The aim of this chapter is not to give a description of all the structures and fault belts which formed during this period, not is it intended to give a detailed description of the rock "pseudotachylite" which occupies so much of the crush zones. I mainly intend to summarise briefly aa few points about the thrust zone and then to discuss their relevance to the characters of the Outer Isles Thrust as a whole. A certain amount of rather tedious description is unfortunately necessary and because of the lack of space many interesting points have had to be summarised.

The chapter will first discuss the rock types involved in the thrust zones. Second, it will describe and as far as possible acoount for some of the small-scale structures formed during this period. Third, the large-scale fractures will be mentioned and very brief descriptions given of the main thrust zones and fault blocks, and fourth, there will be a few comments on the rock pseudotachylite and its possible origin in South Uist.

9.2 The rocks of the crushed zones

The rocks which constitute the "Western Gneiss" and "Eastern Gneiss" of Dearnley (1962) and Jehu and Craig (1925) have been previously described and here only the rocks which are supposed to have formed or been given their essential characters by the period of crushing will be examined.

9.2A Crushed rock and pseudotachylite

Much of the rock close to the major thrusts or faults in South Uist is made up of a structureless mass of crushed gneiss. If the rock had any original fabric then it is either re-orientated or completely broken up. In the Western Gneiss, this leads to the formation of crush breccias on all scales, but in the Eastern Gneiss, where any foliation was originally very weak, all sign of structure tends to disappear.

Within the crushed gneiss most minerals show evidence of internal deformation by cold-working and by cataclasis. The plagioclase grains are often packed with strain twins which themselves may be deformed, and the mafic minerals, hornblende or pyroxene, often show beautiful examples of kink bands, chevron folds, etc.. Most minerals are fractured and granulated at the edge.

Locally, over much of the area but especially in the north round Loch Skiport, this deformation was accompanied by variable amounts of syn-tectonic recrystallisation. Incipient phyllonitisation was common, and thin veinlets of chlorite and/or epidote, often with sericite, show up localised movement planes. Along some of the dislocations south-east of Loch Skiport (839378) more extensive syn-tectonic recrystallisation produced flaggy phyllonitic/mylonitic rocks. There appears to be an inverse relationship between the amount of mylonite/phyllonite and the amount of pseudotachylite. Many of the smaller faults in the north of South Uist are dominantly phyllonitic while to the south phyllonite is less common but the faults are rich in pseudotachylite. Along the Outer Isles Thrust itself, the amount of phyllonitisation decreases from north to south, while the amount of pseudotachylite and mylonite sensu

strictu is greatly increased.

The pseudotachylite is a dark semi-isotropic mass of powdered (and/or de-vitrified) rock containing coarse rock powder, rock and mineral grains or breccia. There is a series of rock types from the isotropic pseudotachylite to coarse rock breccia with thin pseudotachylite vein ets. In the field the pseudotachylite occurs as a brown or more usually black rock, well-jointed and extremely hard and brittle. The maximum thickness of pure pseudotachylite occurs on the main thrust plane at Ben Kenneth (805194) where there is over seven metres of pure pseudotachylite without any visible grains of recognisable rock. This layer can be traced intermittently for over three kilometres along strike and probably ranks as the largest exposed mass of pure pseudotachylite anywhere in the world. Impure pseudotachylite, i.e. pseudotachylite with small rock grains or fragments, reaches thicknesses of over thirty metres on Ben na Hoe (813285) to the south of Ben Mhor, and on this hill the whole mass of pseudotachylite-bearing crushed rock is far thicker, being somewhat in the region of one hundred and fifty metres.

Tehu and Craig (1923 and 1925) considered pseudotachylite to be the product of frictional melting and thought that much of it was originally in the form of a glass, which has since devitrified. Kursten recorded an occurence of still undevitrified glass from the thrust plane on Ben Kenneth. I have found quite a bit of pseudotachylite which has a vitreous lustre, but when sectioned it always appeared as a dull, semi-isotropic mass. One has no evidence of the amount of pseudotachylite which may have been molten and which then produced a glass. Much of the rock may only have been a fine-grained powder or "ultra-mylonite". This problem will be returned to later, when more of the thrust zone structure

has been discussed.

9.2B Metamorphic rocks associated with the crushing

In the area round Loch Skiport, in both Western and Eastern Gneiss, the rocks suffered a late period of low-grade, greenschist or low amphibolite facies, metamorphism. Along the dislocation, this metamorphism caused a syn-tectonic recrystallisation, producing thin phyllonite/mylonite/folia and in the rest of the rocks there is often evidence of static recrystallisation producing chlorite, epidote and sericite. This late static retrogressive alteration is most common in rocks which hade already suffered some cataclastic deformation. In the Western Gneiss, the resulting rock is often a fabric-less messy brown (due to iron staining) gneiss.

In the south, a similar alteration can often be found close to some thrust planes; beneath the main thrust zone on Ben na Hoe there is a region of iron-stained and slightly altered rock. Alteration also occurs along sets of small fractures, locally abundant on the Western Gneiss, causing the rock close to the fracture to stand out as a well-marked 'rib'. The fact that alteration in the south is confined to faults or fractures, and the fact that there is always a well-developed iron staining along these fractures, would indicate localised metasomatic activity and not a regional metamorphic episode.

The widespread metamorphism of crush zone age is therefore confined to the northern region and occupies an area similar to that of the late Laxfordian greenschist/low amphibolite facies retrogression (see section 3.3D). In many areas it is impossible to distinguish between these two phases of retrogression and it is very tempting to think that they may be one and the same. If this is the case then the thrusting would have

occured at a time when the northern rocks were still cooling, but when the rocks to the south were in a cold and brittle condition. This has the implication that the thrusting occurred at some period very late in the Laxfordian sequence, before all the Laxfordian metamorphism had terminated, and would therefore put the thrust in the period of 1600 - 1000 m.y.

9.3 Small scale structures associated with the fault deformation

9.3A Visible faults or thrusts

In almost all the crush belts, there are dislocations across which the structures and/or petrology do not match, and sometimes a component of movement can be measured. These faults vary in width from a fraction of a centimetre upwards, with a similar variation in amount of apparent displacement.

In the fairly homogeneous Eastern Gneisses, these small-scale faults occur as regular, sub-parallel and often closely spaced structures. In the Western Gneisses, however, the small scale structures have a more nearly random orientation and grouping. The foliation plane often tends to act as a plane of weakness and thus the gneissose banding is often followed by crush planes, pseudotachylite crush breccia layers and phyllonite folia. Within crush zones in both Western and Eastern Gneisses, marked lithological boundaries almost always act as dislocation surfaces. It would appear, therefore, that the orientation of the minor faults and shears is partly controlled by the main anisotropies in the rock as well as by the principal stress orientations.

9.3B Folds

Where well-banded Western Gneisses are involved in crush zones,

small scale folds are common. Their wavelength varies from a few centimetres upwards. Most of the main faults and thrusts are accompanied by a re-orientation, often an overturning, of the foliation, and these folds can be used to examine the complex deformation history of this period. From their appearance, the folds seem to be fairly brittle structures and are often associated with small fault and fracture planes, sometimes accompanied by pseudotachylite, and it is not uncommon to find folded mylonites and sometimes folded earlier pseudotachylite veins. Some of the folds may have formed due to compression during brecciation, but others appear to be due to fault drag, and often curve round into the fault plane. It ought to be possible to tell the sense of movement along the fault planes using these folded structures, in the same way as one can tell the sense of movement from shear zones. Sometimes, however, the answers are not as one would expect. In the Parc district of Lewis, for example, small scale folds from the main thrust zone give a movement sense which would indicate a sense of movement downhill on the thrust plane. This sense is inconsistent with the other evidence of thrust movement. The folds, therefore, only show a component of fault movement at a particular time in the deformation sequence; they do not necessarily indicate the main movement direction.

9.3C Crush breccias

The gneiss within the crush belt is usually broken up into blocks, the blocks sometimes separated by crushed gneiss, but more usually by pseudotachylite. Blocks occasionally show some signs of corrosion by the pseudotachylite, but often they remain relatively fresh. They have often been rotated within the breccia, and commonly appear to be exotic relative to the adjacent gneiss. It would appear, therefore, that many of the blocks have suffered a certain amount of transportation along the

thrusts, and from the irregular disarray of the blocks there would also probably have been a localised volume increase.

The breccias are often localised when seen on individual fault planes. They often do not have a very wide extent and are separated by regions where the crush zone is very thin and marked only by a mylonite or thin pseudotachylite layer. Projecting from these thin layers, one often finds thin arms of pseudotachylite which appear to have been injected into fractures in the adjacent gneiss. This would indicate that the pseudotachylite has either moved of its own free will, or has been squeezed out of the movement plane. There therefore appear to be sections of the fault plane carrying breccias, along which there has been a localised volume increase or low normal stress, alternating with sections in which most of the pseudotachylite has been squeezed out, indicating a volume decrease or higher stress.

Such variations could exist if one considers the fault planes as originally having an irregular or curved surface. At any point, a period of very high stress normal to the fault plane would be followed by a period of relative stress release. During the period of high normal stress, sliding friction would be high, resulting in localised high temperatures and the production of mylonites and possibly pseudotachylite. Much of the pseudotachylite would be squeezed out from these regions of high stress. In the following period of low normal stress, blocks within the breccias could suffer rotation and transportation.

Such a mechanism could be important in the production of the crush breccias and large pseudotachylite masses on the major crush zones. A slight curvature of the major thrust planes would cause extra high stresses to develop locally, causing extreme brecciation and possibly

the squeezing out of large masses of pseudotachylite into adjacent areas of lower normal stress. This point will be returned to later, when the large crush belts are discussed.

9.3D Pseudotachylite veins and intrusions

Pseudotachylite may occur as an integral component of the crushed rock, or as a separate intrusive rock. It can often be seen to cross-cut structures in the gneiss, to cross-cut earlier crush breccias and to cross-cut earlier pseudotachylite. Intrusive pseudotachylite often has a flow banding distinguishable under the microscope, and this can often be used to distinguish intrusive pseudotachylite from that formed in situ, which generally is more streaky, being mixed or interlayered with mylonitic or occasionally phyllonitic material.

In many cases, however, it is very difficult to be certain whether or not the pseudotachylite was locally intruded. This problem becomes more acute when one considers the pseudotachylite veins away from visible dislocation planes. For distances of up to 10 metres on either side of the smaller crush zones, one often finds thin veinlets of pseudotachylite, occupying irregular shaped fractures in the rock. These fractures show no evidence of any lateral displacement and could not have produced pseudotachylite by sliding friction. A possibility is that the pseudotachylite was injected into these cracks. For this to be the case, the pseudotachylite must have been fairly mobile, and it is debatable as to whether a fluid composed only of the melt from normal quartzo-feldspathic gneisses would be able to be intruded for such long distances and still consolidate as glass throughout the whole system.

9.3E Small-scale fractures in the gneiss and pseudotachylite

Though these structures, the small-scale fractures and ribs, appear not to be the direct result of pseudotachylite-producing major dislocations, they still throw some light on the history of this late phase of brittle deformation and so will be considered here.

a) fractures or "joints" in the pseudotachylite and crushed gneiss

The pseudotachylite always carries several sets of fractures which intersect to give small rhomb-like blocks of rock. The shape of these rhombs varies with the spacing and the intersection angle of the different fractures. The orientation of the fractures is fairly consistent within local areas, but appears to be highly variable on a regional scale. There has been no regional systematic study of them, but there are a couple of points which ought to be mentioned:

(i) The fractures appear to be of several sets and have probably not all formed under the same stress conditions. One finds that sets of fractures which are subordinate in some areas are the dominant sets in adjacent areas.

(ii) At first sight many of the fractures would appear to have formed as joints during the late uplift of the area. Certain features on Ben Kenneth, however, suggest that this is not the case, and that the fractures formed during the sequence of thrust deformation. Within the mass of almost pure pseudotachylite, which forms a well-marked pavement on the northern slopes of the hill (808203) there is a set of closely-spaced parallel fractures, resembling a widely-spaced slaty cleavage. These fractures are themselves folded, and show a variety of fold shapes from open warps to chevron folds and kink bands. These folded fracture planes indicate that

not only did the fractures form well before the end of the deformation sequence, but also that the pseudotachylite has suffered a certain amount of strain after its solidification. Such strains would aid any processes of devitrification and it is no wonder that very little glass has been found.

b) Ribs

Small fractures and joints are common in all the Western Gneisses. Some may be related to the various Laxfordian deformation episodes, whilst others may be of thrust plane age or later. There is, however, one set of notable fractures along which there has been localised metasomatic activity, producing a rock of a different appearance and different mineral composition, a "rib". Occasionally second prder fractures can be found, associated with these ribs, or there may be a small amount of displacement, enabling one to get the sense of "movement". In many areas, the ribs occur in vertical conjugate sets, and from the few localities where observations on the sense of movement can be made it would appear that the fractures formed under compressive stress conditions.

Large numbers of observations on rib orientations were taken from different areas (Map 9) to see whether there was any systematic variation or any correspondance, throughout South Uist, or whether one could determine anything of the principal stress orientations.

It was found that the ribs occur in marked sets, which often give good peaks to the histogram (Map 9). When one set of ribs is developed preferentially to others, it is often at a high angle to the gneissose foliation, and in areas with a well-developed foliation, ribs do not often form close to the foliation orientation.

Apart from a few areas, notably near Ormaclett, the ribs measured were all close to vertical, and yet when plotted for the whole of South Uist they have an almost complete spread of strike directions, with small concentrations at N.70° and N.140°.

Their frequency, though related largely to the rock type and local stress state, is also variable on a regional scale. They are concentrated to the west and south-west of Loch Eynort, and also round Loch Skiport. They are absent from many parts of northern South Uist and from the hilly interiors. The ribs appear therefore to be concentrated in areas where there are the most topographical features (as seen from maps and aerial photographs) and probably the most large scale fractures and faults. The local rib orientations also appear to compare favourably with the orientations of the major fractures and so it is reasonable enough to claim that they are both related.

One can conclude therefore that the ribs are small-scale fractures formed under the same stress conditions as those which gave rise to the larger-scale vertical faults and fractures in South Uist. There may be more than one generation pf such fractures, formed under differing stress conditions, though a large number of fractures have orientations close to N.70° and N.140°. If one assumes that these sets form a conjugate pair, then one could make some estimation of the principal stress directions.

9.4 Large-scale structures

9.4A Large-scale folds

Beneath the thrust zone on the southern slopes of Beinn Mhor, and on the southern slopes of Stulaval, the foliation is re-orientated into a

large-scale fold (Map 1). I would suggest that this structure has been formed by the movement of the main thrust plane, as is the case with similar smaller-scale examples described previously. In both the examples mentioned above, the foliation was folded from a vertical north-west striking plane, through a synclinal structure to a north-easterly striking plane. This would indicate that the thrust had a certain component of movement which caused fault drag from south th north. Thus the thrust probably had a component of sinistral strike slip movement and the total movement was by no means up the dip of the thrust plane.

Within the Eastern Gneiss, the foliation often has an orientation parallel to the thrust planes when in the vicinity of the larger crush belts (Map 6) and it would appear that the foliation has been folded into this orientation.

9.4B Large-scale crush belts

The large crush belts, as shown on Map 6 and Fig. 9.1, divide the island into a number of different structural blocks, telescoping much of the Laxfordian structure and often producing thick belts of crushed rock and pseudotachylite. The most notable crush belt, as recognised by Jehu and Craig (1925) is that occupying the chain of hills; on this, crushed rock and pseudotachylite reach to thicknesses of over 200 metres. From its low dip, Jehu and Craig considered this crush belt to be a simple thrust, thrusting the Eastern Gneiss westwards. More detailed work has shown both that the movement may not be simply up-dip (see previous section 9.4A) but may be partially strike-slip, and also that the Jehu and Craig "Outer Isles Thrust Plane" is only one of a whole series of crush belts, which themselves are made of numerous smaller planes of deformation.

In this section, I shall therefore briefly examine some of the features of these large crush belts, taking as principal examples the features of the Meall Mhor and the Outer Isles Crush Zones (Fig. 9.1) which together form the main belt of crush along the South Uist hills. Following this, the large crush belts and their intervening blocks of uncrushed rock will be described.

Within many of the major crush belts in the Eastern Gneiss, one can see that movement was not confined to one plane, but occured along several closely spaced crush planes and this largely accounts for the great thickness of some of these zones.

The large thickness of crushed rock and pseudotachylite on the main crush belt on Stulaval, Beinn Mhor and Corrodale is probably due to a similar cause. Within the thick crush zone on the southern shoulder of Beinn Mhor and on Ben na Hoe (Map 6) several individual crush zones can be recognised, which coalesce to the north and south into a thick mass of crushed rock (the Meall Mhor Crush zone of Fig.9.1). Elsewhere along this main crush zone, individual smaller crush planes can only be recognised near the upper and lower edges, where they become more widely spaced. If one extends this observation, one could suggest that this Meall Mhor Crush Zone, and presumably most of the others, are made up of a large number of individual smaller crush planes which have crushed and re-crushed the rock, leaving a structureless but pseudotachylite-rich zone.

Along part of this main crush zone, to the south, between Stulaval and Ben Kenneth, one can trace a layer of almost pure pseudotachylite. On Ben Kenneth, this pseudotachylite has sharply defined edges and may have intruded the local crushed rock. However, on the southern slopes

of Stulaval, this layer is "pinched-out" and the whole thrust zone is reduced from a thickness of 50 metres to only 6 metres. The reasons for this are either:

(i) that the whole local thrust zone has been concentrated along a narrow plane

or (ii) that most of the crush zone has been faulted out by later movement(s) along a more gently inclined surface.

To the north, close to the summit of Stulaval, the crush zone widens again, though it is complicated by a set of east - west late faults. To the north of Stulaval and east of Loch Snigisclett, the outcrop of this main crush belt reaches a width of a half kilometre; the belt has a thickness of about 200 to 300 metres.

It is not only the small individual crush planes which coalesce or fault each other; the large crush belts appear to do the same. Kursten recognised that on Ben Kenneth a belt of crushed rock branched off from the main crush zone; similar features have been observed to the north. On Beinn Mhor the two principal crush zones coalesce. The lower of these two crush zones (the Meall Mhor-Crush Zone) has the steeper dip, and carries a very thick sheet of pseudotachylite and pseudotachylite crush breccia. This is separated by a zone of crushed Western Gneiss from the higher crush zone (the "Outer Isles Crush Zone") which has a gentler dip and is also the one which carries the Eastern Gneiss over the Western Gneiss.

One can conclude that the movement pulses of the Hebridean fault zones did not concentrate along a single plane as in many other fault systems. It has long been known that in most modern fault systems movement is not a long and continuous process, but takes place by a series

of short intermittent pulses, and generally occurs along a well-defined or a small group of well-defined dislocation surfaces, along which frictional resistance may be reduced by the high pore water pressure. In the Hebridean faults however, one is dealing with crystalline rocks where pore water pressure would be low or non-existent and the frictional resistance would therefore be very high. It is conceivable that if there was a large enough amount of fluid pseudotachylite, it would act as a relatively incompressible fluid and give its own "pore pressure", hence reducing frictional resistance. This process would only act locally and as soon as movement ceased, pseudotachylite would tend to solidify to form a hard and brittle rock. There would therefore be no clean dislocation plane on which a next movement pulse could act; the old fault plane would be "sealed up". New planes of failure would have to be opened up, their position and orientation depending partly on the heterogeneities within the crushed rock. Thus there would be produced a whole zone of crushed and pseudotachylitised rock.

As the various dislocation planes need not be parallel, small and large crush zones would be displaced and faulted out, and thus one can gain some idea of the sequence of the major crush belts and crush belt orientations in South Uist. However, before making any more comments about this sequence, the major crush zones and thrusts will be described. For descriptive purposes, South Uist can be divided into a series of crush zones and relatively uncrushed fault blocks as in Fig. 9.1. From west to east they are as follows:

I. The "Foreland" block

This area which lies to the west of the main thrust zones contains two principal groups of crush belts:

a) North-west - south-east trending, steeply dipping belts.

b) East - west trending belts with a gentle dip to the south

a) The north-west - south-east belts

These vertical crush belts show internal features very similar to those of the large "thrust" crush zones. They are made up of many small belts of crushed rock and pseudotachylite-rich belts, locally concentrated to make a crush zone. One such zone mentioned by Jehu and Craig (1925) is seen at Eudha Arvule (711311), where six individual crush belts, with a thickness of between 20 centimetres and several metres occur within a half-kilometre of coast section. This crush zone can be seen again along strike at Askernish and on the western slopes of the hill Askervein. Other similar crush zones occur on the north-west coast of Benbecula, near Baliwanich; to the west of Loch Eynort , from Loch Ollay to Arnaval; and on the hill Beinn Bheag Dheas. None of these structures have been observed displacing or crossing the thrust planes, and from the way the north-west - south-east trending structures are cut off by the thrusts, they can be considered to be earlier.

b) The east-west zones

Many of the east - west features with southerly dips seen on hills of central South Uist can probably be attributed to a group of faults. They are characterised by altered rock and occasionally by patches of mylonitic material, though no associated pseudotachylite has been found. It is usually difficult to match up the lithologies and the structures on either side, and this could indicate a certain amount of movement. They seem to be associated with the Eynort Crush Zone and may be similar in age to the faults with the same orientation on Maol Martaig, which will be discussed below.

A) The Eynort Crush Zone

This major crush zone can be traced from Stulaval in the south to Loch Skiport in the north, where is dies out in a splay of features bearing crushed rock or mylonite (as on Fig. 9.1). In the south it has a general strike of N.10°, and an easterly dip of about 60°, but in the north it has a strike of N.25° and a 45° to 50° dip. It is therefore curved and slightly "spoon shaped".

Along with this change in strike and dip, it also changes some of its other characteristics. In the area round Loch Eynort, there is a zone some 150 metres wide, of crushed and altered rock, about 20 metres of this being intensely crushed and pseudotachylitised. This zone makes a strongly marked topographical feature from Stulaval to Corrodale; and on the west side of Hecla it forms the base of a series of gentle crags, some 175 metres high, which in places are shot through with pseudotachylite and crushed rock. On the northern arm of Hecla, at 827355, there is over 100 metres of intensely crushed rock. This point, however, marks the break-up of the single crush zone into a large number of branching crush belts (Map 1 and Fig. 9.1). These smaller belts sometimes carry pseudotachylite, but are more usually associated with mylonites or phyllonites.

Some of these small belts curve round to form east - west trending features which appear to cross and displace the main thrust zone on Maol Martaig (Map 6), indicating a later age for this group of faults.

If my interpretation of these features is correct, then it would appear that the Eynort Crush Zone breaks up into a suite of branching and possibly second-order fractures. This enables one to obtain a sense of

movement for this crush zone and from this sense, a component of dextral strike - slip movement in the north is indicated.

It is difficult to ascertain the total amount of displacement. The Laxfordian structures usually do not match up across the crush zone, except in the area of Stulaval, south of which the crush zone dies out. On Hecla, it can be seen that one is dealing with totally different parts of the f_3 structure on opposite sides of the thrust (Map 1) and on Beinn Mhor there is no immediate correlation. I would suggest therefore that in the centre of the crush zone there has been a large amount of movement.

II The Eynort fault block

Apart from a few areas in Coire Rudale, Hecla, this block consists largely of relatively uncrushed Western Gneiss, which shows structural features very similar to those to the west. No major north-west - south-east crush zones have been recognised, though aerial photographs of the southern slopes of Beinn Mhor reveal a suite of closely spaced north-west - south-east fractures.

B) The Meall Mhor Crush Zone

This is the most striking crush zone in South Uist, and possibly in the whole of the Hebrides. It carries an enormous amount of crushed rock and pseudotachylite, over 200 metres thick, and is best developed on the peninsula of Meall Mhor at the eastern end of Loch Eynort. To the north, it is cut out by the gentler dipping Outer Isles Thrust Zone, but then reappears as an inlier in Glen Hellisdale. To the south, it merges and becomes indistinguishable from the above-mentioned Outer Isles Thrust. Many of the features in this zone have been discussed previously, at the beginning if this section.

III. The Meall Mhor block

This is a mass of highly crushed Western Gneiss, involved in the main crush zone, but separating the massive pseudotachylite-bearing Meall Mhor Crush Zone from the thinner crush zones and the Eastern Gneiss above.

No Laxfordian structures can be recognised.

C. The Outer Isles Thrust Zone

Following the nomenclature of Jehu and Craig (1925) and Dearnley (1962), the crush zone which separates the Eastern Gneisses from the Western Gneisses has been called the "Outer Isles Thrust". It has a gentler dip and a more nearly north-north-east - south-south-west strike than the adjacent crush zones, and so appears discordant on both maps and sections (Map 6). In the north, on Maol Martaig, it consists of about six or seven metres of crushed rock, shot through with phyllonitic mylonites, altered rock and locally a little pseudotachylite. On the hills of Ornish it makes a very poor topographical feature, and if it were not for the difference in lithology and structure above and below, it could easily be mistaken for a very minor crush belt, or even ignored. If the amount of pseudotachylite or crushed rock is related to the amount of thrust movement, then it could be considered that this feature played only a minor part in the telescoping of the eastern part of South Uist, much of the movement having been carried by the earlier steeper crush zones.

To the south of Loch Eynort the crush belts of this zone join forces with those of the lower zone, to give a much simpler but locally thick "Outer Isles Thrust" on the hills of Stulaval and Truirebheinn, though as has already been mentioned, on Ben Kenneth, some crush belts

again diverge. At some stage in the history of this crush zone, a large mass of pseudotachylite appears to have been partially injected into the crushed gneiss. This pseudotachylite is relatively pure, free from inclusions and must have resulted from either a period of very intense mylonitisation or from a region where the pseudotachylite breccia was re-worked many times. It is probably, therefore, not a coincidence that this homogeneous mass of pseudotachylite should occur in a region where many of the crush belts are localised , where the rocks have suffered an enormous number of crush phases.

IV. The Ornish fault block

This relatively uncrushed belt of rock forms only a thin strip in the Beinn Mhor area, but to the north it forms a wide zone of uncrushed Eastern Gneiss.

D. Corrodale Crush Zone

Most of the hill Corrodale and the eastern slopes of Hecla are made up of crushed Eastern Gneiss, which in the Glen Corrodale area reaches a thickness of over 300 metres. In this region, the crush belts have only rarely coincided to give any major crush zone, but appear to have been fairly widely spaced. The zone appears to die out to the north-east of Hecla and to the south, round Loch Eynort, occurs as a much narrower group of crush belts.

V and VI. The Hellisdale and Liadale blocks

These are regions of relatively uncrushed Eastern Gneiss, in which one can recognise the main Eastern Gneiss structures.

E. The Liadale Crush Zone

This crush zone separates the above two regions of Eastern Gneiss

(see Map 9). It has a north-east - south-west strike with a southerly dip of about 25°. It is probably that the deflection across the strike of the "Usinish Mylonites" and the main f_3 antiformal axial trace, between Usinish Bay and Rudha Hellisdale, may be associated with this crush zone.

To summarise:

1. The crush zones are made up of a large number of smaller crush belts which themselves are made up of a number of dislocation planes. These planes may be parallel but are more commonly oblique and hence later movement planes dislocate earlier planes. Similarly, later large crush zones dislocate the early zones.

2. Much of the eastern part of South Uist has suffered from crush deformation, small crush planes are relatively common. Generally, however, the major movement planes were concentrated into a few major crush zones, and so the most fault movement was confined to the large crush zones which now form the Meall Mhor - Outer Isles Zone.

3. If one considers the variations in amount of crush deformation along, instead of across, the strike, then the central part of South Uist appears to have suffered more deformation than the north or south. Both the Eynort Crush Zone and the zones in the Eastern Gneiss are best developed here, and the Meall Mhor - Outer Isles Zones are better developed here than anywhere else in the Hebrides.

9.4C Large-scale fractures

As has been mentioned previously, aerial photographs show many of the large scale fracture patterns throughout the island. From a study of these photographs one can see that there are two dominant sets of

fractures, one striking at between N.65° and N.90° and the other between N.130° and N.150°.

These fractures are not restricted to South Uist, but can be seen from aerial photographs, in the crushed rocks of eastern North Uist and the Parc district of Lewis. In all areas the same sets can be recognised, possibly indicating similar stress conditions throughout the Hebrides.

The fracture patterns change or stop against the various crush zones, and it would appear that they were earlier in origin than these zones.

9.4D Late faults

In the region of Maol Martaig, the thrust zone is deflected by a group of east - west vertical faults, and it is probable that these late structures are responsible for many of the east - west or west-north-west - east-south-east topographical features which cross this part of the Uists. In the region of Loch Eynort, the outcrops of the main crush zone are displaced by such faults, and since there is no corresponding displacement of the near vertical structures, it is supposed that the east - west faults had a dominantly vertical movement direction.

9.5 The sequence of large-scale fault deformation

C_1. The earliest structures to form appear to have been the north-west - south-east faults and crush zones in the 'Foreland' block. The large fractures as recognised from aerial photographs have possibly a similar age and similar orientations, and from the association mentioned previously it would appear that the ribs also date from this period. This

early deformation phase therefore involved fracturing on a widely variable scale, from tiny rib-producing fractures to large crush zones. It may be noted that almost identical structures can be found in the Lewisian of the Scottish mainland (Peach et al, 1907, Sutton and Watson, 1951, Park, 1964) where they pre-date the Torridonian sediments and therefore presumably formed during the late stages of the Laxfordian period.

C_2. The next crush zones had a north-east - south-west orientation and gentle dips (15° to 25°) to the east. They form most of the crush zones of the Uists. From the shape of the associated folds, it would appear that these thrusts suffered an important sinistral component of strike-slip movement from south to north and movement does not appear to have been directly up dip. Great thicknesses of crushed rock and pseudo tachylite were built up during this phase. Crush zones with a similar orientation occur on Tiree and on Iona and these could possibly be of the same age.

Later stages of this phase of crush deformation produced crush belts with a more nearly north-north-east strike and a gentle dip. These belts, in the form of the Outer Isles Crush Zone, cut through the earlier zones.

It is difficult to date the Eynort Crush Zone, relative to the gentler-dipping zones to the east, but some of the branching features on Maol Martaig may represent second order faults belonging to the Eynort Crush Zone, and these branching features appear to be later than the Outer Isles Crush Zone. This zone also would appear to have a dextral strike slip movement component, different from the earlier

crush zones, and it may well be that the principal stress orientations were changing in direction during the deformation. An alternative hypothesis is that the strike slip movement on the Eynort Crush Zone may not have been consistent for the whole of its length. It may have been dextral only in the northern part and sinistral in the southern part, giving maximum thrust movement, dominantly up-dip in the central part of the zone, at the curve in the spoon-shaped thrust zone. The steep dip of this crush zone relative to those to the east could be due to curving stress trajectories as first illustrated by Hafner (1951) and Hubbert (1951).

C_3 The last phase of movement led to the production of east - west to north-west - south-east normal faults, with downthrows in either direction. There is no evidence of pseudotachylite on any of these faults and they may be of a very much later date than the above mentioned thrust zones.

9.6 The problem of pseudotachylite production

I do not propose to go into this problem in any detail here, and I shall only summarise a few points which can be established in South Uist. The origin of pseudotachylite has been debated for years. Certain authors have held to the belief that the pseudotachylite was produced by gas fluxion, and that the pseudotachylite crush breccias have a mode of origin similar to many tuffisites and volcanic gas breccias (e.g. Roberts, 1966). Other authors, (e.g. Jehu and Craig, 1925), believed that the pseudotachylite was produced by frictional melting of part of the rock to produce a glass. Some of the evidence for pseudotachylite origin which is given in South Uist is summarised below:

1, The pseudotachylite is usually associated with crushed gneisses, cataclastic mylonites and zones of movement. Such movement or such cataclasis, could produce enough frictional energy, under certain conditions, to cause localised frictional melting. The amount of frictional energy could be locally increased if the movement planes were slightly curved.

Large thicknesses of crushed rock and pseudotachylite, however, may have been produced by a multitude of small localised movements rather than by one large thrust.

2. The pseudotachylite often appears to have been squeezed away from the crusi planes and generally appears to have been fluid enough to be able to intrude the adjacent gneiss. There is a problem, however, as to whether some of the pseudotachylite which occurs at some distance from the crush planes formed in situ or was intruded.

3. Country rock may or may not be assimilated into the pseudotachylite. The amount of assimilattion would probably have been related to the temperature of the pseudotachylite and the melting piont of the country rock and also to the time spent by the country rock within the pseudotachylite melt.

Blocks of gneiss, small rock fragments and mineral grains many be found as inclusions. Of the mineral grains, quartz and feldspar predominate.

4. Glass has been reported from the pseudotachylite of the Hebrides (Jehu and Craig, 1925; Kursten, 1957). I have found no evidence for the amount of this glass originally. The pseudotachylite, when free from inclusions, appears homogeneous throughout, even throughout the masses six metres thick. I find it very hard to believe that these

thick masses of pseudotachylite could have formed as a homogeneous glass throughout and it seems more likely that these masses of pseudotachylite consisted of a homogeneous paste of powdered material and possibly of molten rock. The ratio of powder to melt would vary from area to area depending on the conditions of formation, rock composition, etc.

5. No gas bubbles have been recorded from the pseudotachylite and the only evidence of the presence of gas is from localised patches of metasomatised rock close to some of the thrust planes, especially in the north. If the rock was only a powder, however, a gas phase many have been necessary to explain its apparently high mobility.

6. There is an inverse relationship between the amount of pseudotachylite and the amount of altered rock and phyllonite. Since both can be found in different positions along the strike of the same crush plane, then it is debatable whether such effects as rate of movement or depth of cover were wholly responsible for the variation. Such variation may also be due to differences in temperature or pore water pressure of the country rock.

7. When one examines the occurences of pseudotachylite on a world-wide scale, one is struck by the fact that it only occurs in any quantity in crystalline rocks where the rocks have or had a low pore water pressure. The effects of friction in these rocks would therefore be much greater than in rocks of relatively high pore water pressure (Rubey and Hubbert, 1959).

I would therefore suggest that pseudotachylite owes its origin to:

(i) low pore water pressures in the rocks

(ii) possibly to slight curvatures in the movement plane, thus

increasing the normal stress.

(iii) the brittle nature of the rock, causing the rock to deform by brittle failure and by a large number of small rapid movements, rather than by semi-brittle or plastic creep.

One can see that the problem of the origin of pseudotachylite is still a long way from being solved. The above discussion is really most unsatisfactory, as it only takes a superficial look as the issues involved. Francis, working on Barra, intends to discuss the problem in far more detail and so rather than cause a repetition of work I shall leave it to him.

9.7 Discussion

Comments on the thrust zone in the northern Hebrides will be published elsewhere and here it will be sufficient to say that the thrust zone in the northen part of the Hebrides contains relatively little pseudotachylite and is generally similar to the thrust zone in the northern part of South Uist, north-east of Hecla. It usually contains many phyllonite/mylonite veins and is associated with a great deal of low grade metamorphic alteration. As in South Uist, the thrust zone is made up of a great many individual movement and crush planes. Sometimes the thrust zone is wide, as north of the Parc area of Lewis, sometimes it is narrow and poorly defined as on Burrival, North Uist. The north-west - south-east trending faults can be recognieed throughout the Hebrides (Myers, 1968; Francis, 1969) as can the late east - west trending faults, and the latter form the main east - west trending features in the Parc area of Lewis.

Much of the thrust deformation and movement has, however, been

concentrated in the South, on South Uist and on Barra. The amount of movement is unknown, though from the structures adjacent to the crush zone an east - west shortening of at least two to three kilometres is indicated. Though no detailed work has been done in the northern part of the Hebrides, it may be that there the east - west telescoping is much less.

The age of the thrust is purely speculative. The thrust zone is parallel to the other major British thrust zone, but as mentioned in section 6.2, there are many other structures of various ages which have the same trend. The fact that the thrust is parallel to the Moine Thrust does not mean that they are both of exactly the same age. The only evidence for the age of the thrust zone has been given in section 9.2B where it was shown that the area of late alteration on the thrust zone is the same as the area of late Laxfordian retrogression and slow cooling. It may be that the thrust occured at a time when the northern rocks were still cooling sown, i.e. the thrust may be a very late Laxfordian event (? 1500 - 900 m.y.?)

CHAPTER X

SUMMARY OF RESULTS AND DISCUSSION

At the end of the main pre-Laxfordian period of deformation, the rocks of South Uist consisted of a fairly uniform group of gneisses, possibly increasing in competence in the lower structural levels (see Chapters III and VI). There were then two phases of igneous activity (see Chapter VII); the first produced the early amphibolites, the second, the Scourie-dykes. The number of these igneous bodies locally increased downwards into the lower structural levels and was partly responsible for a further increase in competence (see Chapter VI). At the beginning of the Laxfordian period, the rocks of South Uist were markedly heterogeneous and the history of the Laxfordian deformation hinges on these heterogeneities and their effects on the amount of deformation.

The variations in the amount of f_1 and f_2 deformation appear to have been due to variations in the homogeneity of the rock and the deformation was largely affected by the number, the orientation and the structure of the pre-Laxfordian intrusives. The f_2 folds originally formed about flat-lying axial planes and generally "faced" to the north-east. There is no major inverted f_2 limb in the northern part of South Uist (though rocks are inverted by the f_2 phase on the south coast of South Uist and on Eriskay (Francis, 1969) and on the western part of North Uist (Graham, 1969).

The f_3 phase produced large cross-folds and the upper and less competent parts of the f_2 structural pile were folded in the form of pinched-in synforms into the lower and more competent

rocks. Heterogenous deformation also affected the underlying "basement" and produced a major plunge culmination trending in a north-north-east - south-south-west direction across South Uist.

In the west the f_4 structures often have an axial plane orientation similar to that of the f_3 folds and the number of f_4 folds and the intensity of f_4 deformation is directly related to the intensity of late Laxfordian migmatisation; f_4 folds are absent over much of North Uist and Lewis where there was no migmatisation. The f_4 deformation phase on South Uist and Barra may therefore be co-incident with the later parts of the f_3 deformation elsewhere and may be due to the orogenic forces which produced the f_3 structures but under stress conditions which changed due to the large scale competence variations in the underlying migmatised and granitised rocks.

Some of the ideas put forward in this thesis may have some bearing on the interpretation of the Lewisian rocks of the mainland. The Laxfordian structures on the mainland generally have a plunge to the south-east, while those of the Outer Hebrides generally plunge to the north-west (Dearnley and Dunning, 1968, Coward et.al., 1970). The Minch, therefore, may be occupied by a plunge culmination very similar in age and origin to that found on South Uist.

Similarly, the form of the f_3 structures may find a counterpart on the mainland. On the mainland, there is a central mass of dense, pyroxene bearing rock, where Laxfordian deformation is largely confined to local zones of intense deformation: the "shear zones" of Sutton and Watson (1951). To the north, this central mass is bounded by a zone of intense deformation

while to the south, there are zones of relatively low deformation separated by synformal zones of higher deformation. The structure of the mainland could therefore be considered as analogous to that of the Hebrides, in that the regions of low deformation are bounded by tight synformal zones, trending north-west - south-east. Thus the main phase Laxfordian structures of the mainland as found in the Laxford Bridge, the Gairloch and the south Torridon areas, may perhaps be correlated with the f_3 structures of the Outer Hebrides.

The centres of the "f_3" antiformal regions in Barra and in the central zone of the mainland are occupied by dense, metasediment free, pyroxene bearing rocks. These "basement" rocks underlie the normal quartzo-feldspathic, sediment bearing rocks of the Western Gneiss and exposures of them in south-east Barra and in the central mass of the mainland are probably located where f_3 antiformal regions and plunge culminations cross. Where these "basement" rocks do not quite reach the surface, then the antiformal regions are occupied by Western Gneiss which has suffered only slight f_3 deformation. Such areas include the northern part of South Uist and Benbecula, central Lewis and probably part of the Glenelg area.

In the light of the above discussion, it may be worth reconsidering the Minch-fault hypothesis as put forward by Dearnley (1962) in which the central block of the mainland is correlated with a central block of the Hebrides. Dearnley considered that this central block of the Hebrides was composed dominantly of Scourian structures but it has been shown in this thesis that the area includes a great many Laxfordian structures of f_2 and f_4 age,

and also forms the core of a large f_3 antiform. It may be possible to correlate f_3 antiformal regions across the Minch, but I can see no reason why the central mass of the mainland should correlate with the antiformal region on South Uist and Benbecula, when there are other similar antiformal regions to the north (Coward et. al. 1970 Watson, 1969) With the present amount of evidence, I personally feel that a Minch-fault hypothesis is unnecessary for a correlation between the two areas of Lewisian.

Structures similar to the f_3 antiforms occur in basement areas elsewhere in the world and have sometimes been considered to be due to a gravity updoming (eg. Eskola 1949). It is difficult to consider the Lewisian structures as being the result of such a mechanism as in almost every example it can be shown that the f_3 antiformal core is occupied by the more dense material. The antiformal regions appear to have formed irrespective of local gravity tectonics and not because of them. It has also been shown (section 6.5) that gravity tectonics played no obvious part in the development of the f_4 structures even though f_4 deformation was associated with migmatisation.

The main theme of this thesis, however, concerns the heterogeneity of deformation within a basement area. Variations in amount of deformation have been shown to be due both to competence and compositional variations which existed prior to the deformation phase and also to syn-tectonic variations caused by metamorphic changes. Similar strain variations have

been noted from other basement areas, from the Scottish mainland and from south-east Greenland (Bridgewater and Gormsen, 1969) and appear to be an important feature of any basement regions. This thesis, therefore, has tried to show the importance of an analysis of such strain variations, not only for the understanding of the local structure but also for a possible insight into the origin or cause of the deformation. Gneissose regions, because of their general homogeneity and lack of major discontinuities (nappes, thrusts etc.), appear to be the ideal place to study such problems as "the amount of crustal shortening" and "the nature of orogenesis" and I believe that it will not be until more is known about the structural geology of basement regions that one will be able to understand and account for some of the large structures in the high level and more recent mountain belts.

REFERENCES

Badoux, H. (1963) Les belemnites tronconnees de Leytron (Valais), Bull.Lab.Geol.,Mineral.Geophys.Musee Geol.Univ.Lausanne, 138 : 1-7.

Biot, M.A. (1957) Folding instability of a layered viscoelastic medium under compression, Proc.Roy.Soc.(London), Ser.A, 242 : 444-454.

Biot, M.A. (1961) Theory of folding of stratified viscoelastic media and its implications in tectonics and orogenesis, Geol. Soc.Am.Bull., 72 : 1595-1620.

Biot, M.A. (1964) Theory of viscous buckling of multilayered fluids undergoing finite strain, Am.Inst.Phys.Mag., 7 : 855-859.

Biot, M.A. (1964) Theory of internal buckling of a confined multilayer structure, Geol.Soc.Am.Bull., 75 : 563-568.

Bowes, D.R. (1968) An orogenic interpretation of the Lewisian of Scotland, Report of 23rd Int.Geol.Congress, 4 : 225-236.

Bridgewater, D. & Gormsen, K. (1969) Geological reconnaissance of the Precambrian rocks of South East Greenland, G.G.U. Report, Nr. 19, 43-50.

Buddington, A.F. (1963) Isograds and the role of H_2O in metamorphic facies of orthogneisses of the Northwest Adirondack area, New York, Geol.Soc.Am.Bull., 74 : 1155-1182.

Cloos, E. (1947) Oolite deformation in the South Mountain Fold, Maryland, Geol.Soc.Am.Bull., 58 : 843-918.

Cobbold, P.R. (1969) Unpublished B.Sc. structural geology thesis, Imperial College, London.

Coward, M.P., Francis, P.W., Graham, R.H., Myers, J.S., Watson, J.V. (1969) Remnants of an early metasedimentary assemblage in the Lewisian complex of the Outer Hebrides, Proc.Geol.Assoc. 80 : in press.

Coward, M.P., Francis, P.W., Graham, R.H., Watson, J.V. (1970) Large-scale Laxfordian structures of the Outer Hebrides in relation to those of the Scottish mainland, in press.

Dearnley, R., (1962) An outline of the Lewisian complex of the Outer Hebrides in relation to that of the Scottish mainland, Q.Journ. Geol.Soc.Lond., 118 : 143-176.

Dearnley, R. (1963) The Lewisian complex of South Harris, Q.Journ. Geol.Soc.Lond., 119 : 243-307.

Dearnley, R., & Dunning, F.W. (1968) Metamorphosed and deformed pegmatites and basic dykes in the Lewisian complex of the Outer Hebrides and their geological significance, Q.Journ. Geol.Soc.Lond., 123 : 335-378.

Elliot, D. (1965) The quantitative mapping of directional minor structures, J.Geol., 73 : 865-880.

Eskola, P.E. (1939) in Barth, Correns, Eskola, Die Entstehung der Gesteine, Berlin, p.339.

Eskola, P.E. (1949) The problem of mantled gneiss domes, Q.Journ. Geol.Soc.Lond., 104 : 461-476.

Evams, C.R. (1965) Geochronology of the Lewisian basement near Lochinver, Sutherland, Nature, London, 207 : 54-56.

Flinn, D. (1962) On folding during three-dimensional progressive deformation, Q.Journ.Geol.Soc.Lond., 118 : 385-433.

Flinn, D. (1965) On the symmetry principal and the deformation ellipsoid, Geol.Mag., 102 : 36-45.

Flinn,D. (1965) Deformation in metamorphism in W.S.Pitcher and G.W.Flinn eds. Controls of metamorphism, Oliver & Boyd, London.

Francis,P.W. (1969) Unpublished Ph.D. thesis, Univsity of London.

Fyfe,W.S.,Turner,F.J.,Verhoogen.J. (1958) Metamorphic reactions and metamorphic facies, Geol.Soc.Am.Memoir, 73.

Giletti,B.J.,Moorbath,S.,Lambert,R.St.J., (1961) A geochronological study of the metamorphic complexes of the Scottish Highlands, Q.Journ.Geol.Soc.Lond., 117 : 233-264.

Graham,R.H. (1969) Unpublished Ph.D. thesis, University of London.

Goguel,J. (1962) Tectonics W.H.Freeman & Co, New York.

Green, D.H. & Ringwood,A.E. (1966) An experimental investigation of the gabbro to eclogite transformation and its petrological applications, Aust.Nat.University, Dept.of Geophys.& Geochem. Publication no. 444 : 1-59.

Hafner,W. (1951) Stress distributions and faulting, Geol.Soc.Am. Bull., 62 : 373-398.

Hubbert,H.K. (1951) Mechanical basis for certain familiar geologic structures, Geol.Soc.Am.Bull., 62 : 355-372.

Huddleston,P.J. (1969) Unpublished Ph.D. thesis, University of London

Jehu,T.J. & Craig,R.M. (1923) Geology of the Outer Hebrides, Part I, The Barra Isles, Trans.Roy.Soc.Edin. 53 : 615-641.

Jehu,T.J. & Craig,R.M. (1925) Geology of the Outer Hebrides, Part II, South Uist and Eriskay, Trans.Roy.Soc.Edin. 54 : 615-641.

Kursten,M. (1957) The metamorphic and tectonic history of parts of the Outer Hebrides, Trans.Edin.Geol.Soc. 17 : 1-31.

O'Hara,M.J. (1961) Petrology of the Scourie dyke, Sutherland, Min. Mag. 32 : 848-865.

Peach,B.N. and others, (1907) The geological structure of the north-west Highlands of Scotland, Mem.Geol.Surv.G.B..

Ramberg,H. (1959) Evolution of ptygmatic folding, Norsk.Geol.Tidsskr. 39 : 99-151.

Ramberg,H. (1960) Relationships between length of arc and thickness of ptygmatically folded veins, Am.J.Sci., 258 : 36-46.

Ramberg,H. (1963) Evolution of drag folds, Geol.Mag., 100 : 97-106.

Ramberg,H. (1964) Selective buckling of composite layers with contrasted rheological properties; a theory for simultaneous formation of several orders of folds, Tectonophysics, 1 : 307-341

Ramsay,J.G. (1962) The geometry and mechanics of formation of "similar" type folds, J.Geol., 70 : 309-327.

Ramsay,J.G. (1963) Structure and metamorphism of the Moine and Lewisian rocks of the Northwest Caledonides, in Johnson and Stewart (eds.), "The British Caledonides", Oliver & Boyd Ltd., Edinburgh and London.

Ramsay,J.G. (1967) Folding and fracturing of rocks, McGraw-Hill Book Company, New York.

Ramsay,J.G. & Graham,R.H. (1970 in preparation).

Roberts,J.L. (1966) The emplacement of the main Glencoe fault intrusion at Stob Mhic Martain, Geol.Mag., 103 : 299-316.

Rubey,W.W. & Hubbert,M.K. (1959) Role of fluid pressure in mechanics of overthrust faulting, II., Geol.Soc.Am.Bull., 70 : 167-206.

Sherwin,J. & Chapple,W.M. (1968) Wavelengths of single layer folds: a comparison between theory and observation, Am.J.Sci., 266 : 167-179.

Sutton,J. & Watson,J.V. (1951) The pre-Torridonian metamorphic history of the Loch Torridon and Scourie areas in the north-west Highlands, and its bearing on the chronological classification of the Lewisian, Q.Journ.Geol.Soc.Lond., 106 : 241-307.

Talbot,C. J.(1966) Unpublished Ph.D. thesis, University of Leeds.

Tan,B.K. (1969) Unpublished Ph.D. thesis, University of London.

Turner,F.J. & Verhoogen,J. (1960) Igneous and metamorphic petrology, McGraw-Hill Book Company, New York.

Waard,D.de, (1965) A proposed subdivision of the granulite facies, Am.J.Sci., 263 : 455-461.

Watson,J.V. (1965) Lewisian, in G.Y.Craig (ed.) The geology of Scotland, Oliver and Boyd, Edinburgh, p.49-77.

Watson,J.V. (1968) Post-Scourian metadolerites in relation to Laxfordian deformation in Great Bernera, Outer Hebrides, Scott. J.Geol., 4 : 53-67.

Watson,J.V. (1969) The Precambrian gneiss complex of Ness, Lewis, in relation to the effects of Laxfordian regeneration, Scott.J.Geol. 5 : in press.

Winkler,H.G.F.(1965) Petrogenesis of metamorphic rocks, Springer-Verlag, Berlin.

Yoder,H.S.Jr. & Tilley, C.E. (1962) Origin of basalt magmas: an experimental study of natural and synthetic rock systems, J.Petrology, 3 : 342-532.

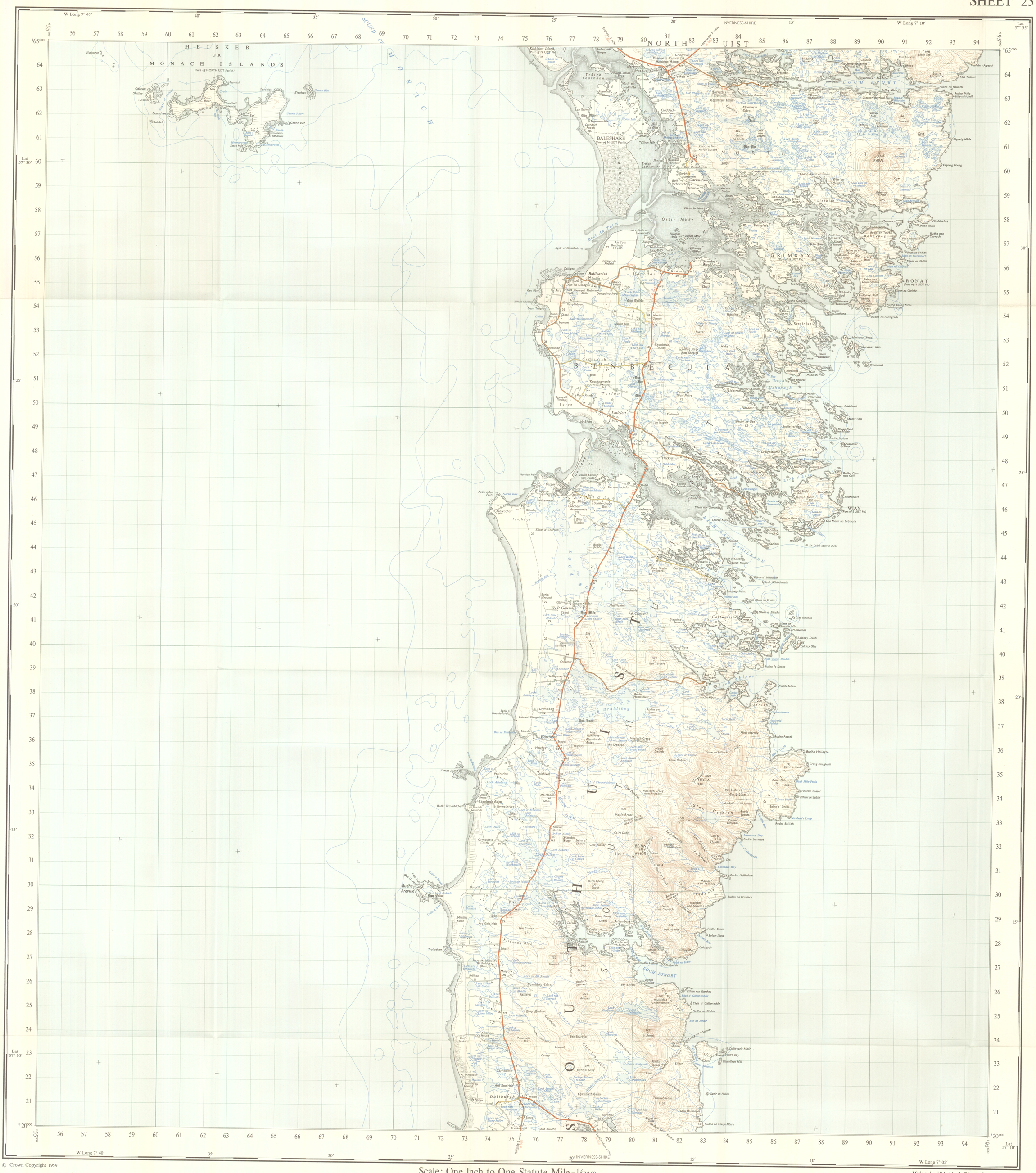
M. P. COWARD D.I.C. 1969

ORDNANCE SURVEY of GREAT BRITAIN
One Inch to One Mile Map (Seventh Series)

SOUTH UIST

MAP 2

SHEET 23



© Crown Copyright 1959

REVISION DIAGRAM

Fully Revised 1956
Road revision with later corrections
Revised from Air Photos

Index to adjoining Sheets. Hatching denotes overlaps

Contours in Lochs are given in feet and are taken from the Bathymetrical Survey of Fresh Water Lochs of Scotland
The Submarine Contours are given in Fathoms and are taken from the Soundings of Admiralty Surveys

PRICE:—3/6 net.

This map is also available in Folded or Mounted and Folded styles.

Difference from Grid North

(1) True North	
4° 52′ E	N W Corner
4° 19′ E	N E ,,
4° 48′ E	S W ,,
4° 15′ E	S E ,,

(2) Mag North
about 8¾° W in 1959
decreasing by about ⅓′ in four years

Magnetic North, Grid North, True North

Scale: One Inch to One Statute Mile = 1/63360

1 ¼ ½ ¼ 0 1 2 3 4 5 Miles

1,000 0 1,000 2,000 3,000 4,000 5,000 6,000 7,000 8,000 9,000 Yards

1 0 1 2 3 4 5 6 7 8 Kilometres

TO GIVE A GRID REFERENCE CORRECT TO 100 METRES

The incidence of grid letters and numbers on this sheet

NF (06)

EXAMPLE School
See diagram on left for Grid Letters. They are NF

East		North	
Take west edge of kilometre square in which point lies and read the large figures printed opposite this line on north or south margins.	74	Take south edge of kilometre square in which point lies and read the large figures printed opposite this line on east or west margins.	27
Estimate tenths Eastwards	4	Estimate tenths Northwards	4
	744		274
Full 100 Metre Reference NF 744274			

The above Full Reference is unique. For many purposes the first grid letter can be omitted, giving a reference F 744274 which recurs at intervals of 500 Kilometres. If both grid letters are omitted, the resulting reference 744274 recurs at intervals of 100 Kilometres. When the area concerned is sufficiently restricted, as will usually be the case with maps on scales of one inch to the mile and larger, both the grid letters are normally omitted.

Made and published by the Director General of the ORDNANCE SURVEY, CHESSINGTON, SURREY, 1959

Roads: Ministry of Transport, Trunk A 87 (T); Class 1 A 865; 2 B 891; 14 ft of Metalling & over (not included above); Under 14ft of Metalling, Tarred TOLL; Untarred Gate; Minor Roads in towns, Drives and Unmetalled Roads (Unfenced Roads are shown by pecked line); Steep Gradients 1 in 5 or steeper; 1 in 7 to under 1 in 5; Footpaths and Tracks

Railways: Multiple Track; Single; Narrow Gauge; Principal Station; Bridge over; Tunnel; Cutting; Footbridge; Viaduct; Station; Station (closed); Level Crossing; Embankment; Bridge under; Mineral Lines, Sidings & Tramways

Wood; Orchard; Parks and Ornamental Grounds; National Trust (always open) for Scotland NTS; (opening restricted) NTS

Electricity Transmission Lines (with pylons spaced conventionally); Pipe Lines (water); Arrow indicates direction of flow; Highest point to which Tides flow; Sand Hills; Cliffs; Marsh; Glasshouses; Canal; Aqueduct; Lock; Weir; Sluice; Flat Rocks; Beacon; Rough Pasture; Quarry; Lake; Bridge; Ferry; For Vehicles (V) or Foot Passengers (P); Boundaries, National; Open Pit; Contours at 50 ft intervals; Submarine Contours in Fathoms; County; Parish; County with Parish, Co of City; Heights in feet above Mean Sea Level (surveyed by levelling 275; not surveyed by levelling (109); Triangulation Station; Intersection, Lat & Long at 5′ intervals (not shown where it confuses important detail); Youth Hostel; Bus & Coach Stations

Church or Chapel with Tower; ,, ,, ,, Spire; ,, ,, ,, without either; Wireless Aerial Mast; Windpump; Windmills (in use) (disused); Lighthouse; Lightship; Post Office P; Telephone Kiosk P O T, A A A, R A C R

Antiquities: Roman AD 43 to AD 420 VILLA; Other Antiquities Tumulus, Castle; Site of Antiquity; Site of Battle

The representation on this Map of a Road, Track, or Footpath is no evidence of the existence of a right of way.

SHEET 23
A

FOR REVISED PRICE SEE CURRENT PRICE LIST.

SHEET 23

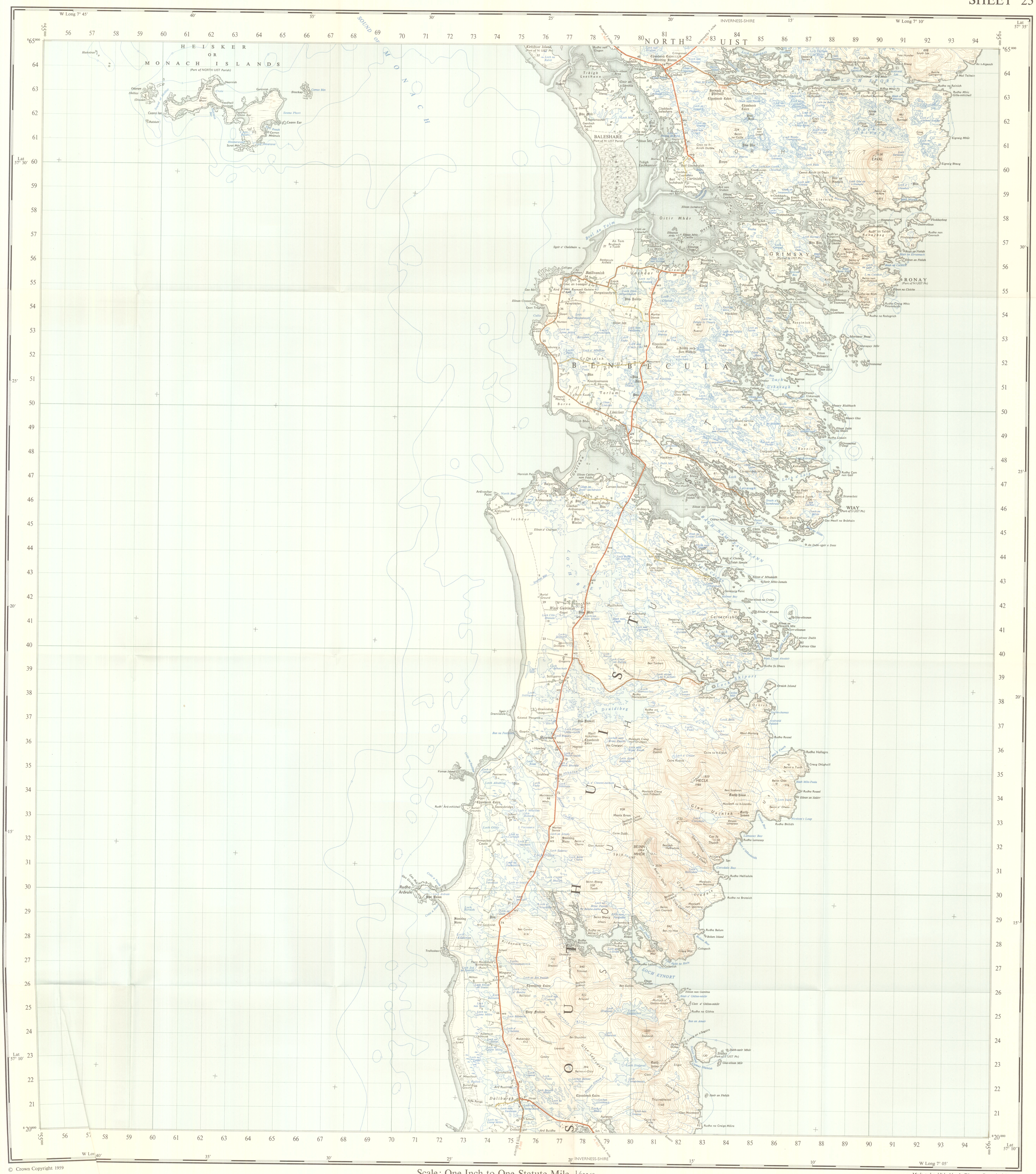
M.P. COWARD Ph.D. 1969

MAP 2

ORDNANCE SURVEY of GREAT BRITAIN
One Inch to One Mile Map (Seventh Series)

SOUTH UIST

SHEET 23



© Crown Copyright 1959

REVISION DIAGRAM

Fully Revised 1956
Road revision with later corrections
Revised from Air Photos

Index to Hatching & Sheets, overlaps

Contours in Lochs are given in feet and are taken from the Bathymetrical Survey of Fresh Water Lochs of Scotland
The Submarine Contours are given in Fathoms and are taken from the Soundings of Admiralty Surveys

PRICE:—3/6 net.

This map is also available in Folded or Mounted and Folded styles.

SHEET 23
A

REVISED PRICE 5s.
CURRENT PRICE LIST.

Difference from Grid North

(1) True North	
4° 52′ E	N W Corner
4° 19′ E	N E ,,
4° 48′ E	S W ,,
4° 15′ E	S E ,,

(2) Mag North
about 4¾° W in 1959
decreasing by about ¼° in four years

True North, Grid North, Magnetic North

Scale: One Inch to One Statute Mile = 1/63360

1 2 3 4 0 1 2 3 4 5 Miles

1,000 0 1,000 2,000 3,000 4,000 5,000 6,000 7,000 8,000 9,000 Yards

1 0 1 2 3 4 5 6 7 8 Kilometres

TO GIVE A GRID REFERENCE CORRECT TO 100 METRES

The incidence of grid letters and numbers on this sheet

NF (08)

EXAMPLE

See diagram on left for Grid Letters. They are NF

East		North	
Take west edge of kilometre square in which point lies and read the large figures printed opposite this line on north or south margins. Estimate tenths Eastwards	744 4 744	*Take south edge of kilometre square in which point lies and read the large figures printed opposite this line on east or west margins. Estimate tenths Northwards*	27 4 274

Full 100 Metre Reference NF 744274

The above Full Reference is unique. For many purposes the first grid letter can be omitted, giving a reference F 744274 which recurs at intervals of 500 Kilometres. If both grid letters are omitted, the resulting reference 744274 recurs at intervals of 100 Kilometres. When the area concerned is sufficiently restricted, as will usually be the case with maps on scales of one inch to the mile and larger, both the grid letters are normally omitted.

Made and published by the Director General *of the* ORDNANCE SURVEY, CHESSINGTON, SURREY, 1959

Roads — Ministry of Transport, Trunk A 87 (T); Class 1 A 865; 2 B 893; 14 ft of Metalling & over (not included above); Under 14ft of Metalling, Tarred TOLL; Untarred Gate; Minor Roads in towns, Drives and Unmetalled Roads (Unfenced Roads are shown by pecked lines); Steep Gradients 1 in 5 or steeper (1 in 7 to under 1 in 5); Footpaths and Tracks

Railways — Multiple Track; Single; Narrow Gauge; Viaduct; Principal Station; Station; Station (closed); Bridge over; Level Crossing; Tunnel; Embankment; Cutting; Bridge under; Footbridge

Wood; Orchard; Parks and Ornamental Grounds; National Trust (always open) for Scotland (NTS); (opening restricted) (NTS)

Electricity Transmission Lines (with pylons spaced conventionally); Pipe Lines (water); Arrow indicates direction of flow; Highest point to which Tides flow; Sand Hills; Cliffs; Marsh; Glasshouses; Canal; Aqueduct; Lock; Weir; Flat Rocks; Beacon; Rough Pasture; Quarry; Open Pit; Bridge; Ferry; For Vehicles (V) For Foot Passengers (P); Boundaries, National; County; Parish; County with Parish, Co of City; Contours at 50 ft intervals; Submarine Contours in Fathoms; Bus & Coach Stations

Church or Chapel with Tower; ,, ,, ,, Spire; ,, ,, ,, without either; Wireless Aerial Mast; Windpump; Windmills (in use) (disused); Lighthouse; Lightship; Post Office P; Telephone Kiosk P O T, A A A, B A C R

Antiquities: Roman AD 43 to AD 420 VILLA; Other Antiquities Tumulus, Cairn; Site of Antiquity; Site of Battle

The representation on this Map of a Road, Track, or Footpath is no evidence of the existence of a right of way.

SHEET 23

MAP 1

NORTHERN SHEET.

Country-rock gneiss

Metasediment

Early Amphibolite

Scourie-dyke

Late Granite and Pegmatite

Thrust or fault zone

F_3 Axial-trace

F_4 Axial-trace

Scale - 1-10560

N

Ardmore

Gashernish

LOCH CARNAN

Carnan

Holmar Bay

Tonacheois

LOCH BEE

Calternish

LOCH SHEILAVAIG

East Geirinish

Ath Cumhang

Caolas Luirsay

LOCH SKIPORT

Rueval

Skiport

Ornish

Loch Teanga

Loch Bein

Loch Spotal

LOCH DRUIDIBEG

Maoil Daimh

Coire Rudale

Haarsal

HECLA

GLEN USINISH

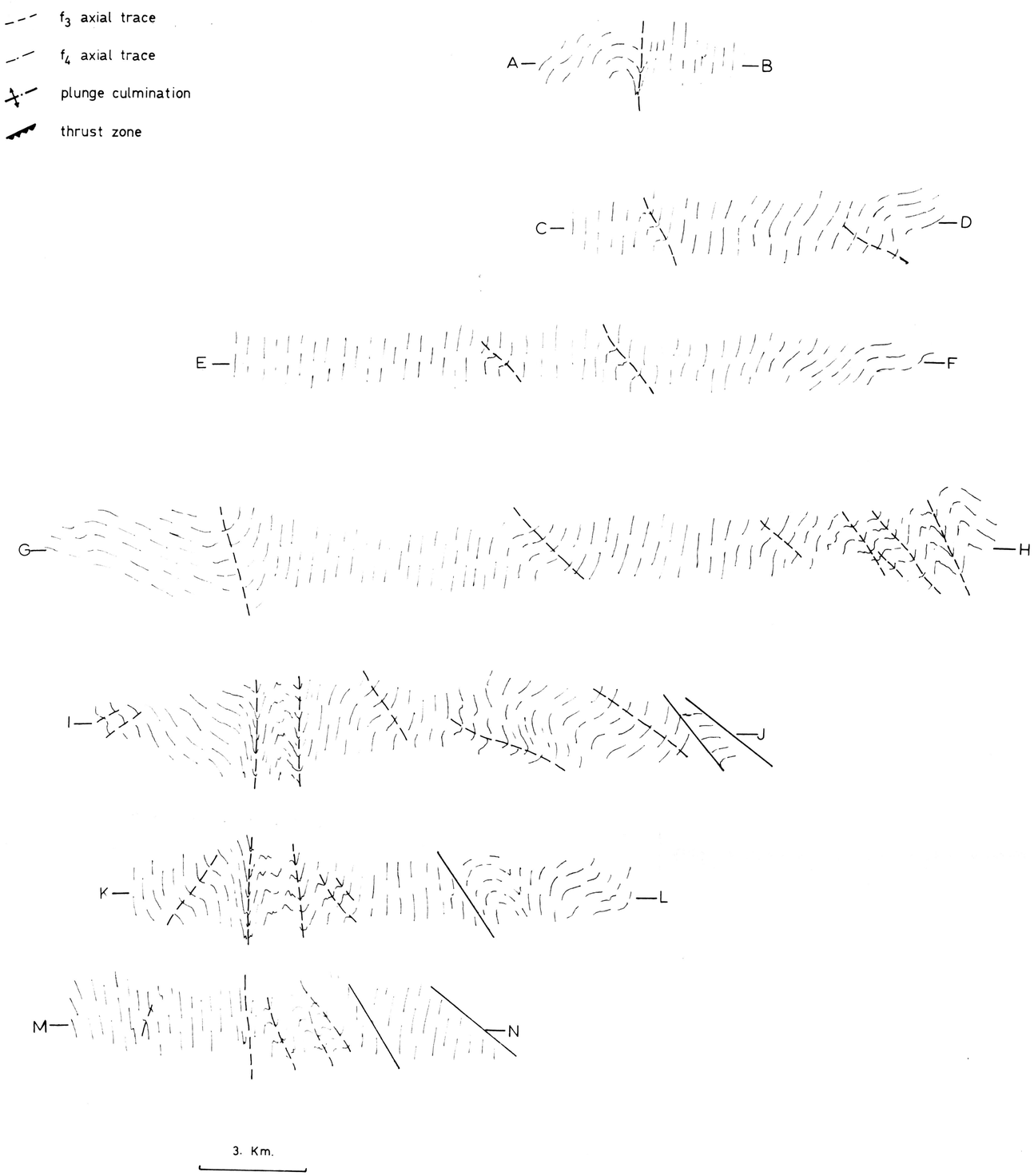
1 km.

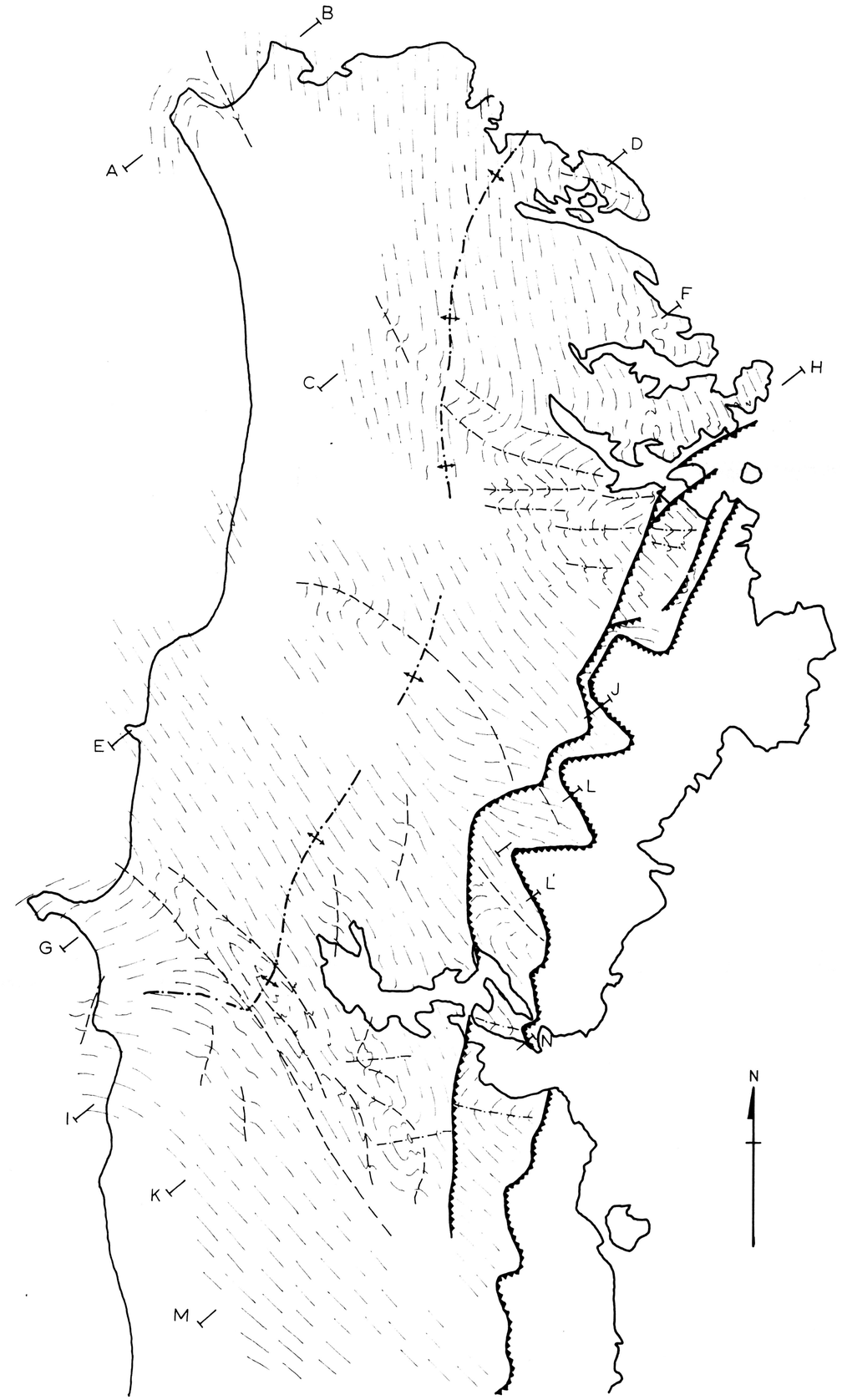
M.P.COWARD Ph.D. 1969

MAP 1

SOUTHERN SHEET

STRUCTURAL PROFILES THROUGH SOUTH UIST



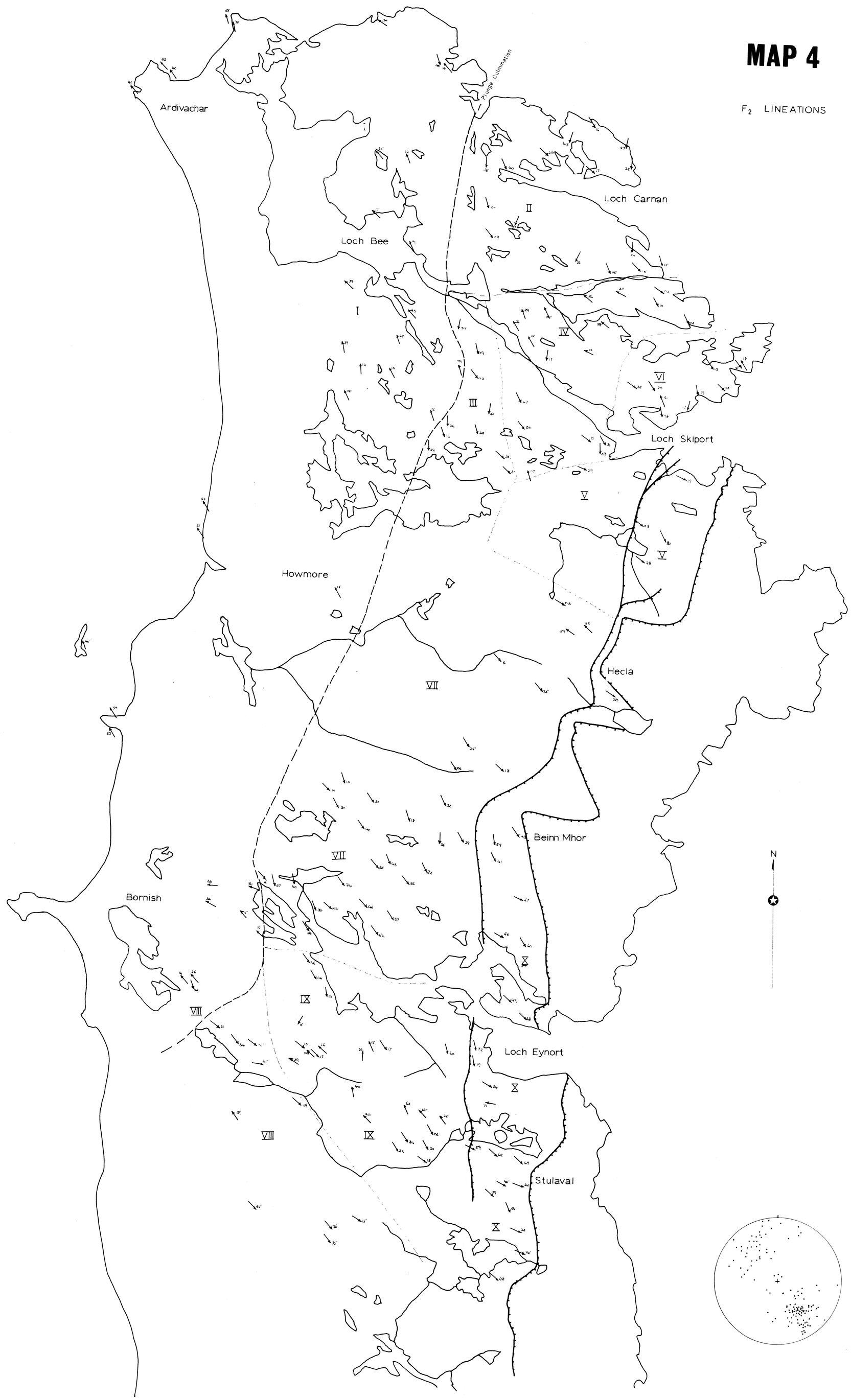


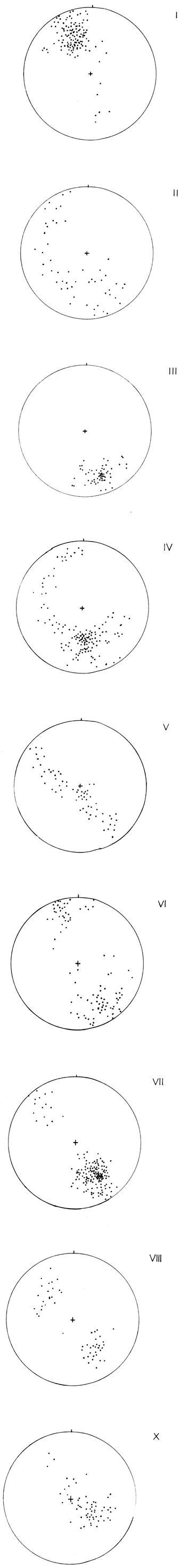
MAP 3

M.P.COWARD Ph.D. 1969

MAP 4

F_2 LINEATIONS

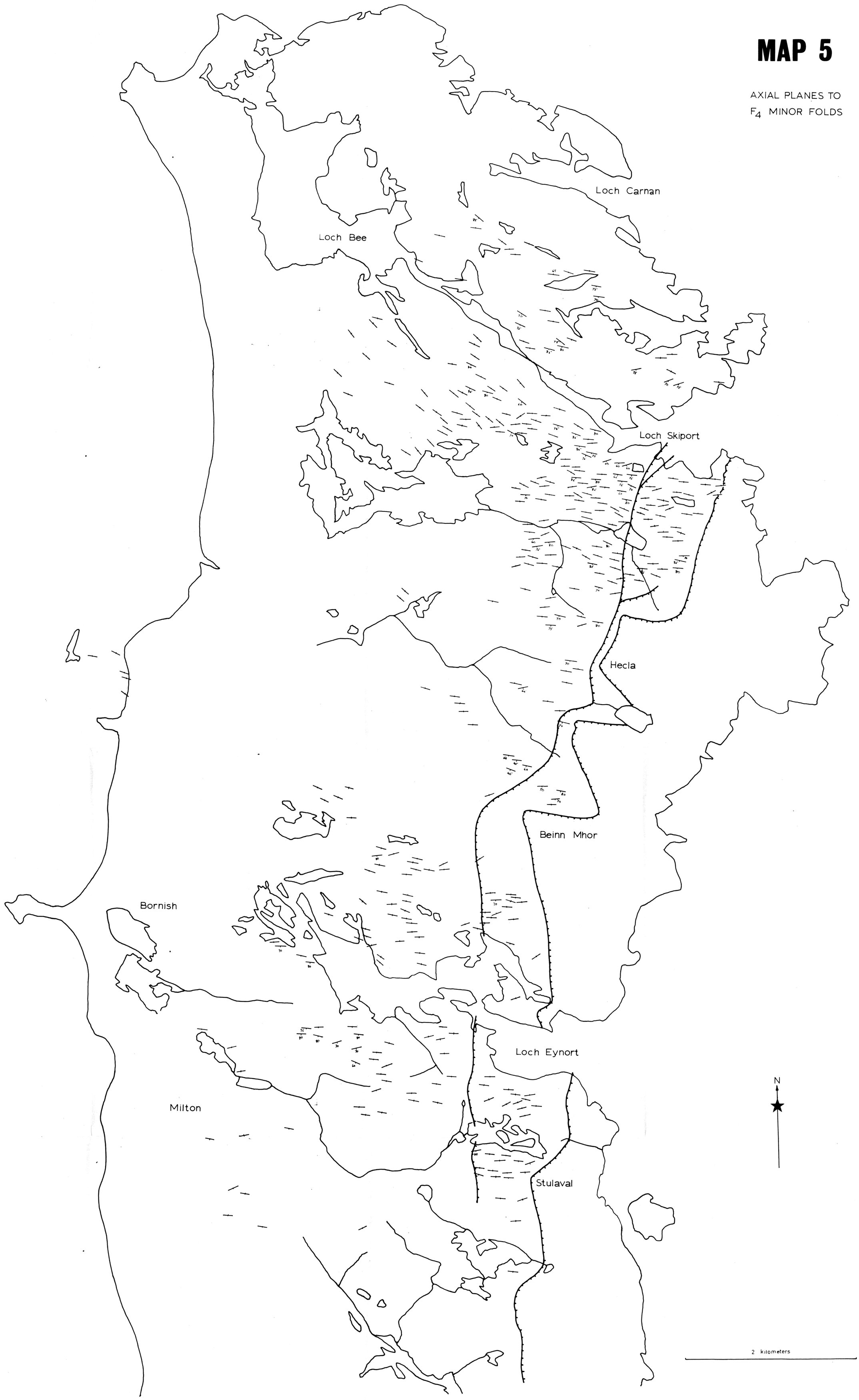




M.P. COWARD Ph.D. 1969

MAP 5

AXIAL PLANES TO
F_4 MINOR FOLDS



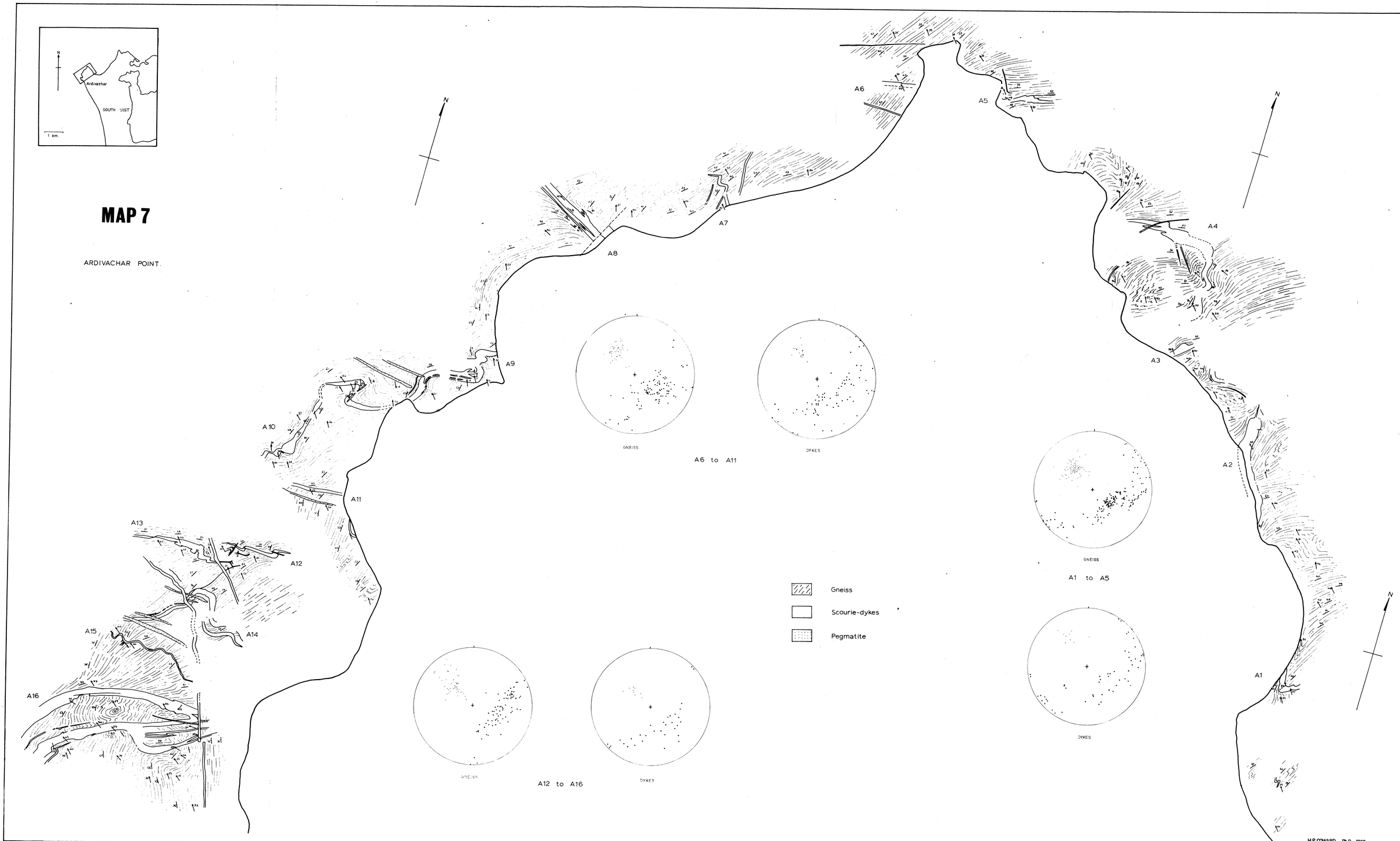
M.P.COWARD Ph.D. 1969

MAP 6

Thrust zone
Crushed and altered gneiss.
Pseudotachylite.
Usinish "mylonite"
Foliation.
N
ORNISH
MAOL MARTAIG.
RUDHA ROSSEL
COIRE NA H-EITICH
COIRE RUDALE
USINISH
HECLA
BEN SCALAVAT
GLEN USINISH
LOCH CORODALE
CORODALE HILL
BEINN MHÓR
RUDHA HELLISDALE
GLEN LIADALE
RUDHA NA BRATAICH
BEN NA HOE
RUDHA BOLUM
MEALL MÓR
1 kilometer
v.s. = h.s.
M.P.COWARD Ph.D. 1969

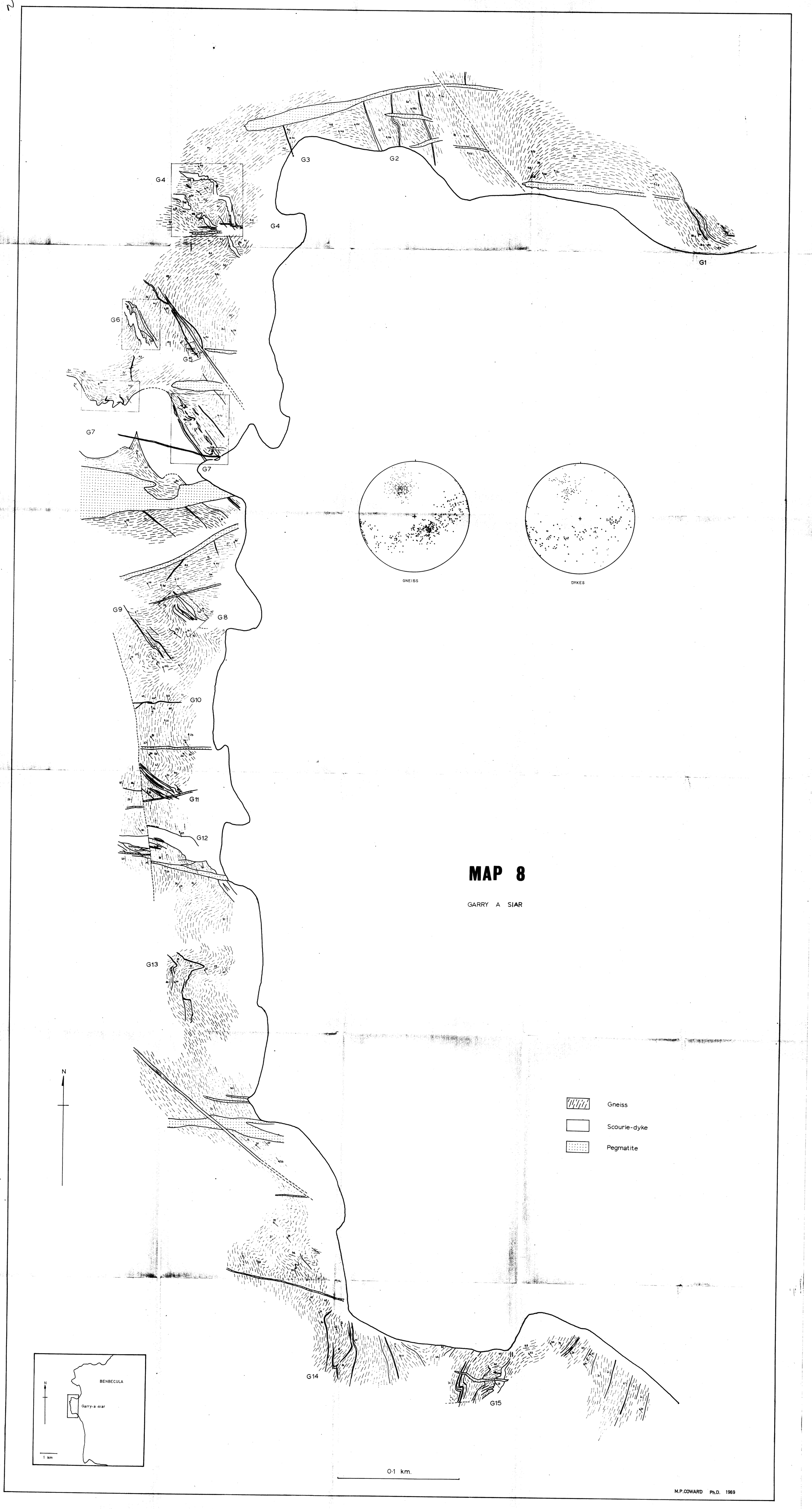
MAP 7

ARDIVACHAR POINT.



M.P. COWARD Ph.D. 1969

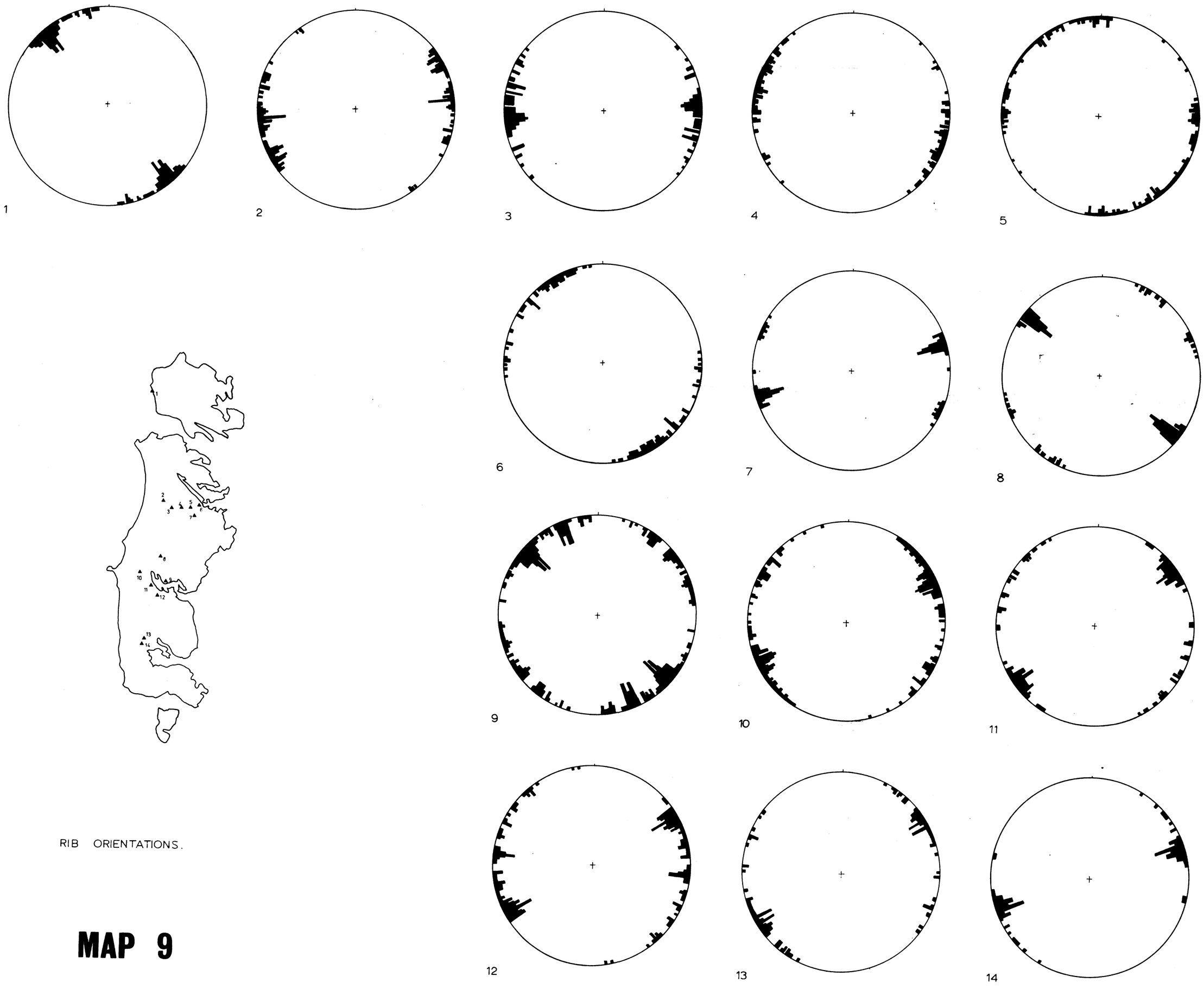
No 8.



MAP 8

GARRY A SIAR

M.P.COWARD Ph.D. 1969



RIB ORIENTATIONS.

MAP 9

M.P. COWARD Ph.D. 1969