A structural analysis of Lewisian rocks in parts of North Uist and the Sound of Harris, Outer Hebrides.

A thesis submitted for the degree of Ph.D. in the University of London

by

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October 1969.

ABSTRACT

After the intrusion of a suite of basic dykes, the Lewisian rocks of the North Uist-Sound of Harris area suffered ductile deformation in three phases, F1, F2 and F3. This Laxfordian deformation was inhomogeneous because of lithological inhomogeneities of the gneisses. The largest-scale strain variations are of F3 age and are related to the major structures of the area, a series of broad antiforms in which F3, and sometimes earlier deformation, is slight, and tighter synforms in which F3 deformation is significant.

Once a state of inhomogeneous deformation is recognised, quantitative methods are necessary for its complete analysis. The most useful method of quantitative strain assessment proved to be measurement of the orientation of deformed veins in apparently homogeneous sub-areas. Strain measurements were useful in analyses of the fold structures of the North Uist coast, and of the deformed lineations of the Sound of Harris.

Flinn's symmetry concepts suggest the possibility of a connection between fabric evolution and the superimposition of successive finite strains. Spatial variations in fabric on the North Uist coast are explained in this way.

Water content was important in both the deformation and the metamorphism of the gneisses, and many metamorphic reactions are associated with hydration. An association between water, deformation and recrystallization is seen throughout the area, especially in South Harris.

ACKNOWLEDGEMENTS

This work was carried out during the tenure of a Natural Environment Research Council Research Studentship. It was supervised by Dr. J.V. Watson, whose interest and critical comment were a constant source of encouragement.

Thanks are also due to Professor J.G. Ramsay for valuable discussion on certain aspects of the work, and to other staff of Imperial College for the use of technical facilities. Miss B. Evans typed much of the original manuscript and provided other assistance of various sorts. Mr. B. Donnelly read and corrected the whole text and Mrs. M.E. Queyrane typed the final copy.

Discussions with fellow research students, notably Dr. M.P. Coward, who worked in the adjacent area of South Uist, were always interesting and often fruitful.

Finally I would like to thank those people in the Hebrides whose friendship and boundless generosity made my spells of fieldwork so enjoyable, especially Miss M. Macmillan of Bornish House, South Uist, The Maclellans of Tigharry, Mr. E. Macaskill, Newton Ferry, and Mr. and Mrs. I. Macaskill of the Newton Hotel, North Uist.

FRONTISPIECE

Weakly deformed Scourie dyke amphibolites, Udal, North Uist.

Strongly deformed Scourie dyke amphibolites in the Tigharry Isocline, North Uist.

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INTRODUCTION

A. The aims of the work

This thesis is a contribution to research in the Lewisian rocks of the Outer Hebrides, Scotland. It is one of a series of studies by various members of Imperial College, and deals with the structural geology of an area between Borve Lodge, South Harris and Balranald, North Uist (see map 7).

The areal emphasis of the work is uneven. Detailed structural analyses are only possible in areas of good exposure, therefore in this study emphasis is given to two main sub-areas; the north-west coast of North Uist and the larger islands of the Sound of Harris, while the intervening ground receives scanty attention. The initial disadvantage of poor exposure between these well exposed sub-areas, and of general poor inland exposure was compensated for by the possibility of an extension of the work onto the mainland of South Harris, so that the rather unusual structures there might begin to make sense in the context of the geology of the Southern Hebrides.

The approach was thus a conventional areal one. Detailed mapping was necessary, partly because the areas concerned were previously inadequately described and partly because detailed structural studies of the kind now usual in orogenic terrains have been relatively rare in basements. Also I believe that the analysis of particular topics in structural geology is often best achieved if such topics emerge naturally in the course of orthodox fieldwork, for only then can they be examined in the context of their environment.

In the early stages of this research it became apparent from purely qualitative observations that there were, on all scales, variations in the intensity of deformation of the rocks.

An analysis of this feature became essential and it forms a dominant theme of this thesis. Two chapters are given over to possible semi-quantitative methods of strain analysis, and from this work, perhaps, there may come some points of interest in understanding the geometry and mechanics of basements in general.

B. Lewisian Chronology

The Lewisian outcrops of north-west Scotland and the Hebrides form the largest area of basement rocks exposed in Britain. Like other basement terrains, the Lewisian complex contains a variety of high grade metamorphic gneisses which are usually banded and thoroughly deformed.

A complex geological history in the Lewisian of the Scottish mainland was demonstrated long ago by the officers of the Geological Survey of Great Britain and is presented in the authoritative North-West Highland memoir (Peach et al., 1907). Sutton and Watson, (1951, 1961), following the Survey, used a suite of basic dykes (the Scourie Dykes) as time markers and gave the names "Scourian" and "Laxfordian" to the episodes of deformation and metamorphism which could be separated using the variable metamorphic state of these dykes. These episodes were regarded by Sutton and Watson as orogenies and were dated by Gilletti et al (1961) and later workers at 2,200-2,600 million years and 1,200-1,600 million years respectively. The division was found to have a spatial as well as a temporal significance, for a central zone from Scourie to Gruinard Bay showing few effects of Laxfordian regeneration is flanked by areas of considerable Laxfordian re-working.

Later work on the Lewisian rocks of the mainland has extended the chronology, and another "orogeny" (the Inverian) has been recognised (Evans and Lambert 1964).

Dearnley (1962) was the first to suggest that amphibolite bands in the Hebrides were equivalent to the Scourie Dykes of the mainland, and thus the first to bring any Lewisian chronology to the Hebrides. His reconnaissance study (1962) was in many ways the stimulation of the present Imperial College work. Although his hypothesis of a three-part division of the Hebrides comparable to the spatial division of the mainland has since become exceedingly doubtful, it was this concept which suggested the possibility of interesting geology in the Sound of Harris area - the northern boundary of Dearnley's Scourian zone.

Direct evidence concerning the early history of the complex is not apparent in most of North Uist and the Sound of Harris; it is almost always obscured by Laxfordian deformation and recrystallisation. The chronology presented in this thesis, therefore, begins with the intrusion of a suite of basic dykes into a pre-existing gneiss complex - an event for which there is, in some places, abundant evidence. All the phases of deformation and metamorphism described in the following pages are of Laxfordian age.

C. <u>Lithology in the North Uist - Sound of Harris area.</u>

Acid gneiss is the principal rock type in the North Uist - Sound of Harris area. It is, however, of variable aspect and composition and some of these variations have important structural implications. Other components of the complex include minor concordant basic and ultrabasic bands, patches of metasediment (paragneiss), deformed granite and, in South Harris, large deformed and metamorphosed masses of basic igneous rock. Everywhere, except in the part of South Harris with which this thesis is concerned, there are abundant amphibolite bands. Usually these are concordant, but in the areas of low deformation (e.g.Udal, see below) they show cross-cutting relationships with the acid-gneisses, and clearly represent pre-Laxfordian intrusives. On this evidence they are correlated with the "Scourie Dykes" of the mainland and are termed "Scourie dyke" amphibolites (c.f. Dearnley 1963).

D. Polyphase Deformation

Rocks which have suffered repeated deformation can be adequately described only when the effects of each individual episode of deformation have been separated. Here the usual methods of structural analysis, the recognition of re-folded folds and associated minor structures, folded boudins, minor fold senses etc., has been used to identify three phases of ductile deformation (F1, F2, and F3).

The three sets of structures are almost always co-axial on the North Uist coast, though not in the Sound of Harris. They are all earlier than the semi-brittle and brittle deformation associated with the Outer Isles Thrust (see Coward 1969).

E. The regional structure of the area

The large scale structure of the gneisses of the North Uist - Sound of Harris area is produced by large scale F3 folds (see map 7). Major F3 synforms with "S" asymmetry occur in Berneray and on the north-west coast of North Uist (hereafter known as the "North Uist coast"). These synforms are separated by a much more open antiform whose long flat limb occupies the poorly exposed north coast of North Uist. The antiform is not give a name, but the synforms will be known as the <u>North Uist coast Synform</u> and the <u>Berneray Synform</u>. North-east of the Berneray Synform another major antiform, the <u>Sound of Harris Antiform</u> is inferred, and the steep limb common to the two structures is known as the <u>Sound of Harris steep zone</u>. The gently dipping north-east limb of the Sound of Harris Antiform passes beneath a large mass of metamorphosed igneous rock on the island of Ensay.

Although the structures described above are the largest visible folds in the area, their asymmetry suggests the existence of an even larger scale fold, presumably in the southern part of South Harris, a possibility which is discussed in chapter 8.

The North Uist coast, the Sound of Harris and South Harris are all areas in which significant F3 deformation can be recognised, whereas in the antiformal domain of the north coast of North Uist, and near the hinge of the Sound of Harris Antiform in Pabbay there is little evidence of F3 deformation At Udal on the north coast of North Uist, and also in north-east Pabbay, areas where earlier Laxfordian deformation was also slight, there are "deformation lows" where the Scourie-dyke amphibolites preserve cross-cutting relationships with the acid gneiss banding. These areas are comparable with those in South Uist and Benbecula which are described by Dearnley and Dunning (1968) and Coward (1969). The Udal area is not described in detail in this thesis because it is so like these areas, and because its boundaries are inadequately exposed.

F. <u>Structural summary</u>

The structural history given below forms the skeleton of this work. The comments are necessarily generalizations, intended only to aquaint the reader with the general descriptive scheme.

<u>LAXFORDIAN DEFORMATION AND METAMORPHISM</u>

Brittle and semi-brittle deformation associated with the Outer Isles thrust.

(F4 in South Uist (Coward, 1969)

F3 The deformation which produced the present structural architecture of North Uist and the Sound of Harris. Folds are upright about N.W.-S.E. trending axial surfaces.

F2 Often the most important deformation phase. Characteristically there are short intensely folded zones and long unfolded "limb-zones". A "gneissic cleavage" is developed axial planar to F2 folds in suitable lithologies. The approximate orientation of the axial surfaces of F2 folds before being re- by F3 folds was N.E.-S.W. with a moderate or gentle dip to the N.W. F2 folds probably originated as neutral folds.

F1 The first fabric in the Scourie-Dyke amphibolites is usually at least partly an F1 structure. F1 folds are rare. They are difficult to identify or differentiate from Scourian folds in the acid gneiss. Scourie-Dykes probably do not show F1 folds because the XY plane of the F1 deformation was parallel to the original orientation of the dykes so that they were flattened, not folded.

(continued overleaf)

INTRUSION OF THE SCOURIE-DYKES ON A N.W.-S.E. TREND

SCOURIAN GNEISS COMPLEX: The banding (also called foliation) of the acid gneiss is of this age, as are many lithological features of the complex.

G. Outline and discussion of Previous Work in the Area

Published accounts of the geology of North Uist are conspicuous by their absence. Macculoch passed through the island and made several observations; Heddle (1888) includes some geological comment amid a description of scenery. Beveridge's book on the Archaeology of North Uist (1924) includes a section on the topography of the island, but the first systematic geological study was presented by Jehu and Craig in 1926 (Geology of the Outer Hebrides, part III, North Uist and Benbecula). The geological observations of these authors are generally sound but the work is entirely descriptive.

In their description of the North Uist Coast, they do little more than point out the "irregularity" of the foliation.

Dearnley's reconnaissance study of the whole Hebrides (1962) placed North Uist in a central zone of Scourian structure, although Laxfordian metamorphic events were recognised and an area of Laxfordian "migmatization" was thought to exist on the north-west coast of the island. In the unexposed central part of North Uist, Dearnley described a Scourian chevron fold (the Lochmaddy synform) whose E-W or ENE-WSW axial trace runs from near Bayhead to Lochmaddy. He considered that both limbs of this fold dip northwards and therefore regarded the northern limb as overturned.

Dearnley and Dunning (1968), in a detailed study of parts of South Uist, made further observations on the North Uist Coast. They recognised fairly large scale folding in the "basic facies of the grey gneiss", and suggest that such folding is "probably younger than the small scale plastic folding in the acid facies of the grey gneiss" though (the axes of) both types of structure have plunges "similar in direction and amount". They recognised increasing felspathization northwards along the coast and suggested that "within a short distance (the gneiss) becomes a true migmatite with ptygmatic folds", until near Griminish Point" the grey gneiss is a Laxfordian migmatitic granite gneiss". This sort of observation is possibly valid as far as it goes, it simply does not go very far. Dearnley and Dunning were not specifically concerned with the North Uist coast.

The major conflict between my own work and the published work of Dearnley and of Dearnley and Dunning stems from their implication of a Scourian age for the main folding of North Uist. In my view this is not a valid interpretation and field evidence does not support it.

Geological accounts of the Sound of Harris are even rarer than those of North Uist. Jehu and Craig visited many of the islands and Dearnley includes a description of the rocks of Ensay in his paper on South Harris (1963). Livingstone (1965) describes the petrography of an ultra-basic rock from Berneray; otherwise there is nothing.

South Harris has fared better, presumably because of the greater petrological interest of its rocks.

Macculloch and Heddle paid visits, and Jehu and Craigs' account is more substantial than their complementary studies in North Uist. Davidson's lengthy petrological descriptions (1943) are now almost classic and Dearnley's paper (1963) is based on detailed fieldwork. Dearnley's lithological map of South Harris is used as a basis for some of the discussion in chapter 8 of this thesis, but there are significant differences between Dearnley's interpretation of the geology of South Harris and that presented here. These differences are fully discussed in chapter 8 and need not be mentioned at this stage.

H. <u>The lay-out of the thesis</u>

Initial chapters describe the structural geology of the North Uist coast and the Sound of Harris in a conventional way. Folding and the fold structures occurring in the area are then discussed, and later the possibilities, difficulties and methods of measuring strain in basement terrains are described. The metamorphic history and factors influencing the metamorphic state of the North Uist-Sound of Harris rocks are briefly considered, and more space is given to the relationships between metamorphic mineral fabrics and strain conditions, especially changing strain conditions. Chapter 7 considers the usefulness of strain measurements in interpreting two specific problems. Chapter 8, a sort of appendix, describes the geology of part of South Harris, an area of initially rather dull structural geology, but which exemplifies several important aspects of deformation in basements. The conclusion of the thesis tries to show how all the important aspects of this study may be related in a very simple general pattern of basement deformation.

I. Terminology

The terminology used in this thesis is more or less that used by Ramsay (1967). Fold description uses the terminology suggested by Fleuty (1964). Some terms which require definition at this stage are:-

(i) Foliation: used here as a "sack term" for the dominant planar structure in all sorts of gneisses. It includes lithological structures (which are also called "banding" or "layering") and mineral alignments in homogeneous rocks which do not possess such lithological structures. In general the word "foliation" applied to acid gneisses refers to the banding, a structure which is probably of Scourian age, while "foliation" in Scourie-Dyke amphibolite is a planar fabric of Laxfordian age.

(ii) Gneissic cleavage: a secondary planar or lineated planar fabric developed in some acid gneiss lithologies (especially metasediment-derived rocks) by the dimensional orientation of discrete biotite or, occasionally, hornblende crystals. It cannot be called a cleavage because of the coarseness of grain of the minerals involved, and the use of the word "schistosity" is misleading since rocks containing gneissic cleavage are not schists. Gneissic cleavage often forms parallel to the axial surfaces of F2 folds.

(iii) Domain : The ground occupied by rocks with some feature in common, e.g. a region of homogeneous strain, or the rocks belonging to a major fold. It is used descriptively, e.g. "F1 minor folds are uncommon in the domain of the Scolpaig Synform".

CHAPTER 1

THE STRUCTURAL GEOLOGY OF THE NORTH UIST COAST

1:1 INTRODUCTION

The low cliffs and wave-cut platforms of the North Uist coast provide limited, though often very good, three-dimensional exposure. Neighbouring inland exposure is negligible. Mapping was done on enlarged aerial photographs and the entire coast was mapped twice in successive field seasons. A useful mapping technique involved tracing out accurately all the major amphibolite bands (see map 1).

As pointed out in the introduction to this thesis, three phases of ductile deformation can be recognised on the North Uist coast. In all but a few localities where slight (up to 10°) angular discordances exist between certain linear elements, the structures of the three deformation phases are exactly co-axial. Mineral lineations, the longest axes of boudins, rodding structures, and fold axes of all ages are all parallel. They plunge north-westward at moderate angles and their orientation is consistent (Fig.1a), hence I offer no map of the linear structures of the North Uist coast.

1:2 FOLD ORDER

Part of this description of the structures of the North Uist coast is given in terms of fold wavelength order. Since fold wavelength depends on the thickness of competent units involved in buckling and the viscosity contrast between these units and their matrices, one might expect a random

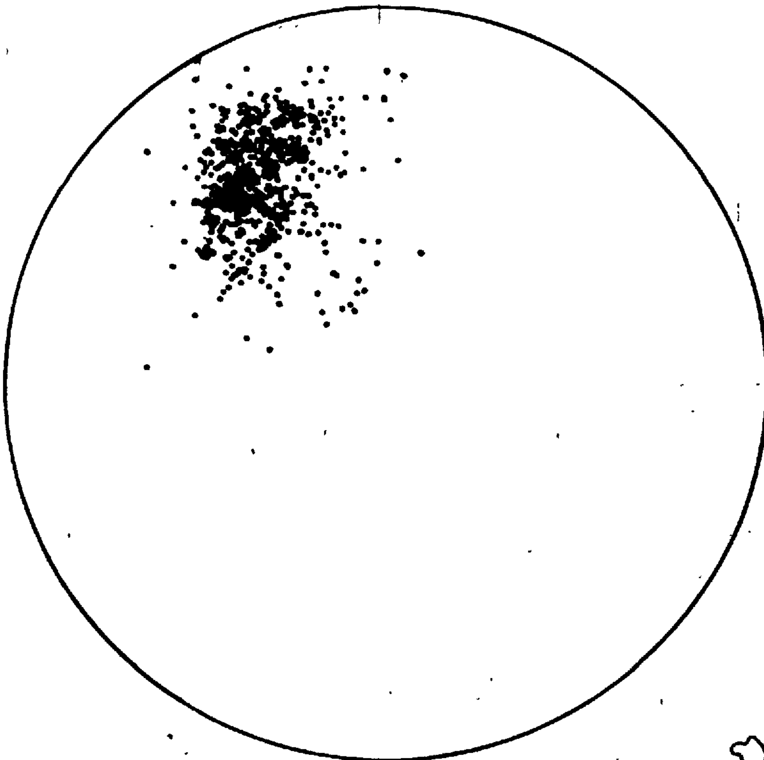


Fig 1a

Linear elements of the North Uist coast

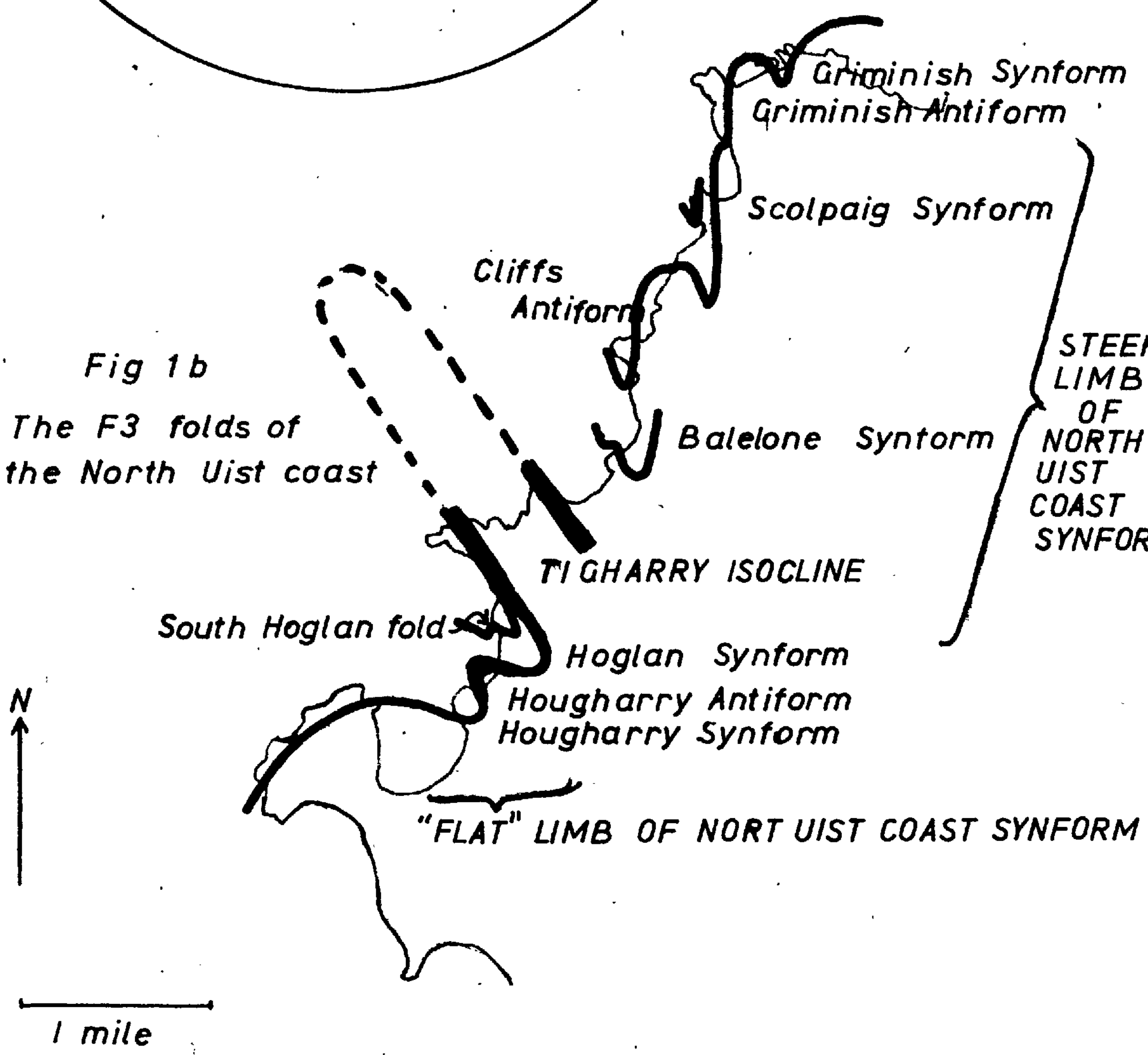


Fig 1b

The F3 folds of the North Uist coast

family of wavelength orders in naturally deformed banded rock sequences.

Through an examination of fold asymmetry and axial surface fanning relationships, however, five general orders of F3 folds can be recognised on the North Uist coast, and though the groupings are grossly oversimplified (especially as far as the fourth and fifth orders are concerned) they provide a powerful tool for descriptive purposes.

	1st Order	-	The largest structures	Presented non-diagram-matically on the maps.
	2nd Order	-	Large scale folds	
	3rd Order	-	Smaller folds parasitic on both those above	
Gradational	4th Order	-	Small folds up to 2mts wavelength	
	5th Order	-	Crinkles	

1:3 A CLASSIFICATION OF THE ACID GNEISS ON THE NORTH UIST COAST

Leaving aside any obvious metasediments, it is possible to divide the acid gneisses of the North Uist coast into two main lithological types, called for convenience "rough gneiss" and "smooth gneiss".

"Rough gneiss" is coarse-grained, coarsely banded quartzo-felspathic gneiss containing subordinate hornblende and biotite whose appearance is occasioned by abundant deformed pegmatite veins within it. The veins which are most likely ultimately of Scourian age for they carry fold closures which are either Scourian or F1 in age.

"Smooth gneiss" is vein free, unbanded, homogeneous and finer grained than "rough gneiss". It is composed of quartz and felspar with subordinate biotite and lacks hornblende.

Pink microcline augen up to an inch in diameter are distributed throughout the rock. They may be recrystalized early phenocrysts or porphyroblasts.

Field evidence suggests that before any Laxfordian deformation the present rough gneiss was already a banded gneiss while the smooth gneiss was a nearly homogeneous and perhaps porphyritic granite.

The lithological differences between rough and smooth gneiss have important implications in the study of rock fabric (see chapter 6). Homogeneous smooth gneiss with a relatively small proportion of platy minerals is not amenable to the imposition of planar fabrics, and smooth gneiss in an appropriate structural position (. see chapter 6.) can become a true "L" tectonite. An expression of linearity, however, is not easily created in strongly banded biotitic rocks, and rough gneiss maintains some visual planarity in almost all structural situations.

The two sorts of gneiss deform differently in another way. Minor folds, common in rough gneiss, are almost absent in the smooth variety. "Smooth gneiss" is nearly homogeneous and therefore suffered fairly homogeneous deformation. The banding of the "rough gneiss", however, provided competence discontinuities sufficient for heterogeneous deformation and the development of small scale folds. In a gneiss succession of amphibolite bands within smooth gneiss, the amphibolite bands were the only competent units and the thickness of these bands determined the lowest wavelength order of the folds, hence in areas where all the amphibolite bands are thick (e.g. the North Uist Cliffs) only fairly high order folding occurs and the visual effect is startling.

The rocks of the North Uist coast were a varied assemblage before the onset of Laxfordian deformation, and included granites, quartzo-felspathic gneisses, local paragneisses of various types, and basic bands of possible igneous affinities (the "Older basics" of Sutton and Watson, 1951). These old basics are common within sequences of rough gneiss, but absent from areas of smooth gneiss. Scourie-Dyke amphibolites occur in considerable numbers everywhere on the coast and are always concordant with the gneissic banding.

1:4 THE F3 STRUCTURES

Though the North Uist Coast is an area of complex polyphase deformation, the finite fold pattern is dominated by the youngest (F3) structures, and it is logical to describe these third structures before discussing the older ones. It is also convenient to give names to the most important F3 structures, and these are listed below and shown on Fig. 1b.

1st Order North Uist coast Synform (with the Tigharry Isocline) and its antiformal complements to the north-east and south-west.

2nd Order Hougharry Synform
Hougharry antiform
Hoglan synform
(South Hoglan fold)
Balelone Synform
Cliffs antiform
Scolpaig Synform
Griminish Antiform
North Griminish Synform

3rd Order Asymmetric folds, parasitic on the above

4th Order
5th Order Distinction seldom made, if one exists.

Two of the second order antiforms (the Cliffs and the Griminish antiforms) are broad, double hinged, and box-like folds whose two axial surfaces are inclined towards one another. The Hougharry antiform, on the other hand, resembles the synforms in being close and upright about a north-west trending axial surface.

1:4:1 <u>The First Order Structures</u>

a) <u>The North Uist Coast Synform</u> is an open (interlimb angle about 90°) asymmetrical fold whose long steep limb occupies the coast from Hosta Beach to Griminish Point and whose short "flat" limb is exposed near Hougharry. It is divided into two parts by the antiformal Tigharry Isocline so that both structures are developed about the same axial surface (dip 80° towards 230° - 240°). It will be shown later that this phenomenon of an open synform containing a central isoclinal antiform is a common feature of the present area, and the present case is merely a large scale manifestation of it.

The domain of the North Uist Coast synform ends at Griminish in the north and near Hougharry in the south. In these areas are the broad hinges of unnamed first order antiforms complementary to the North Uist Coast Synform. The northern one is complicated by second order folds but has an axial trace running generally 140°, while the axial trace of the southern hinge runs N-S through "Hougharry Bay". It must be noted here that the North Uist Coast Synform is not in any way comparable with Dearnley's (1963) Lochmaddy Synform, for I have been unable to recognize Dearnley's structure in the field.

b) The Tigharry Isocline is seen in the field as a belt of intensely planar rocks which outcrops along the coast from the N.E. edge of Hoglan Beach to the south edge of Hosta Beach. There are relatively few minor folds in this region, boudinaged pegmatite veins and rather problematical fractures being the dominant minor structures. The planar zone is bounded to the south-west and the north-east by two amphibolites which, though composite, (i.e. containing interbanded acid material) are uniquely thick. They will be called the Hoglan and Hosta amphibolites respectively and it will later be contended that they are two outcrops of the same unit. Most of the rock between the amphibolites is smooth gneiss interbanded on all scales with amphibolite, though in the centre of the planar belt there is a group of metasediments interbanded with ultrabasic rocks and Scourie - dyke amphibolites. The hinge of the Tigharry Isocline has not been located, the description of the rocks between Hoglan and Hosta beaches as in isoclinal fold is partly interpretative, and the reasoning is set out in full in chapter 7.

1:4:2 Second order structures

The flat (southern) limb of the North Uist coast Synform contains three second order folds which are, from south to north, the Hougharry Synform, the

Hougharry Antiform and the Hoglan Synform. A large and important, though strictly speaking third order, fold the "South Hoglan fold" is also described in this section. All the structures are close and moderately plunging about steeply inclined axial surfaces. The axial surfaces of the three second order folds form a fan which converges upwards, viz;

Hougharry Synform	Axial surface dips 80° towards 253°.
Hougharry Antiform......	Axial surface dips 85° towards 55°.
Hoglan Synform	Axial surface dips 58° towards 32°

(see also fig 20 in chapter 7).

a) Hougharry Synform

The hinge region of this fold contains smooth gneiss together with a fairly thick Scourie-dyke amphibiolite. The fabric of the "smooth" acid gneiss is truly linear, and in the amphibolite an F3 crenulation cleavage cuts the earlier foliation at right-angles so that a marked linear appearance analogous to "pencil cleavage" in slates, is produced.

The amphibolite band was boudinaged before the development of the Hougharry Synform (F2 deformation ?), and an isoclinal antiform, presumably also of F3 age lies between the two deformed boudins in the middle of the synform (Fig.2). Structures such as this are known in this thesis as "shoot through" structures. It is contended that the Tigharry Isocline (described above) is a large scale example of one of these structures.

b) The Hougharry Antiform

The exposed part of the hinge of the Hougharry antiform

Fig 2

Hinge of the Hougharry Synform

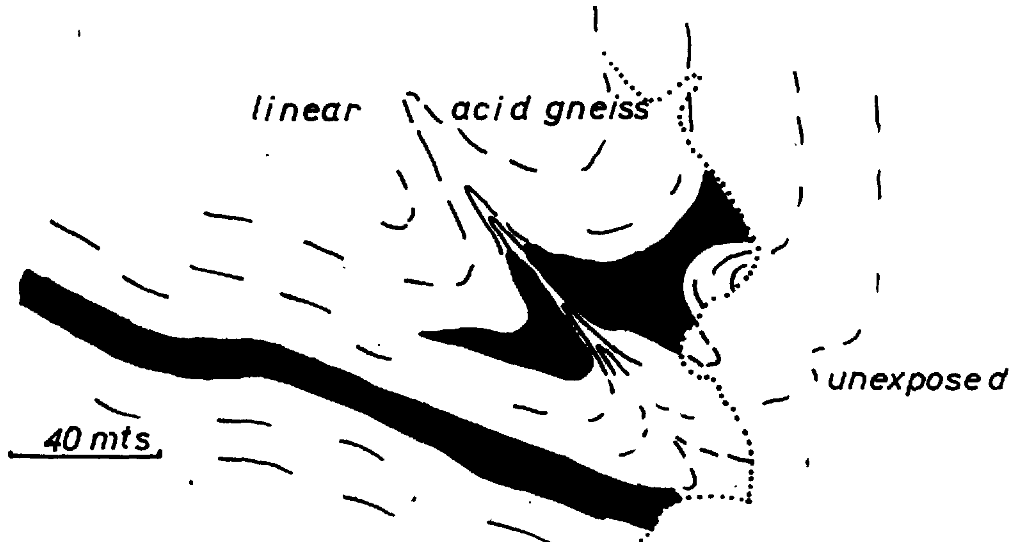
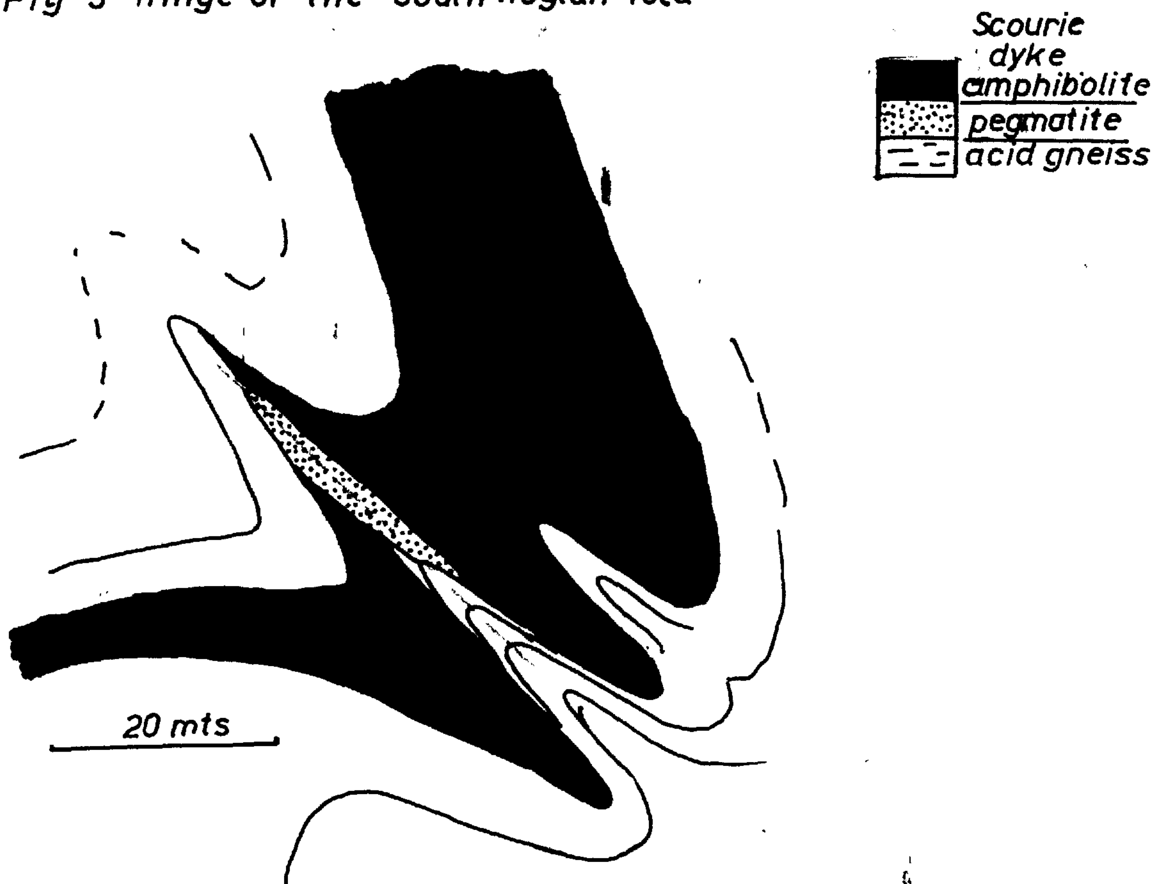


Fig 3 Hinge of the South Hoglan fold



is contained in a large amphibolite unit which may well be continuous with the Hoglan and Hosta amphibolites. This unit possesses a well-developed lineated foliation of F1/F2 age which is intensely folded by minor F3 folds.

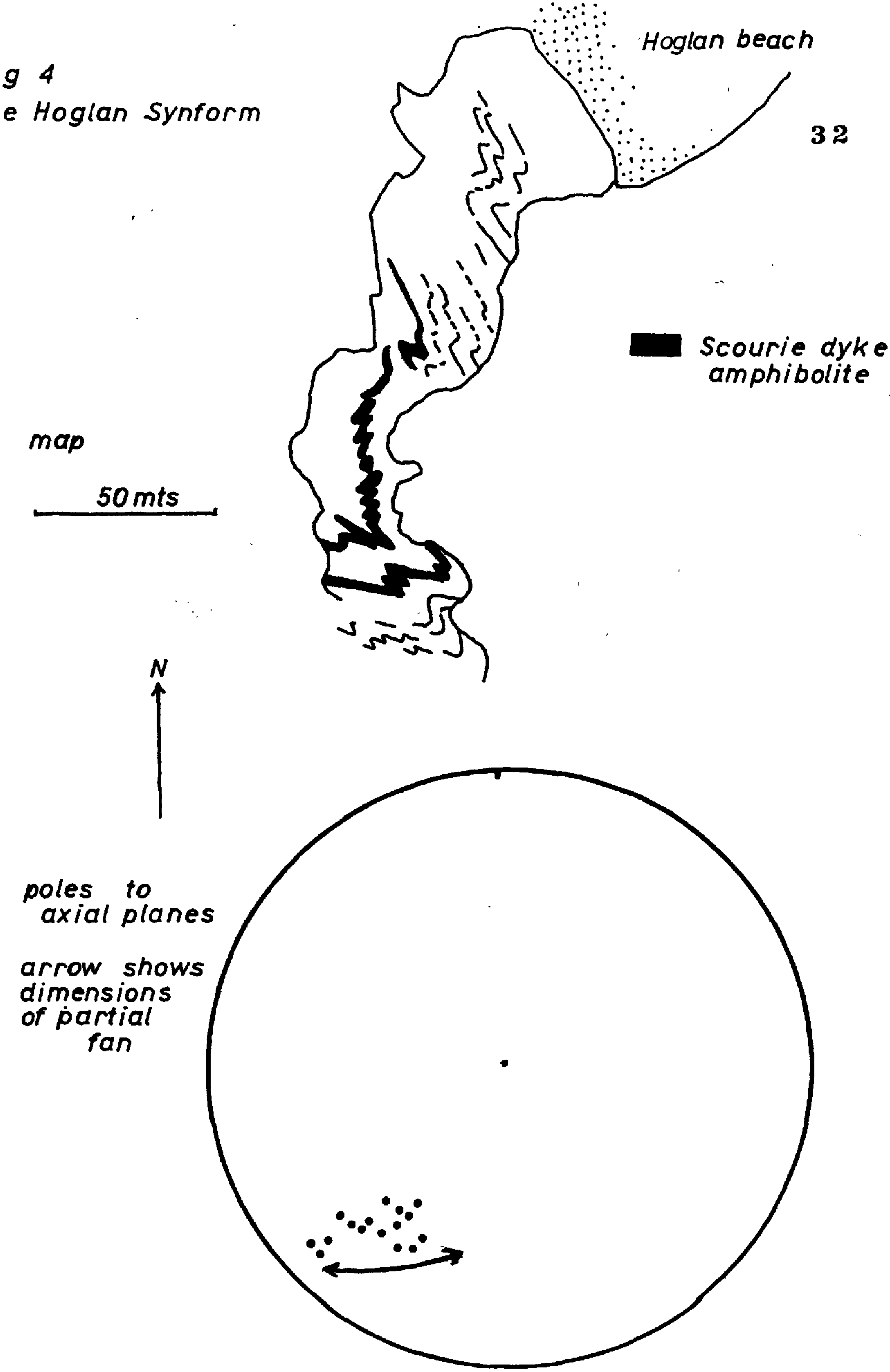
c) Hoglan Synform

Unlike the Hougharry folds, the limbs of the Hoglan Synform contains a large number of third order parasitic folds. One of the most important of these is the South Hoglan fold (fig 3) which, like the Hougharry Synform, contains a central isoclinal antiformal "shoot through structure". Measurements of the true thickness of a single hornblendic layer showed that it was thickened by 8 or 9:1 in the hinge of the antiformal "shoot through structure", but only 2.5 or 3:1 in the adjacent synformal hinges. This marked difference is probably not due to variations of the viscosity of the layer along its length, and the antiform probably represents a local area of intense strain.

The hinge of the Hoglan Synform itself (fig 4) is seen in an amphibolite band of some 2 mts thick. No very thick competent bands are seen in the part of the fold now exposed, and the fold might owe its large scale either to some thicker, unexposed competent unit (perhaps the Hoglan amphibolite) or to some other mechanism (see 3:6:2, and 7:3:8).

The parasitic third order folds of the Hoglan Synform possess axial planes which, though consistently dipping in a north-easterly direction, fan somewhat about the major structure (fig 4). Third order folds lying in the north-eastern limb of the Hoglan Synform become fewer towards the Hoglan Amphibolite, which contains no such

Fig 4
The Hoglan Synform



structures and marks the beginning of the domain of the Tigharry Isocline.

North-east of the Tigharry Isocline, in the steep limb of the North Uist coast Synform the axial planes of all the minor folds dip to the west, thus a large scale fan exists about the Isocline itself. This phenomena receives attention in chapter 7.

Second order folds in the steep limb of the North Uist coast Synform.

Five second order folds lie in the steep limb of the first order synform. They are, from south to north, the Balelone Synform, the Cliffs Antiform, the Scolpaig Synform, the Griminish Antiform and the Griminish Synform.

Immediately north of Hosta beach, "smooth gneiss" gives way to banded "rough gneiss" and consequently low (fourth) order folds become abundant. This area is part of the western limb of the Balelone Synform, and the third order folds here have "S" asymmetry. On Raikinish headland F2 minor folds are co-axially refolded by these structures, (Fig.5).

d) Balelone Synform

The hinge of this fold is exposed in a fossil wave-cut platform near Balelone farm. An ill defined contact between rough and smooth gneiss can be traced around the fold, though the bulk of the exposed part of it contains smooth gneiss. Several of the thick Scourie-dyke amphibolite bands involved in the Balelone synform are branched. Though this branching is mostly a modified original feature, at least one earlier (F2?) fold hinge

Fig 5 Co-axial interference of F2& F3 folds, Raikinish

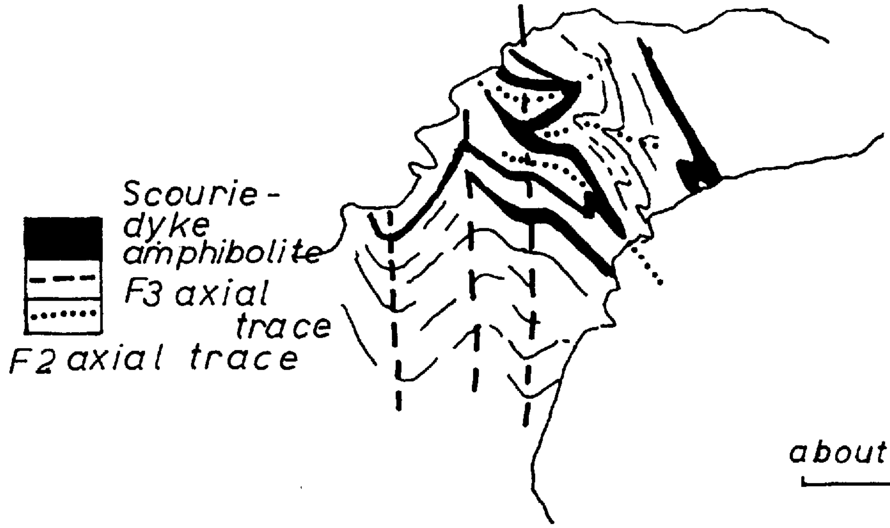
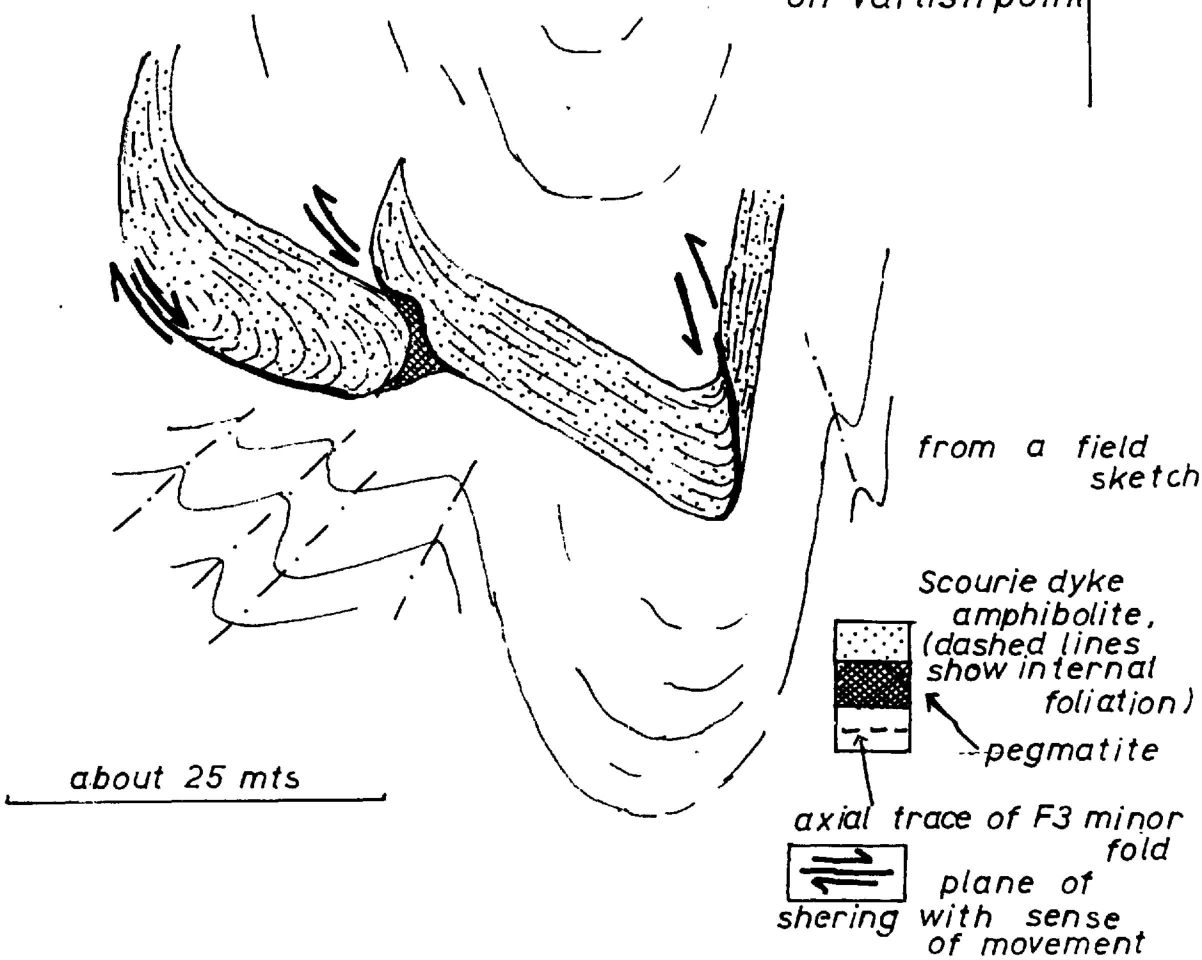


Fig 6 The hinge of the Balelone Synform
on Varlish point



is recognisable. Clearer, but smaller-scale examples of interference are given by (F2) minor folds which, folding the F1 foliation of the amphibolites, are themselves folded by the Balelone synform. This is taken as good evidence of the F3 age of the latter structure.

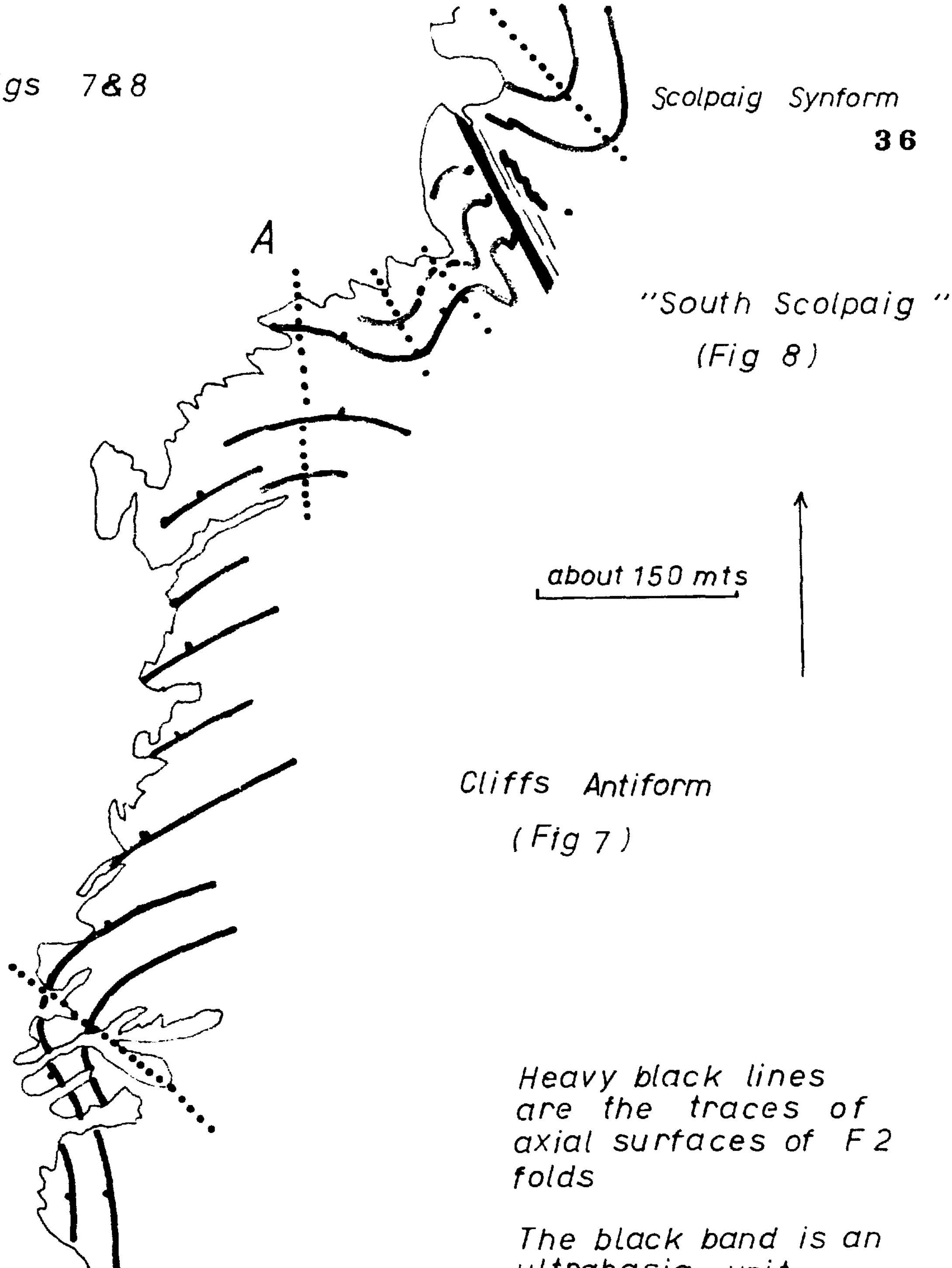
Though running in a general N-S direction, the axial trace of the Balelone synform curves. The axial plane dips steeply (60^{o}) and the hinge plunges moderately. The fold is close, the average interlimb angle being about 45^{o}.

Northerly exposures of the fold (on Varlish Headland) show the hinge developed in an amphibolite band which was boudinaged prior to folding. Two boudins have been forced together again by the F3 movements, and local shearing has occurred. The sense of movement of the amphibolite masses (Fig.6) is a good indication that an important flexural slip component was present in initial fold formation. Intense shearing occurred along major discontinuities between the amphibolite unit and its acid gneiss matrix.

e) Cliffs Antiform (Fig.7).

The eastern limb of the Balelone synform strikes N-S and dips steeply. Tight minor F2 folds within it have axial surfaces of similar orientation. Near a natural arch, however, (Fig.14) the pile of F2 folds is refolded, and their axial surfaces take on a moderate, north-westerly dip. F2 axes thus come to lie near the dip-direction of their axial surfaces, and the F2 folds become effectively neutral. In the neighbourhood of the arch this refolding can be traced in individual F2 folds. It represents the western hinge of the box-like F3 Cliffs Antiform. The pile of north-westerly dipping rocks exposed in the cliffs north of the arch lies in the broad hinge of this structure.

Figs 7&8



Heavy black lines are the traces of axial surfaces of F 2 folds

The black band is an ultrabasic unit

Dotted lines are F3 axial traces

see maps 1&2 for more detail

The axial trace of the western hinge of the Cliffs Antiform runs 170° and the axial surface dips about 60° to the north east. The western hinge is not quite co-axial with the earlier structures and F2 fold axes and mineral lineations are slightly deformed across it. Evidence of F3 deformation is rare within most of the hinge zone, however, and only one minor F3 fold is seen.

Unlike its western counterpart, the eastern hinge of the Cliffs Antiform is not simple. At (A) on Fig. 7 the neutral F2 folds cross a third fold hinge and become more upright. Almost immediately, however, they are involved in a zone of "minor" F3 folding, and the resulting interference is very complicated, and has not been analysed perfectly. The only consistent feature (as everywhere on the North Uist coast) is the plunge of the linear elements. Broadly speaking, this zone of complexity is synformal. It is separated from the domain of the <u>Scolpaig Synform</u> by a narrow N.W.-S.E. trending nearly vertical, unfolded belt of rocks, the most important of which is a parallel sided ultrabasic band, along which there has evidently been considerable shearing. The two synforms have no intervening antiform, only this narrow belt of vertical rocks which might concievably represent an antiformal "shoot through" structure in the middle of the complex synformal zone (Fig.8).

f) <u>Scolpaig Synform</u>

For practical purposes, the Scolpaig Synform may be considered as the direct complementary fold of the Cliffs Antiform. The hinge of the synform is exposed in

Fig 9 The curving axial surface of the Scolpaig Synform

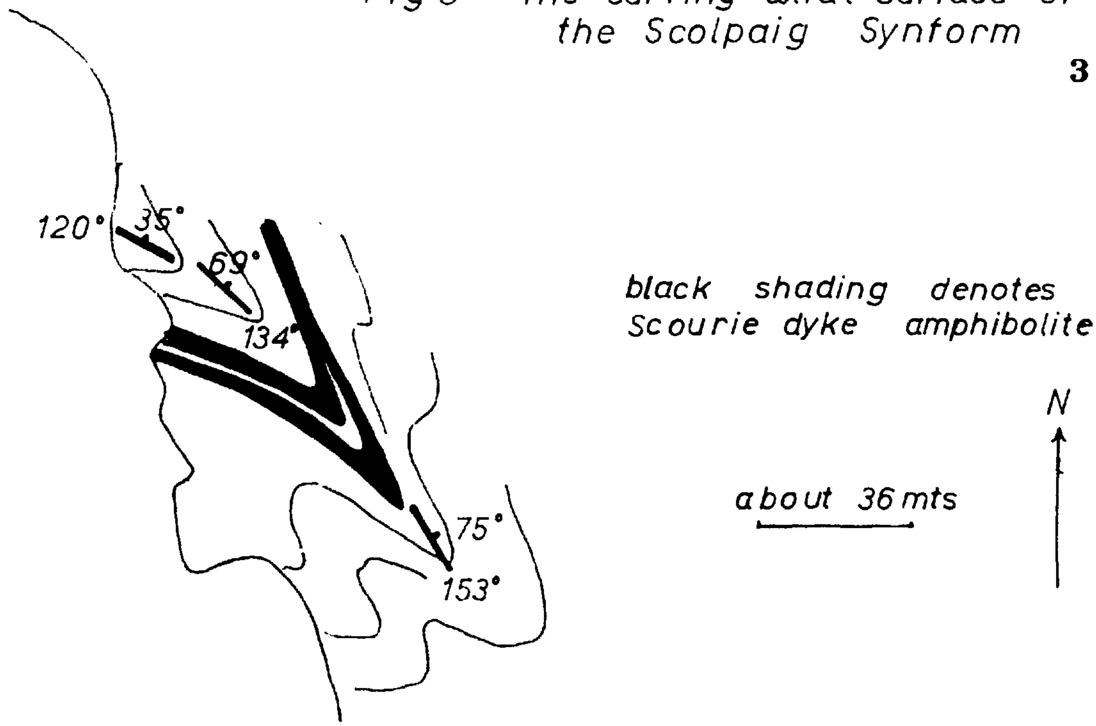
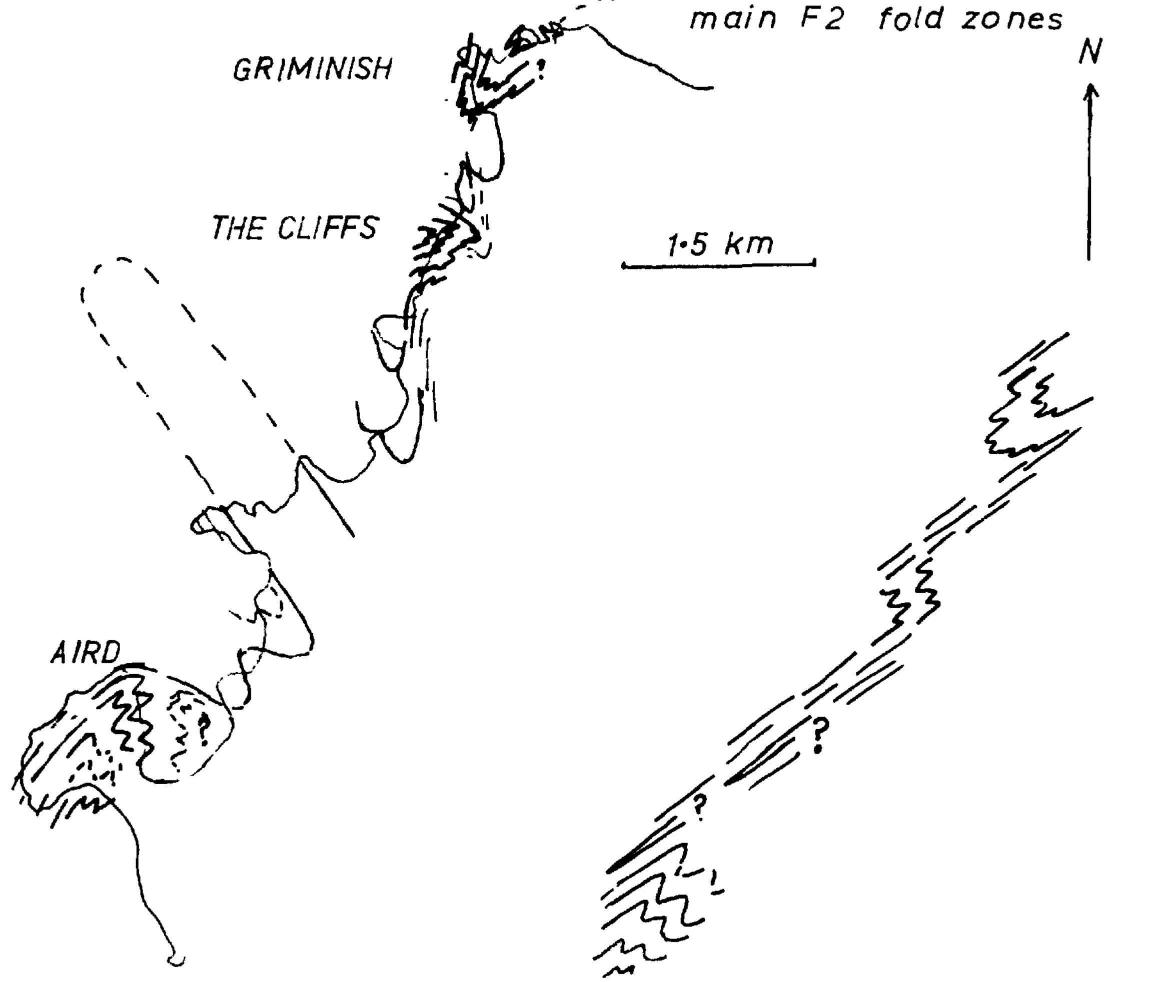


Fig 10 F2 structure of the North Uist coast showing the location and possible original interrelationship of the main F2 fold zones



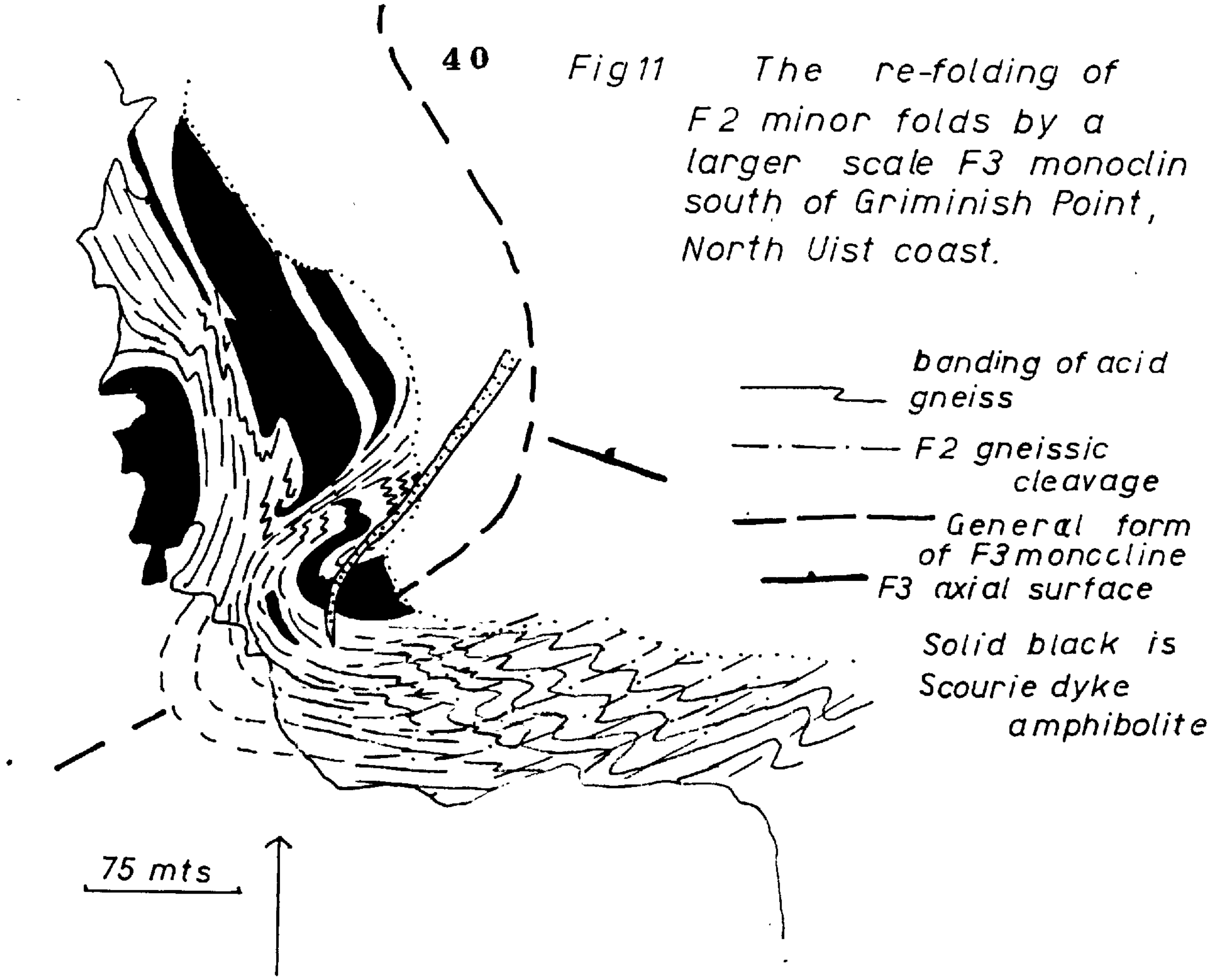
the wave-cut platform of the headland west of Scolpaig Bay. It can be traced in amphibolite bands, in a biotitic basic mass which may or may not be of Scourie dyke age, and in a distinctive series of anthophyllite-quartz-felspar rocks. The interlimb angle varies along the axial trace (see chapter 3, Fig.10) and the synform is tightest in the central part of its outcrop. This variation may be a function of variable ductility contrasts throughout the fold.

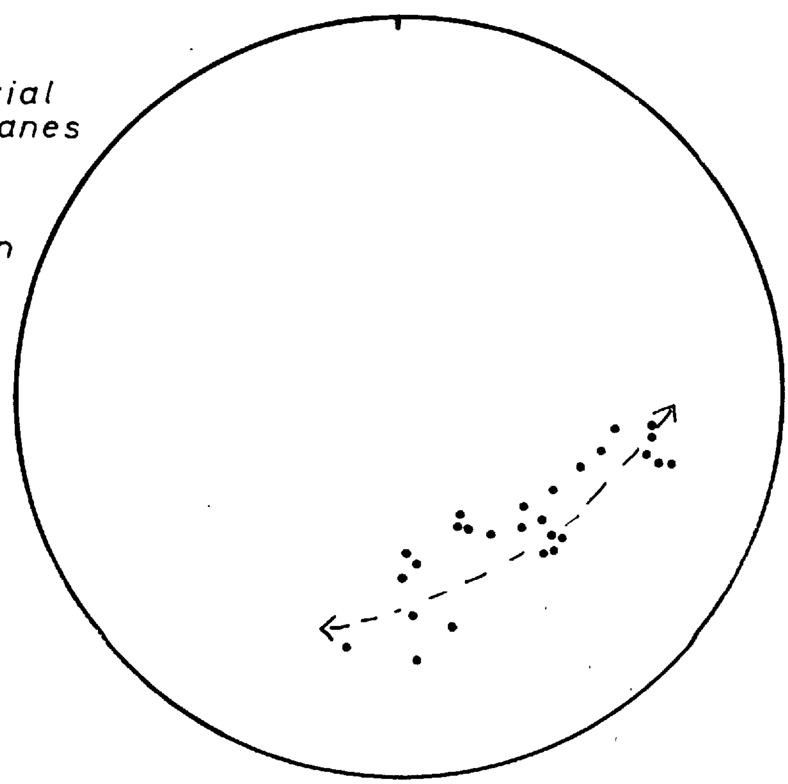
Generally speaking the Scolpaig Synform is upright and moderately plunging. Its axial trace curves and the dip of its axial surface varies somewhat (Fig.9). It differs from other second order structures (except the Hoglan synform) in having large numbers of third and lower order folds parasitic on it. Especially in the hinge of the major structure, the axial surfaces of these minor folds fan dramatically, and a detailed discussion of this feature is given elsewhere. It will be suggested that tangential longitudinal strain (Ramsay 1967 p.397) was more important in the development of this structure than in the development of any other on the coast.

g) <u>North Scolpaig and Griminish</u>

The steeply dipping eastern limb of the Scolpaig Synform occupies the coast from Scolpaig Bay to Griminish Point. A fault disrupts the structural unity a little, though the displacement along it is probably slight. There are two asymmetrical third order folds on Griminish Point itself, and two rather larger, though still third order, folds to the south of that area. These folds are also

Fig 11 The re-folding of F2 minor folds by a larger scale F3 monoclin south of Griminish Point, North Uist coast.





Poles to F2 axial planes

showing great circle distribution caused by co-axial re-folding asymmetrical with short "flat" limbs (where minor F2 folds are neutral about gentle north-westerly axial surfaces) and long steep limbs where minor F2 folds are upright on steeply dipping N.W.-S.E. striking axial surfaces. The hinge of the southern-most of these four folds is faulted out.

The axial surfaces of the F3 folds dip moderately or steeply to the north-east. Except in one locality (Fig.16) the F3 folds are significantly larger than the F2 folds which they re-fold. Such differences in the relative orders of interfering fold sets are a common cause of the complex appearance of some basement areas. (Fig.11).

h) Griminish Antiform

The northern complement of the Scolpaig Synform is the Griminish antiform. Like the Cliffs Antiform this fold has box-like geometry, and gentle northerly dipping rocks occupy its broad hinge zone. This is exposed on a headland called Blaaskie but henceforth known by the field name of 'Dreary Point'.

The western hinge of the Griminish antiform lies beneath the sea. Its axial trace appears to trend about 150 and its axial plane probably dips steeply north eastwards. The open easterly hinge of the fold has an axial trace which trends about 20° and a north westerly dipping axial plane. Around it, the Dreary Point gneisses swing into a north-westerly strike and reach a maximum dip of 70° in the eastern limb. This eastern hinge zone of the Griminish antiform is partly occupied by a major basic unit which is exposed on the headland of Castell Odair (see chapter 5). This rock mass was boudinaged before the F3 deformation and the hinge of the fold is probably located here because of the existence of the old boudin neck (a common occurrence, see chapter 3).

There are no minor folds or any other evidence of F3 deformation in the hinge of the Griminish antiform. Its eastern limb is short and quickly runs into the hinge of the complementary Griminish synform. This is a close, upright structure whose axial plane dips 85° towards 257°. It is slightly asymmetric, the limb which it shares with the Griminish antiform being its steep limb. The dip of the rocks forming its eastern limb lessens even more as they swing into the northerly dipping orientation which marks the end of the domain of the North Uist Coast Synform.

1:5 THE F2 STRUCTURES

1:5:1 The general F2 structural pattern

The major F2 structure of the North Uist coast consists of large scale very asymmetrical folds whose long limbs contain very few minor folds and whose short limbs are extremely strongly folded on a small scale, (Fig.10). For convenience the long limbs will be called "limb zone" and the short limbs "fold zones". Three fold zones lie within the domain of the North Uist coast Synform; they are exposed on part of Aird an Runaird, (near Hougharry), in the Cliffs, and in the Griminish area, though not on Griminish point itself. The intervening limb zones sometimes carry F2 boudinage structures. Each major fold zone is now described.

1:5:2 Aird and Runaird

Exposure on the west shore of "Hougharry Bay" is not very good. Few fold hinges are actually visible and the accuracy of the mapping is somewhat doubtful. There appear, however, to be eleven close or tight, F2 folds plunging moderately to the north-west, more or less parallel to the dip direction of their axial surfaces. As in the cliffs, this neutral geometry is due to the position of the folds in a gentle, northerly dipping F3 antiformal hinge area.

The F2 fold zone of "Hougharry Bay" may also be seen further south at the northern end of Traig Iar, but inland exposure does not exist between the two localities.

1:5:3) The Cliffs

F2 folds are common throughout the Cliffs area of the North Uist Coast. Most of them lie in the hinge of the F3 Cliffs antiform and therefore have neutral geometry (see above). In most of this region a lack of banding has prohibited the development of very small scale folds within the acid gneiss and most of the visible F2 folds have a fairly high wavelength order occasioned by the relatively thick competent amphibolite bands in which they are developed. Their size is broadly equivalent to that of the third order folds of the F3 deformation.

In the most northerly part of the cliffs, the scale of the minor F2 folds decreases as the amphibolite bands decrease in thickness, and very small-scale folds occur in the banded ("rough") acid gneiss of this region. At "South Scolpaig" there is complex co-axial interference with small scale third folds (see above) but minor folds of obvious F2 age are uncommon in the domain of the Scolpaig Synform, further north.

1:5:4 Griminish

The north-east shore of Scolpaig Bay is part of an F2 limb zone as far as the fault mentioned above (1:3:3,g). Here a steep F3 limb is replaced by a gentle F3 limb and an F2 limb zone by a zone of F2 minor folds. North of the fault, therefore, there are, as in the cliffs, numerous F2 folds lying in a northerly dipping succession. There is, however a significant difference between the F2 folds of these two areas, for the acid gneiss of Griminish contains a well developed secondary planar structure (a gneissic cleavage) parallel to the axial planes of minor F2 folds. Even where minor F2 folds are not well developed, the gneissic cleavage is a conspicuous structure oblique to the gneissic banding. Its occurrence in this area might be a function of a greater proportion of biotite in the rocks here compared with those further south; it may record more intense F2 strain in this area, or it may simply be an inhomogeneous fabric development (see chapter 6) induced by the strong banding originally present in the rocks of the Griminish area. East of Dreary Point the gneissic banding is often completely transposed, the acid gneiss has a stronger linear appearance, and the gneissic cleavage is no longer an obvious structure, features which may indicate a real increase in the intensity of the strain.

In the neighbourhood of Sloc Roe minor F2 folds are re-folded by larger scale F3 folds, and the axial plane poles of the F2 folds lie on a stereographic great circle (Fig.11). Most of Griminish Point is made up of an F2 limb zone, though F2 folds occur in its extreme south-eastern corner, and in this area a strain ellipsoid was constructed from the orientation of veins deformed by F2 strain (see chapter 4).

1:5:5 F2 Boudinage

On the North Uist coast it is sometimes difficult to decide the age of a particular example of boudinage. Some of the excellent exposures of boudinaged Scourie-dyke amphibolites on Griminish Point are undoubtedly of F2 age, however, because the F1 foliation of the amphibolites is deformed co-evally with the boudin formation (Fig.12), and the boudins themselves are involved in F3 folds. Boudinaged amphibolites are more common in the Griminish area than anywhere else on the North Uist coast. Since lithological combinations similar to those of Griminish also occur in more southerly parts of the coast this distribution, like the gneissic cleavage and the transposed banding referred to earlier, may indicate more intense F2 strain in this part of the coast.

1:5:6 The original orientation of the F2 structures

Unless there has been massive rotation of the F1, F2 and F3 linear structures (and therefore the whole foliation pile) by late F3 deformation, the orientation of the F2 folds in the broad hinge zones of the second

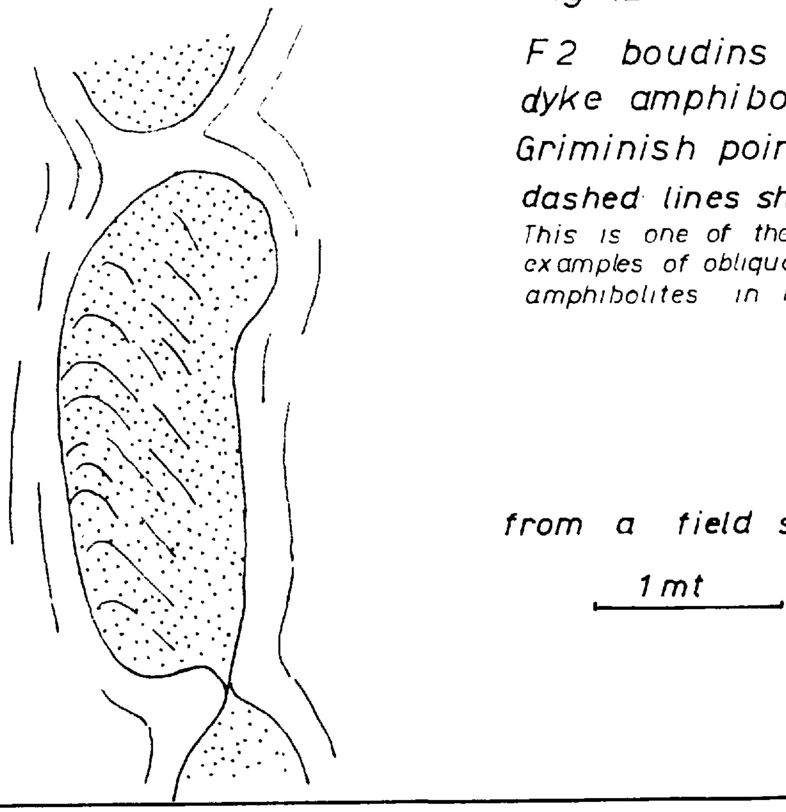


Fig 12

F2 boudins in Scourie dyke amphibolite (stippled)
Griminish point

dashed lines show foliation
This is one of the extremely rare examples of obliquely foliated amphibolites in North Uist

from a field sketch

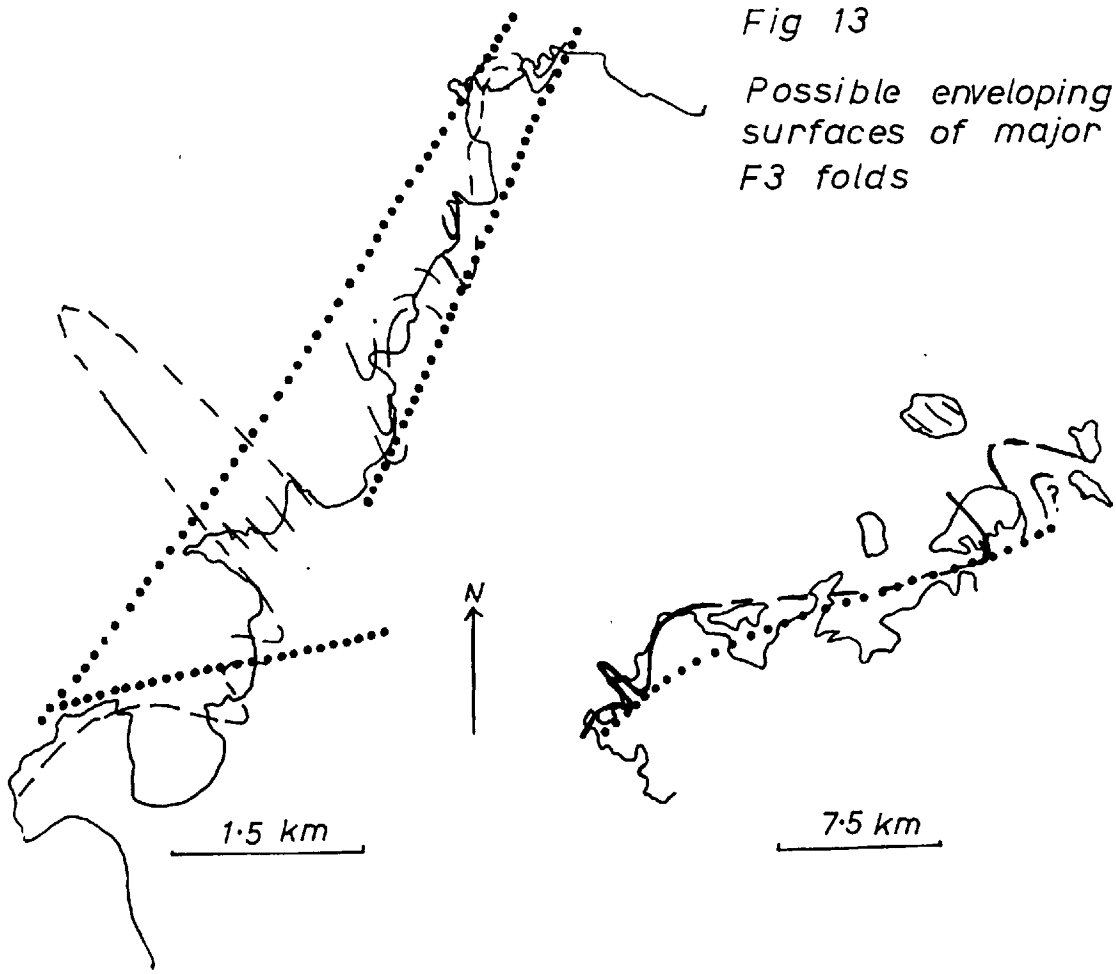


Fig 13

Possible enveloping surfaces of major F3 folds order F3 antiforms is probably close to the orientation in which they formed (i.e. before the F3 deformation took place). Similarly the enveloping surfaces of F3 major folds must reflect the general orientation of the foliation pile after the F2 deformation had taken place.

In both cases (even though several F3 enveloping surfaces may be constructed depending upon exactly which F3 major folds are chosen (see Fig.13))it seems that F2 folds developed with neutral geometry on axial surfaces dipping to the north-west at moderate angles. The axial surfaces of the F2 and F3 folds developed at very high angles to each other.

1:6 THE F1 STRUCTURES

Folds of indisputable F1 age are seen in only a very few places on the North Uist coast. The most conclusive examples of pre-F2 folds are in felsic veins near Sloc Roe (see Fig.16), but even these may be Scourian. Highly modified F1 boudins are seen on Dreary Point (Fig.17), and in the Cliffs (Fig.14), but the most important structure of F1 age is the fabric of the Scourie-dyke amphibolites. Although the F1 fabric may be severely modified by later deformation (see chapter 6) its ultimately F1 age is apparent when we see it folded in F2 folds. In both North Uist and the Sound of Harris F1 is defined as the first deformation to produce fabrics in the Scourie-dyke amphibolites.

1:6:1 The original orientation of the F1 structures

The deformed boudins of Dreary Point (Fig.17) suggest that the XY planes of the F1 and F2 deformations were superimposed at near right angles to one another, and a similar impression is conveyed by the deformed felsic veins near Sloc Roe (Fig.16). It is not possible to construct an accurate enveloping surface for the F2 major folds, so that we are left with only two lines of reasoning (i) Although we can identify F1 boudinage in the Scourie-Dyke amphibolites, there are no F1 folds in these bodies, also the F1 planar fabric of the amphibolites is almost always exactly parallel to the margins of the bands, never oblique to them in any way (as we might expect if the fabric of the bands were axial planar to folds in the same bands). This all suggests that during the F1 deformation the Scourie dykes were flattened rather than folded (i.e. the XY plane of the F1 deformation lay very close to the original trend of most of the Scourie-dykes. (ii) If the observation that the XY planes of the F1 and F2 deformations were superimposed at near right angles is correct, it follows that when the XY plane of the F2 deformation is returned to its original north-westerly dipping orientation, the XY plane of the F1 deformation takes on a north-west trend and a nearly vertical dip. This, then is the postulated original orientation of the F1 XY plane and the original orientation of the Scourie-dyke amphibolites on the North Uist coast. It is the same as that suggested for the dykes and F1 structures of the Sound of Harris (where the orientation of some almost undeformed dykes provides compatible evidence) and it is the same as the orientation of completely undeformed parts of the Scourie-dyke swarm on the mainland.

Fig 14

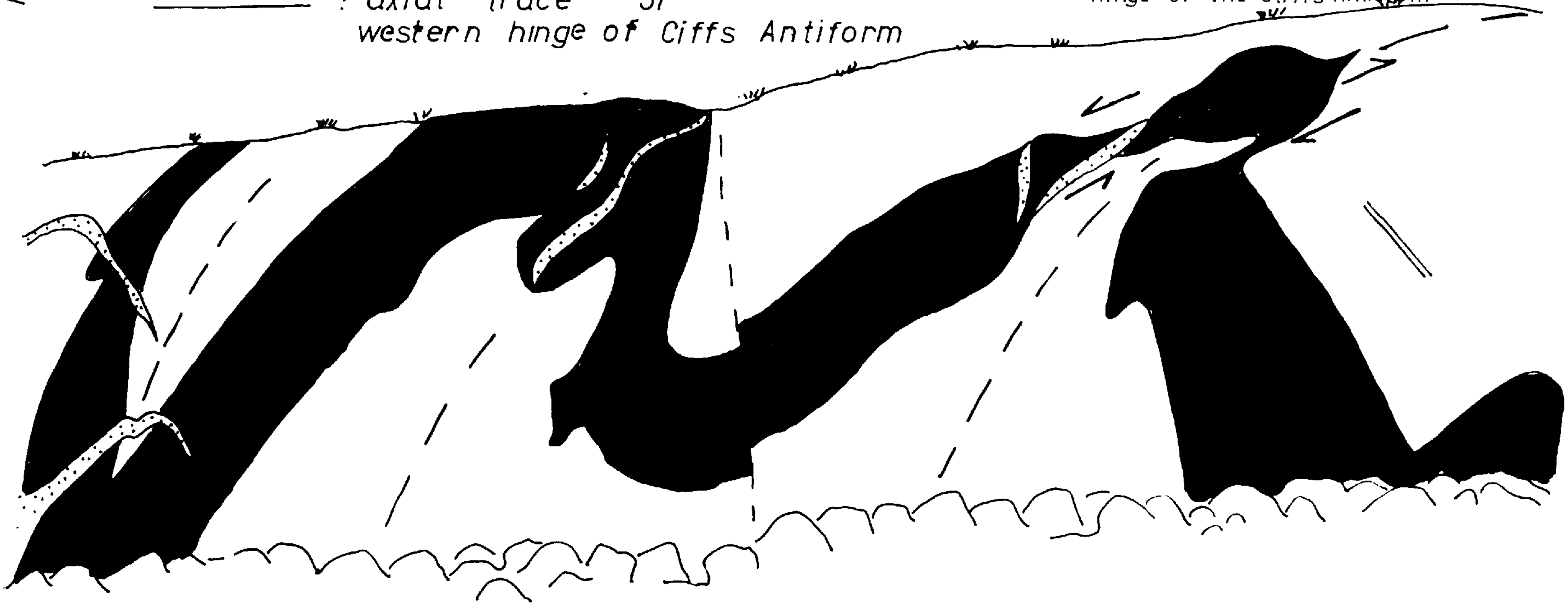
F2 folds in Scourie dyke amphibolite: cliff face exposure near a natural arch on the North Uist coast. The F2 axial surfaces are folded by the western hinge of the F3 Cliffs antiform

Some pegmatite is developed in intensly strained areas, some is post kinematic.

The complex geometry below is probably the result of

1) F1 boudinage
2) F2 folding of the boudinaged layer
3) shearing in the same sense as flexural slip towards the hinge of the Cliffs Antiform

══ : axial trace of western hinge of Ciffs Antiform



about 10 mts

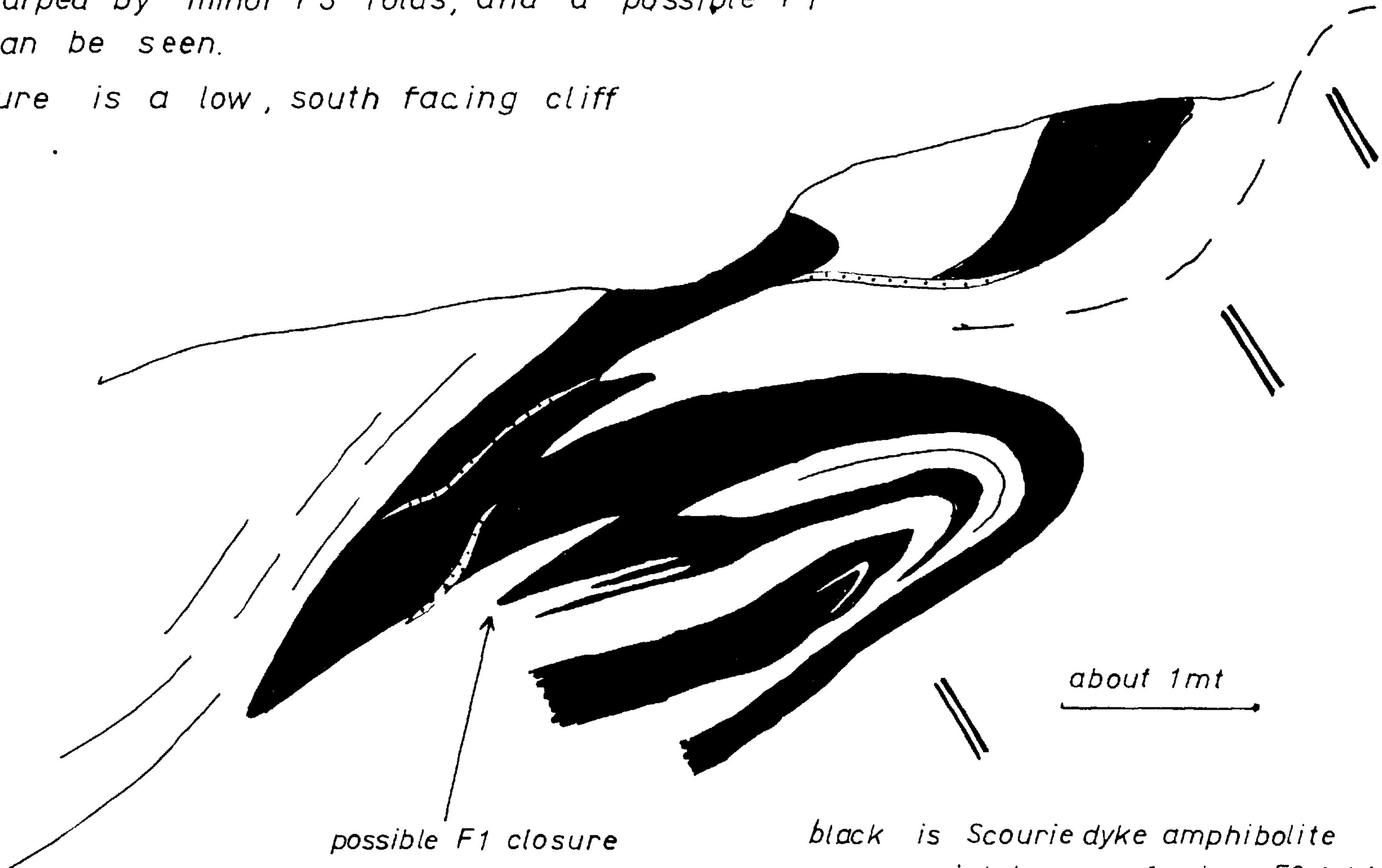
looking north

pegmatite is stippled

Fig 15

F2 folds with nearly neutral geometry in the hinge zone of the Cliffs Antiform. The axial surfaces are warped by minor F3 folds, and a possible F1 fold can be seen.

Exposure is a low, south facing cliff



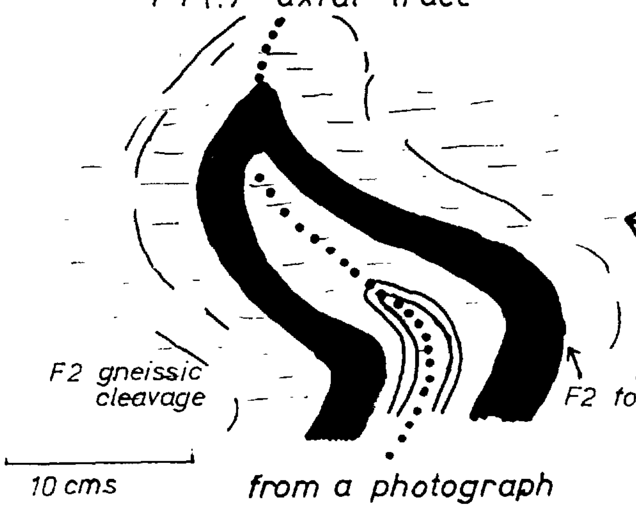
black is Scourie dyke amphibolite

═══ axial trace of minor F3 fold

pegmatite is stippled

Fig 16 a) Pre-F2 folds in an amphibolite band, Sloc Roe, Griminish

The folds are F1 if the amphibolite is a Scourie dyke amphibolite



b) General relationships (from a sketch) Most of the F1 folds are contained in felsic bands. Low order F3 folds are present

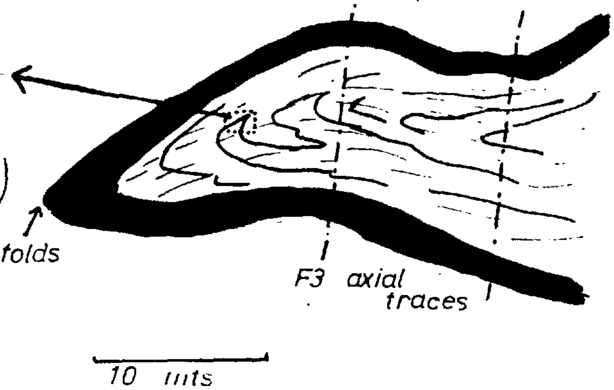
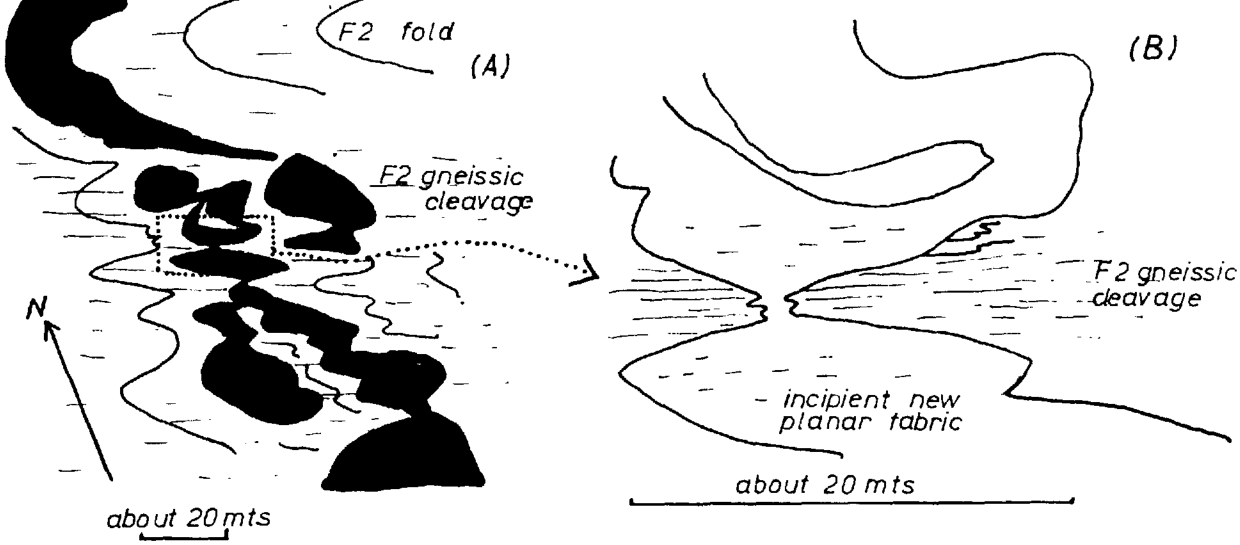
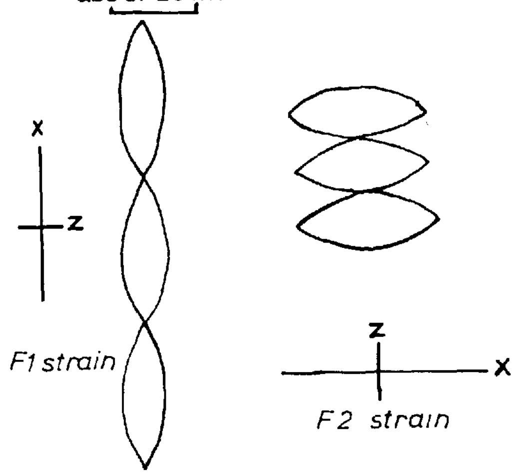


Fig 17

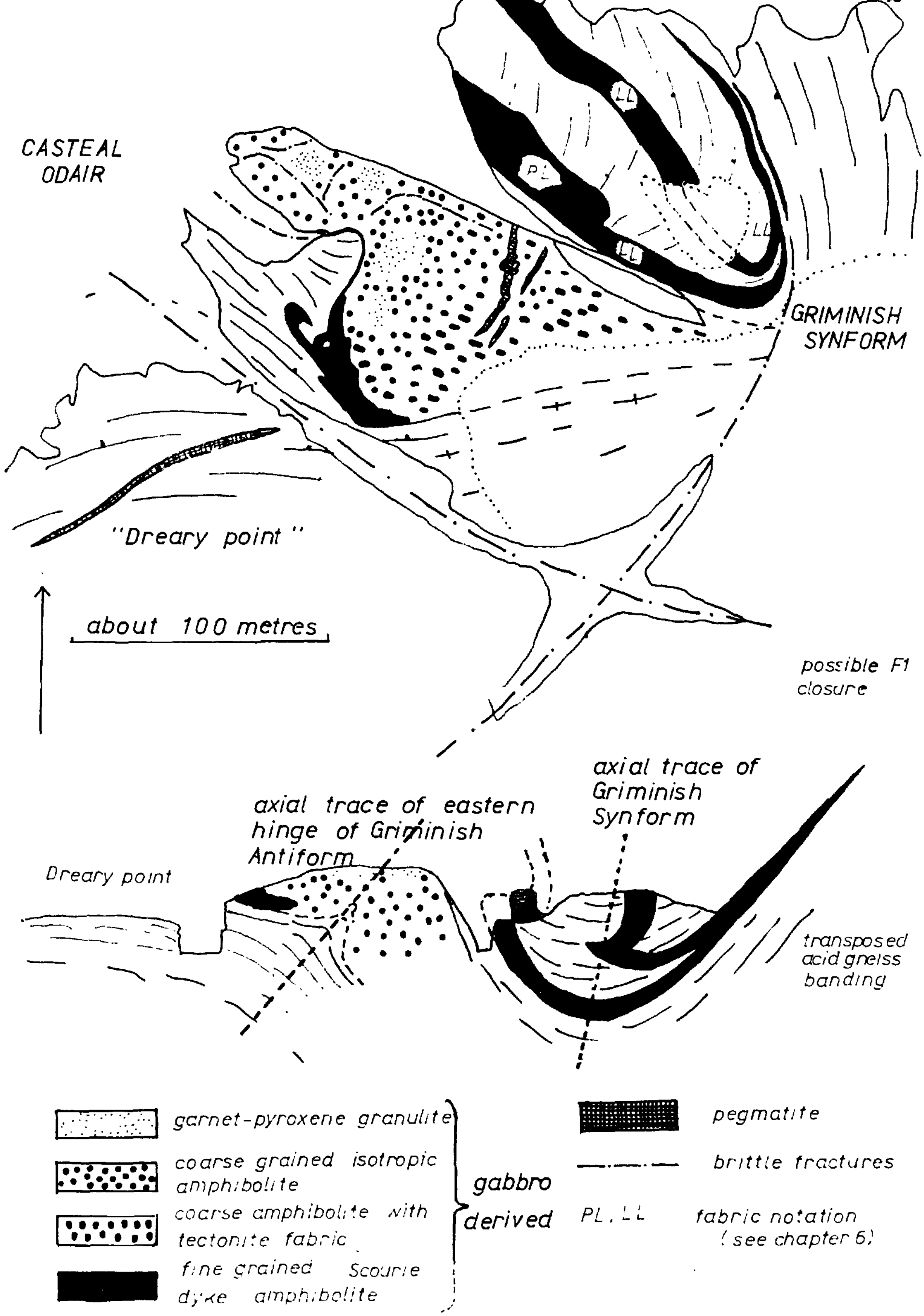




Deformed boudins in Scourie dyke amphibolite "Dreary Point", Griminish, North Uist (A) & (B) with an explanation of their origin (C)

(A) is an enlarged field map
(B) is a field sketch

Fig 18 Map and sketch-section of the Griminish Synform and Casteal Odair.



CHAPTER 2

THE REGIONAL GEOLOGY OF THE SOUND OF HARRIS

2:1 INTRODUCTION

The Sound of Harris is a maze of small islands, but I have worked only on the four largest - Pabbay, Berneray, Killegray and Ensay (see maps 4, 5, and 3). As on the North Uist coast, three phases or Laxfordian deformation can be recognised, and as in that area F1 is defined as the first deformation to produce fabrics in the Scourie-Dyke amphibolites, while the major structural pattern is produced by folding of F3 age. (Fig. 2)

A major F3 synform (the Berneray synform) crosses Berneray, and a complementary major antiform is thought to lie in the Sound of Harris, the hinge lying close to the north-east coast of Pabbay. The presence of this "Sound of Harris Antiform" is indicated by large scale changes of foliation orientation and, more significantly, by the opposed senses of the F3 minor folds in the two limbs, and the axial surface orientation of these minor folds. The minor F2 folds of Berneray and Pabbay on the one hand, and Ensay and Killegray on the other, have the same "Z" asymmetry within foliation of opposite dip, and have suffered higher order re-folding by the antiform. The steep limb common to the Berneray Synform and the Sound of Harris antiform passes through central Berneray and western Pabbay. It is the dominant geological feature of these islands, and will henceforth be known as the "Sound of Harris Steep Zone". The strike of this zone changes from 150^{0} - 160^{0} in Berneray to 125^{o}- 135^{o} in Pabbay, but the reason for this swing is imprecisely understood, and may be due either to different positions on an open F3 fold parasitic on the Berneray Synform, or to later folding altogether.

2:2 LITHOLOGIES OF THE SOUND OF HARRIS ROCKS

2:2:1 Pabbay and Berneray

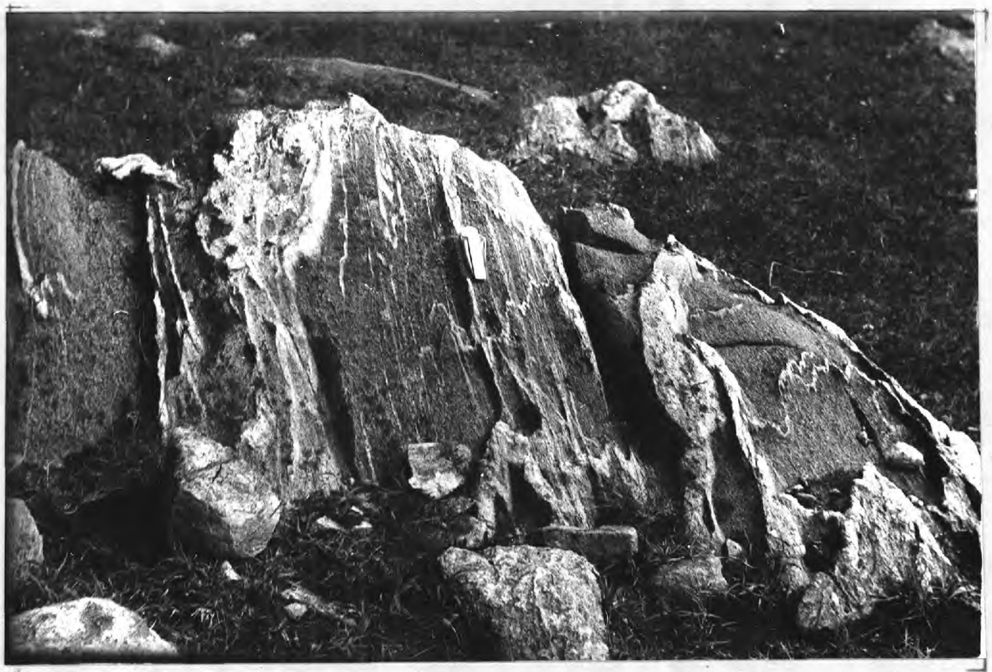
The flaggy, quartzo-felspathic gneisses of the southern shore of Berneray are so ordinary that they defy adequate description, and Dearnley, in coining the expression "grey gneisses" has given us the best term for them. Similar rocks occur on the north shore of "Sand Hill beach" in Berneray and in the eastern half of Pabbay, but most of the rocks of the Sound of Harris steep zone are of markedly different appearance, being fine grained biotitic or hornblendic planar rocks identical with some of those in the Langavat belt of South Harris (Dearnley, 1963; Myers, 1968; and chapter 8 of this thesis). The Berneray rocks differ from those of Langavat only in that they contain no rocks of distinctive metasedimentary composition, unless some laminated hornblendic rocks be so considered. On the north coast of Pabbay, however, near Alarip Bay there is a significant array of true metasediments including quartzites and garnetiferous semi-pelites. These rocks occur adjacent to the largest basic mass in Pabbay (map reference P4), indeed, this body is completely surrounded by a narrow band of garnetiferous psammite. Such a distribution suggests a significant relationship between basic rocks and the metasedimentary relics, and the possibility that the relics represent fossil thermal metamorphic aureoles cannot be dismissed (see Coward et al., 1969 and chapter 8 of this thesis).

Two distinctive lithologies can be matched on either side of the Sound of Pabbay. The first is a group of rocks which it is convenient to refer to by the field name of "pink and blue gneisses", for they contain abundant pink felsic veins set in a homogeneous biotite rich matrix which usually exhibits a good gneissic cleavage. The veins could originally have been either pegmatite bands or the hightly modified quartzitic layers of a one-time metasediment. The pink and blue gneisses are of special interest because they provide the best material in the southern Hebrides for the determination of strain from the orientation of deformed veins in a homogeneous matrix (see plates 1).

The second rock unit common to both Pabbay and Berneray is an ultrabasic body first noted by Jehu and Craig (1926) and later described again by Livingstone (1965). Mineralogically it is almost identical with the ultrabasic bodies scattered throughout the Langavat belt of Harris (also studied by Livingstone), for the central parts of the mass contain areas of pure olivine rock which is increasingly retrogressed outwards in a zonal manner and minerals such as anthophyllite, talc, calcite and (marginally) actinolite occur. There are two exposures of ultrabasic rock on Berneray, but though identical, they are not continuous along strike. The ultrabasic mass on Pabbay is imperfectly exposed and disappears in totally unexposed ground. The original nature of these rocks is obscure, they may have been originally intrusive ultrabasic dykes or sills, but there is no evidence of their ultimate age.

In northern Pabbay the large basic mass referred to during comment on the metasediments is divided by a major low angle fracture. The upper parts of the body are composed of highly foliated, almost laminated, amphibolite while below the fracture are large areas of isotropic rock undeformed save for crushing

Plate 1 "Blue and pink" gneisses, S.E. Pabbay.



and pseudotacylite veining associated with the fracture. The age of the body is problematical - it may be the same age as the Scourie-Dyke amphibolites (although it does not quite resemble them), or it may be earlier.

In Berneray there is a deformed granite - the Berneray Granite - which forms the rough ground of Borve Hill. It is a coarse-grained grey granite with a strong F2 gneissic cleavage produced by the planar orientation of the abundant biotite crystals. Banding is more or less absent in the central parts of the exposure, but where present it is sometimes oblique to (earlier than) the gneissic cleavage. Towards the margins of the mass, however, the lithological banding becomes increasingly important and the granite passes transitionally into the surrounding gneiss. There are a few deformed Scourie-Dyke amphibolites within the mass; some are paired in outcrop and are probably tightly folded (F2), while the largest possesses a coarse-grained isotropic centre. The Berneray Granite is most likely of Scourian age.

2:2:2 Ensay and Killegray

Most of Killegray is composed of quartz-biotite gneisses identical with those of the Sound of Harris steep zone, though "grey gneiss" is exposed in the southern parts of the island. Most of Ensay is made of quartzo-felspathic gneisses (which Dearnley (1963) curiously regarded as metasediment) though on the east coast, adjacent to a large igneous mass is a thin strip of garnetiferous biotite-rich gneiss of true metasedimentary aspect. This large igneous mass is the same as, and probably continuous with, Dearnley's "pyroxene graunlite" on South Harris, and will be examined in detail in the chapter on South Harris. The rock forms two headlands in eastern Ensay.

2:2:3 Summary

(a) The gneisses of the Sound of Harris steep zone and those in much of Killegray are parts of the same belt of rocks lying on opposite limbs of the Sound of Harris Antiform, the Killegray rocks lying in the gentle dipping N.E. limb, the rocks of the steep zone forming the steep S.W. limb.

(b) The similarity between these rocks and those of the Langavat belt of Harris is startling and may be regionally important, for no other Hebridean rocks look quite like this. This matter is examined fully in chapter 8.

TABLE I

LITHOLOGIES IN THE SOUND OF HARRIS

ROCK TYPE	OCCURRENCE
Fine grained quartz-biotite gneisses (including "pink and blue" gneisses) possibly of metasedimentary origin.	The Sound of Harris steep zone and much of Killegray.
True metasediments (Psammites and garnetiferous semi-pelites)	Small patches in North Pabbay and East Ensay.
Large masses of basic rocks (other than Scourie dyke amphibolites).	North Pabbay Ensay
deformed granite	Berneray
Scourie Dyke amphibolites	Universal
Ultrabasic bands	Locally in the Sound of Harris steep zone
"Grey gneiss"	All remaining areas

2:3 STRUCTURAL GEOLOGY

2:3:1 Introduction

Unlike the North Uist coast, the Sound of Harris area has no simple system of fold ordering by which precise description can proceed. Here the Sound of Harris Antiform, the Berneray Synform, and the principal F2 fold zones (2:3:3) are termed "major folds", and all lower order parasitic structures are called "minor folds".

The regional structure of the Sound of Harris is a product of major F3 folding, therefore the F3 structures will be described first.

The structures of the three deformation phases are not always co-axial. F1 and F2 linear structures are often deformed, and their orientation varies. F3 structures plunge gently to either the north-west or the south-east.

2:3:2 The F3 structures

(a) Berneray (Map 4)

The hinge of the Berneray Synform lies in poorly exposed ground around the north-east shore of Loch Borve. The general NW-SE trend of its axial trace is shown on map 4; its axial surface dips moderately to the north-east. Parasitic minor folds are uncommon in the steep limb of the major fold (i.e. in the Sound of Harris steep zone) and exposures of them are only to be found in the following localities

(i) South of Borve (axial surfaces dip about 60° to the N.E.).
(ii) Sand Hill (very open structure with E-W axial trace).

(iii) Near the schoolhouse (exposure imperfect).
(iv) Smaller scale structures on the N.E. coast.
The sense of these folds is consistent "Z" and their flatter limbs are shorter.

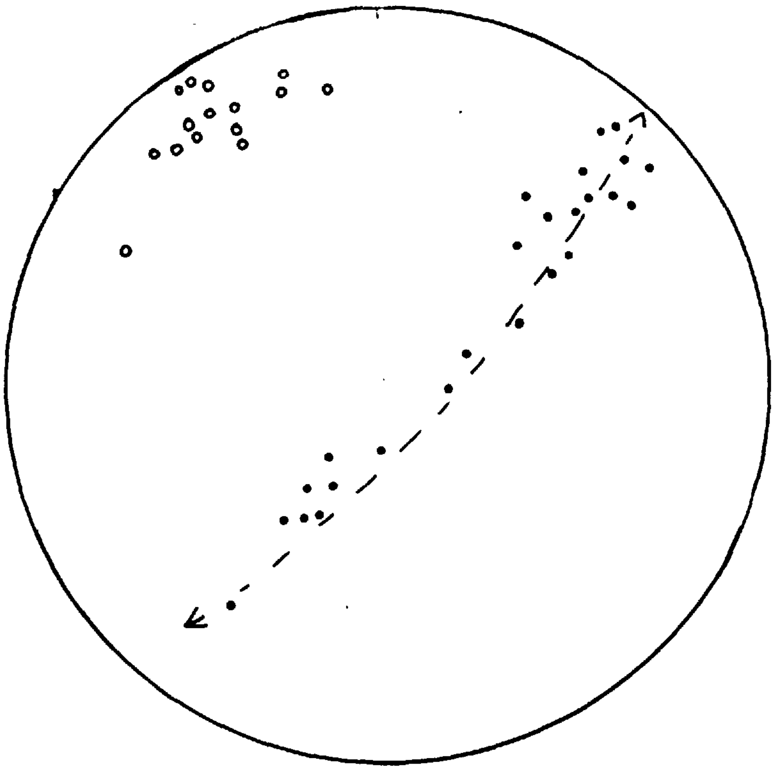
The flat limb of the Berneray Synform forms the whole of the southern shore of the island and differs from the steep limb in that it carries a large number of minor folds of several different orders. These folds are close or tight about axes plunging north westward in westerly dipping axial surfaces, and the largest ones (those directly parasitic on the Berneray Synform) have "S" asymmetry. On Rovoisinish there is simple co-axial interference between F3 minor folds and minor F2 isoclinal folds, and FIG 1 shows a girdle of F2 axial plane poles about the F3 axes, and the cluster of F3 axial plane poles.

Although the flat limb of the Berneray Synform is more strongly folded than the steep limb it need not have suffered more F3 strain. We shall see later (chapter 7) that it is difficult to assess the amount of F3 strain in the Sound of Harris steep zone. A further point of passing interest at this stage is the lack of tectonite fabrics in the amphibolite bands of the flat limb, contrasted with the well developed fabrics of similar rock units in the steep limb. This may be related to different strain states in the two limbs, but the field evidence suggests more profound post-kinematic crystallization took place in the flat limb, and the earlier tectonite fabrics were destroyed.

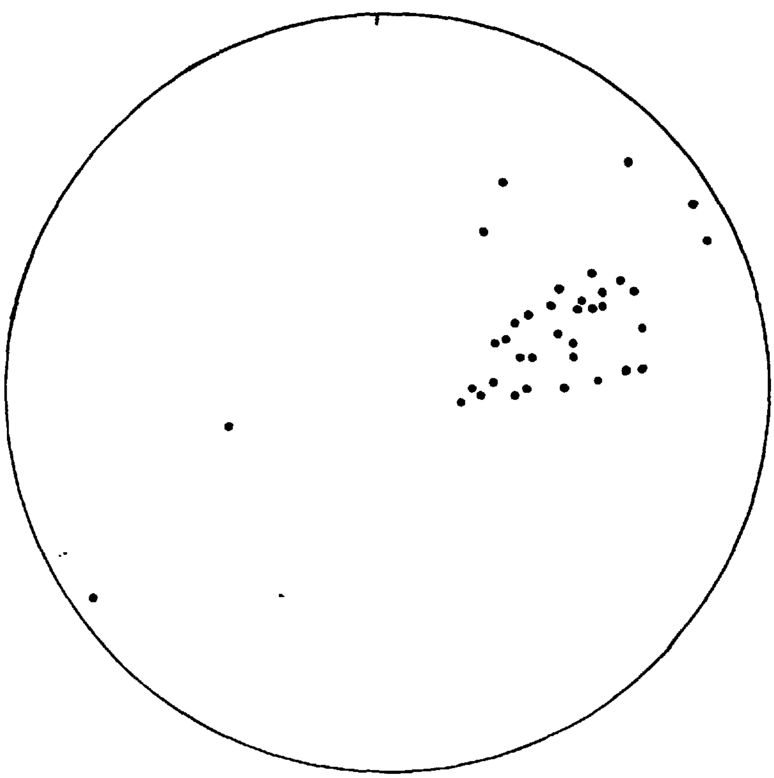
(b) <u>Pabbay</u>

The entire island of Pabbay lies within the Sound of Harris steep zone i.e. within the steep limb of the Berneray Synform. Two minor F3 folds are exposed on the north coast and are large

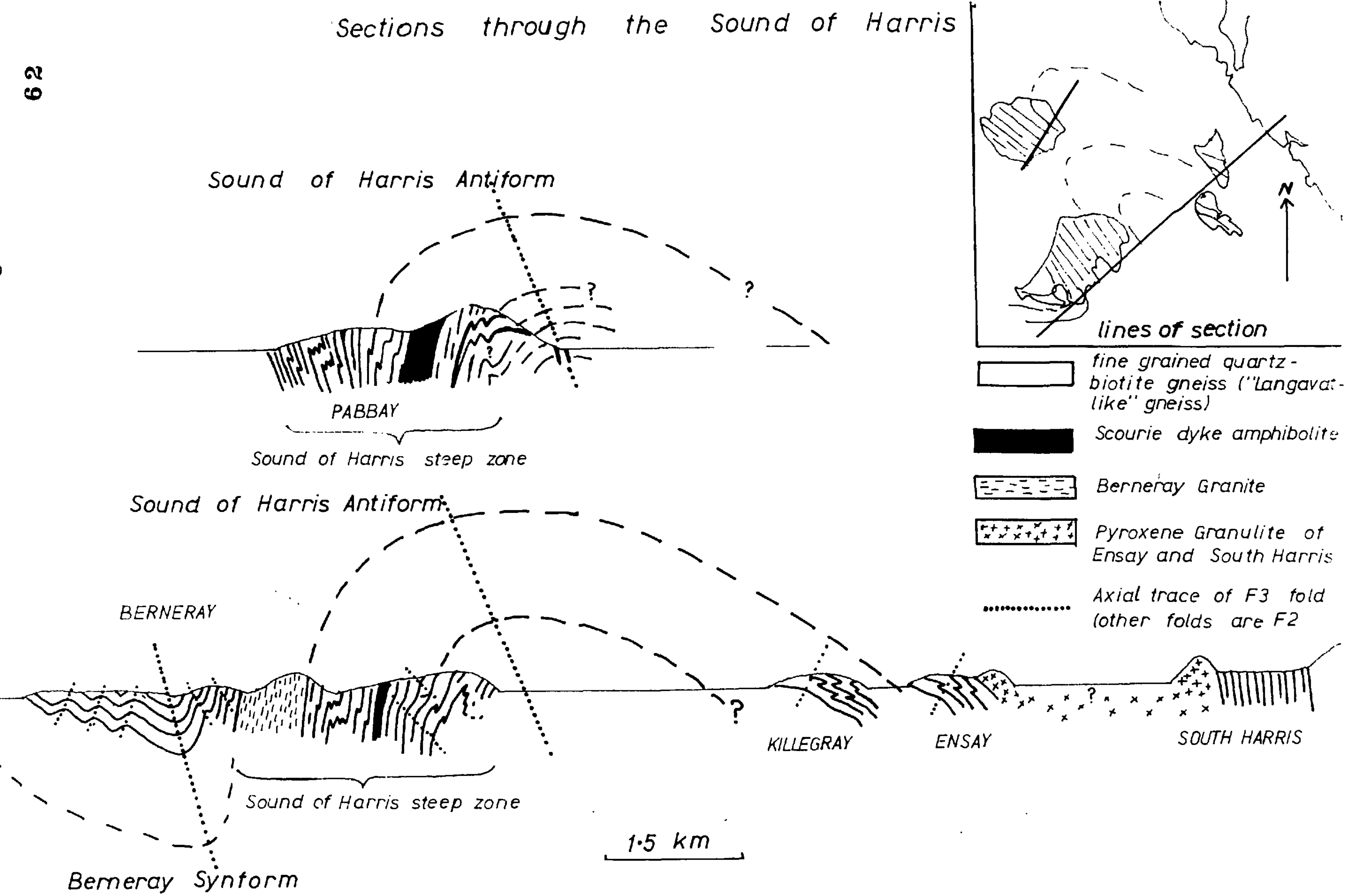
Fig 1 Rovoisinish and the south coast of Berneray



A girdle of poles to F2 axial planes about F3 fold axes



poles to F3 axial planes



Sections through the Sound of Harris
Sound of Harris Antiform
PABBAY
Sound of Harris steep zone
Sound of Harris Antiform
BERNERAY
KILLEGRAY
ENSAY
SOUTH HARRIS
Sound of Harris steep zone
1·5 km
Berneray Synform
N
lines of section
fine grained quartz-biotite gneiss ("Langavat-like" gneiss)
Scourie dyke amphibolite
Berneray Granite
Pyroxene Granulite of Ensay and South Harris
Axial trace of F3 fold (other folds are F2

Fig 2

enough for accurate presentation on the map; their axial traces run slightly north of east, and their axial surfaces dip moderately northwards. The major effect of F3 deformation in Pabbay, however, (also visible in Berneray) is the fanning of both the foliation and the F2 axial planes; a change in the dip direction of the Sound of Harris steep zone from steeply S.W. in the north east to steeply N.E. in the south-west. The fanning is produced by the gradual change in dip towards the hinge of the Sound of Harris Antiform. (FIG 2).

(c) Ensay and Killegray

These islands lie in the flat north-eastern limb of the Sound of Harris Antiform, and the foliation and F2 axial surfaces dip fairly gently to the north-east. Very open fairly large scale F3 folds with "S" asymmetry and north-south trending axial traces occur in both islands, and some smaller scale F3 monoclines occur on the south coast of Ensay.

(d) Late, possibly, F3 lineation

In several localities within the Sound of Harris (e.g. Bruista, Berneray) a late, nearly horizontal, lineation in hornblende or biotite overgrows deformed rodding structures, fold axes and mineral lineations of F2 age. Though often barely discernible amid randomly oriented crystals it is a significant structure, and is examined more closely in chapter 7.

2:3:3 The F1 and F2 structures

(a) Introduction

There are no very large-scale F2 folds in the Sound of Harris area. The regional pattern of the F2 folds is similar to the pattern of the F2 structures of the North Uist coast, namely long limbs with relatively few minor folds (limb zones), and shorter, intensely folded limbs (fold zones), although in the Sound of Harris the sense of these structures is different to the ones seen in North Uist, it is "Z" not "S". There are three main "fold zones" in the Sound of Harris steep zone, one badly exposed in central Berneray and two in the western half of the north coast of Pabbay (map 6 FZ1 and FZ2).

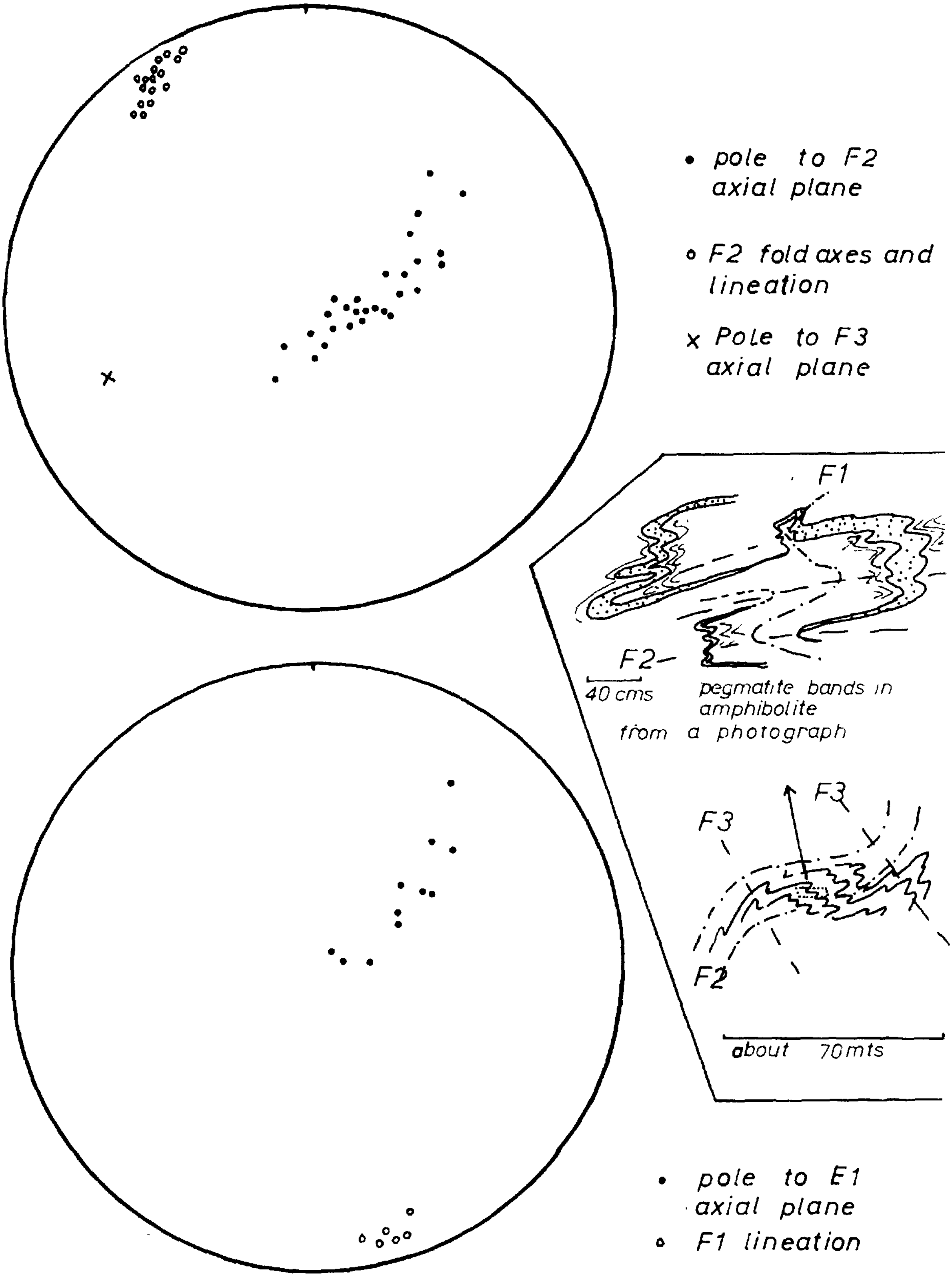
In north-east Pabbay there is a dearth of F2 structures and the Scourie-Dyke amphibolites cross-cut the banding of the acid gneisses and are virtually undeformed. It is a "deformation low" comparable with those at Udal, North Uist, Garry a Siar, Benbecula and Ardivachar Point, South Uist, and a detailed discussion of it appears later in this chapter.

(b) <u>F1 and F2 structures in Ensay and Killegray</u>

Minor F2 folds are tight, a gneissic cleavage is developed parallel to their axial surfaces in suitable lithologies, and the F1 foliation of the Scourie-dyke amphibolites is usually crenulated. The axes of F1 and F2 folds and boudins, together with other early linear structures are deformed on a large scale in the plane of the foliation, and they form a partial great circle on a stereographic plot (see Fig. 1, chapter 7). Deformed linear elements also occur in the Sound of Harris steep zone and the first part of chapter 7 examines their origin.

Pegmatite bands in the acid gneisses in both islands are commonly boudinaged. Sometimes boudinaged bands are folded by F2 minor folds and therefore pre-F2 in age, sometimes it seems likely that they are F2. Isolated (boudinaged?) ellipsoidal masses of garnetiferous amphibolite whose internal

Fig 3 Re-folding relationships, northern end of "Sand Hill beach", Berneray.



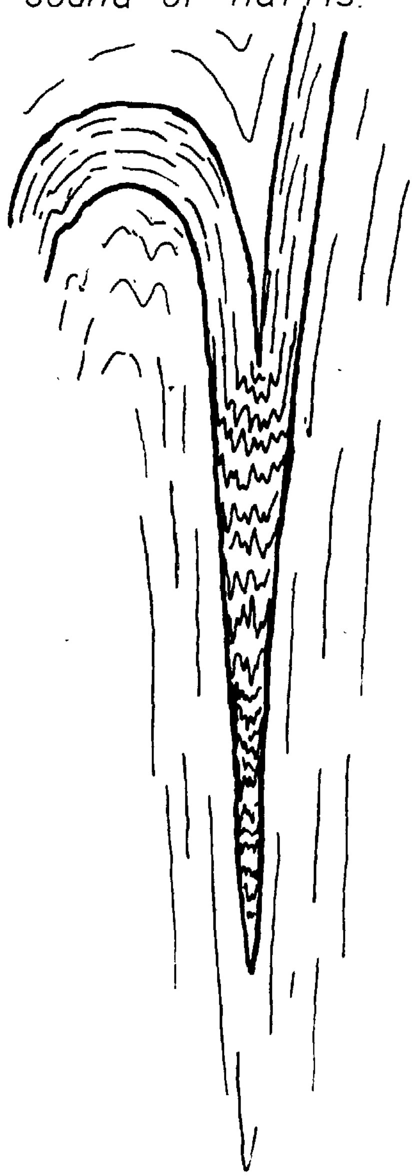
foliation is oblique to the foliation of the acid gneisses occur occasionally. These detached masses are identical lithologically, and in their structural relationships, with the garnetiferous amphibolites distributed throughout the Leverburgh Gneisses of South Harris (see chapter 8 where a more thorough description is given).

(c) Berneray

F2 minor folds are tight with "Z" or "M" profiles over the whole of the island except the north east coast where "S" profiles occur locally in the north eastern limb of a larger scale F2 fold which closes at the northern end of Sand Hill beach. In this area there is also interference between F2 minor folds and earlier folds, which, because of their development in Scourie-Dyke amphibolites, are probably of F1 age. FIG.3 summarizes these re-folding relationships, (the girdle of F2 axial plane poles is due to an open F3 monocline in this area). Though appearing co-axial on the figure they are not quite so, the F1 lineation in the foliation planes of both acid gneisses and amphibolites plunges about 5° towards 166° while the F2 lineation, produced by crinkling of the early foliation, plunges north-westward (e.g. 4° towards 333°).

An F2 gneissic cleavage occurs throughout central Berneray; in the major limb-zones it is usually parallel to the banding, and obliquity of the two structures is only seen in fold hinges and fold zones. Gneissic cleavage is not well developed in the "grey gneisses" of the extreme southern and northern parts of the island, and best developed in "pink and blue" gneisses near Bruista. In central Berneray and in much of Pabbay amphibolite bands with intensely crenulated internal foliation can be seen (FIG.4). Bands with this type

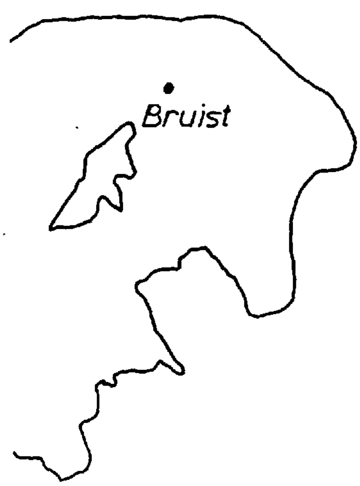
Fig 4   crenulated foliation in folded (F2) scourie dyke amphibolites in the Sound of Harris.

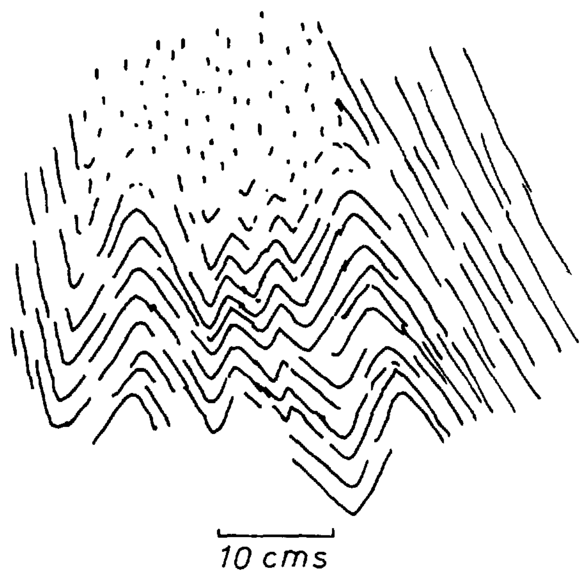


(left) Schematic representation of occurrence

(below) detail of crenulated foliation, Bruist, Berneray

The sketch also shows a linear fabric (dotted) derived from the old (F1) foliation. This is known as homogeneous fabric development (see chapter 6)





of internal structure die out along their lengths and it is clear that they are the flattened hinges of very tight or isoclinal F2 folds whose shape is exaggerated in the flat exposure by gently plunging hinges.

(d) Pabbay (Map 5)

(i) Description of the structures

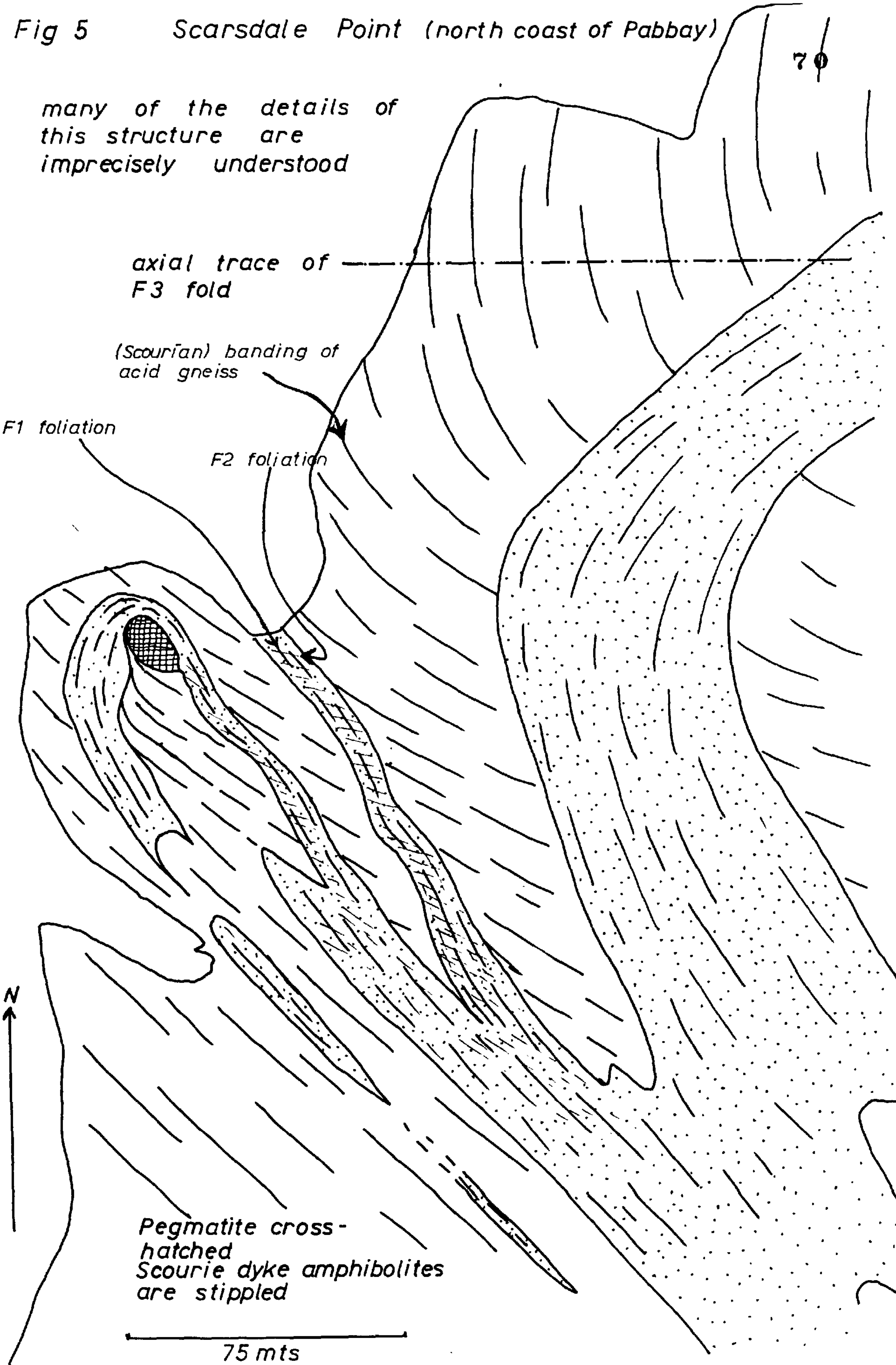
The western half of Pabbay contains fine grained, very planar biotitic gneisses of "Langavat type" and, as in central Berneray, the F2 folds are tight and F2 gneissic cleavage is almost universally developed. Scourie-Dyke amphibolites are always concordant with the banding of the acid gneisses, and the main F2 fold zone occurs where a thick amphibolite unit is exposed (FZ1). The F1 fabric in the amphibolites is strongly folded or crenulated by F2 deformation and the banding of the acid gneisses is also strongly folded (some of these folds are figured in chapter 3). The geology of this part of Pabbay is identical with that in central Berneray.

A dramatic change occurs north-eastwards across the strike of the rocks, however. In the neighbourhood of Alarip Bay on the north coast the F2 folds are no longer tight structures, but open monoclines, and gneissic cleavage is absent. Also, the rock type changes in this area, - east of Alarip Bay the rocks are "grey gneisses" quite different from the biotitic "Langavat - like" rocks of the west. Two thick Scourie-Dyke amphibolite units are involved in (and probably initiated) the major F2 fold zone FZ2, and traced inland they become progressively thinner and eventually join, (an original junction, probably, not an F1 fold). The acid gneiss banding is not folded in this area, but is markedly discordant to the amphibolite bands and occupies a position close to the axial surfaces of the folds in the bands. Inland, as the amphibolites become thinner, they become progressively nearer to parallel with the banding of the acid gneiss.

At Scarsdale Point on the north coast, a branch of one of these major amphibolites itself divides into two. Both branches carry an F1 fabric parallel to their margins and both show a 10°-15° discordance with the acid gneiss banding. Both also carry a later planar fabric in a direction similar to, though slightly refracted from, that of the acid gneiss banding. The more westerly of the branches is folded and both fold limbs are gently warped. FIG 5 is a sketch map of the Scarsdale point area.

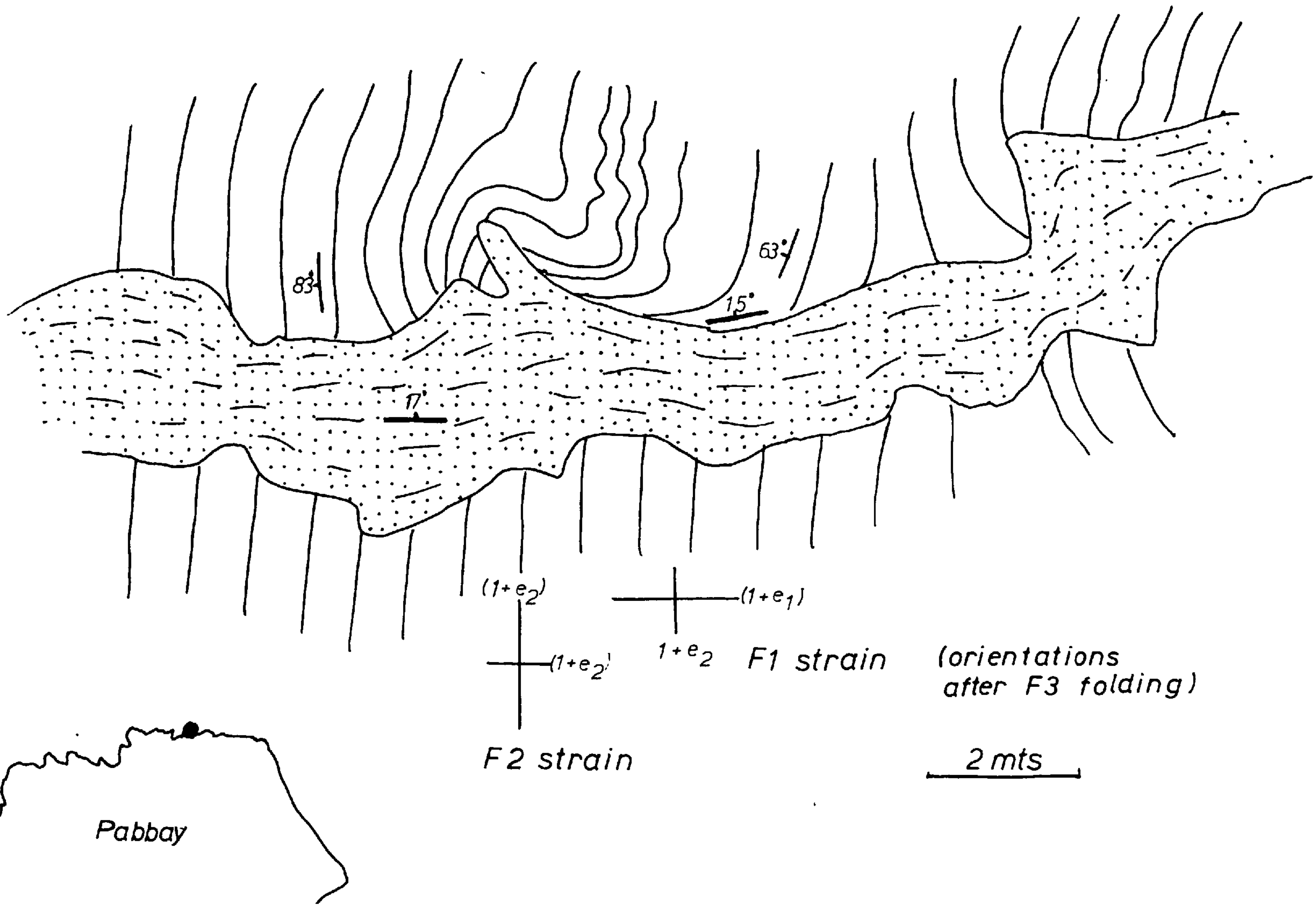
East of Scarsdale point the Scourie-Dyke amphibolites on the north coast are not folded at all, and are very strongly oblique to the banding of the acid gneiss, so that the Scourian age of the banding becomes obvious (FIG.6). The amphibolites are internally foliated parallel to their margins, and these margins are usually rather irregular. There is a progressive variation in the angular relationships of the amphibolites and the gneiss along the north coast. Around map reference P1 the amphibolites dip gently while the banding is steeply dipping, but towards the east the relationships progressively change until at P2 the amphibolites are nearly vertical and the banding nearly horizontal. This change in relative inclination is attributed to F3 folding around the broad hinge of the Sound of Harris Antiform, and we may view it, together with the two minor F3 folds in the area, as an extra complication in the present discussion.

Fig 5 Scarsdale Point (north coast of Pabbay)



Slightly deformed Scourie dyke in the area of low deformation on the north coast of Pabbay

Fig 6



The relationships described above strongly suggest that F2 deformation was slight in the north eastern corner of Pabbay. Crenulations of the Fl foliation in the amphibolites, and perhaps some of the irregularities at the margins of these bodies are the only visible manifestations of F2 strain. In eastern Pabbay the margin of this area of low deformation is well seen. It is found to occupy an antiformal domain complementary to a south easterly plunging, fairly large scale F2 synform (P3) which, unlike the fold zone FZ2 involves acid gneiss banding instead of amphibolite bands only. This synform is a fairly tight structure, and the amphibolite bands in its hinge sometimes show a new axial plane fabric, suggesting that F2 strain was significant in this position. The tailing off of F2 deformation occurs, therefore, in less than 500 metres.

(ii) The original orientation of the structures of the three deformation phases.

Before we can proceed with a discussion of the origin of the "deformation low" in Pabbay we must have some idea of the relative orientations of the three strains which have affected the Sound of Harris rocks. This can be deduced roughly from the orientation of the enveloping surfaces of the various folds, the enveloping surface of the major F3 folds giving the original orientation of the F2 structures, and the enveloping surfaces of the F2 fold zones in their original orientation giving the orientation of the first foliation of the Scourie-Dyke amphibolites (i.e. the XY plane of the Fl strain). As on the North Uist coast, the XY plane of the F2 strain and the axial surfaces of the F2 folds seem to have originated with a NE-SW strike, though in the

Sound of Harris the dip of these surfaces seems to have been lower than in North Uist, and may have been nearly horizontal. Again as in North Uist, the XY plane of the F1 strain seems to have originally had the same orientation as that of the F3 deformation, namely NW-SE and steeply dipping. This is parallel to the most likely original trend of the Scourie-Dyke amphibolites, and explains the consistancy of F1 foliation parallel to the margins of these bodies, and the general absence of F1 folds involving amphibolite bands.

(iii) Explanation of the structures in north-east Pabbay.

The absence of F2 deformation in north-eastern Pabbay enables us to examine the F1 structures in a way impossible elsewhere, but perversely there is relatively little F1 deformation here to examine. There is the internal foliation of the Scourie-Dyke amphibolites, and there are one or two folds in the acid gneiss whose axial surfaces are more or less parallel to the general orientation of the dykes (and the orientation of their internal fabrics), otherwise there is nothing. The original cross-cutting relationships of the amphibolites remain very pronounced. It seems that either the F1 strain was very weak in this area (it will be shown in chapter 6 that tectonite fabrics can be developed in originally isotropic rock in responses to a very small amount of strain), or that most of the F1 strain in the acid gneisses was taken up in homogeneous layer shortening without the development of folds (the relative orientations of the amphibolites and the acid gneiss banding are such that flattening of the amphibolites normal to their length should have been accompanied by folding of the banding). The first of these explanations is the most attractive because original irregularities remain in the amphibolites. It seems that north-east Pabbay was an area of low deformation during each of the phases.

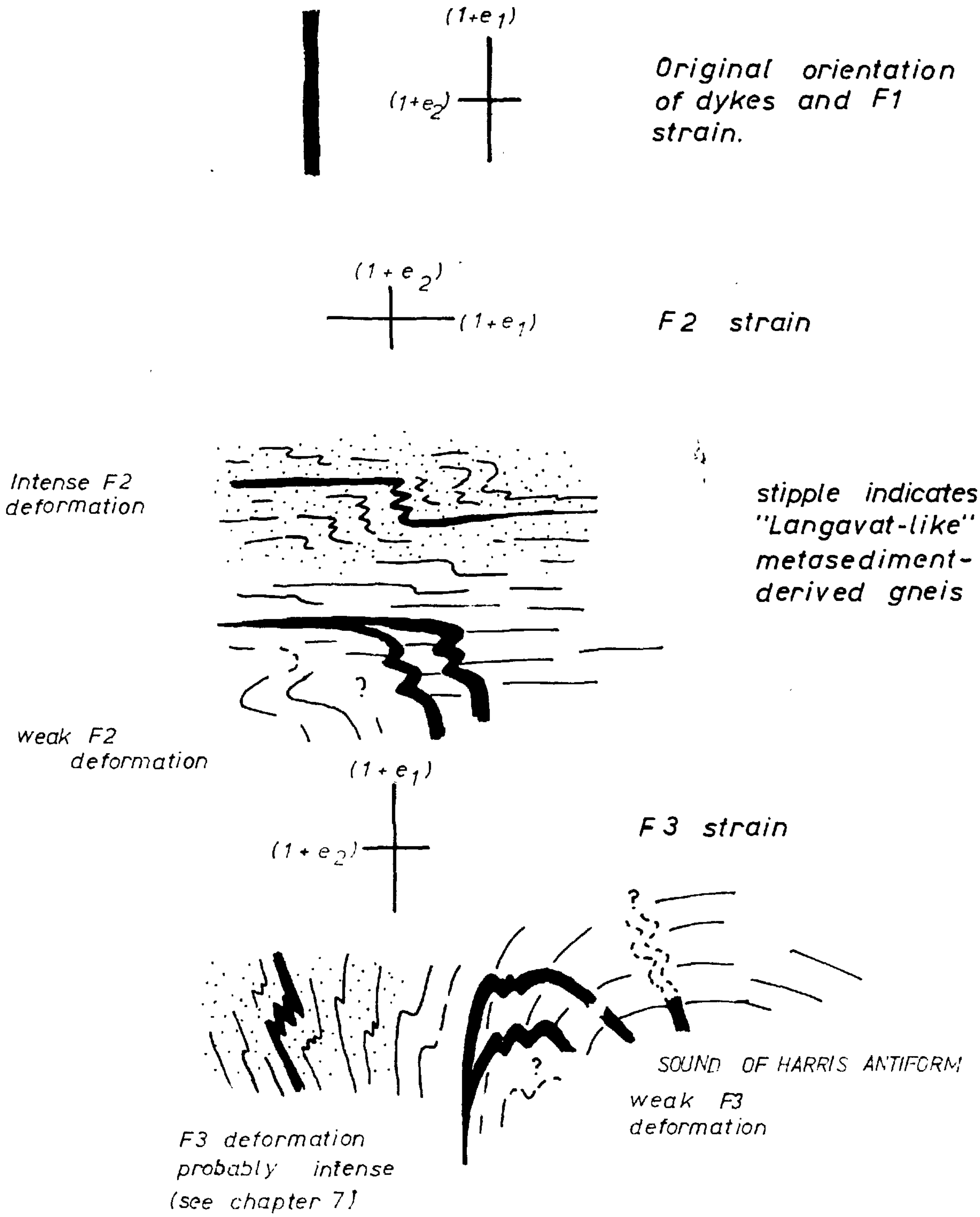
The fact that the acid gneiss banding is folded by F2 folds in the western part of Pabbay and also by the large F2 synform P3 in the south east part of the island, but not folded in the F2 fold zone FZ2 is a difficult problem. In the fold zone FZ2 the position of the banding parallel to the axial surfaces of F2 folds in the amphibolite bands suggests that it originally lay close to the XY plane of the F2 strain and was flattened, not folded by that deformation. Why then was its orientation different in western Pabbay and in the region of the synform at P3? The two possible answers are firstly that the re-orientation of the gneissic banding was considerable in these last two areas i.e. the F1 deformation was more intense there, or secondly that the F1 deformation was uniformly weak and the different orientations of the banding are Scourian phenomena. The first explanation is favoured since it lends further weight to the idea that north-east Pabbay was an area of relatively low deformation throughout the Laxfordian time interval.

Fig. 7 presents a diagramatic interpretation of the origin of the structures in Pabbay.

(iv) The reason for the strain variations.

It is probably not mere co-incidence that the astonishing drop in the amount of F2 (and probably also F1) deformation coincides with a lithological change in the acid gneisses

Fig 7 Possible evolution of the Pabbay "deformation-low"



from find grained, biotitic, probably originally meta-sedimentary rocks in western Pabbay to less biotitic, quartzo-felspathic "grey gneiss" in the east. We shall see later that the presence of biotite is one of the most important factors influencing the competence of rock, and it may be that the "Langavat-like", possibly metasedimentary rocks were relatively incompetent, and were therefore more strongly deformed.

(v) Pabbay deformation low, concluding remarks.

The identification of the Pabbay deformation low is an example of how purely qualitative structural geology can reveal variations in the finite strain of basement rocks. In Pabbay the change from very strongly deformed rocks to gneisses which have hardly suffered Laxfordian deformation occurs in a very short distance and coincides with an important lithological variation. It represents a virtual absence of F2 and probably also F1 deformation, and it occupies the hinge region of the F3 Sound of Harris Antiform. We might speculate on a possible connection between these facts. The Sound of Harris Antiform is an enormous fold, and there are no exposed competent units thick enough to have been responsible for it by a buckling mechanism - the Langavat-like rocks probably go over the fold, but are almost certainly incompetent as a unit.

The hinge of the Sound of Harris Antiform is situated in rocks containing almost undeformed Scourie-Dyke amphibolites at high angles to the banding of the acid gneiss, whereas the limbs of the fold contain well banded biotitic gneisses in which the Scourie-Dyke amphibolites are thoroughly concordant. The F1 and F2 deformations (especially the latter) produced the concordance between the amphibolites and the biotitic gneiss, and created a banded succession capable of folding anew when the F3 deformation began. In the area unaffected by F1 and F2 deformation, however the strong discordances which still existed between the amphibolites and the gneissic banding might have produced a rigid structural lattice in which the rocks were effectively locked and incapable of folding in any but the gentle manner seen in north-east Pabbay. Thus the F3 deformation folded a highly strained succession of one-time metasediments around a local rigid block.

Coward (1969) reached similar conclusions from a study of the "deformation lows" of South Uist and Benbecula.

CHAPTER 3

FOLDING AND RELATED TOPICS

3:1 INTRODUCTION

Folding is one of the most important processes by which strain is taken up in inhomogeneous rocks, indeed many structural analyses are entirely concerned with the geometry of folds and the age relationships of phases of folding. Some emphasis is given to these subjects in Chapter I and 2; here three other subjects are reviewed

a) the use of folds in estimating strain.

b) The relative competence of folded layers.

c) Axial surface fanning.

3:2 THE USE OF FOLDS AS STRAIN MARKERS

Ramsay (1967 Chapter 7) has erected a classification of folded layers based on the relative curvature of their bounding surfaces and the consequent variable form of the dip isogons (lines of equal dip, Elliot 1965) which may be drawn between these surfaces.

The classification is as follows.

Class 1
: curvature of inner arc is greater than curvature of outer arc
: convergent isogons

Class 2
: curvature of inner arc is the same as curvature of outer arc
: parallel isogons (similar folds)

curvature of inner arc is less than curvature of outer arc

Class 3

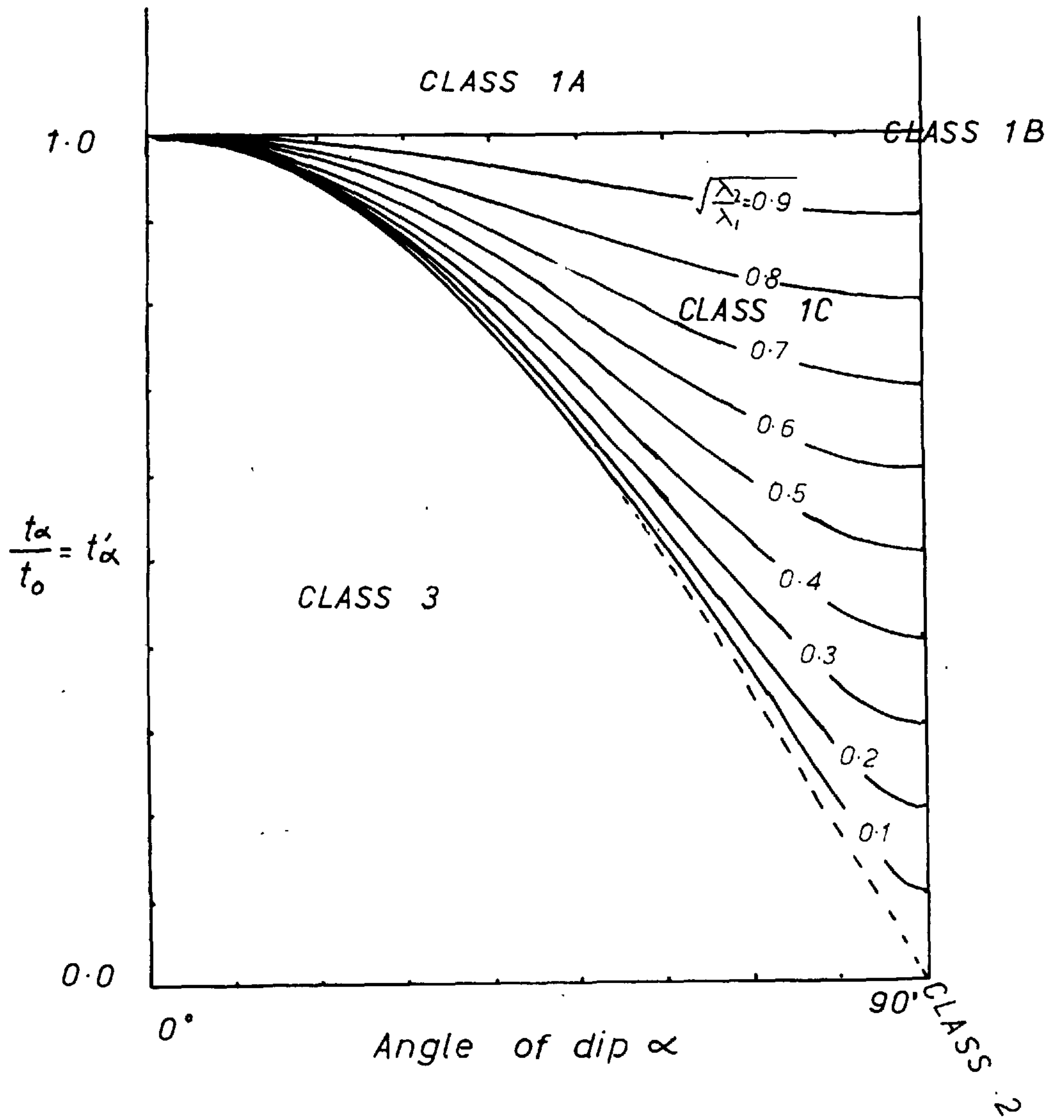
divergent isogons

Class 1 is further divided into: 1A strongly convergent isogons
1B parallel folds
1C weakly convergent isogons

All these classes and sub-classes occupy specific positions on Ramsay's graph (Fig. 1)(op.cit.pps.361,366, 370 and 413) of the change in orthogonal thickness around a fold $\frac{t_\alpha}{t_o}$ against α, where α is the variable angle made by successive tangents to each surface with the axial trace of the fold, t_α is the thickness of the layer for different values of α, and t_o is the thickness parallel to the axial trace). In the North Uist - Sound of Harris area folds in competent layers usually belong to classes 1B/ or 1C and in incompetent layers to classes 2 or 3.

Folding mechanisms have concerned structural geologists for many years. Several different kinds have been proposed, of which the buckling of competent layers in a multilayered succession is probably the most important. Buckling by flexural slip or flexural flow (Ramsay op. cit. p.392) produces parallel folds. The geometry of the other classes is most realistically explained by modification of the buckled layers either by intense tangiential longitudinal strain (Ramsay 1967 p.339) or by the addition

Fig 1 "Values of t'_α in flattened parallel folds with variation in $\frac{\lambda_2}{\lambda_1}$" (from Ramsay, 1967)



of a contemporaneous or post-buckle flattening (homogeneous strain) component.

The geometry of folded layers may be analysed by drawing dip isogons and measuring layer thickness variation on photographs of profile sections (i.e. sections normal to fold axes) of natural folds. The amount of homogeneous flattening in individual layers can be determined by plotting the thickness variation measurements on Ramsay's graph (Fig.1), and the flattening component removed by unflattening the folds on grids of the appropriate dimensions. This process gives a fold with parallel geometry, and direct measurements of the final and original lengths of the buckled layer $\frac{(l_o - l_1)}{l_o}$ give the amount of buckle-strain. The buckle and flattening components are then multiplied together to give an estimate of the shortening along the layer. Strain ratios obtained in this way, however, do not represent the total shortening, because they do not take into account any homogeneous shortening which may have occurred prior to buckle. The amount of this homogeneous shortening depends on the viscosity ratio between the buckled layer and its matrix: if there is no viscosity contrast, folding does not take place. Viscosity ratios can be determined from the wavelength: thickness ratios of parallel folds (Ramsay 1967 p.377, Sherwin and Chapple 1969) and once they are determined the pre-buckle shortening (and thence the total shortening) could be found. If, for example the combined values of shortening due to buckle and shortening due to late homogeneous strain (post-buckle flattening) equal 10:1, and the initial homogeneous layer shortening was 50% (4:1) the total shortening along the folded layer would equal 40:1. Estimates of total shortening have not been made in the present thesis, however. All strain determinations made here, including those done on folds, probably provide values significantly lower than the real value of the strain. This is especially true since the wavelength - thickness ratios calculated from some important lithological combinations in the North Uist-Sound of Harris area (see below) indicate viscosity ratios which are low (less than 10:1) and it is evident that significant shortening occurred prior to buckle and that high strains were necessary to initiate buckle. In Biot's theories of fold development (Ramsay, 1967 p.379) these lithological combinations should not have undergone folding at all.

The following measurements of wavelength: thickness ratios were taken from isolated competent layers (wavelength: thickness ratios are of debatable significance in multi-layered successions where harmonic folding occurs).

a) Scourie-Dyke amphibolite in homogeneous (smooth) acid gneiss, North Uist: 8 values range 5.5 - 3 average 4.1.

b) Felsic veins in Scourie-Dyke amphibolite: 12 values range 5-2 average 3.5.

c) Felsic veins in "Blue and Pink gneisses" - Sound of Harris: 8 values range 8 - 2 average 5.2.

These data are crude and insufficient but they are a reasonable guide. Their consistancy may indicate that the error in all the strain measurements is a fairly constant one.

Systematic studies of possible regional variations in the ammount of late flattening on minor folds have not been attempted because local inhomogeneities would almost certainly overprint any such variations. We can, however, make two general observations.

(i) Buckle components were recognised in all the competent layers of folds from the North Uist-Sound of Harris area.

(ii) The most nearly similar (most strongly flattened) competent layers occur in minor F3 folds in the Tigharry isocline and in minor F2 folds in the Sound of Harris steep zone (and Killegray), where associated incompetent layers often have truly similar, or class 3 geometry.

The minor F2 folds of the Sound of Harris often have a gneissic cleavage developed parallel to their axial surfaces, a relationship especially well seen in the fold zone FZ1 in north-west Pabbay. In the field there seemed to be in this area a relationship between the amplitude: wavelength ratio of these folds and the extent of gneissic cleavage developed in the incompetent layers. A direct relationship between these two features would exist only if amplification were entirely the result of homogeneous flattening. As this is not the case (since some buckle component is always present in the competent bands), it was hardly surprising that the field impression did not stand up to analysis in the laboratory.

For what they are worth the field observations are as follows:

1) Where the amplitude:wavelength ratio is low, fold hinges in competent bands are not thickened and an old foliation in the incompetent biotitic parts of the rock (the "blue" of the "pink and blue gneisses") is merely crinkled.

2) As the amplitude:wavelength ratio increases these crinkles tighten, and biotite flakes begin to be orientated parallel to the axial surfaces, i.e. a true 'strain slip' cleavage is produced.

3) The crinkles are no longer in evidence and a penetrative gneissic cleavage is developed in the incompetent bands. This appears between amplitude:wavelength ratios of 0.5 - 1.24:1 and between ratios of 2.5 - 5:1 for the maximum: minimum thicknesses of folded incompetent layers.

3:3 COMPETENCE

Isogon construction on folds provides a useful way of determining the relative competence of folded layers. The biotite content of rocks seems to have been the most important factor affecting competence; very biotitic rocks, whether acid or basic are always relatively incompetent. Coarse grained felsic veins and pegmatite bands, on the other hand, are always competent in any matrix. The "old basics" (see chapter 1) are invariably

Fig 2 Competent felsic bands in a Scourie dyke amphibolite., N.W.Pabbay

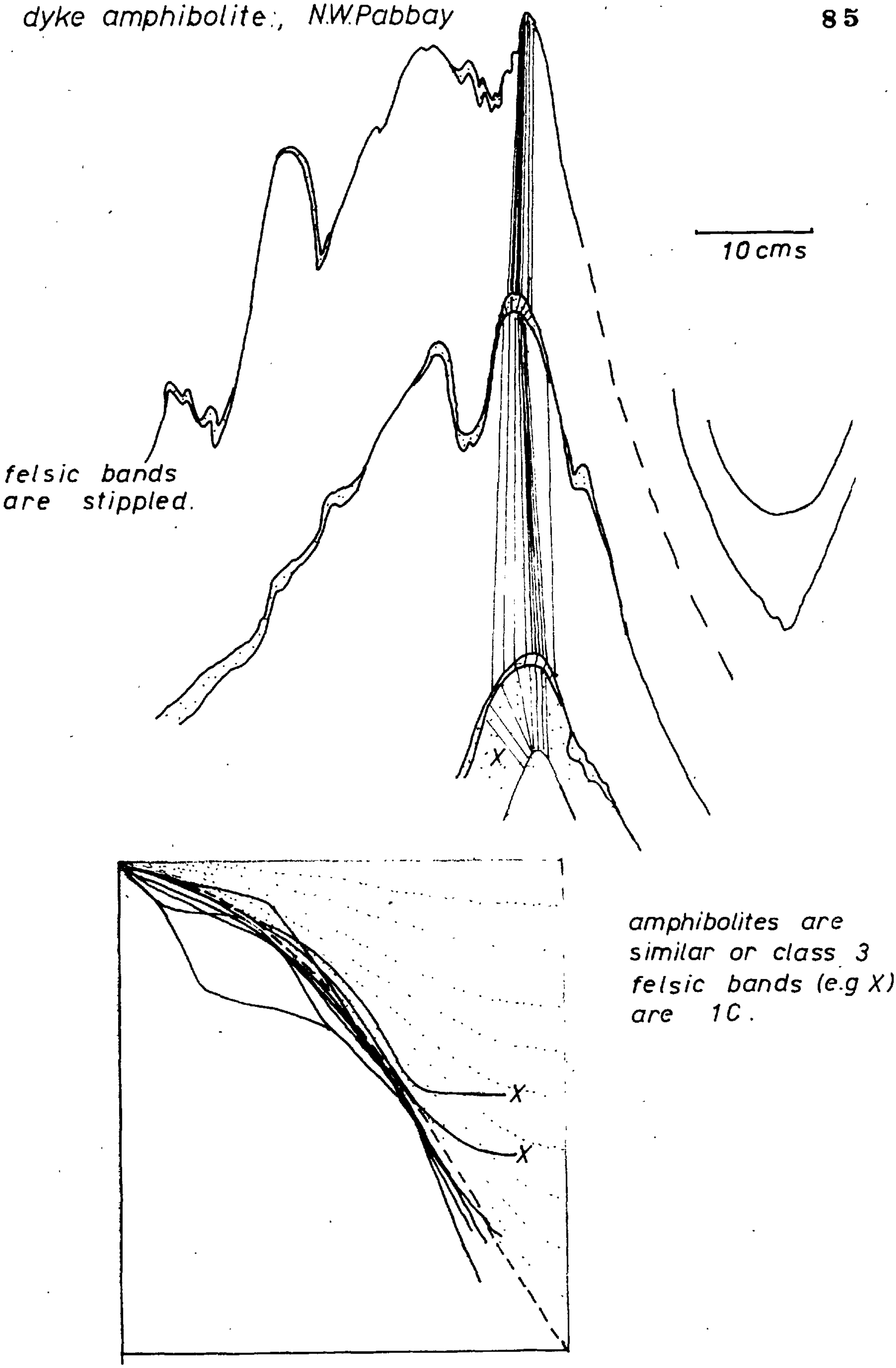
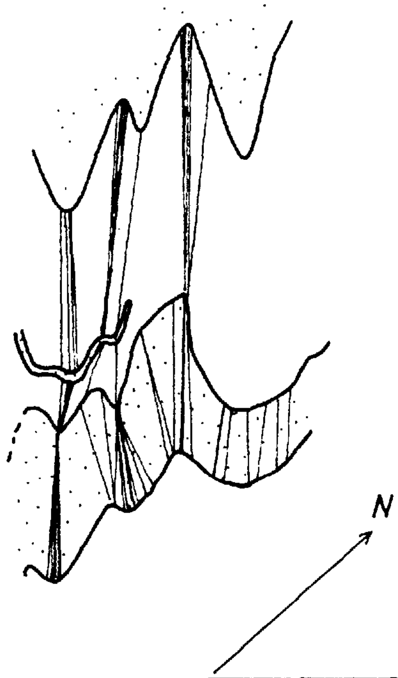


Fig 3

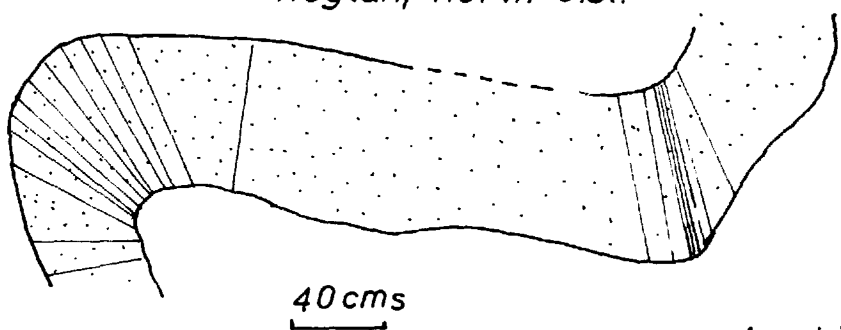


Part of the fold zone FZ1, N.W. Pabbay (Scourie dyke amphibolite is stippled)

Dip . isogons show the relative competence of the amphibolite.

50 mts

Fig 4 Third order F3 fold in amphibolite band, Hoglan, North Uist.



Amphibolite band is competent (1C)

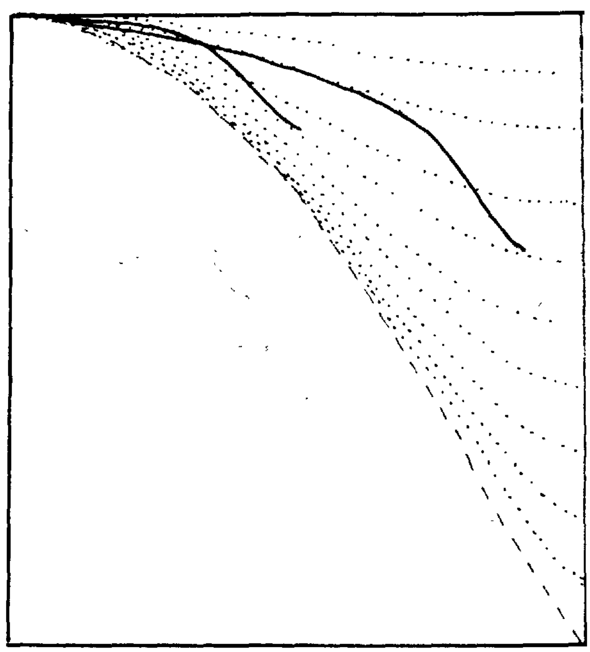
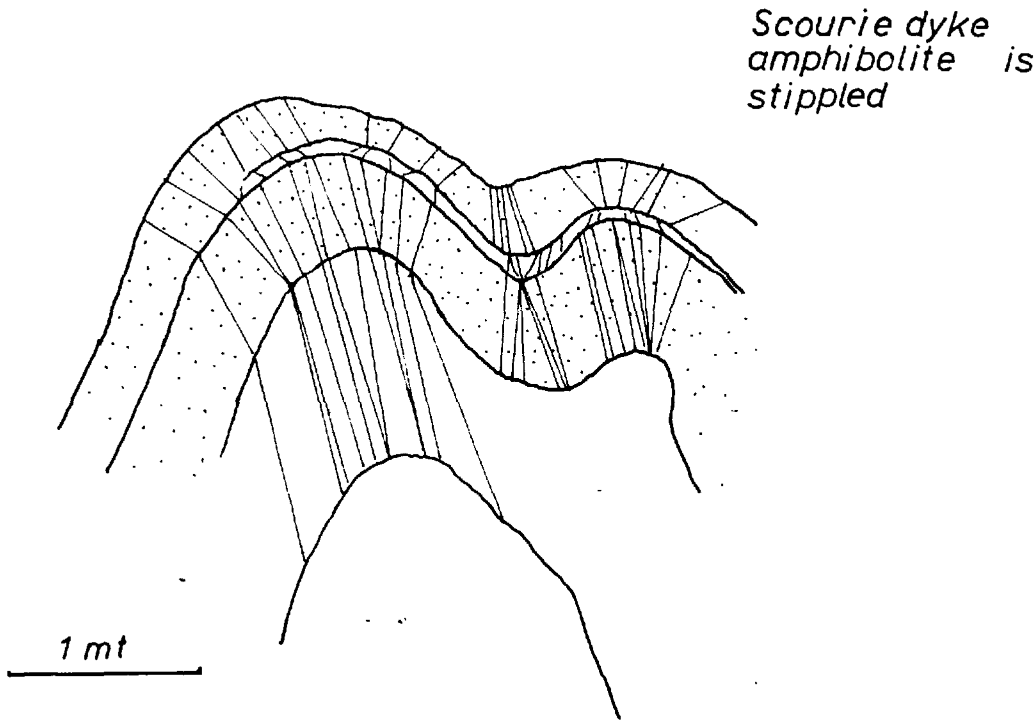


Fig 5 F2 fold from the Cliffs area of the North Uist coast



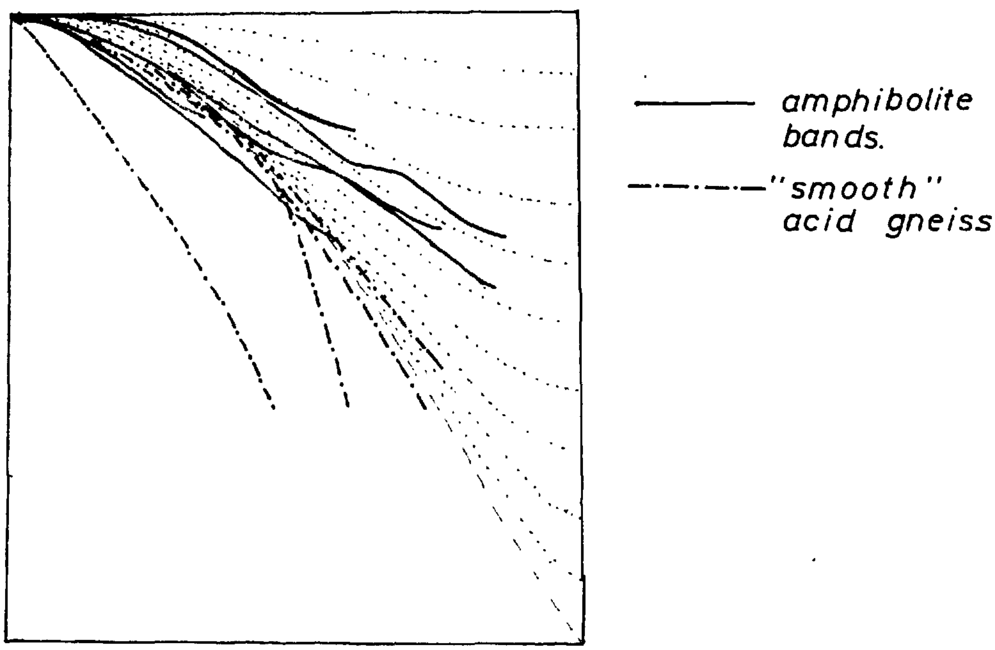
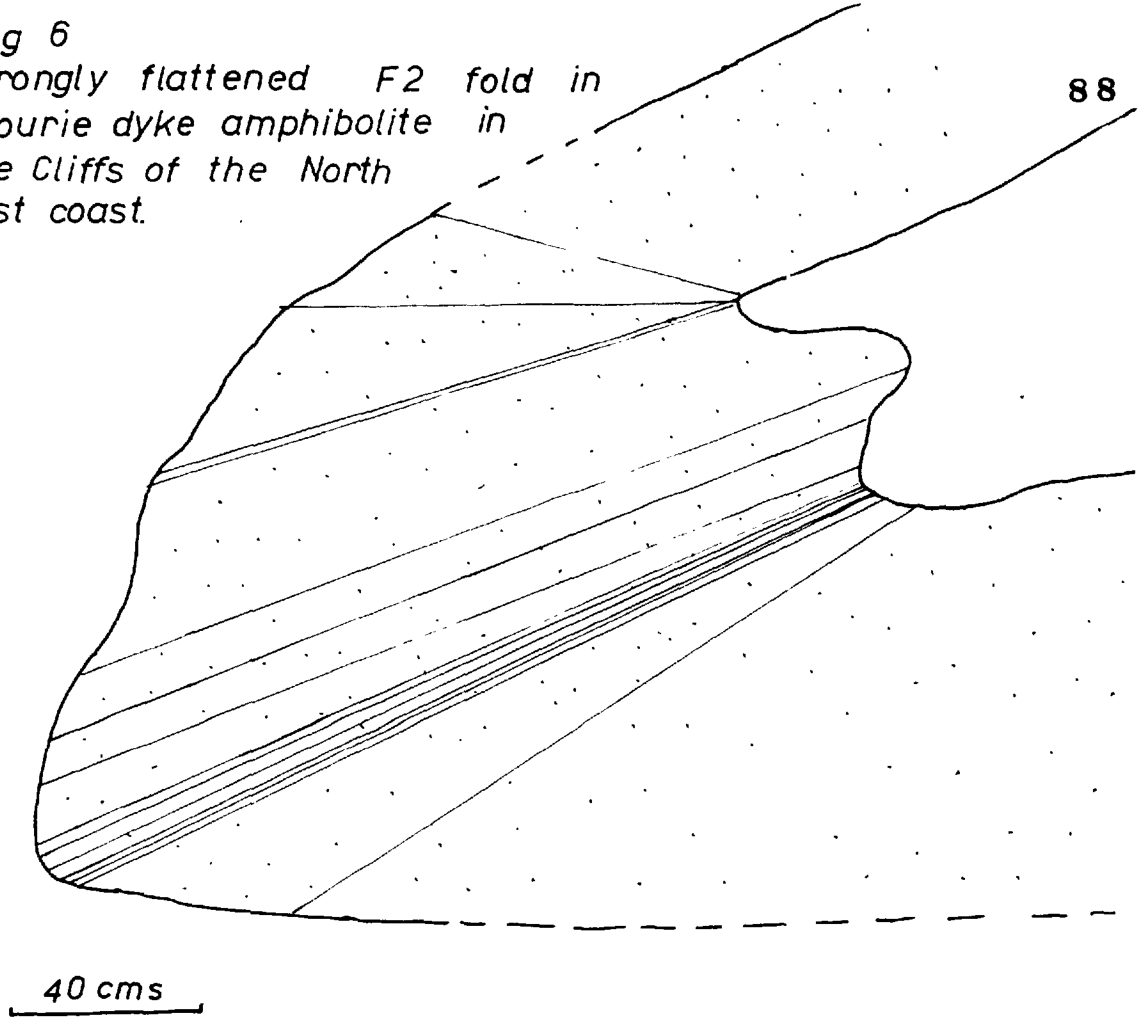
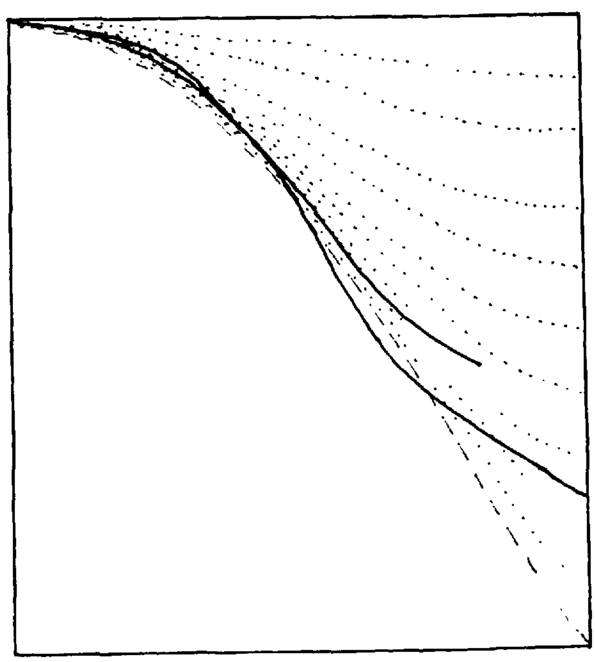


Fig 6
Strongly flattened F2 fold in Scourie dyke amphibolite in the Cliffs of the North Uist coast.





biotitic and almost always less competent than acid gneiss, whereas the "Scourie-Dyke" amphibolites are biotite-free and isogon construction on folds from the North Uist Coast reveals their relative competence within both rough and smooth gneiss. The Scourie-Dyke amphibolites behave as competent units in most of the folds in this area, and determine the fold wavelength throughout the coast. (Figs. 4, 5, and 6). The sensible system of fold ordering described in Chapter 1, especially the fairly uniform size of the third order F3 parasitic folds is due in part to a reasonable consistency in the thickness of the amphibolite units. In areas of smooth gneiss where there are no thin amphibolite units minor folds (lower than third order) do not exist; fourth and fifth order folds are restricted to areas where fine banding occurs within the acid gneiss.

Many of the amphibolite bands in F2 folds of the Sound of Harris area are strongly flattened and isogon convergence is exceedingly restricted, As on the North Uist coast, felsic veins within Scourie-Dyke amphibolites are relatively competent,(Fig.2) whereas on a larger scale the amphibolite bands are relatively competent within the acid gneiss. Fig.3 shows an amphibolite layer in the F2 fold zone FZ1 of N.W. Pabbay. The general relationships described above apply to folds of both F2 and F3 age - F1 folds are so rare that it is difficult to comment on them.

Examination of boudinage structures also provides a means of investigating the competence differences of different rock types. Boudinage is better developed in the Tigharry Isocline of the North Uist coast than anywhere else in the present area (except South Harris); also, the Tigharry Isocline shows well developed interbanding of various lithological units, so that a considerable number of observations can be made. The main points are as follows:

Pegmatite bands lying in or close to the foliation (the XY plane of the F3 deformation in the Tigharry Isocline) are frequently boudinaged, and the separation between individual boudins is often large, with a maximum indicating at least 50% extension: (note that this figure may not represent anything like the total extension suffered by the rock because any homogeneous elongation of the layers cannot be calculated - see chapter 4). Pegmatite bands are boudinaged in any matrix. Scourie-Dyke amphibolite bands within "smooth gneiss" (the dominant variety of acid gneiss in the Tigharry Isocline) are only infrequently boudinaged, there is usually no complete separation of individual boudins and the length:thickness ratio of the boudins is very large (up to 100:1). If the amphibolite is slightly biotitic, however, these relationships are dramatically reversed. Acid gneiss bands within biotitic amphibolite are boudinaged with moderate separation and a length:thickness ratio of 10:1. Pegmatite, acid gneiss and amphibolite may all suffer boudinage within a biotitic matrix.

The existence of boudinage depends on several interrelated variables, including the symmetry and intensity of the strain, the strain rate (i.e. the ductility of material at a given strain rate) and the competence difference (ductility contrast) between the boudinaged layer and its matrix. The Tigharry Isocline can be regarded as a domain of overall strain homogeneity (chapter 4, chapter 7) so that variations in the observations recorded above are probably direct functions of the different relative competence of different rock types. Amphibolite and "smooth gneiss" apparently have similar competence so that extension of an interbanded succession of these rocks usually took place homogeneously. Rare examples of boudinaged amphibolite suggest that amphibolite was marginally the more competent of the two rock types. Rocks containing a significant amount of biotite, whether acid or basic, are the least competent of all. In boudinage formation, as in fold development, biotite content is the most important factor governing rock competence, a fact which might also explain the different behaviour of quartzo-felspathic pegmatite and quarto-felspathic acid gneiss.

The significant contrast between the competence relationships deduced from boudinage and those deduced from folding is that the Scourie-Dyke amphibolites, though loath to boudinage within "smooth gneiss" form the competent units of folded successions of "smooth gneiss" and amphibolite. This contrast probably indicates that the minimum ductility contrasts necessary to induce buckle are less than those necessary to induce boudinage.

3:4 CUSP-STRUCTURE

If an interface between two media of different viscosities is deformed so that the principal compressive strain lies in the surface, the interface becomes folded into a series of broad rounded folds separated by much tighter, pinched-in folds where the broad folds face the less viscous medium (Fig.7). These structures have been produced experimentally (Ramsay 1967, p.383) and Ramsay (op.cit. pps. 383 -386) draws analogies between them and geological structures such as load casts and mullions, and on a much larger scale, certain basement - cover relationships in the Alps and in Africa where "pinched-in" synforms of cover sediments lie between broad antiforms of gneissic basement.

Many of the folds in the North Uist - Sound of Harris area have this alternating pinched-in and broad geometry. Most of the second order F3 antiforms of the North Uist Coast are considerably broader than their complementary synforms,(see maps 1-2 and Fig.8) and numerous third order structures also have this form (Fig.9). F2 folds in the Scourie-Dyke amphibolites of the Sound of Harris sometimes display the same feature (e.g. in N.W. Pabbay) and sometimes the synforms are enormously extended. On a larger scale the North Uist Coast synform and the Berneray sunform are both tighter and show considerably more evidence of F3 deformation than the broad antiformal domain which separates them. Indeed, this pattern is seen on the very largest scale throughout the southern Hebrides (Coward et al. 1970, and conclusions of this thesis). The implications of these large scale phenomena are discussed elsewhere, for the moment note two significant points:

a) The sense of the F3 cusp structures is consistent, regardless of the scale of the structures. The broad folds are always antiforms (i.e. they

Fig 7 Deformed interface between two materials of different viscosities

(from Ramsay, 1967 p. 383)

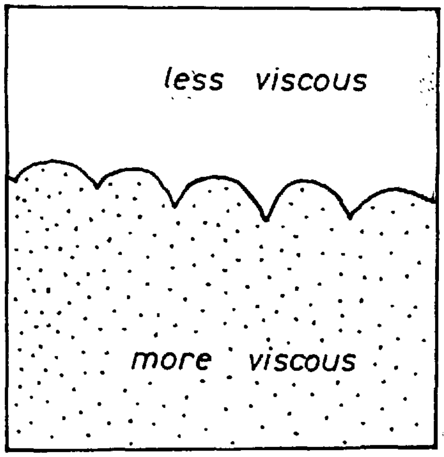


Fig 8 Deformed interface between "rough" gneiss and "smooth" gneiss (stippled) on the North Uist coast.

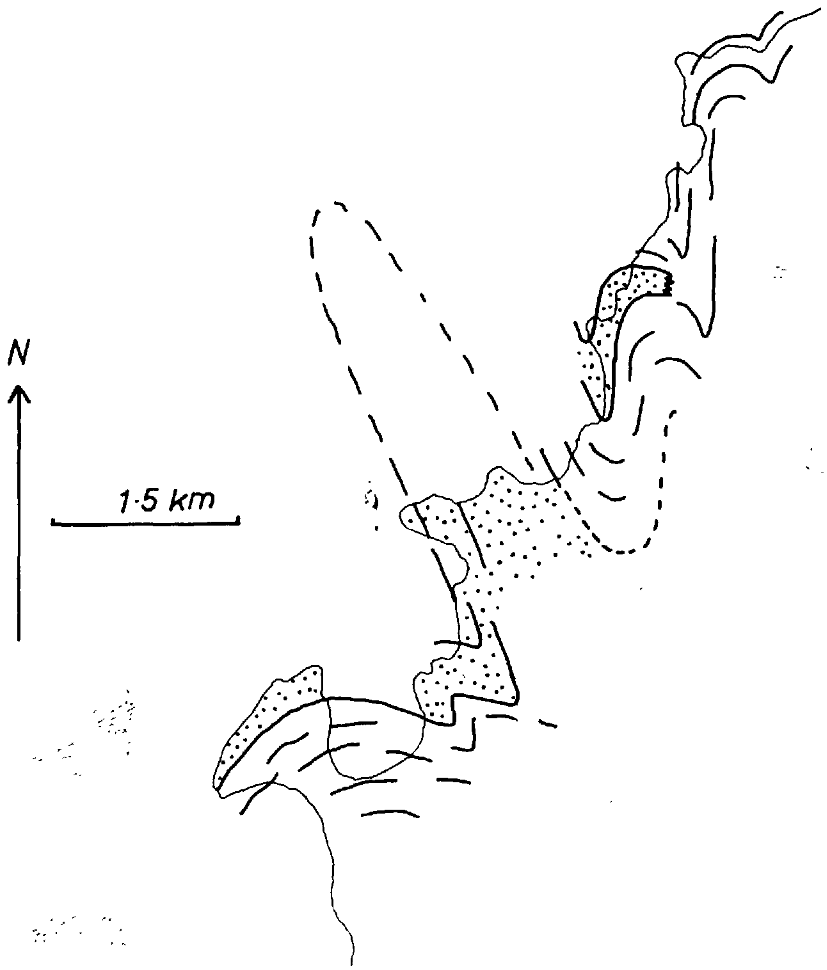


Fig 9

Cusp structure in third order F3 folds parasitic on the Scolpaig Synform.

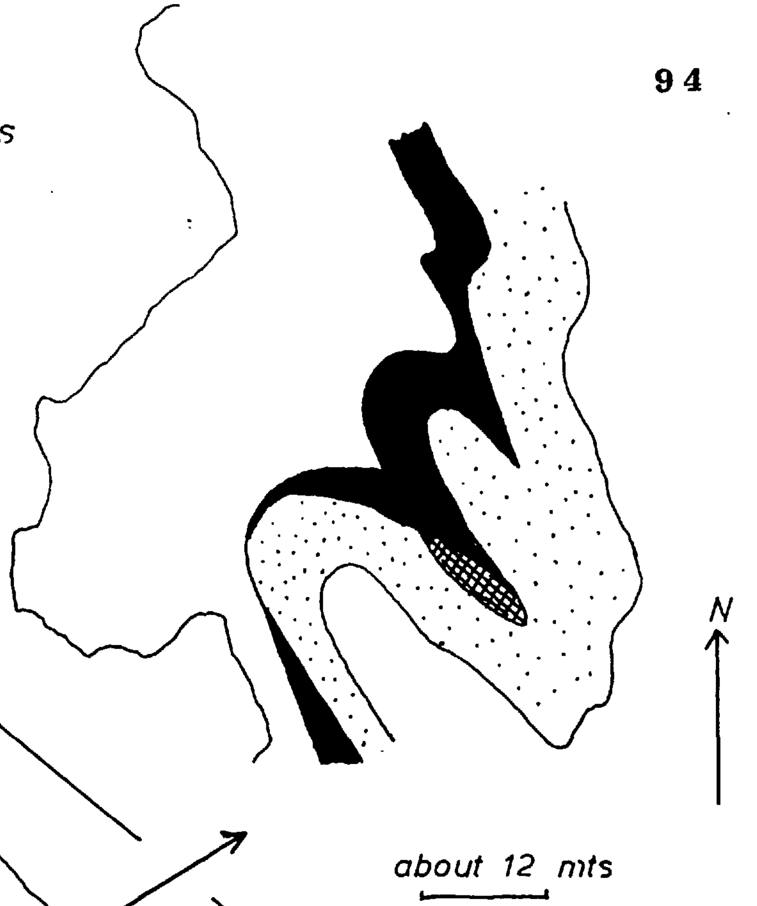
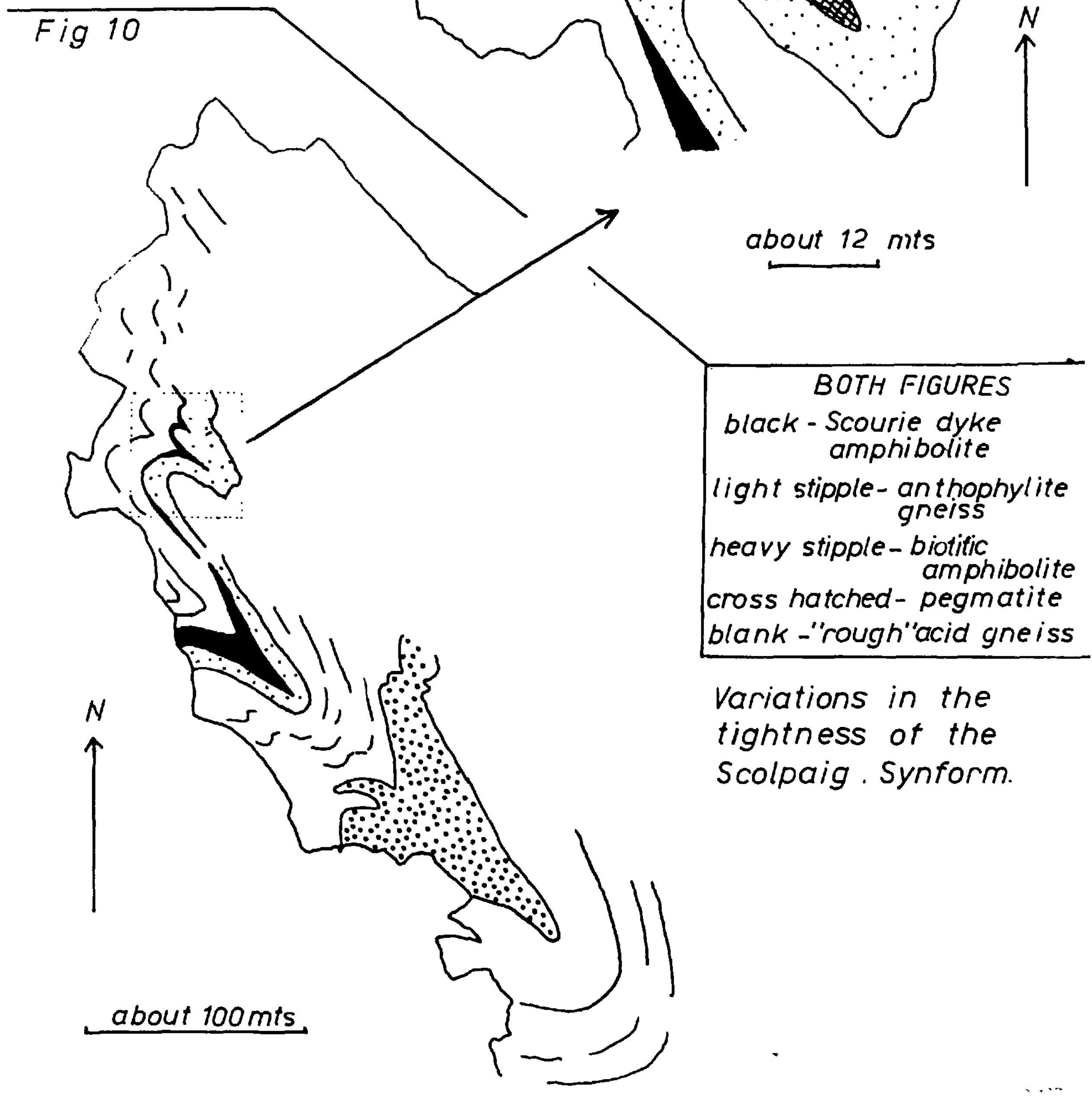


Fig 10



Variations in the tightness of the Scolpaig . Synform.

always face upwards), the pinched-in folds are always synforms.

b) Both surfaces of a single lithologic unit show the same relationships (e.g. the surfaces of amphibolite bands in F2 folds in the Sound of Harris, and 3rd order folds from the Scolpaig Synform Fig.9).

Both these points make an explanation of the structures of lower orders in terms of the deformation of an interface rather difficult to apply. There is no exposed visible lithologic boundary anywhere on the North Uist Coast which could account for the cuspate form of the second order F3 folds, except perhaps the interface between rough and smooth gneiss (Chapter 1 and 2). The distribution of the two rock types in relation to the structure, however, is neither simple nor consistent (Fig.8). The cuspate structure shown by the two surfaces of the same Scourie-Dyke amphibolite unit at Scolpaig (Fig.9) might be explained because the amphibolite lies along a contact between rough acid gneiss and anthophyllite-gneiss, two rock types which might possess different ductilities i.e. amphibolite is more competent than rough gneiss but anthophyllite-gneiss is more competent than amphibolite. Similar arguments might be invoked to explain the variation in tightness of the Scolpaig Synform itself, (Fig.10). We cannot, however, use any argument involving relative competence to explain the "pinched-in" F2 folds of the Sound of Harris, the general consistency of sense of the cuspate F3 folds on the North

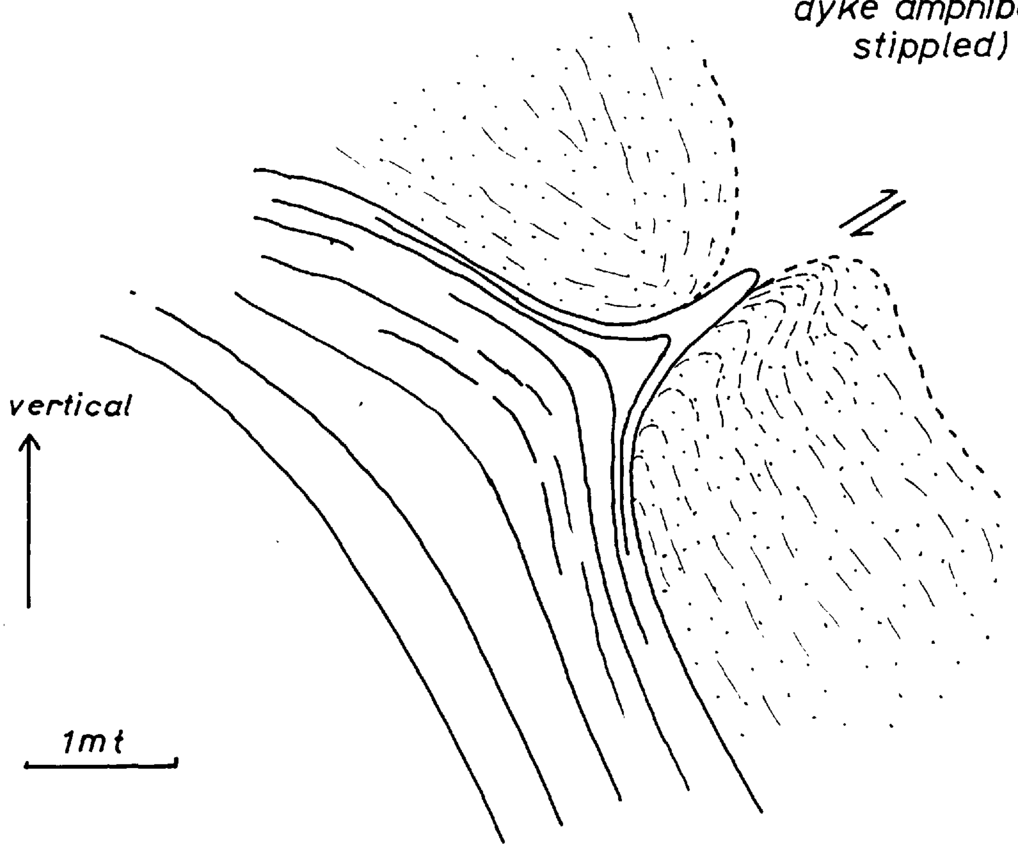
Uist Coast, or the form of the first order structures of the North Uist - Sound of Harris area. A general explanation of the phenomena of pinched-in synforms and broad antiforms may involve the relative upward translation of rock in the antiforms, and ductility contrasts may only modify this basic process. This concept of upward movement of material is compatible with another common fold style in the area - namely, the occurrence of antiformal "shoot throughs" and will be discussed at a later stage.

3:5 "SHOOT THROUGH" STRUCTURES

The occurrence of isoclinal antiforms in the middle of more open synforms has already been noted, the most conspicuous example being the Tigharry isocline. We have seen (Chapter 1) and it will be shown again (Chapter 7) that these isoclines represent local areas of very intense deformation. A possible explanation of the origin of "shoot through" structures might appear from a study of small scale representatives of the phenomena. Often we can see that some "shoot through" structures are abnormally well developed versions of "bend folds" (Ramberg 1955) into boudin necks (Fig.11).

In the North Uist-Sound of Harris area we often see that if a boudinaged layer is later compressed normal to its length, the original bend folds increase their amplitude and become highly flattened i.e. as the boudins are forced together, incompetent material effectively flows into and through the neck as a series of, or a single tight or isoclinal fold (Fig.11). The hinges of new folds initiated by the later compression

Fig 11 "Shoot - through" structure, Boreray (Scourie dyke amphibolite stippled)



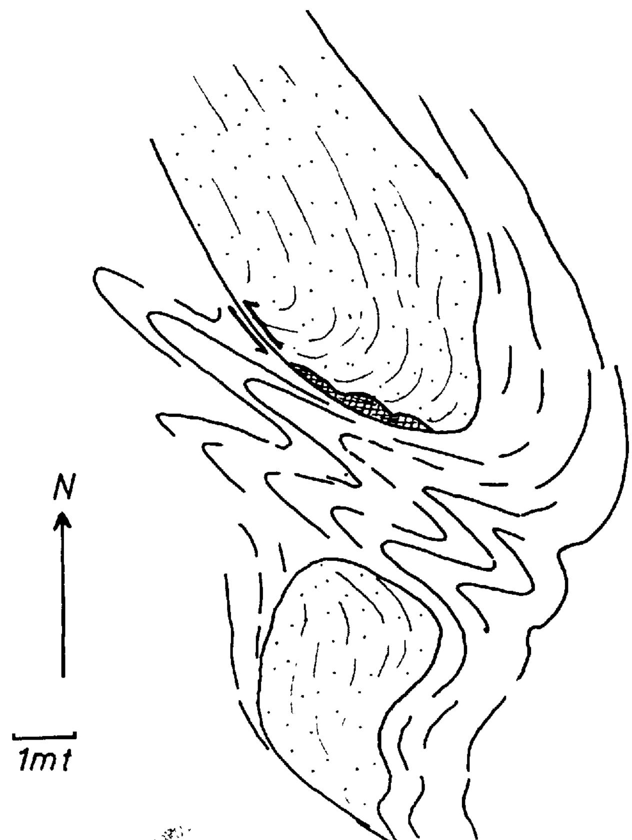
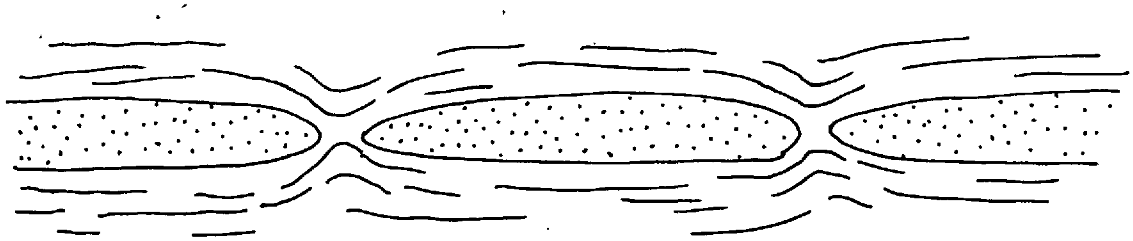


Fig 12

"Shoot through" structure in third order F3 synform, Scolpaig Scourie dyke amphibolite is stippled.

Both figures show development of "shoot through" structures in old boudin necks.

Fig 13 Possible mode of development of "shoot through" structures



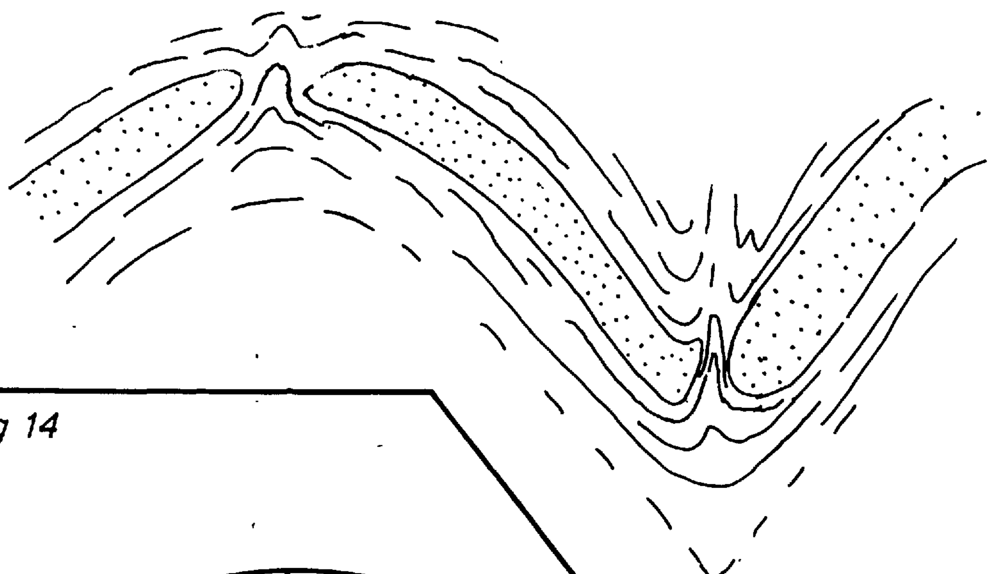
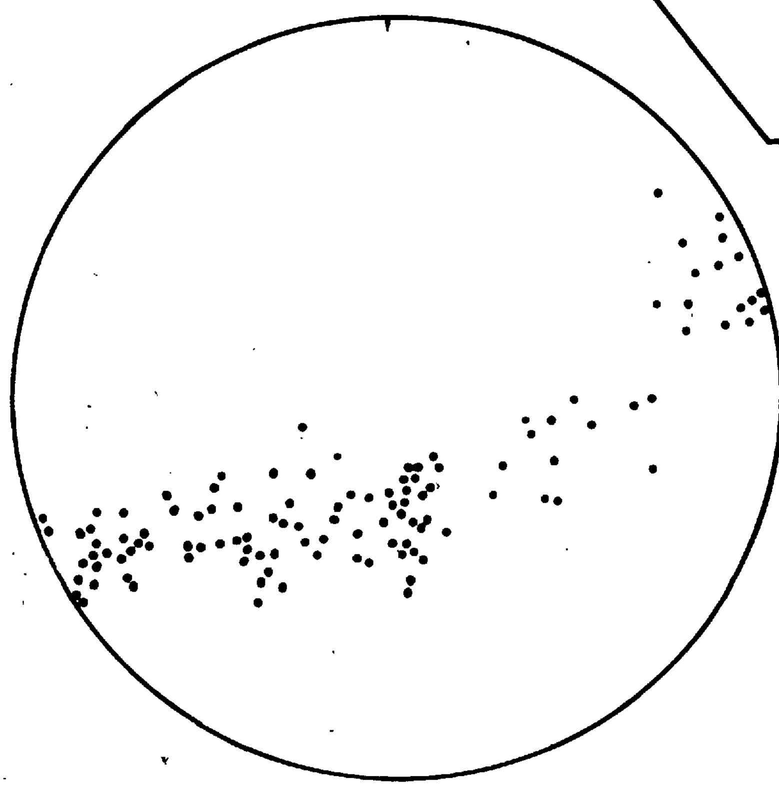


Fig 14



Fan of axial planes of minor (third and fourt order) F3 folds within the Scolpaig Synfor (omitting those shown in figs 15 &18)

tend to be located in the boudin necks; the weak spots in a competent layer. If the new fold is a synform, and therefore tighter and probably more highly strained than its complementary antiforms, the development of the central isocline progresses (Fig.12).

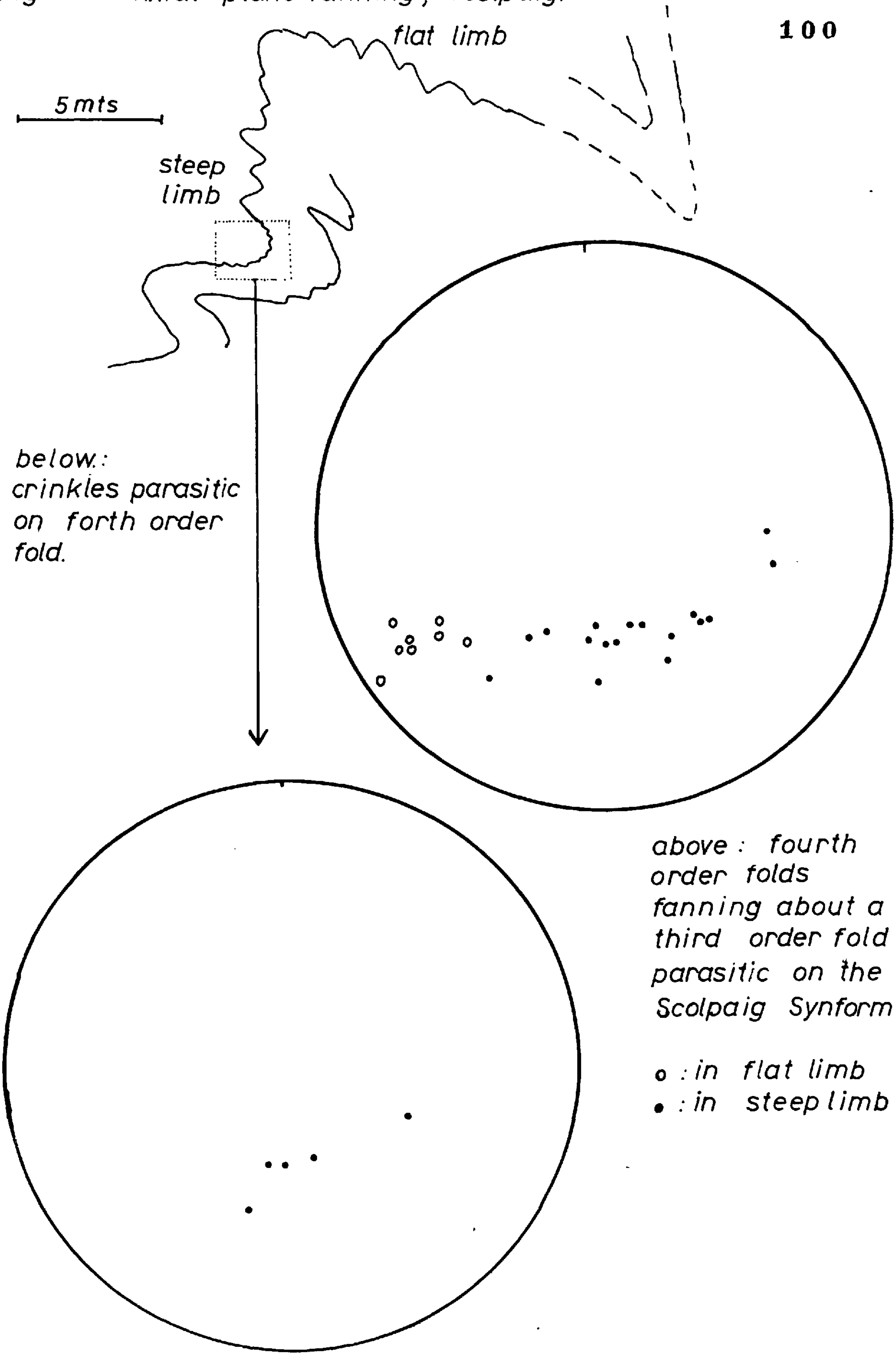
Sometimes central isoclines are visible in synforms when no previously boudinaged layer exists. Here it is probably correct to attribute the phenomenon to a flow of highly strained material in a direction of easiest release. In the North Uist - Sound of Harris area, almost all the modified boudin bend-folds and central isoclinal antiforms are results of F3 deformation and they always face the same way - upwards. This is, of course, in the same direction as the relative translation of the broad F3 antiforms, and thus the two sorts of structures which initially seem incompatible, are probably manifestations of the same phenomenon, (Fig.13).

3:6 AXIAL SURFACE FANNING

3:6:1 The Problem

A common feature of F3 folds on the North Uist Coast is the fanning of minor fold axial surfaces around higher order folds. Sometimes this fanning is extreme, so that almost complete girdles of axial plane poles occur on a stereographic projection of data from a single high order fold (Figs. 14 and 15). Usually, however the fan is not complete (Figs. 16, 17 and 18 and also Hoglan Synform, Chapter 1).

Fig 15 Axial plane fanning, Scolpaig.



minor folds around a third order F3 fold parasitic on the Hoglan Synform

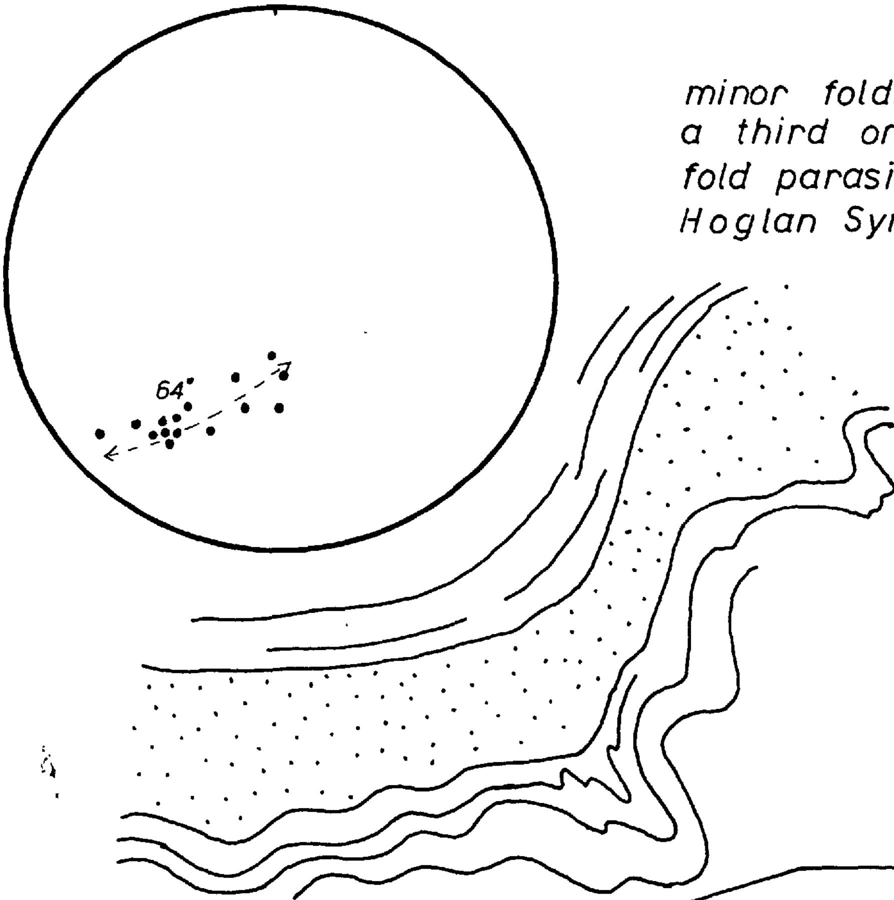


Fig 16

Partial axial plane fans

Fig 17

As fig 16 (see top right)

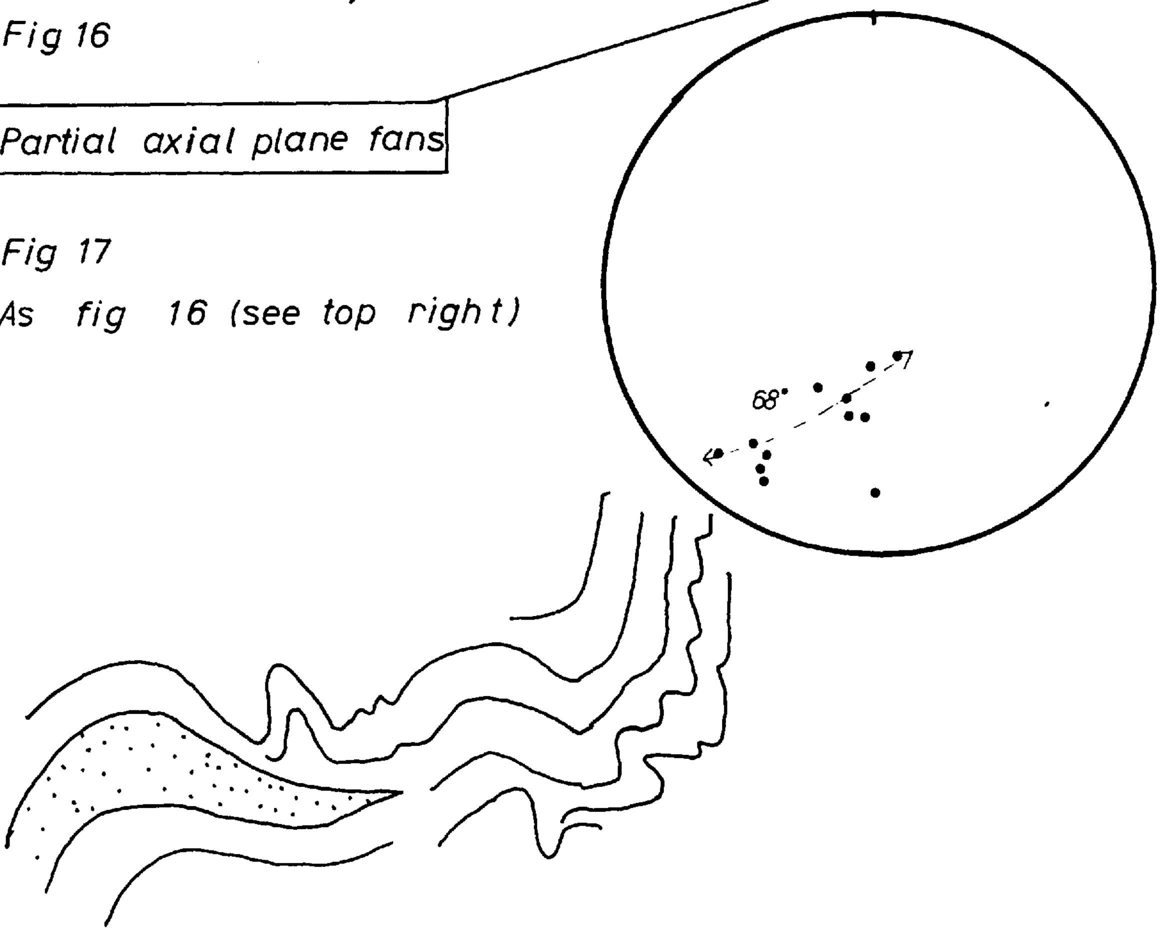
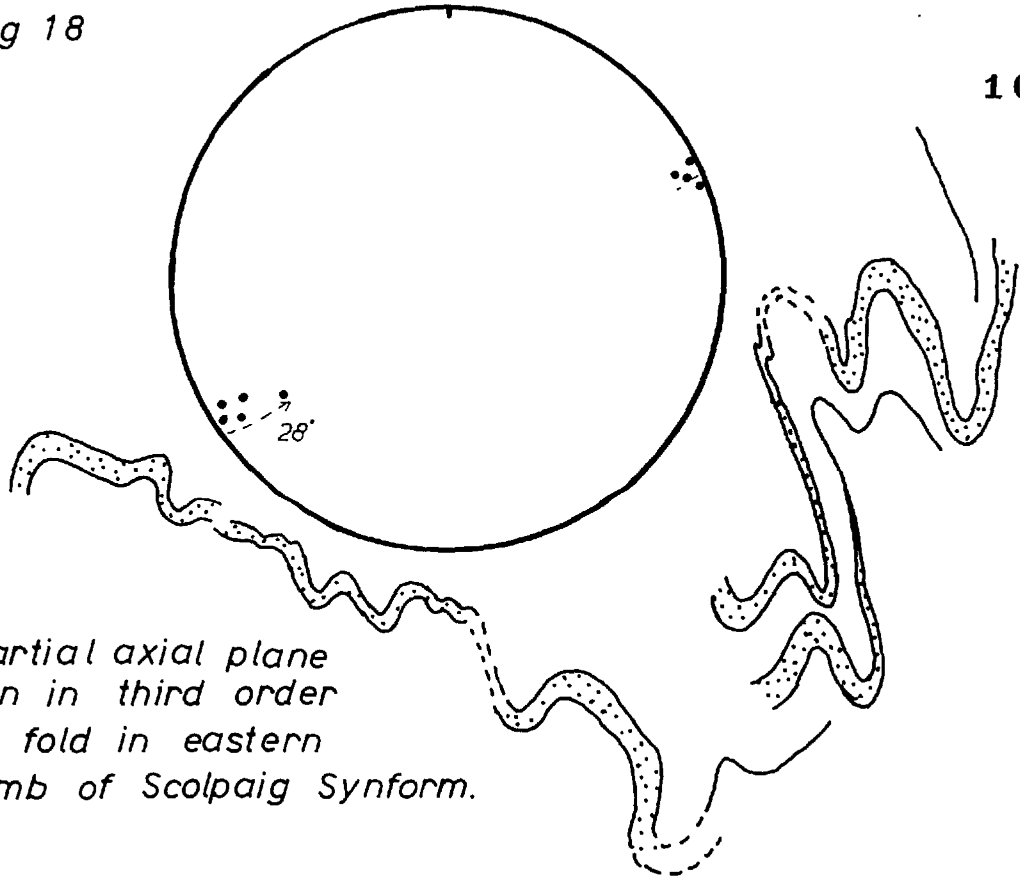
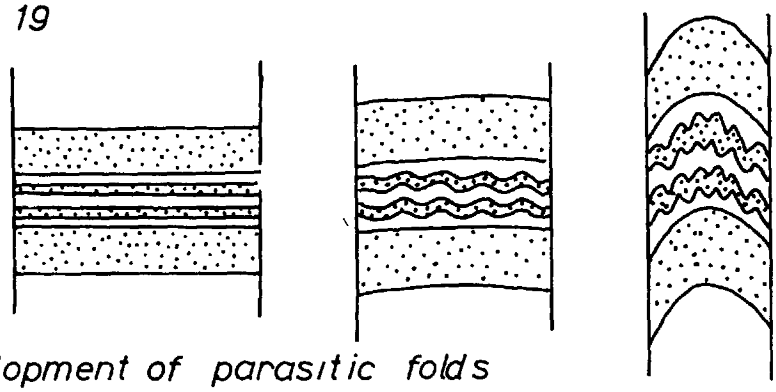


Fig 18



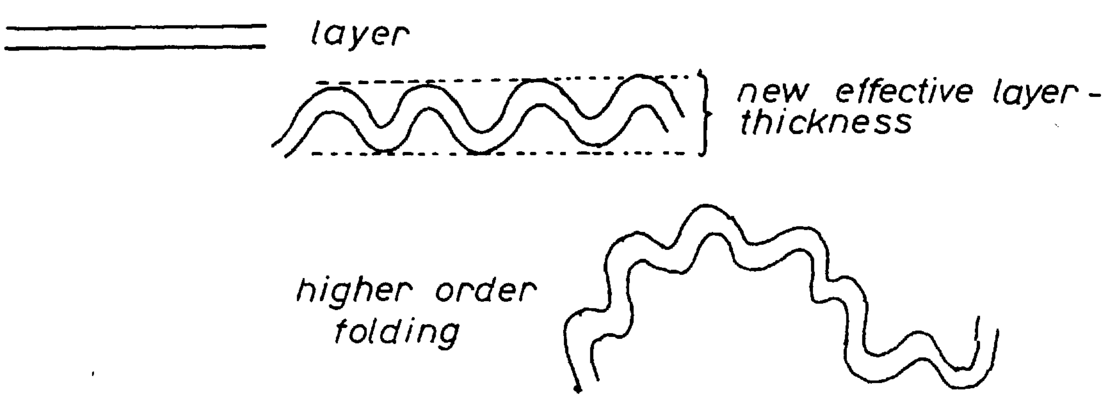
Partial axial plane fan in third order F3 fold in eastern limb of Scolpaig Synform.

Fig 19



Development of parasitic folds (after Ramberg, 1963b. From Ramsay 1967 p.397)

Fig 20



3:6:2 The Explanation

Fanning of this type is probably not analogous with contemporaneous cleavage fans where the cleavage directly records the finite strain trajectories through folded layers. The best explanation of the fanning of minor fold axial planes stems from Ramberg's (1963a) theory of the formation of parasitic ("drag") folds by the modification of early symmetrical buckles by the later development of a fold of greater wavelength in a thicker competent unit. (Fig.19). If a thicker unit is not present it may be that a layer completely buckled on a wavelength appropriate to its thickness may buckle anew on a wavelength determined by the amplitude of the original buckles (Fig.20).

If the major fold forms by flexural slip (Donath, 1962; Ramsay, 1967, Chapter 7) and the axial surfaces of the minor folds formed normal to the layering (and to the maximum compression which lay in the layering), the shear component in the limbs of the major fold reduces the maximum possible angle of the fan of the minor axial planes to some 66°. This maximum angle is obtained in major folds whose limbs are parallel to their axial trace (Fig.21). If on the other hand, the major fold forms by tangential longitudinal strain (Ramsay 1967, 397-400, 463 - 465), the absence of a shear component in its limbs enables complete (180°) fans of minor axial surfaces to develop in major folds whose limbs are parallel to their axial trace (90° from their original position) (Fig.22). Where refolding is by tangential longitudinal strain alone, there is no asymmetry of the minor structures; asymmetry in parasitic folds is produced by the shear component in the limbs of the major fold, and is therfore restricted to folds which formed with some component of flexural slip. Some measurement of the asymmetry (e.g. relative limb

Fig 21 Formation of parasitic folds by higher order folding by flexural slip.

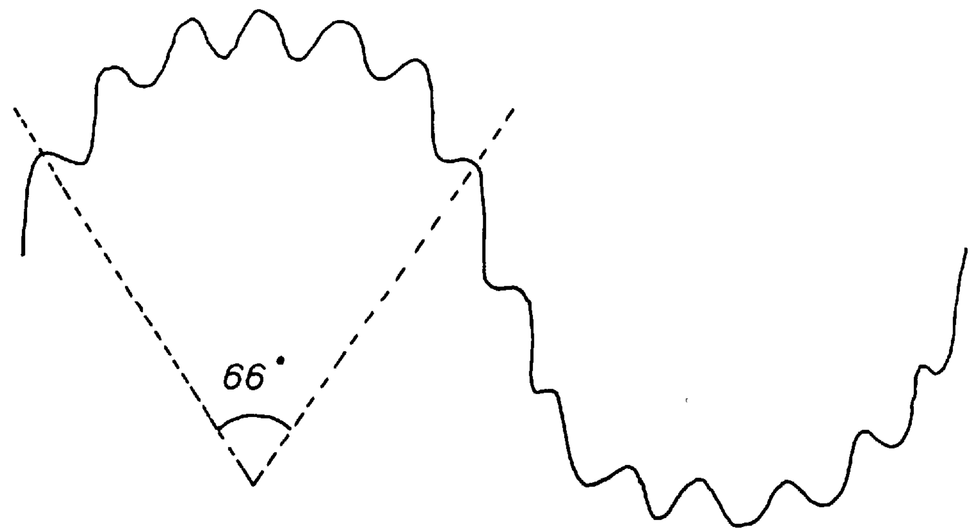
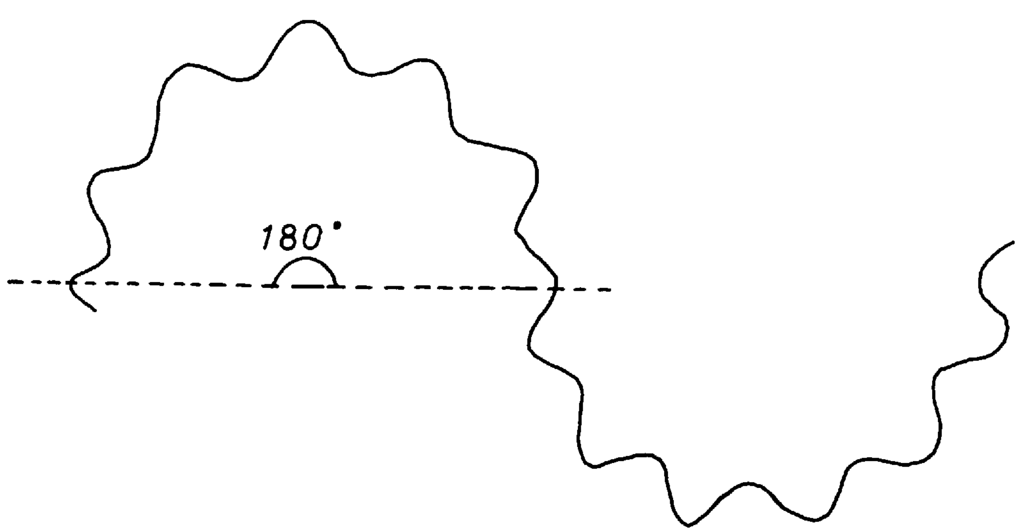


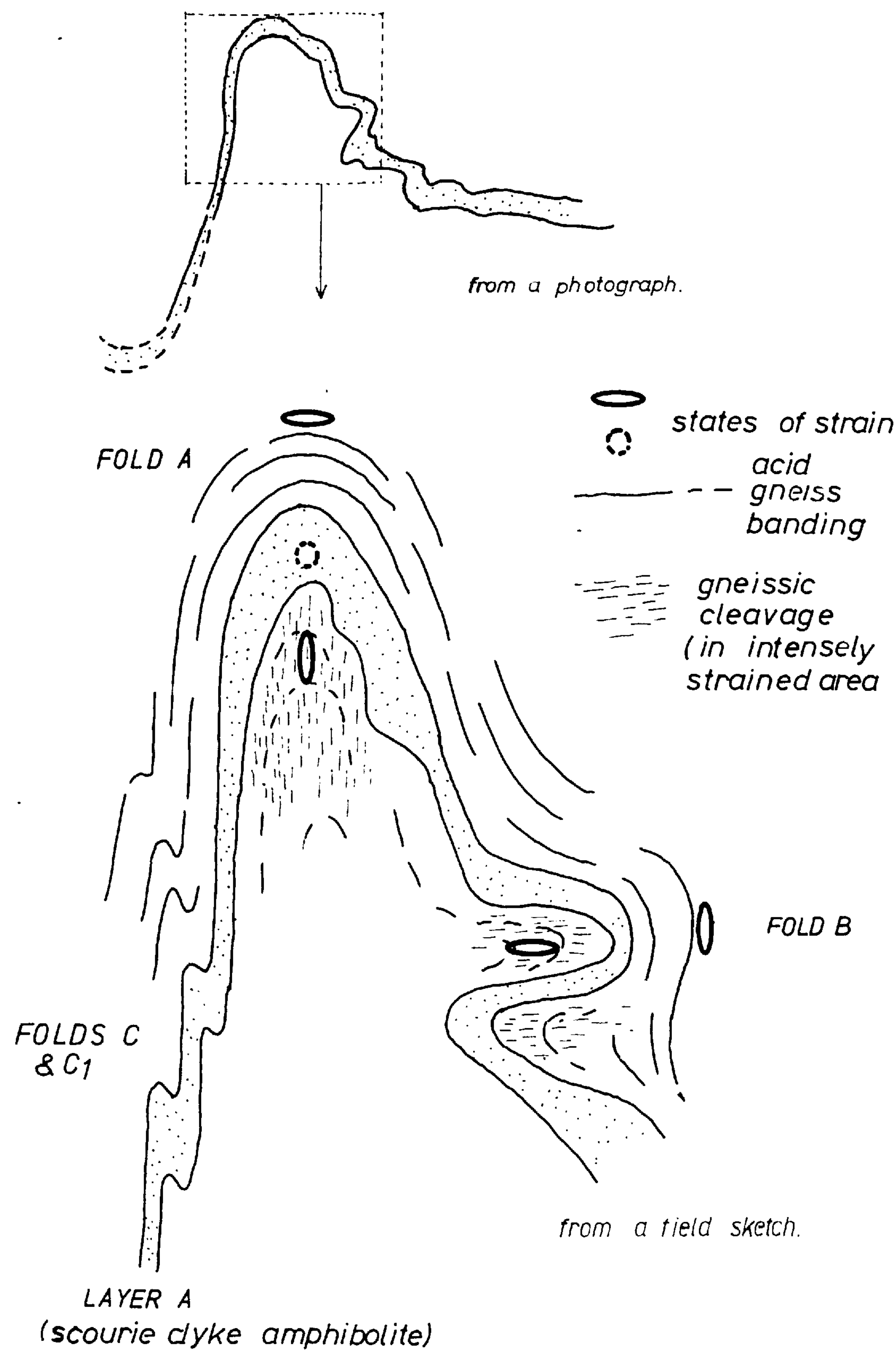
Fig 22 Formation of parasitic folds by higher order folding by tangential longitudinal strain.



lengths) of minor folds on the limbs of major folds might provide an indication of the mechanism by which the major fold formed.

Minor folds re-orientated by larger scale folding by tangential longitudinal strain might perhaps be expected to show some shape modification related to the states of strain within the major buckle, while no matter what the mechanism of formation of the major fold there ought to be a difference in tightness of the minor folds on the inner and outer arcs of the major structure. Minor folds lying beyond the outer arc of the major buckled unit are effectively unfolded, those lying in the inner arcs are probably tightened. (Figs. 21 and 22). Any homogeneous flattening during or after the development of the buckles will reduce the angular dimensions of the fan in the way demanded by homogeneous strain theory (e.g. Ramsay 1967, p.67). Where the major fold formed by tangential longitudinal strain and was later flattened we might expect the minor folds in the most extreme part of the fan to be re-folded. The absence of such re-folding in the field is probably explained by the inadequate lengths of the minor fold limbs; the thin bands have already developed buckles of appropriate wavelength and smaller scale folds cannot develop in them. A more homogeneous modification of minor folds, however, is commonly visible. Fig.23 shows an extreme case of re-folding by tangential longitudinal strain and the effects of later flattening can be seen in layer A (a Scourie-Dyke amphibolite band). Layer A is thickened in the hinge of fold A but thinned in the hinge of fold B, so that in fold B it has Class 3

Fig 23 Buckled layer, Scolpaig. See text for explanation.



geometry. The pattern of the planar fabrics (pecked lines) on fig. 23 itself suggests that both fold A and fold B formed by tangential longitudinal strain. It appears that fold B was re-folded about fold A even though the two are developed in the same amphibolite band (Layer A). The very slight fan of the axial planes or folds C and C_1 indicates that either:

1) Rotation of the major limb and the minor axial planes by homogeneous strain was very important,

or 2) considerable shear took place on this limb of the major fold, and that the development of the major structure was complex, with the steep limb containing a much greater shear component.

3:6:3 Conclusions

The extent of minor fold axial plane fanning is a useful guide to the mechanism by which folding occurred, or to be more precise, it gives some indication of the relative importance of flexural slip and tangential longitudinal strain in the formation of a particular fold, (most natural folds probably form by a combination of these mechanisms). The relationship is complicated by the effect of any homogeneous deformation, so that this rather rough guide is difficult to treat quantitatively. The most important general conclusion is that if a minor axial plane fan exceeds 66^{o}, tangential longitudinal strain was obviously important in the formation of the major fold.

In the North Uist - Sound of Harris area the greatest development of tangential longitudinal strain is in the domain of the Scolpaig Synform, a second order F3 fold on the North Uist Coast. Figs. 14, 15, 18 and 23 show axial plane fanning from this region. Fig.14 shows the total fanning of minor folds within the major structure, Fig.15 the fanning of fourth order folds about a third order monocline, and of fifth order crinkles about a fourth order fold, os that a complete sequence is demonstrated. It has already been noted (Chapter 1) that the axial plane fanning of minor folds in the hinge zone of the Scolpaig Synform is extreme. It decreases in the limbs of the structure and it is probable that even the Scolpaig Synform formed by a combination of flexural slip and tangential longitudinal strain. The girdle of axial plane poles is probably made complete by readings from the core of the structure where shearing was negligible. The asymmetry of third order structures in the limbs of the major fold probably indicates some component of flexural slip.

3:6:4 <u>Alternative explanation</u>

An important alternative explanation for axial plane fanning is that it may be a function of constrictional deformation. In the North Uist area the orientation of linear elements is always constant, and where complete fanning occurs (e.g. the core of the Scolpaig Synform and "South Scolpaig" nearby) there is an effective axial symmetry. The dominance of linear fabrics in the core of the fold might also be taken as evidence of constrictional strain, and in Chapter 5 it is suggested that the finite strain ellipsoid in the hinge of the

Scolpaig Synform is made constrictional by the superposition of oblate finite strain ellipsoids.

The two explanations are not entirely incompatible. A multilayered sequence subject to constriction is perhaps most likely to buckle by tangential longitudinal strain so that if the longest axis of the ellipsoid lies parallel to the fold axes (i.e. is added in an original surface) the finite neutral surfaces of the buckled layers may represent a series of cross-sections through prolate ellipsoids. The main objections to compatibility between the explanations are:-

1) The necessity of the X direction being added in the original plane, or of the complete rotation of all elements towards that direction.

2) At Scolpaig, as elsewhere in North Uist, the prolate symmetry of the fabric is due to superposition of F3 strain on an earlier fabric, and not to a simple phase of constrictional deformation.

It is, therefore, better merely to record that in the Scolpaig Synform there is an association between complete axial surface fanning, and finite constriction in the fold hinge, and between incomplete fanning, minor fold asymmetry and lineated planar fabrics in the limbs.

CHAPTER 4

THE DETERMINATION OF STRAIN IN BASEMENTS

4:1 INTRODUCTION

Major tasks of the structural geologist are

1) To assess the strain state of naturally deformed rocks.
2) To study any variation in this strain state and analyse its relationship to the visible structures.

The first aim is usually achieved by the measurement of deformed objects such as fossils and ooids whose original shape is known. Lewisian gneisses contain no such objects and other methods must be sought.

There are several possibilities.

1) Measurement of the extension along boudinaged layers, (e.g. Ramsay 1967, p.250). This method is usually unsatisfactory because any homogeneous elongation cannot be recognised. The measurements are only two-dimensional.

2) Measurement of the shortening along concentric, ptygmatically folded layers. Like method (1) this gives strain ratios in two dimensions only. Its use is limited by the relative rarity of good material.

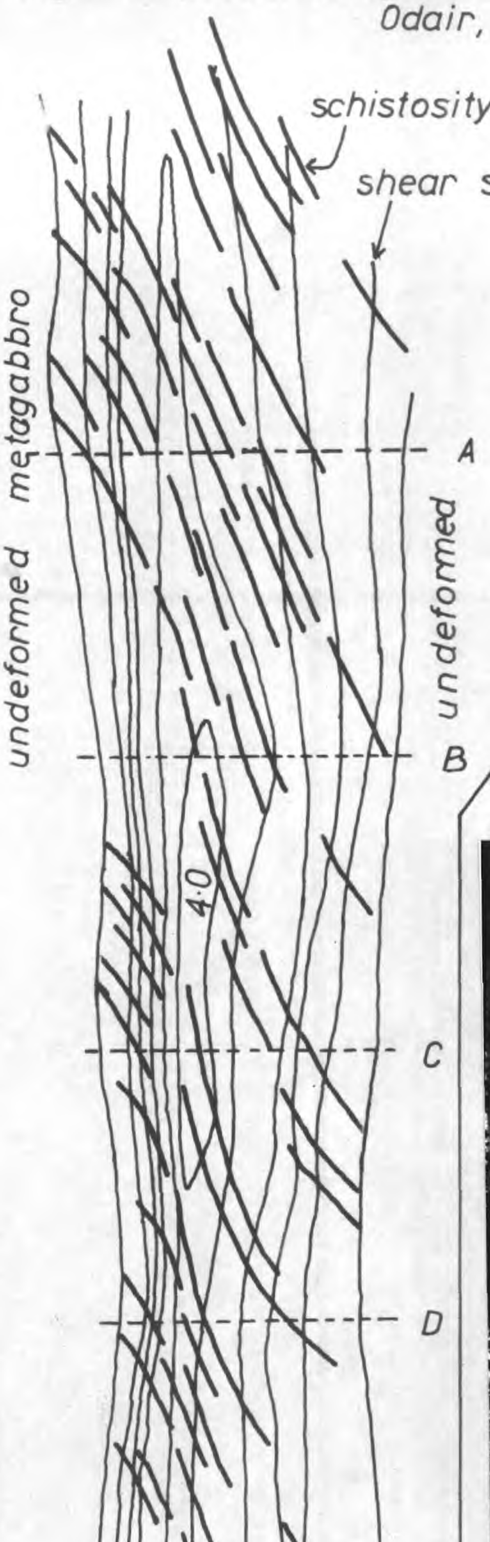
3) By a combination of (2) and geometric techniques on folds including isogon construction and studies of layer-thickness variation, (Ramsay 1967, chapter 7).

Once again this only gives the two-dimensional shortening, but it is otherwise exceedingly useful. It is discussed elsewhere (Chapter 3).

4) A common feature of basement deformation is the localization of intense strain in discrete planar shear zones (plate 1, Fig.1) in which a new penetrative foliation is developed, (e.g. Clough in Peach et al, 1907, Sutton and Watson, 1951, 1961, Teall, 1885). Where these shear zones cut otherwise undeformed rock, and volume change can be ruled out, it can be established that deformation in the zones proceeded by simple shear alone (Ramsay and Graham, in press). Thus the shear strain (γ) at various points within the shear zone can be determined by measuring the angle (θ') between the shear zone foliation (the XY plane of the strain ellipsoid within the zone) and the shear direction ($\gamma = \tan 90 - \theta'$). From this the strain ratios in both two and (since simple shear produces a plane strain ellipsoid) three dimensions can be determined (Ramsay 1967 p.85). This method provides a means of defining quite accurately the strain state throughout a shear zone. Fig. 1 shows a graphical analysis of a shear zone in a basic body on the headland of Castell Odair on the North Uist coast. More detailed discussions of shear zones appear in Ramsay and Graham (op cit) and elsewhere in this thesis.

The four techniques outlined above may be supplemented by the following qualitative and subjective methods of

Fig 1 Graphical analysis of a shear zone from Castell Odair, North Uist (from Ramsay and Graham, in press)



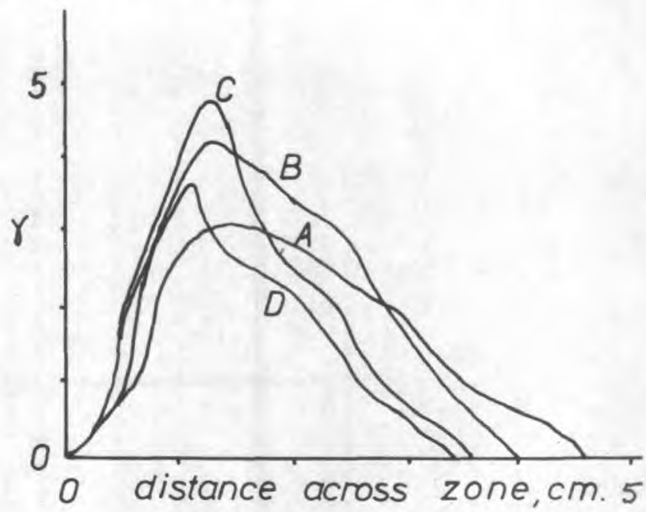
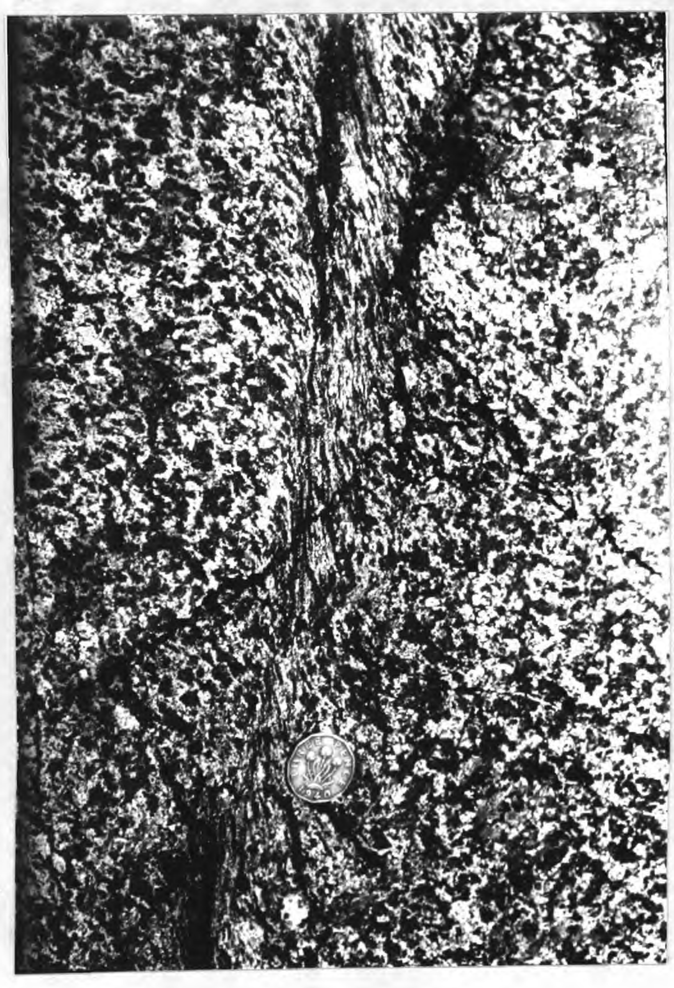


Plate 1 The same shear zone (photograph by J.G.Ramsay)



strain assessment.

1) Given certain conditions (namely that layering is present to provide the necessary ductility contrasts, and that this layering has an appropriate orientation) the presence or absence of folds of a particular generation can sometimes be regarded as an indication of the intensity of deformation.

2) The preservation or obliteration of original cross-cutting structures is sometimes a function of deformation intensity. A qualification here is that deformation can sometimes intensify original cross-cutting relationships (Ramsay 1967, Chapter 9).

An example of the usefulness of these qualitative observations is given in the account of Pabbay in Chapter 2, where mention is also made of the increasing "regularity" of gneisses as they are increasingly deformed. The visual structural geometry of "undeformed" Lewisian (i.e. Scourian acid gneiss cut by undeformed basic dykes) is extremely complex, whereas in highly deformed gneisses it appears much simpler, (see frontispiece).

All the quantitative techniques mentioned above (i.e. 1, 2, and 3) involve making measurements on structures which are essentially the small scale manifestations of inhomogeneous strain. There is, however, a quantitative method of strain assessment which utilizes homogeneous strain theory and the rest of this chapter explores its potential.

4:2 STRAIN DETERMINATION FROM THE ORIENTATION OF DEFORMED VEINS.

During an irrotational progressive deformation, planes rotate towards the XY plane of the strain ellipsoid. Those lying close to XY are elongated, those lying close to YZ are shortened and the bounding surface between shortened and elongated elements is the "surface of no finite longitudinal strain" (Ramsay 1967, Chapter 4) of the finite strain ellipsoid.

Flinn (1962, p.424) first noted the possibility of using the orientation of elongated or shortened veins in a "deformed granite or migmatite" to determine the strain state of the host. Talbot (1967) was the first to develop the possibility into a practical method of strain determination.

The method has the advantage of being three dimensional i.e. it defines the strain ratios in three dimensions and therefore the symmetry of the strain ellipsoid. Further, it is the only method giving the orientations of the principal strains, and for this reason alone it is, if valid, a highly significant tool in the structural interpretation of crystalline rocks. Significant disadvantages, however, are

1) It assumes no volume change during deformation.
2) Deformation is assumed to be irrotational.

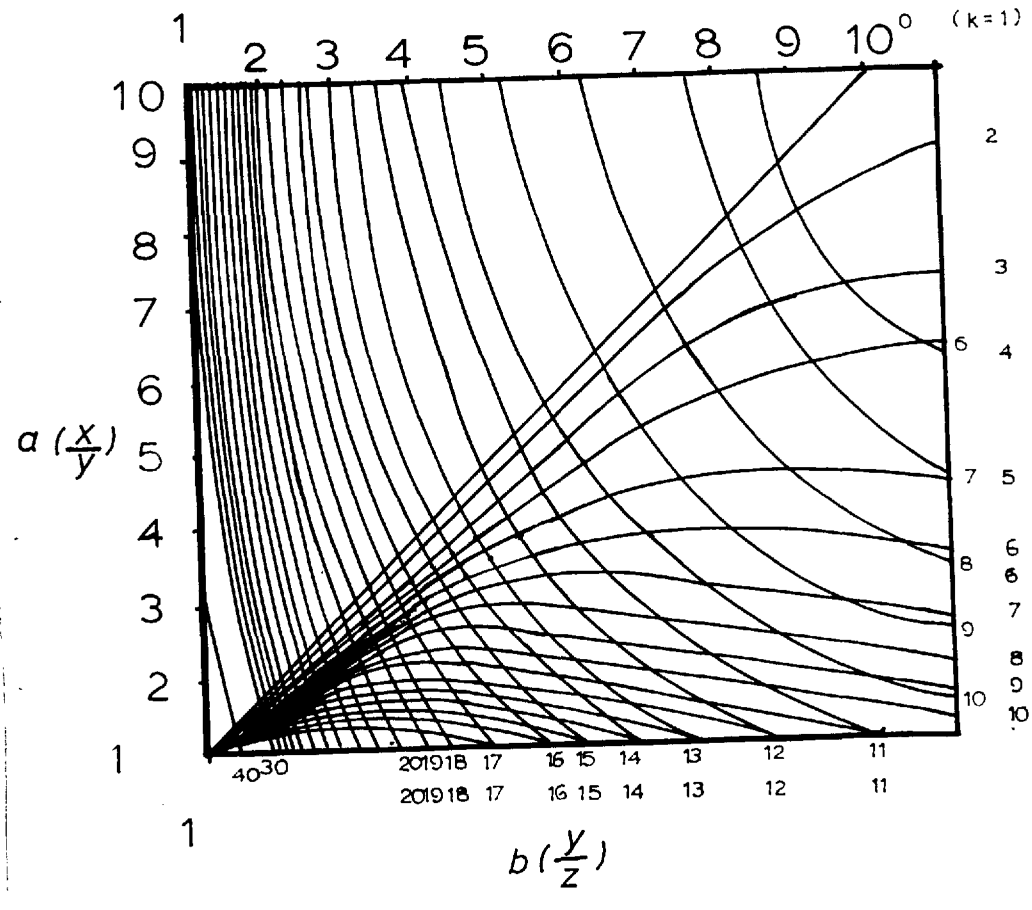
Talbot's thesis contains a rigorous discussion of the method and its usefulness in an African basement area. Here, therefore, it is necessary only to give a general outline of the working technique together with a few general criticisms and some comment on its application to the present atea.

4:3 THE WORKING TECHNIQUE

Immediately there is the problem of the recognition of shortened and elongated veins, shortened veins can only be recognised when they are folded, elongated veins when they are boudinaged, i.e. the method is only useful when there is a ductility contrast between the veins and the matrix, with the consequent inhomogeneous deformation of the veins (see section 4:4:3). Ideally, then, one measures the orientation of the inflexion or enveloping surfaces (Ramsay 1967, Chapter 7) of a folded vein on the one hand, and the orientation of a boudinaged vein on the other. The poles of these planes are plotted on a stereogram and the bounding surface between them is taken to be the projection of the surface of no finite longitudinal strain. The scatter of poles to elongated elements is theoretically a polar projection of the extension field of the ellipsoid, and its centre is the shortest axis (Z) of the ellipsoid. It is circular only where the ellipsoid is uniaxial oblate ($k = 0$), otherwise it is elongated on a great circle (the ZX plane) towards X. X is 90 away from Z in this plane, Y is 90 from X in the plane to which Z is the pole.

The axial ratios of the ellipsoid are found by measuring the half angles (half dimensions) of the projection of the extension field in the YZ and XZ planes respectively, and substituting the values in a graphical plot of Flinn's data for lines of "no finite deformation" (Flinn 1962, Table 1). This graph is shown on Fig. 2.

Fig 2 Plot of data from table 1, Flinn 1962: variation of ψ_2 and ψ_3 (angular dimensions of the extension field of the strain ellipsoid) with variation of $\frac{X}{Y}$ (a) and $\frac{Y}{Z}$ (b).



4:3:1 Domains of Homogeneity

In any assemblage of deformed rocks it is possible to recognise areas of overall strain homogeneity on a variety of scales. Homogeneous strain probably exists, for example, within individual laminae in the limb of a microfold, but not across the microfold. A suite of such microfolds may exist within the limb of a larger scale fold, and then the strain in that limb would be considered homogeneous on the scale of the larger structure.

Thus, as Talbot demonstrates, an area of deformed basement rocks contains a whole spectrum of rock domains which are effectively homogeneous on a particular scale. A technique derived from the theory of homogeneous strain is valid only if it is applied to one of these domains.

4:4 THE SHORTCOMINGS OF THE METHOD

These fall into four categories, "statistical", theoretical, practical and "geological".

4:4:1 "Statistical" shortcomings.

Readings must be taken from subjectively chosen areas of apparent strain homogeneity. The rocks must not be strongly banded (see below) and there must be no later refolding on a scale to interfere with the homogeneous aspect. Such areas are usually of limited size and consequently there is often a deficiency in the number of readings which can be made. The maximum number of readings used in a single determination here is 300, and most are based on half this number or less (see appendix sheet 1).

More important than absolute numbers of readings, however, is any subjectivity in the definition of the boundary of the extension field (the surface of no finite longitudinal strain) on the stereographic plot. Vagaries in this may be introduced by a lack of readings in the critical zone (see 4:4:4), but even given sufficient readings, the position of the boundary are made uncertain by a degree of overlap of the two facets of the data. In the present work, several possible boundaries are drawn on each plot.

If the projection of the extension field is large (i.e. the strain ratios are small) the errors introduced by this uncertainty are not important. If it is small, however, (as in the present examples) the errors are highly significant, and ellipsoids with ratios as different as 8:4:1 and 10:10:1 can be obtained from the same data (see appendix sheet 1). In cases like this, one must merely rule out the extreme results and take as significant either:

1) The boundaries which subjectively seem the most appropriate.

or 2) An average of several appropriate boundaries

or 3) An average of all possible boundaries.

All methods are unsatisfactory. Strain ratios used in this discussion are obtained by the second method, and are the "favoured results" shown on Appendix sheet 1, and used in chapter 7.

4:4:2 Practical Shortcomings

The graph (Fig. 1) is sensitive only if the dimensions of the extension field are small. When the half angle in ZX about 25° and the half angle in ZY about 17(i.e. at low strain values) it ceases to be useful. Thus where rocks are moderately or slightly strained and errors in the field measurement and the siting of the bounding surface on the stereogram are slight, the conversion graph is insensitive. It is sensitive in areas where these errors are far more significant.

These difficulties could be overcome by

1) the use of a logarithmic plot of Flinn's data.
2) A mathematical conversion of the measurements.

These treatments are not presented here, however, because of the geological considerations which make the very application of the method dubious in mildly strained rocks (see below).

4:4:3 Theoretical Shortcomings

An anachronism in the method is that a bulk strain ellipsoid is defined from the manifestations of inhomogeneous strain i.e. from deformed competent veins. Ideally these veins could be regarded merely as passive markers; their inhomogeneous deformation is then irrelevant and their sole significance is that they record the areas of shortening and the areas of elongation in the host rock. Even if we accept this in principle, however, we cannot entirely dismiss the significance of ductility contrasts between the veins and their matrix, especially if these are variable.

It is apparent, for example, that lower ductility contrasts are necessary to initiate buckling than to initiate boudinage (Chapter 3, section 3). Thus, at high strains at least, shortened veins are almost always folded, but elongated veins are not necessarily boudinaged. It is therefore necessary to assume that any planar veins at high angles to symmetrically (M) folded veins have suffered homogeneous elongation and in practice one measures the orientation of the inflexion surfaces of folded veins on the one hand, and the orientation of apparently undeformed (i.e. uniformly extended) veins on the other.

Coward (personal communication 1969) has pointed out a significant disadvantage in this procedure, especially where ductility contrasts between the veins and their matrix are small, or (an exactly similar implication) where the deformation is low. Under these conditions some veins would shorten homogeneously, and in the critical area around the surface of no finite longitudinal strain, they would be indistinguishable from homogeneously elongated veins. Thus the projected area of elongation would be too large and the strain ratios significantly too low.

In highly strained rocks where all shortened veins are folded to some extent this uncertainty is replaced either by one produced by a dearth of veins around the surface of no finite longitudinal strain or by the existence of a family of highly asymmetrically folded veins in the area. In these circumstances the position of the bounding surface between shortened and elongated elements is still speculative, but the importance of Coward's specific criticism is minimized.

Nevertheless, any strain within the host rock which was taken up in homogeneous shortening of the veins cannot be recorded by this method. Thus, if the veins were shortened only about 30% (approximately 2:1) before they buckled (a perfectly reasonable amount - see Sherwin and Chapple 1968 and Hudleston 1969) the strain recorded by this method would be about half the real strain value, (see Chapter 3).

Harmonic folding

Since the basic theory of the method demands free rotation of the veins within their matrix, it is theoretically essential that all individual veins behave as separate entities. The method is unreliable if the distance between individual competent veins is small enough for the contact strain between them to initiate harmonic folding, (Ramberg 1963a).

4:4:4 "Geological" shortcomings

a) Original Orientation

Theoretically, the method is valid only if the original orientation of the veins was random. Such an original orientation would mean folding about a random family of axes, and therefore the projection of the extension field would be completely surrounded by folds to pole inflexion surfaces. The boundary between the two elements could, with the previous reservations, be established with some confidence. In the North Uist-Sound of Harris area, however, there is no evidence that any deformed veins were originally randomly distributed, for there is a marked degree of parallelism of all fold axes. The incomplete mantling of the extension field by the inflexion surface poles which occurs with parallel fold axes introduces yet another subjective element into the construction of the surface of no finite longitudinal strain, but need not completely invalidate the technique. If it did do so this method of strain determination would be of extremely limited use; completely random sets of veins are surely rather rare in nature.

b) Polyphase deformation

In areas of polyphase deformation there can sometimes be doubt of the exact meaning of the strain ellipsoids defined by this technique. They may record the total finite strain or they may only record the strain incurred during one or more of the deformation phases which make up that finite strain. As a general rule it is probably true that if ellipsoids are constructed from veins folded and extended during a particular deformation they dominantly record the strain suffered during that deformation phase. Each individual locality must, however, be interpreted separately and in each there may be particular complications, e.g. strain ellipsoids in the Sound of Harris steep zone have been constructed from veins folded during F2 yet they probably also contain a significant F3 component (see Chapter 7).

The veins in all rocks used for strain ellipsoid construction in this thesis are probably ultimately of Scourian age.

4:5 ORIENTATION OF THE PRINCIPAL STRAINS

The orientation of the axes of the finite strain ellipsoid given by this technique is accurate if the surface separating the poles of the folded and boudinaged elements really is the projection of the surface of no finite longitudinal strain. Although gross uncertainties may occur in measurement of the size of the projected extension field, there is much less possible variation in its shape and orientation or in the position of its central point. Thus a high degree of certainty can usually be assigned to the predicted orientations of the principal strains.

It must be noted that the constructed orientations are not necessarily the original orientations of the principal strains. They may have been bodily re-orientated by, for instance, large scale later folding of the whole homogeneous domain, (see Chapter 7).

4:6 CHECKS ON THE ACCURACY OF STRAIN MEASUREMENTS BY THIS METHOD

The accuracy of the results might be checked by comparing them with results from some other method of strain determination. As we have seen, however, few other methods are available, and the only practical comparison is with values of the total shortening along the folded veins themselves.

There are two reasons why these two sets of data should be different:

1) Competent veins may not have suffered as much strain as their matrix.
2) Strain ratios from folded veins are two dimensional in a plane normal to the fold axes, and fold axes are not necessarily parallel to any

strain ellipsoid axis. Thus a direct comparison of results is difficult.

Also, the thinness of veins in many localities within the North Uist- Sound of Harris area made it difficult to obtain reliable estimates of the shortening along folded layers. Nevertheless some independent control of the results of the vein orientation method was thought desirable, and it is presented on Table 1. The localities are those of appendix sheet 1. Neither method takes into account any homogeneous shortening of the layers. Strain ratios from constructed strain ellipsoids should be significantly higher than strain ratios from folded competent veins, since they represent the strain suffered by the incompetent matrix. The fact that they are rarely greater probably means that the strain ratios determined from veins are unduly low.

The agreement between the two sets of data is considered good for Griminish, and fairly good for the Tigharry isocline and for Berneray. It is very bad in N.W. Pabbay (2) and worse in N.W. Pabbay (1). These two localities gave strain ratios which are probably much too low, perhaps because some of the folding in these areas is harmonic.

4:7 THE USE OF THE METHOD

The various shortcomings noted above limit the use of the method.

1) Two factors make the strain ratios derived from its application in moderately strained rocks of limited value. They are:-

a) Complications introduced by ductility contrasts within the rocks (see above).

b) General rock irregularity in conditions of low strain (the 'regularity' of gneisses increases with increasing deformation). This gives an undue proportion of "background noise" on a stereographic projection.

Together these features can increase the size of the projection of the extension field to ridiculous proportions; it is possible to obtain negative strain ratios by a mathematical conversion of data from mildly strained rocks!

2) Simple, measurable distributions of data are produced only if those data are collected from a domain of homogeneous deformation. The method cannot be used in areas where later folding interrupts this homogeneity.

Great care must be taken in the interpretation of results. A strain ellipsoid determined from a small area of overall strain homogeneity may bear no relationship to the strain state of rocks even a small distance away; neither need it have any regional significance since the small area of strain homogeneity probably lies at a special position within some major structure.

Considering all these various limitations it is wise to use results from this method in two ways only.

1) As an aid in the interpretation of the geometry of strained rocks. This method provides reasonably accurate knowledge of the symmetry (k value) of the finite strain ellipsoid and gives the orientation of its principal axes. There is as yet no other way of obtaining this information in rocks not containing deformed objects of known primary shape unless we assume that the information is provided directly by the rock fabric (Flinn 1965).

2) In a comparitive way, if the shortcomings of the technique apply everywhere with equal significance it would be valid to compare the quantitative strain estimated in different areas. This assumption is, however, somewhat unjustified, and if such comparisons are made it must be with great care. The principles of, for example, the effects of ductility contrasts may be universal, but the significance of this feature will vary with different rock types. Nevertheless these two courses are the ones followed in this thesis. Detailed examples of the use of constructed strain ellipsoids are given in Chapter 7.

A final general remark concerning strain determination in basements stems from the manner in which strain increments multiply together to produce the finite strain state.

Two strains of say 10:10:1 superimposed co-axially in the orientation

$$\begin{matrix} X & x \\ Y & y \\ Z & z \end{matrix}$$

produce a finite strain with ratios 100:100:1, not 20:20:1. Since folds are effectively similar after being flattened by 10:1, and angular discordances are effectively no longer visible, one wonders whether the effects of higher strains than this could be discerned, and whether to differentiate between ratios of say 20:1 and 60:1 is slightly academic. In the realms of very high strain we almost lose the value of quantitative structural geology for its own sake.

4:8 APPLICATION OF THE METHOD IN THE NORTH UIST - SOUND OF HARRIS AREA

This section is limited to a general discussion of the data presented on appendix sheet1 . Two detailed examples of the use of the results are given in analyses of the strain history of the North Uist Coast and the Sound of Harris steep zone in Chapter 7, and briefer references are made in the chapter on the interpretation of fabric in gneisses.

Strain ellipsoids were determined from vein orientation in eight localities. Each of these in an area of up to several hundred square metres and represents a domain in which the strain may be considered homogeneous on this scale. The actual dimensions of the localities are shown on the appropriate maps. All of them occupy known positions in the major structure of the area. The Hougharry data (4) are from the hinge zone of and F3 antiform, that from Tigharry (1, 2, 3,) from the limbs of the F3 Tighary Isocline. The Griminish sampling site (5) lies in the limb of the F3 Griminish antiform near the margin of an F2 fold zone, similarly, the data from Pabbay and Berneray (6,7,8,9) come from the limb of the F3 Sound of Harris antiform, those from N.W. Pabbay (1) being from an F2 limb zone while the other three sets of data are from approximately the same F2 fold zone. The strain ellipsoid from Hougharry probably dominantly records the F3 strain whereas elsewhere the effect of the F3 deformation on the finite strain cannot be directly assessed. A more detailed discussion of the meaning of the Tigharry and Sound of Harris ellipsoids appears in Chapter 7. Data from the Tigharry isocline were collected from two sampling sites on opposite limbs of the fold, but in the same structural domain. Data from each site are presented separately and together on appendix sheet 1. Even put together the sets of data are numerically inadequate, and the boundary of the extension field is poorly defined - indeed much of it merely surrounds the scatter of poles to extended elements. This is unfortunate because the strain ratios from Tigharry are probably much more realistic than those from say, N.W. Pabbay, and there is reasonable correspondence between them and two-dimensional ratios from folded veins, (Table 1).

The plots from Berneray and S.E Pabbay appear almost ideal, but it must be noted that the complete mantling of the extension field by inflexion surface poles in these areas is not a function of an initial random distribution of the veins, but rather of "in plane" deformation of the fold axes (see Chapter 7). This deformation of linear elements in the XY plane of the strain ellipsoids is shown in their girdle distributions of stereograms 7, 8 and 9 on appendix sheet 4. It would appear that as the projection of the extension field becomes more nearly circular (i.e. the strain ellipsoids become more oblate) the spread of lineations in the XY plane becomes more complete. Initially this would seem directly in accord with homogeneous strain theory (e.g. Flinn 1962; Ramsay 1967, Chapter 4). Lineations of originally constant direction, however, can only be deformed by inhomogeneous strain in the rock planes, and an explanation of the feature in the simple terms of homogeneous strain must be regretfully abandoned. The problem of these deformed lineations is examined in detail in Chapter 7.

TABLE 1

STRAIN RATIOS DETERMINED FROM DEFORMED VEINS COMPARED WITH 2 - DIMENSIONAL STRAIN RATIOS FROM FOLDS (MEASURED NORMAL TO FOLD AXES).

STRAIN RATIOS FROM VEIN ORIENTATION		STRAIN RATIOS FROM FOLDS
	Pabbay N.W. (1)	
min.	1.5:1.5:1	Similar to N.W. Pabbay (2) below.
max.	2.5:2:1	
Average	2:1.75:1	
	Pabbay N.W. (2)	
min.	2:2:1	Ratios in competent bands vary from 2:1 to 16:1, average values 5:1 to 10:1. Incompetent bands have near similar geometry (i.e. shortening is theoretically infinite). Fold axes are usually near Y i.e. these ratios record the maximum 2-dimensional shortening.
max.	6:3:1	
Average	4:2.5:1	
	Berneray	
min.	7.4:3.7:1	Competent veins 10:1 to 21:1 in sections mid-way between ZX and ZY. 21:1 may be anomolously high because the section is not exactly normal to the fold axis. 10 to 15:1 might be considered realistic.
average	11.7:6.5:1 (favoured).	
max.	10:10:1 17:10:1	

STRAIN RATIOS FROM VEIN ORIENTATION	STRAIN RATIOS FROM FOLDS
Pabbay S.E.	
min 6:4:1 max. 7:5:1(favoured) 6:6:1	No data available
Tigharry Isocline	
min. 9:3.5:1 favoured 18:6:1 max. 10:9.5:1	competent bands 12 - 15:1 section between ZX and ZY.
Griminish	
6:3:1	Average shortening along competent bands 3:1 in sections of maximum 2-dimensional shortening (ZX).
Hougharry Antiform	
3:1.5:1	No data available

CHAPTER 5

ASPECTS OF THE METAMORPHISM OF NORTH UIST AND THE SOUND OF HARRIS

5:1 THE METAMORPHIC HISTORY AND GENERAL METAMORPHIC STATE OF THE COMPLEX

This thesis does not attempt a detailed examination of the metamorphism of the rocks of North Uist and the Sound of Harris but since it does examine the structures produced while the rocks were in a highly metamorphic state, a short chapter on the subject is essential.

It has long been the custom to outline the metamorphic histories of rocks in orogenic and gneissic terrains by reference to the phases of deformation, i.e. from an investigation of the rock fabrics to establish the mineral assemblages which grew before, during and after the structures of a particular phase of deformation, the time relationships being interpreted from the geometrical relationships which the minerals have with those structures. This procedure requires the inhomogeneous development of fabric (see Chapter 6) and is particularly suitable in mica schists such as those of the Dalradian rocks of Scotland or of the Alpine cover. In quartzo-felspathic gneisses such as those of the present area, this sort of analysis is often difficult to carry out, and where it is possible, the coarseness of grain of the rocks often makes direct field observations more useful than thin section examination. Some of the reasons for these difficulties are presented in the next chapter, and include

i) The restricted variety of mineral assemblages

ii) Greater uncertainties as to whether minerals have rotated into a certain position or really crystallized in it.

iii) The common occurrence of a homogeneous fabric development (see Chapter 6) instead of the overprinting of successive separate fabrics.

iv) The dubious geometrical significance of some fabric relationships, e.g. the co-axial super-imposition of sets of structures means that a mineral lineation need not be (and often is not of the same age as the folds to whose axes it is parallel.

As a generalization it may be said that all the post Scourie-Dyke ductile deformation phases in the North Uist - Sound of Harris area took place under conditions of amphibolite facies metamorphism, and that static re-crystallization took place between the successive phases. This conclusion is drawn mainly from an examination of the fabrics in the Scourie-Dykes themselves, and especially from localities where garnet is contained in these rocks. Garnet is the metamorphic time-marker par-excellence in the North Uist - Sound of Harris area because it always overgrows tectonite fabrics in a post-kinematic manner (syn-kinematic garnets have not been recognised in the area) and is always partially or totally retrogressed during subsequent deformation. Two examples follow as illustrations of evidence for metamorphism syn-kinematic with all three deformation phases on the North Uist coast, and of presumably static metamorphism before and after the F2 deformation. Similar examples could be quoted for the Sound of Harris, but the metamorphism of South Harris is more complex. It is briefly considered in Chapter 8.

EXAMPLE 1. F1 - F2 metamorphism - Scourie Dyke amphibolite from Griminish, North Uist.

Most of the implications of the fabric variations within the Griminish Synform are discussed in the next chapter. The important facts at this stage are as follows. A single Scourie dyke amphibolite band is folded around an F2 fold which is itself refolded by the Griminish Synform. In one limb of the F2 fold the amphibolite contains a lineated planar fabric of F1 age which is overgrown post-kinematically by garnets. In the other limb the fabric is totally linear and almost all the garnet is retrogressed, drawn out hornblende-felspar aggregates after garnet being important components of the linear fabric. The linear fabric is developed by the superimposition of the XY plane of the F2 strain at high angles to the F1 planar fabric.

The metamorphic history may be summarized as follows:

- F1 Amphibolite facies metamorphism, development of a planar fabric.
- Post F1 Garnet growth (amphibolite-facies metamorphism)
- F2 Amphibolite-facies metamorphism, retrogression of garnet and profound geometrical modification of the F1 fabric.

EXAMPLE 2: F2 - F3 metamorphism

Evidence for a static metamorphism between the F2 and F3 phases of deformation is always less convincing than that for post F1, pre F2 metamorphism because of the difficulty of separating the F1 and F2 components of the Scourie-Dyke amphibolite fabrics (see Chapter 6).

Garnets locally overgrow F2 gneissic cleavages in some of the metasediments on Pabbay, while in the Tigharry Isocline of the North Uist coast garnets post-date an F1/F2 fabric in Scourie-Dyke amphibolite bands (See Chapter 7). This fabric is flattened around the garnet crystals to some extent, probably by F3 strain associated with the development of the isocline. Retrogression of the rims of the crystals may have been co-eval with this F3 strain.

Metamorphism post-dating the F3 deformation

All over the North Uist-Sound of Harris area there are large hornblende crystals which probably grew post-kinematic with respect to the F3 deformation. Metamorphism apparently outlasted deformation.

Summary

The evidence recorded above suggests that we may look upon the metamorphic history of North Uist and the Sound of Harris as a continuous amphibolite facies regional metamorphism occupying much of the Laxfordian time interval, and during which there were three important episodes of deformation.

Garnet is a sensitive strain indicator because it is invariably retrogressed as the rocks are strained. If the strain is added to an early planar fabric so that a finite linear fabric is produced (see Chapter 6), the retrogression is almost invariably complete if the original planar fabric is emphasised by the later strain, the retrogression is usually only partial. This observation provides an example of the high degree of interdependence between strain and metamorphism in the rocks of the Hebrides, a subject which is considered to be of fundamental importance, and one which is examined again later in this chapter and elsewhere in the thesis.

An interesting facet of the metamorphism of the North Uist-Sound of Harris area, and one through which we may examine further the association of strain and recrystallization, is the metamorphism of basic rocks of the Scourie-Dyke suite - the alteration of original tholeiites to amphibolitic tectonites. Early stages in this process are not preserved in most of the Scourie-Dyke amphibolites of the area, since they are usually fairly thin and completely recrystallized even where (except for the presence of tectonite fabrics) they appear to be undeformed. Thick basic bodies, however, often possess cores of coarse grained rock where fabrics are isotropic and amphibolite facies recrystallization is not complete. Marginal parts of these bodies are usually thoroughly amphibolized and mineralogical changes can be traced between the cores and the margins.

In the North Uist-Sound of Harris area large basic masses showing these features occur at Castell Odair on the North Uist coast, on Boreray and Oronsay islands, on Crogarry Mor (a hill in the northern part of North Uist), at Hornish on the Newton peninsula, at Veilish Point on Udal, on Borve Hill in Berneray and in northern Pabbay. The Castell Odair body is described below.

5:2 CASTELL ODAIR

5:2:1 Introduction and geological setting

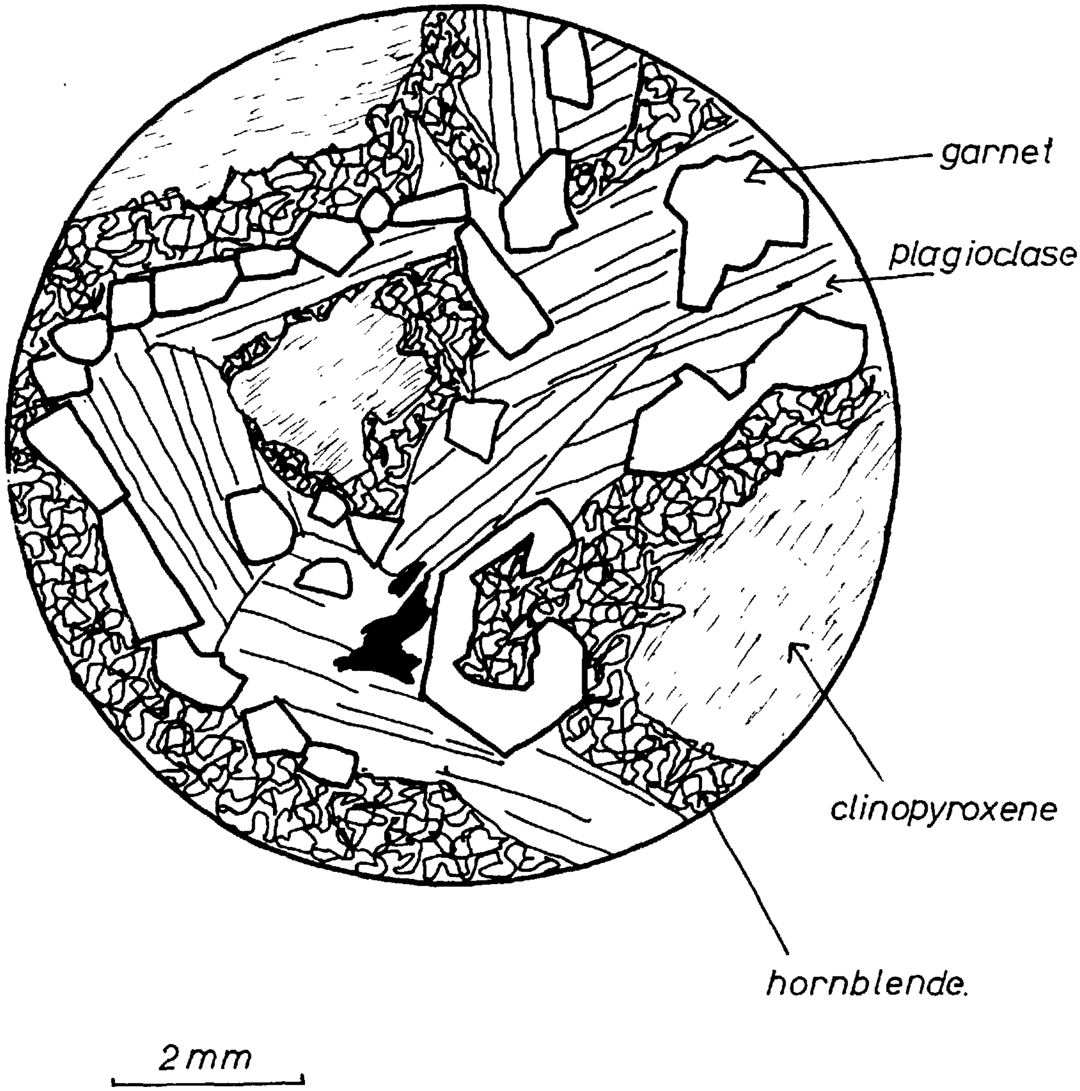
On the headland of Castell Odair on the North Uist Coast is exposed a large basic mass probably belonging to the Scourie-Dyke suite. It has a sheet-like form and is folded around the eastern hinge of the box-like F3 Griminish antiform (see Chapter 1), the hinge of this fold lying in the neck of an older (probably F2) boudinage structure in the sheet. Late low-angle fractures of fairly large scale (see Map 1) are numerous in the Castell Odair area, and one such fracture occurs along the lower contact of the sheet. Wisps and veins of pseudotachylite occur within the body, especially in the neighbourhood of this fracture.

The central regions of the mass are undeformed, the texture of the rock is igneous and at first sight it resembles a gabbro. Away from the centre, however, the body exhibits a coarse tectonite fabric in which both a linear and a planar arrangement of the mineral aggregates is visible. Cutting this fabric, and also present in the undeformed parts of the mass, are numerous small shear-zones in which a very strong foliation is developed, and in which the rock is effectively converted to a hornblende schist(c.f. Teall 1885). The contact zones of the body are highly deformed, and the fabric in these narrow areas is dominantly planar, locally (e.g. along the lower contact) friable amphibolitic "paper schists" are developed.

5:2:2 The Undeformed areas

Much of the central part of the Castell Odair mass is composed of coarse grained amphibolite in which there is no trace of a tectonite fabric. Two small areas within this undeformed central zone are only partly amphibolized, however, and contain the mineral assemblage pyroxene-plagioclase-garnet. Rocks with this mineralogy have been taken as evidence of granulite facies metamorphism by Dearnley (1962). The fabric at Castell Odair is, however, unquestionably igneous, the pyroxene and plagioclase laths are large, and before being corroded were probably euhedral, factors suggesting their primary igneous nature. Garnet is an alteration product and forms fresh sub-hedral or euhedral crystals mantling the pyroxene laths. Inside the garnet rims the pyroxene laths are partly altered to amphibole, though the centres of the crystals usually preserve patches of little altered clino-pyroxene. The amphibole occurs in amorphous masses which are made up of countless tiny hornblende grains. A sketch of a thin section of this pyroxene-plagioclase-garnet rock is shown in Fig. 1.

Amphibolization increases in the Castell Odair body outwards from the two areas of garnet-pyroxene rock. All the pyroxene vanishes, and its place is taken by hornblende in pseudomorphous clots larger in area than the original pyroxene laths. The individual crystals in the hornblende masses increase their size, the large plagioclase laths become sericitized and broken down, smaller more equi-dimensional crystals of less calcic plagioclase develop, garnet disappears and becomes pseudomorphed by masses of opaque material. With the retrogression of garnet, Fig 1 Sketch of thin section of pyroxene-bearing metagabbro from the undeformed centre of the Castell Odair basic mass.



small rounded quartz grains appear within the rock; the quartz is often strained in the "undeformed" rock.

5:2:3 <u>The Deformed rock</u>

i) <u>In Shear-zones</u> (Chapter 4 plate 1)

In the shear-zones there is a reduction in rock grain size. The new foliation in the shear-zones is produced partly by the dimensional orientation of hornblende, and partly by a new lithological banding developed by the alteration of felsic and hornblendic layers of less than 1mm thick. All the felspar is completely sericitized and the opaque masses after garnet disappear. Quartz grains no longer show any signs of straining, presumably because they have been re-crystalized. Sphene is occasionally developed in the felsic layers.

ii) <u>Outside the Shear Zones</u>

The development of tectonite fabrics more generally in the body (i.e. outside the shear-zones) is shown first by the dimensional orientation of the aggregates of hornblende crystals of the undeformed rock. Then the crystals in the aggregates increase in size, the opaque masses after garnet disappear and the quartz loses its strain effects (as in the shear zones it is probably re-crystallized). The main differences between the deformed rocks of the shear-zones and those outside the zones are firstly that the felspar in the shear-zones is far more completely altered and secondly that the fabric in the shear-zones is more perfectly developed than it is anywhere else in the mass except in the contact areas. The alternating layers of the shear-zone foliation may represent enormously modified hornblende aggregates though no real evidence of them remains, whereas their remnants are frequently visible in deformed rocks outside the shear-zones.

5:2:4 Interpretation

The retrogression of the Castell Odair basic mass is essentially a mineralogical adjustment to the introduction of water. All the various mineralogical associations described above are seen as stages in one progressive reaction, their variety being dependent upon the amount of water in the system during recrystallization and on the intensity of co-oval strain. The decrease in quantity of OH bearing minerals (notably of hornblende) towards the centre of the mass is analogous with a situation described by Buddington (1963) where gabbro sheets in a sedimentary sequence are metamorphosed in the amphibolite or "hornblende granulite" facies marginally, but recrystallized in the granulite facies in their centres where water had no access. In the North Uist basic bodies, the central parts contain modified original igneous minerals rather than granulite facies metamorphic assemblages, but the principle is the same. A partly open system is implied in this argument; another argument (Buddington 1963) involving the assumption of a closed system where the mineral assemblages developed depend entirely on the amount of water originally contained in the rock, cannot be applied to the basic bodies of North Uist since they were originally tholeiitic and therefore virtually devoid of H_2O. The assumption of a partly open system suggests the importance of rock permeability in determining metamorphic grade. Original igneous or (e.g. in South Harris, see Chapter 8) granulite-facies mineral assemblages are preserved in the centres of igneous bodies simply because of the impermeability of igneous rock - they cannot be retrogressed because water cannot reach them. This is presumably one reason for the apparent "lagging behind" of basic rocks during retrogressive metamorphism, a phenomenon which has been described by several authors (e.g. Sutton and Watson, 1951).

The centre of the Castell Odair basic mass shows only incipient effects of hydration (retrogression). I would suggest, however that enough water had entered the system to catalyse a reaction between the pyroxene and the calcic plagioclase which led to the formation of hornblende and garnet and the breakdown of the original minerals.

$$\text{PYROXENE} + \text{PLAGIOCLASE} + H_2O = \text{HORNBLENDE} + \text{GARNET} + \text{QUARTZ}$$

Garnet may have formed instead of hornblende because of the limited amount of water present, and with an increasing proportion of water away from the centre of the rock body garnet was replaced by hornblende, plagioclase and opaque masses, and more quartz was thrown out.

After the disappearance of garnet, the metamorphic changes in the mass involved the recrystallization of hornblende with a gradual improvement in the shape of the crystals, and with the crystallization of metamorphic felspar instead of the original igneous plagioclase. Even so, the undeformed amphibolites of Castell Odair are far from completely recrystallized; hornblende still occurs in a wide variety of grain sizes and the crystal shapes are still poor.

The regional deformation of the Castell Odair mass (the development of a tectonite fabric) occurs towards the outside of the fringe of amphibolization. The crystal aggregates become dimensionally oriented and a degree of recrystallization higher than that in the undeformed amphibolite is apparent. The fine grained amphibolites along the contacts of the body, however, are still more recrystallized and carry a stronger tectonite fabric. They are presumably derived from the normal metagabbro by intense planar contact strain in the marginal zones. In the shear zones very intense strains are locally developed (Ramsay and Graham, op. cit.) and as in the contact zones, recrystallization is complete - no vestiges of the early reactions remain.

It seems that the amphibolite facies recrystallization "frozen"at various stages in the undeformed or regionally deformed metagabbro was carried to completion in the highly strained shear zones and contact zones. Reactions which occur only over tens of metres in undeformed rock are telescoped into two or three centimetres at the margins of shear zones.

There are hundreds of examples of this kind of relationship all over the Hebrides; Under conditions of free recrystalization, whenever there is the greatest strain, there also is the most complete recrystallization, and one must conclude that strain is a catalyst of immense importance in metamorphism, without which a metamorphic reaction may never be brought to completion. This relationship is well known; Flinn, for instance, states (1965) "Silicate reactions and phase changes generally take place rather slowly. The grain-boundary migration, nucleation, and strain energy in the lattices during deformation all help to eliminate unstable phases and crystallize new stable ones."

An equally fundamental relationship seen at Castell Odair is the association of water with deformation and recrystallization: wet rocks are usually deformed, dry rocks are usually undeformed (except, perhaps, in discrete zones). The immediate problem is, did the deformation provide pathways for water, or did the presence of water in the rocks enable them to suffer deformation? Though rocks undergoing deformation, or rocks with a well-developed tectonite fabric may well allow water to pass through them more easily than rocks with an isotropic fabric, I contend that at normal temperatures of regional metamorphism water must be present in the rocks before they can suffer ductile deformation. This is supported by the fact that in the Castell Odair body the fringe of intense amphibolization reaches nearer to the centre of the mass than does the area of penetrative deformation. At amphibolite facies temperatures, metagabbro with H_2O in excess will deform in a fairly ductile manner, metagabbro containing only a small amount of water will not; it may behave in a more brittle manner and fail along discrete planes.

Shear-zones are ductile faults (or to be more correct, faults are special, brittle varieties of shear zones): they represent plastic failure in discrete planes. Presumably they developed when the rock mass was not ductile enough (not yet wet enough?) for a more homogeneous type of deformation to occur. The orientation of shear-zones at Castell Odair is apparently random, and the zones probably developed throughout a considerable time interval (some cut the regional fabric of the metagabbro, for instance, while some are probably co-eval with the regional fabric, occurring in the otherwise undeformed rock beyond the limits of the development of the regional fabric). Well developed shear zones do not occur, however, in the least modified (the dryest) metagabbro. Their absence from this area and their irregular orientation might be explained if they have actually been initiated as a result of the hydration of the gabbro mass. If the introduction of water was uneven, there would be a variable volume increase in different parts of the mass, and differential stresses would be set up internally which might initiate shear zones. Alternatively, the shearing might have been taken up along old planes of weakness (e.g. old joint planes) in which, perhaps there was also slightly more water than elsewhere.

In many parts of the Hebrides the deformation state and the metamorphic state of a rock are both dependent upon the water content of that rock, and therefore may depend on the permeability of the rock. Impermeable rock masses, e.g. igneous bodies, or rocks completely free of mica often act as resistant masses within a succession of gneisses. They are often unmetamorphosed, or metamorphosed in the granulite facies, and they are either undeformed or the deformation is restricted to local shear zones in which there has been recrystallization in the amphibolite facies. Rocks outside these impermeable masses suffer amphibolite facies metamorphism and penetrative regional deformation - the "re-working" of basements is often associated with hydrous metamorphism superimposed on an earlier granulite facies metamorphism.

In the writer's view this is the most important feature of basement deformation, and its implications can be carried into almost all areas of naturally deformed rocks. It is discussed further in the conclusions of this thesis after the South Harris area has been discussed.

CHAPTER 6

FABRIC AND STRAIN HISTORY

6:1 INTRODUCTION

The study of crystal behaviour in deformed media is a complex one which falls within the compass of several scientific disciplines. Here it is examined from a strictly geological point of view, and the treatment is non-mathematical.

Flinn (1956, 1965) has advocated the interpretation of fabric in terms of the "deformation ellipsoid". The discussion which follows modifies some of Flinn's assumptions and attempts to examine the relationships between strain history and fabric development, with special reference to superimposed deformation. The first half of the chapter outlines the principles of the study, the second half considers the fabric history of the North Uist Coast as an attempt to apply these principles.

6:2 SHEAR-ZONES - AN EXAMPLE OF A RELATIONSHIP BETWEEN FABRIC AND STRAIN

Flinn's suggested correlation of fabric symmetry with the symmetry of the strain ellipsoid comes partly from a study of the fabrics of a deformed conglomerate involving a comparison of the pebble shapes with the symmetry of the fabric. The association is a special one for most deformed metamorphic rocks do not contain objects from which accurate estimates of the strain can be made. The shear zone described in chapters 4 and 5 are exceptions to this general rule, however. As we have seen, it is possible to define the state of strain quite precisely anywhere in a shear zone providing that simple shear alone is the operating mechanism. From a study of the relationship of the new shear zone fabric to this strain state there emerge two points of fundamental importance.

i) The minerals of an isotropic metagabbro begin to gather themselves together into a foliation trending approximately 45° to the shear direction i.e. given the correct metamorphic conditions, fabric starts to develop in response to very early increments of strain.

ii) The new shear zone fabric becomes more intense as it comes closer to the shear direction i.e. as the strain ratios increase. The manner in which its orientation changes is exactly that in which the XY plane of the strain ellipsoid changes its orientation during simple shear deformation i.e. the shear zone foliation records the XY plane of the strain ellipsoid.

From these points we can see a direct relationship between rock fabric and strain, and that the one evolves along with the other.

Petrofabric analyses were done on quartz grains in a continuous section from apparently undeformed rock at a shear zone margin to highly deformed material in the

Fig 1

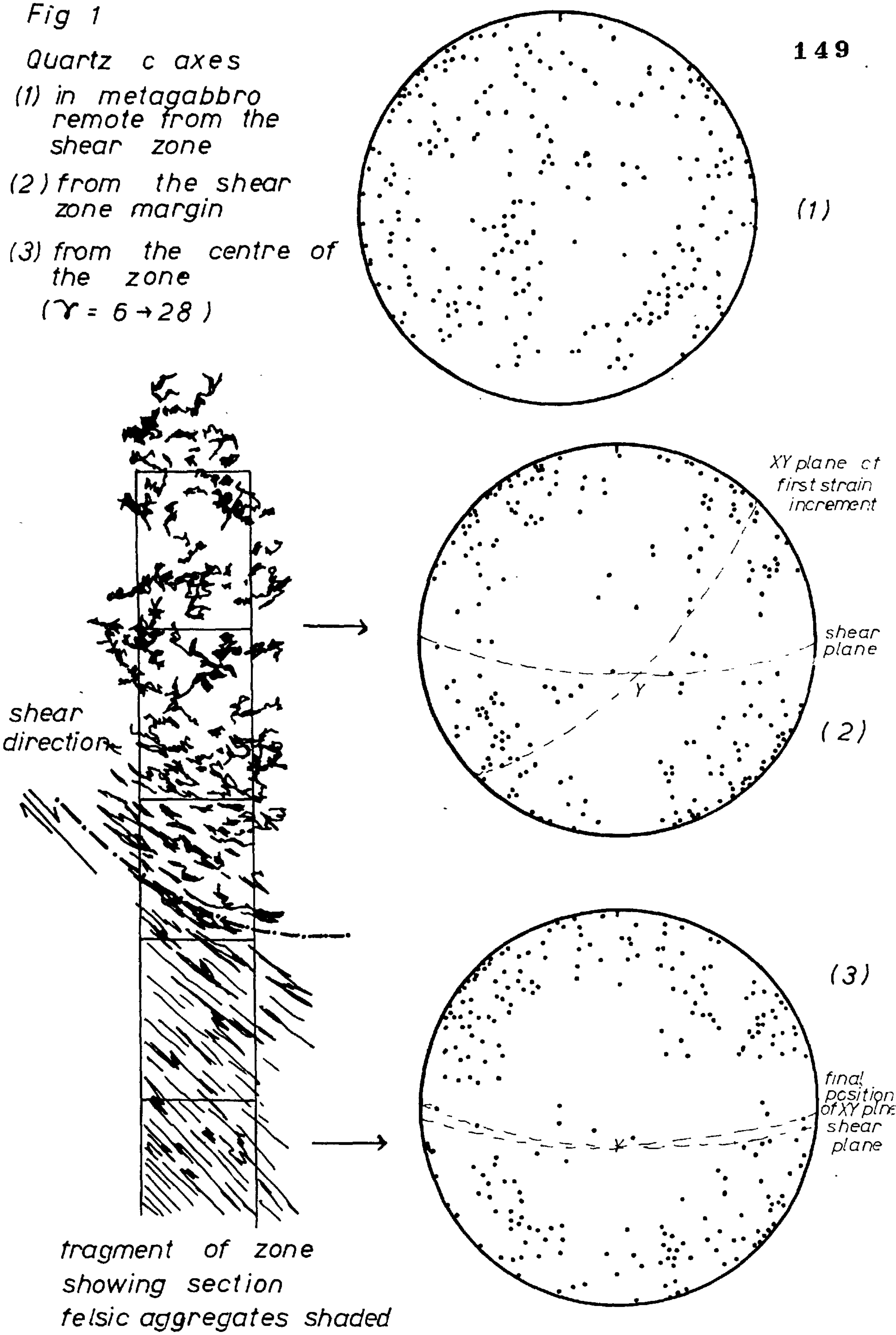
Quartz c axes

(1) in metagabbro remote from the shear zone

(2) from the shear zone margin

(3) from the centre of the zone

(γ = 6 → 28)



centre of the zone, and undeformed rock from areas away from the shear zone was also examined. The results of these analyses are shown in Fig. 1. Undeformed rock remote from the shear zone shows an apparently random distribution of c axes. Apparently undeformed rock from the shear zone margins, however, showed two distinct concentrations of axes, one about the X direction of the very first increment of strain, one around the Z axis of the same ellipsoid. Towards the centre of the shear zone (i.e. as the strain increases) the concentrations become progressively more diffuse until rock from the shear zone centre shows a girdle of c axes (in the plane normal to the final position of the foliation in the shear zone) upon which are only the ghosts of the earlier concentrations. These "ghosts" are probably due to mimetic recrystallization of the quartz about the old crystallographic axes (re-crystallization of quartz is apparent since quartz grains are strained outside the shear zones, but the undulose extinction disappears in the highly strained material), the girdle of axes can be viewed as a response of the crystallizing quartz grains to changing strain conditions.

6:4 PRINCIPLES OF A FABRIC STUDY IN REGIONAL GNEISSES

An intimate relationship between strain and rock fabric is seen in the shear zones, and it can be seen how the orientation and intensity of the fabric varies with the changing orientation and intensity of the strain. Study of the shear zones gives little information concerning the significance of fabric symmetry, and this discussion therefore adopts Flinn's concept of a direct relationship between fabric symmetry and strain symmetry, with the addition (suggested after consideration of evidence from the shear zone) that changes in the symmetry of the fabric occur with changes in the symmetry of finite strain. Superimposition of finite strains (each deformation phase is regarded as a finite strain) is evident throughout the North Uist - Sound of Harris area, and such superimposition must cause changes in the symmetry of the final finite strain. This chapter studies variations in the fabric and attempts to relate such variations to changes in the symmetry of the finite strain, and thereby with the regional structure and structural history of an area (the North Uist coast).

Two initial problems concern the inhomogeneity of deformation in the gneiss complex, and the method of analysis of superimposed deformation. The problem of inhomogeneity is overcome by studying domains of homogeneity which have larger or smaller scale than the overall heterogeneity, while a method of investigating the effects of superimposed finite homogeneous strains is set out below. It involves the direct multiplication of principal strains and follows Ramsay, (1967, p.213).

6:4 THE SUPERIMPOSITION OF FINITE STRAINS

If finite strain ellipsoids are superimposed at random, an infinite number of shapes and orientations of final ellipsoids are possible. If, however, the superimposition is co-axial, there are only six ways in which it can proceed, namely:

X	x	x	y	y	z	z
Y	y	z	x	z	x	y
Z	z	y	z	x	y	x

where XYZ is the earlier finite strain ellipsoid, and xyz is the finite strain ellipsoid which is superimposed upon it. X_1 Y_1 Z_1 is the (final) finite strain ellipsoid. The strain path by which this final strain ellipsoid is achieved (the variation in its geometry with successive increments of the second strain, xyz) can be drawn on a Flinn diagram, and natural fabric variations compared with theoretical strain paths.

Application of this technique to the complexities of deformed rocks might initially seem too great an oversimplification, since there is no necessity for finite strains to be superimposed co-axially in nature. The co-axial style of deformation commonly observed in basement areas does not necessarily mean the co-axial superimposition of strains (see concluding discussion). The geometrical relationships of structures of the three deformation phases of the North Uist-Sound of Harris area (see chapters 1 and 2) suggest, however, that the "XY planes" of the three successive deformations were superimposed at very high angles (effectively at right angles) to one another, and this fact does limit the possible orientations of superimposition considerably. The strain determinations from deformed veins in Pabbay are, if valid, clear indications of the natural co-axial superimposition of strain ellipsoids (see next chapter), for the relative values of the principal strains in that area change while their orientation remains constant.

6:5 EXAMPLES OF THE SUPERIMPOSITION OF STRAINS

Even in the special case of co-axial superimposition of strains, the number of possible combinations is formidable. Figs. 2 - 6 show various strain paths of final ellipsoids produced by the superimposition of ellipsoids of equal k value only; the number of possible paths is limitless where k values are not equal. The examples given are the simplest possible cases.

The superimposition of finite strains in the North Uist - Sound of Harris area so that the successive XY planes were normal to each other limits the possible relative orientations of superimposed ellipsoids to:-

X	x	X	y	X	z	X	z
Y	z	Y	z	Y	y	Y	x
Z	y	Z	x	Z	x	Z	y

The diagrams clearly show that the k value of the final strain ellipsoid (X_1 Y_1 Z_1) may change continually as successive increments of the second strain are added. Furthermore, the axes of the final strain ellipsoid change their relative values along some strain paths, e.g. in the case X y for two triaxial oblate strain
Y z
Z x

where the Z axis of the first strain ellipsoid becomes the Y axis (Y_1) of the final strain ellipsoid after a very small increment of the second strain. A detailed description of all similar changes is not necessary here,

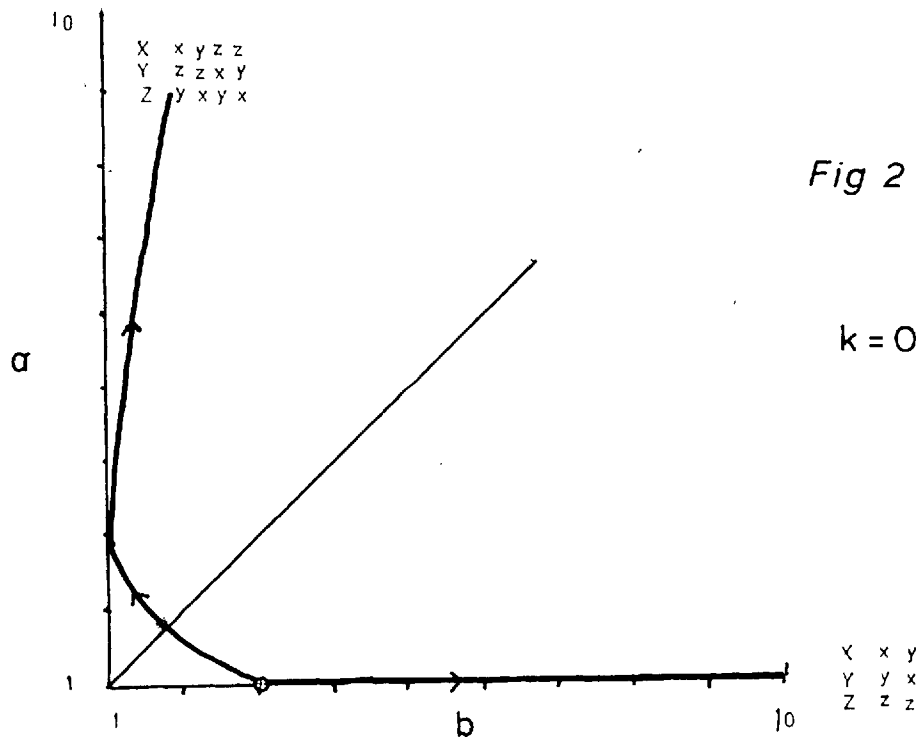


Fig 2

k = 0

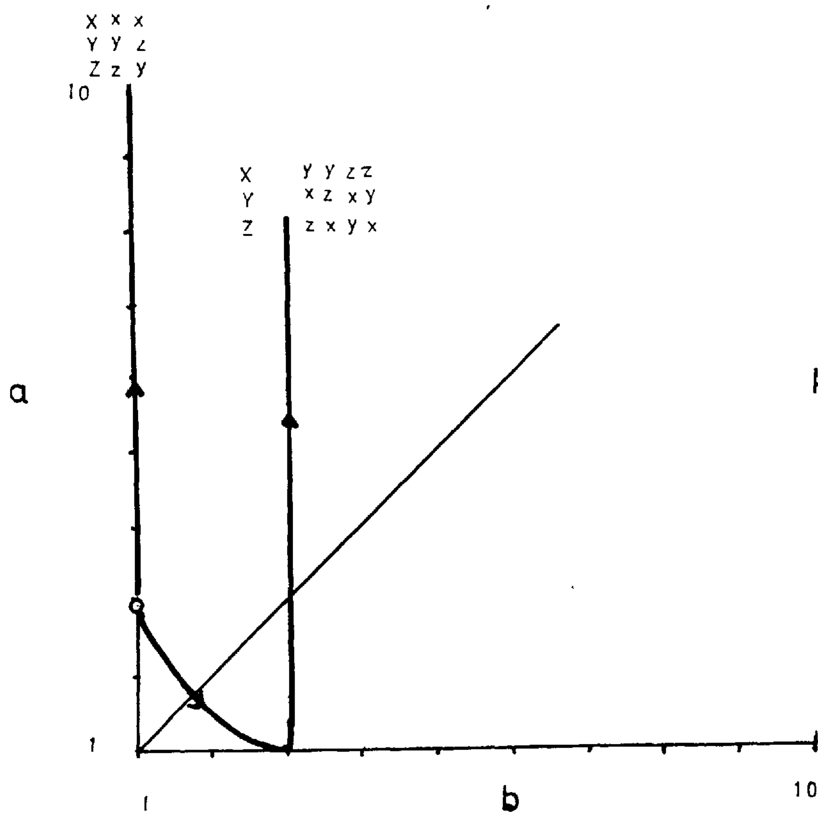


Fig3

k = ∞

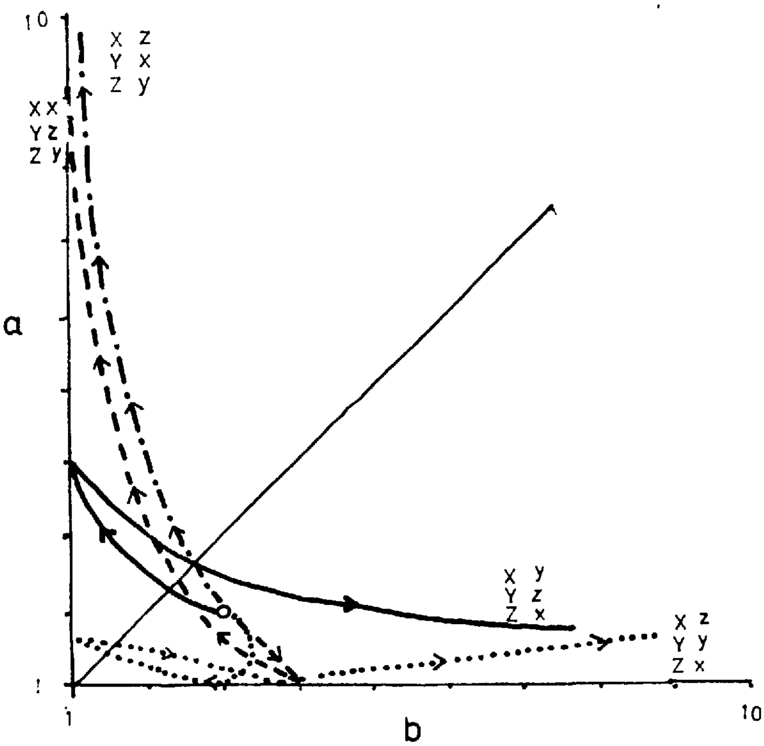


Fig 4

k = 0·5

Fig 5

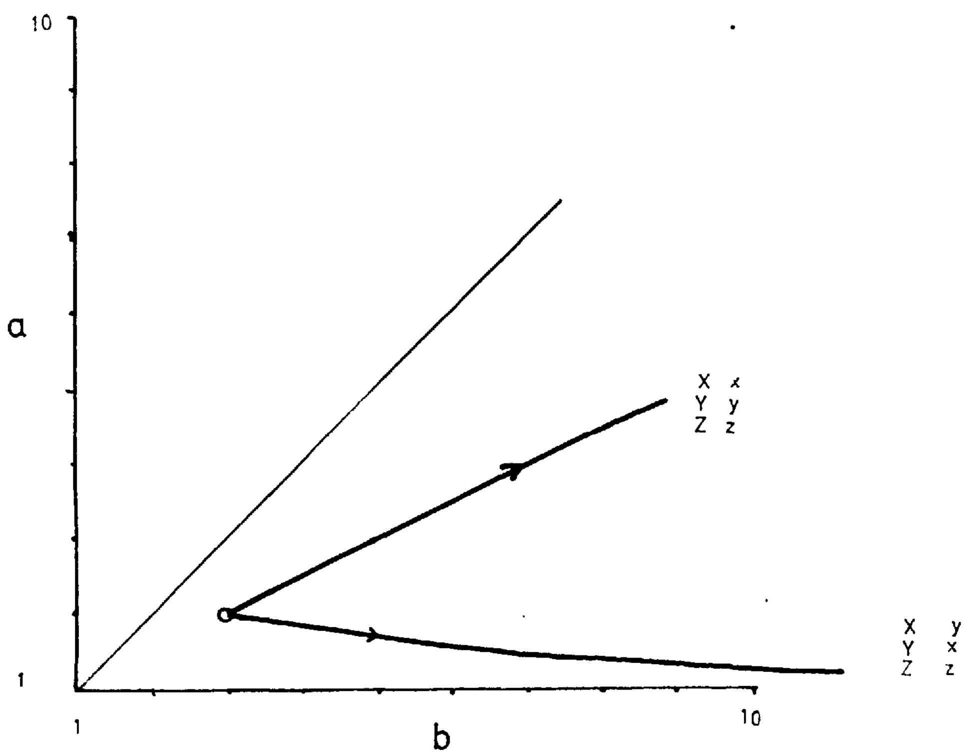


Fig 7

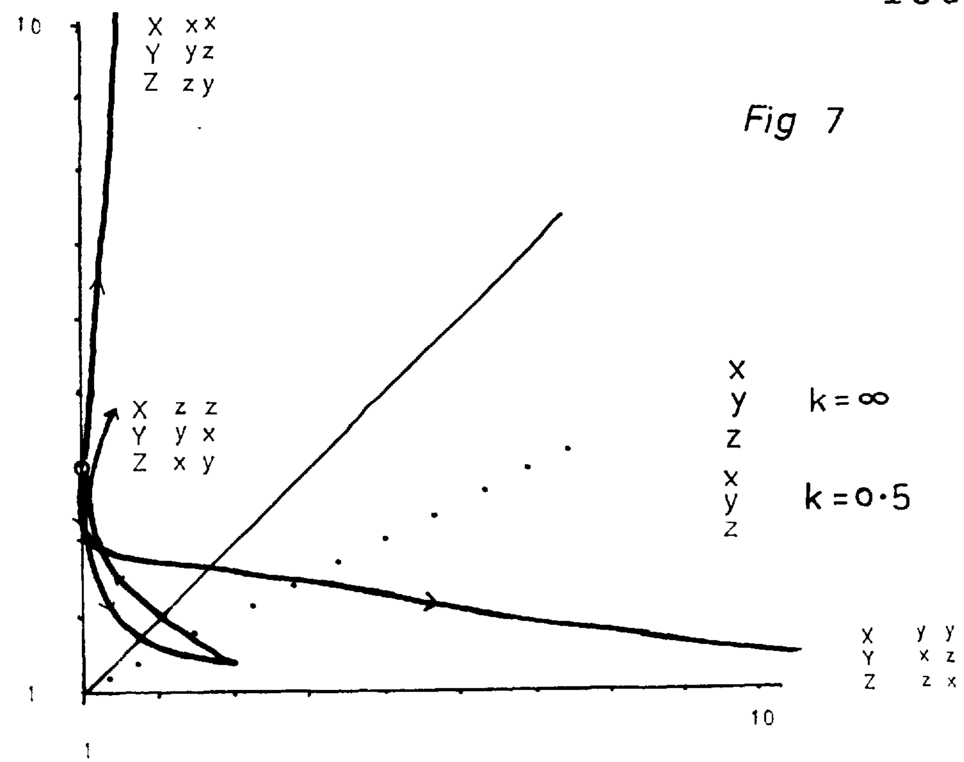
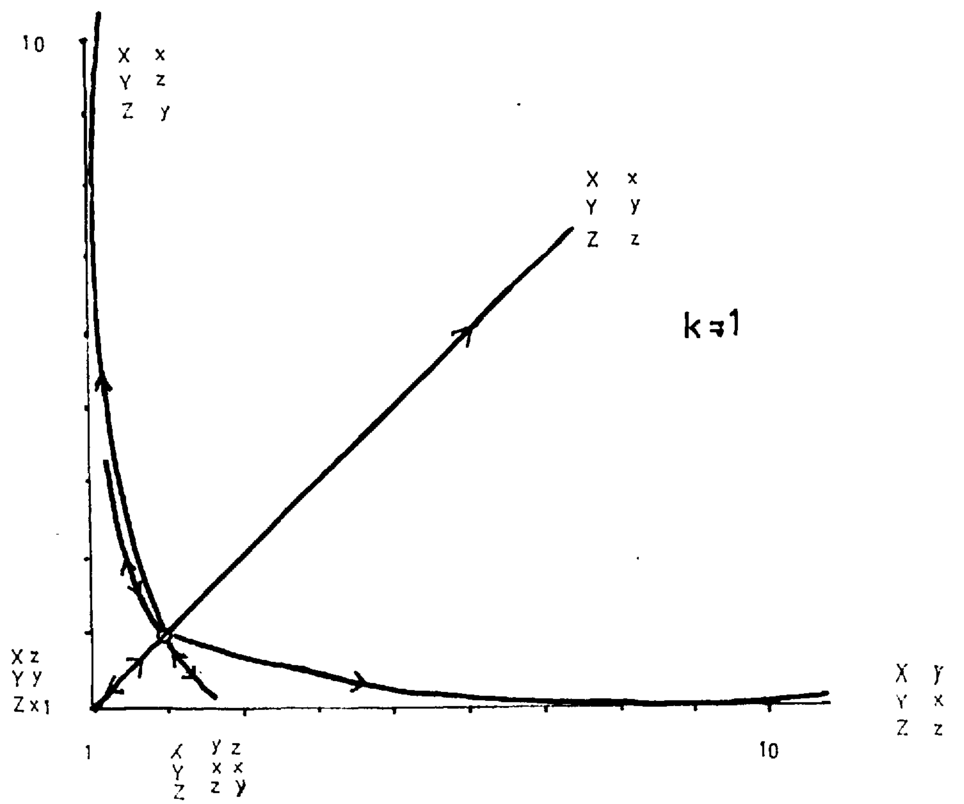

10
X xx
Y yz
Z zy
X z z
Y y x
Z x y
x
y
z
k = ∞
x
y
z
k = 0·5
X y y
Y x z
Z z x
1
10
1

Fig 6


10
X x
Y z
Z y
X x
Y y
Z z
k = 1
X z
Y y
Z x
X y
Y x
Z z
1
X yz
Y xx
Z zy
10

and the reader is referred to Coward (1969) who considers various possibilities in more detail.

6:6 FOLDING

The effect of the folding on the path of the final strain ellipsoid is considerable and complex. I shall consider only one aspect of it, namely the changes in the orientation of strain superimposition which occur as a result of the rotation of the XY plane of the first strain ellipsoid in response to folding induced by the second strain ellipsoid (whose XY plane (xy) is superimposed normal to that of the first).

(A) If, during superimposition in the orientation

$$\begin{array}{cc} X & x \\ Y & z \\ Z & y \end{array}$$

the XY plane is folded and comes to lie parallel to the xy plane of the second strain, superimposition in the fold limbs now occurs in the orientation

$$\begin{array}{cc} X & x \\ Y & y \\ Z & z \end{array}$$

This demands isoclinal folding, if folding is not isoclinal the final strain will take some more complex path. The folding has taken place about the x axis of the later strain.

(B) Superimposition in the orientation

$$\begin{array}{cc} X & y \\ Y & z \\ Z & x \end{array}$$

might produce folding about the y axis of the second strain. In this case, once folding was isoclinal, increments would be added in the limbs in the orientation

X	y
Y	x
Z	z

Similarly, we have

	Hinge		Limbs		
(C)					
X	z		X	y	
Y	x		Y	x	fold about x
Z	y		Z	z	

and

	Hinge		Limbs		
(D) X	z		X	x	
Y	y		Y	y	fold about y
Z	x		Z	z	

Thus, as a result of folding in any of the conditions set out above, the final strain ellipsoid in the fold hinges will evolve along a completely different path from that in the fold limbs. For two strain ellipsoids of $k = 0.5$ the strain path for the final ellipsoid in the fold hinge in (A) travels into the constrictional field and stays there, that for the fold limbs continues along the $k = 0.5$ line (i.e. the final ellipsoid has the same symmetry as the superimposed ellipsoids). In (B) the strain path for the fold hinge is a complex one which travels into the constrictional field (and actually reaches $k = \infty$) with low amounts of superimposed strain, and then returns to the flattening field with further increments of the second strain. The fold limbs in (B) move on a path towards uniaxial flattening. In (C) the strain path for the hinge starts in the flattening field and moves to constriction when the ratios of the superimposed strains are more or less equal, while the final strain in the limbs reaches $K = 0$ and is always in the flattening field. The hinge of (D) follows the most complex path of all, going from flattening to constriction and back again (though it returns only after a great deal of late strain has been superimposed). The limbs follow the $k = 0.5$ path of the superimposed ellipsoids. These various strain paths are the ones shown in Figs. 4 and 5.

6:7 INHOMOGENEOUS FABRIC DEVELOPMENT

In orogenic terrains, superimposed deformations often produce discrete sets of structures, including superimposed sets of mineral fabrics. Most studies of the structural and metamorphic histories of rocks are based on the existence of such structures (e.g. crenulation cleavages developed on earlier slaty cleavages or schistosities). Each superimposed strain is expressed in its own, more or less independent, fabric. This phenomenon appears to be partly a function of the inhomogeneity of the rocks concerned (for they may be strongly banded or laminated), and partly, perhaps, due to a dominance of micas and other platy minerals in the fabrics. A micaceous planarity is perhaps more likely to crenulate than to undergo a more homogeneous mineral reorganisation.

6:8 HOMOGENEOUS FABRIC DEVELOPMENT

In this discussion I shall not consider crenulated (inhomogeneous) fabrics. I shall merely discuss a continuously evolving fabric system which varies between total planarity and total linearity, in response to changing strain conditions i.e. a homogeneous fabric development. It might be supposed that once, say, an amphibolite band is foliated, it becomes comparable with a mica schist and capable of small scale crenulation, but although this type of development is common in the Sound of Harris, it is local on the North Uist coast. The reason for its rarity is probably that the ductility contrast between the amphibolite band and its matrix is greater than that between foliation laminae within the amphibolite; the band deforms inhomogeneously (i.e. it is folded) within its matrix, and though its foliation is folded on a relatively large scale, on a smaller scale it behaves in a homogeneous manner and is continuously modified to reflect the symmetry of the changing strain conditions.

Before proceeding with an analysis of the fabrics of the North Uist coast another theoretical point must be clarified, and both the method of collecting the data and the nature of the fabrics themselves must be described.

6:9 COLLECTION OF DATA

The coarse grained nature of the Lewisian gneisses makes it possible to record the fabric type directly in the field. It is a subjective procedure involving the recognition of the relative importance of planarity and linearity in a rock. Here the data is recorded in a way similar to that suggested by Flinn (1965): "LL" denotes a linear fabric with little or no trace of planarity; "PP" signifies the opposite, namely a planar arrangement containing no linear component. "PL" is used to describe a fabric containing both planar and linear elements (e.g. a foliation defined in by hornblendes which are also aligned as a mineral lineation) P> L and L> P are varieties in which one element is dominant over the other. Fabric data are presented in this form on map 2. Flinn's symmetry concepts allow us to equate an "LL" fabric with a prolate strain ellipsoid and "PP" fabric with an oblate strain ellipsoid. All but uniaxial oblate ellipsoids should be capable of producing some linear element in a fabric since they all contain a specific extension direction, similarly all but uniaxial prolate ellipsoids should encourage some planar element, but to be more specific than this is undesirable because of the subjective aspect of the data.

6:10 CONSTRICTIONAL DEFORMATION

All the constrictional fabrics of the present area occur in rocks which have suffered the superimposition of at least two finite strains at high angles to each-other; they are always intersection phenomena. Usually the oblate symmetry of each of the finite strains is suggested, thus the possibility of single phases of constrictional deformation in the North Uist-Sound of Harris rocks is discounted, and the theoretical discussion of fabric evolution is coined in terms of the superimposition of oblate strains.

Three authors of accounts of regional constrictional deformation (Flinn, 1956, Hossack, 1968 and Watterson, 1968) do not consider the superimposition of oblate strains to be the cause of the finite constrictional strain in their areas, though Hossack (personal communication, 1969) does regard it as a possibility. Two of the studies (those of Flinn and Hossack) were, however, made on deformed conglomerates where, as Ramsay has shown (1967, p.216) an original shape factor or imbrication in the pebbles might give on straining a finite constrictional fabric even though the strain itself has oblate symmetry. Watterson's study was made in a basement terrain which has suffered several important episodes of deformation so that there also the finite constrictional fabrics of the rocks may have been arrived at by strain superimposition.

6:11 THE RELATIONSHIPS OF FABRIC AND LITHOLOGY

The extent to which tectonite fabrics are developed in rocks depends partly on the number of non-equant minerals which the rocks contain. Mica or amphibole-rich rocks take on visible penetrative fabrics more easily than, for example, meta-quartzites. Similarly crystal shape has an influence on the type of fabric preferentially developed; linearity might be somewhat emphasised in some types of amphibolite, but subdued in micaceous rocks where planar fabrics may dominate.

The Scourie dyke amphibolites of the North Uist coast provide excellent material for the study of fabric. Their mineralogy is such that they can accommodate the whole range of fabric types from totally planar to totally linear. Planar fabrics (foliation) in Scourie-Dyke amphibolites is defined by alternating thin (up to 1 mm) and impersistant laminae of hornblendic and felsic material, and by the planar dimensional orientation of hornblende crystals or groups of crystals.

Linear fabrics are produced by the dimensional orientation of hornblende crystals or groups of crystals, and sometimes also in streaked out hornblende-fëlspar aggregates which represent retrogressed garnets.

Total linearity may develop from an earlier foliation with the breakup of the planar arrangement of minerals and the disappearance of the compositional lamination. In only one locality on the North Uist coast (Dreary Point) is there evidence of a new planarity imparted to a previously wholly linear amphibolite. Here the planar fabric is a dimensional orientation of crystals only; there is no lamination since the felsic constituents are randomly distributed through the rock.

The variability of the acid gneiss makes fabric data from it more difficult to use, moreover, inherited Scourian fabric elements such as the banding of "rough" gneiss probably have considerable effect on the finite fabric. Planarity in acid gneiss is expressed in banding and by mica or hornblende orientation, linearity in rodding structures, elongated augen, and poorly defined linear mineral fabrics. The two main sorts of acid gneiss on the North Uist coast, "rough" gneiss and "Smooth" gneiss have been described in Chapter 1. The fabric development in each of them is very different and a description of it takes up part of the next section.

6:12 A FABRIC ANALYSIS OF THE NORTH UIST COAST

6:12:1 Introduction

The earliest fabric in the Scourie-dyke amphibolites of the North Uist coast is a foliation of F1 age. This fabric is deformed by F2 and F3 folds, and the following study is an analysis of the homogeneous modifications of this foliation which occur in different positions in F2 and F3 structures. Because of the co-axial nature of the deformation in North Uist it is not possible to tell whether or not this first foliation contained a lineation of strictly F1 age or whether linear elements of the F1 fabric, even in areas where it seems almost unmodified, are the product of later strain.

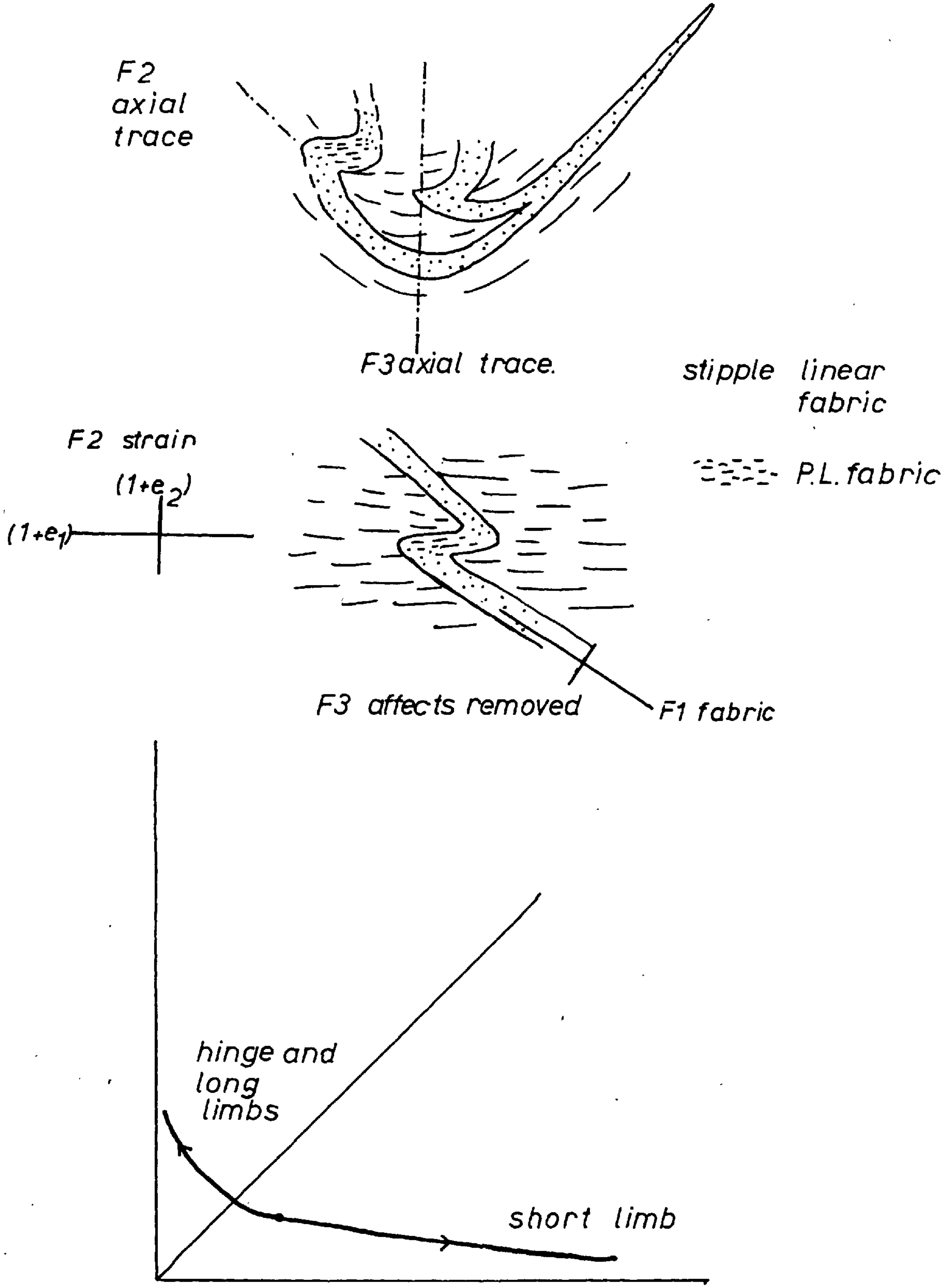
An illustration of the dependence of fabric variations on the superimposition of the finite strains of successive deformation phases is given below. The time relationships of the strains are clearly demonstrated by the presence of garnet (see also Chapter 5 , where the same locality is used as an example of time relationships in metamorphism).

6:12:2 Amphibolites in the Griminish Synform

A Scourie-Dyke amphibolite band containing an asymmetrical F2 fold lies in the limb of the F3 Griminish Synform. In the short limb of the F2 fold, small garnets post-kinematically overgrow and F1 planar fabric. This foliation is flattened around the crystals to some extent, but only the rims of the garnets show signs of retrogression. In the long limb and the hinge of the F2 fold, however, the fabric is totally linear and the garnets are retrogressed to linear felspathic streaks. The time relationships are:

i) F1 deformation; formation of a P.L. fabric in the dykes.

Fig 8 Scourie dyke amphibolites in the Griminish Synform



ii) Post-kinematic growth of garnet

iii) F2 strain; folding, retrogression of garnet and development of linear fabric in the hinge zone and in one fold limb.

It might seem that the short limb has suffered less visible internal strain, and the old fabric elements remain, whereas in the long limbs the superimposition of the F2 strain has modified the old fabric considerably and caused the development of a constrictional fabric. More probably, however, the two limbs have suffered similar amounts of F2 strain, but the manner of its superimposition was different in each, the old planar fabric being re-inforced in the short limb, but destroyed in the steep limb. This effect might be achieved if the short limb was rotated more rapidly towards the XY plane of the F2 strain ellipsoid, while the long limb remained for a longer time in an orientation where superimposition of F2 strain produced a constrictional fabric.

6:12:3 <u>Other parts of the North Uist coast</u>

The discussion is divided into three parts, first a description of fabric variations which are probably the result of superimposition of F1 and F2 strain only, second a discussion of the significance of these variations, and third a discussion of the effects of F3 strain on the fabrics.

The following localities occupy structural positions such that the effects of F3 strain can be discounted (e.g. they lie in the broad hinges of F3 antiforms).

(a) The Cliffs Antiform

In this region the amphibolite bands are deformed by F2 folds (the area lies in a major F2 fold zone). The F1. P.L. fabric of the amphibolites is folded, and there is a tendency for linearity to increase in the fold hinges. The homogeneous "smooth" acid gneiss which surrounds the amphibolites, however, is almost totally linear but it gives way to banded "rough" acid gneiss in which P > L

(b) Aird and Runaird (south-west of Hougharry)

On this headland parts of both an F2 limb zone and of an F2 fold zone are exposed (see chapter 1). Fabrics in the limb zone are P > L, while those in the fold zone have a stronger linear element (L = P).

(c) The western limb of the Griminish Antiform

Although this is not an antiformal hinge, evidence of F3 deformation is not apparent, and the fabric can be considered as a product of F1 and F2 strains only. There is some evidence that the F2 strain was more intense here than in the cliffs, for a gneissic cleavage is developed axial planar to F2 folds in the "rough" gneiss of Griminish, though not in the "rough" gneiss of the Cliffs, and F2 boudinage occurs at Griminish, but is not so well developed in the Cliffs.

A strain ellipsoid determined from deformed veins here effectively records the F2 strain because all the folds in the veins are of F2 age. The strain ratios of this ellipsoid are of the order of 6:3:1 where $k = 0.5$, i.e. the F2 strain is oblate. The amphibolite fabric, however, has a strong linear element (L > P) whereas the rough gneiss is planar (P > L).

(d) Dreary Point (the hinge of the Griminish Antiform

Here the Scourie-Dyke amphibolites are either totally linear (L.L.) or show traces of a new F2 planarity parallel to the axial surfaces of F2 folds (a planar arrangement of hornblendes without alternating felsic layers). The old banding of the rough acid gneiss is highly modified and the rock has a stronger linear appearance than it does anywhere else on the coast. The F2 deformation is extremely intense (see Chapter 1) causing F1 boudins to be strongly flattened normal to their original length.

6:12:4 Discussion of the effects of F2 strain on the fabric

From the variations described above it appears that where F2 deformation is relatively low, the F1 fabric in the amphibolite bands remains more or less unmodified in the F2 fold limbs, though linearity increases in the fold hinges. "Smooth" gneiss is linear under these condition, "rough" gneiss is planar. With increasing F2 strain, fold hinges in amphibolite bands show totally linear fabrics, until with very high strain indeed a new planarity starts to form parallel to the axial surface of F2 folds.

The least modified pre-existing (F1) fabrics seen are PL; the F2 strain in each of the areas above was probably oblate, a suggestion indicated by strain determination in only one area but supported by the existence of F2 boudins and F2 gneissic cleavages in inhomogeneously deformed "rough" gneiss.

If fabric symmetry reflects the symmetry of the final strain ellipsoid these fabric variations could reflect different stages along strain paths produced by the superimposition of oblate F1 and F2 finite strains in either the orientation :-

X	z (a)	or the orientation	X	y, (b).
Y	y		Y	z
Z	x		Z	x

The strain paths are shown in Fig.4.

In both alternatives, linear, and with increasing F2 strain, planar fabrics are produced in fold hinges while planar or lineated planar fabrics would be maintained in fold limbs where superimposition would approach either the orientation

X	x (a)	or the orientation	X	y (b).
Y	y		Y	x
Z	z		Z	z

Similar principles probably apply, on a larger scale, to limb zones and fold zones.

The first possible path above (a) implies that the F1 lineation was parallel to the Y axis of the F1 strain, the second that it was parallel to F1 X. The first path demands a great deal of F2 strain to bring it back into the flattening field, and initially it spends considerable time in the flattening field before it enters the constrictional field. The second spends more time in constriction but also requires a large F2 strain to re-enter the field of flattening. This may be summarised as follows:

(a) $F1 > F2$:- flattening (b) $F1 > F2$:- constriction
$F1 = F2$:- constriction $F1 = F2$:- constriction
$F1 < F2$:- flattening $F1 < F2$:- flattening

When the strain path (a) enters constriction, y (the intermediate axis of the F2 strain ellipsoid) becomes X_1, the longest axis of the final ellipsoid; until then X (z) = X_1. In strain path (b) X (y) is always X_1.

It is difficult to decide which of these paths best fits the field evidence. The second is probably the better because the first F2 increments send this path into constriction, and the first observed fabric modification in response to the F2 strain is an increase in linearity. It might be argued that the effects of early strain increments have been masked by those of the later increments, and that nowhere does evidence remain of an early excursion of the final strain into the flattening field, but even so the second path (b) is preferred.

The $k=0.5$ value used in Fig. 4 for the F2 strain is similar to that recorded in a determination of the F2 strain at Griminish, but there is no similar quantitative information available for the F1 strain, nor, therefore for the final finite strain. The $k=0.5$ value for the F1 strain used in Fig.4 is merely an arbitrary value.

The field evidence presented above shows that fabric variations occur in different rock types as well as in different areas. The fabric of the amphibolites in the Cliffs has not achieved total linearity, whereas that of the adjacent "smooth" acid gneiss is nearly completely linear, and that of the "rough" gneiss planar. At

Griminish a higher degree of linearity in the amphibolites is associated with still planar "rough" gneiss (smooth gneiss is unfortunately absent here). On Dreary Point "rough" gneiss begins to appear linear when the amphibolites are beginning to take on a new planarity. Obviously the fabric development of different rock types takes place at different rates, "smooth" gneiss being the most ready to attain a new linearity, and "rough" gneiss the least ready. Fig.9 summarizes these relationships as stages along the strain path produced with superimposition in the orientation

X y
Y z
Z x

A possible explanation of the differences is that the "smooth" gneiss was poorly banded and fairly homogeneous before deformation commenced, whereas the amphibolites carried a well developed F1 planar fabric, and the "rough" gneiss was extremely strongly banded - indeed it often suffered inhomogeneous fabric development in response to F2 strain (i.e. it developed a new gneissic cleavage).

A further complication in this study concerns the respective orientations of the fabric and the F2 strain ellipsoid at constructed Griminish, for the linear element of the fabric is not parallel to any axis of this ellipsoid. Fig.10 restores the F2 ellipsoid to its most likely original orientation (its orientation before large scale F3 re-folding) and attempts to show how the superimposition of the F1 and F2 strains might have occurred. The diagram is most relevant as an

Fig 9 *Fabric variations due to superimposition of F1 and F2 strains.*

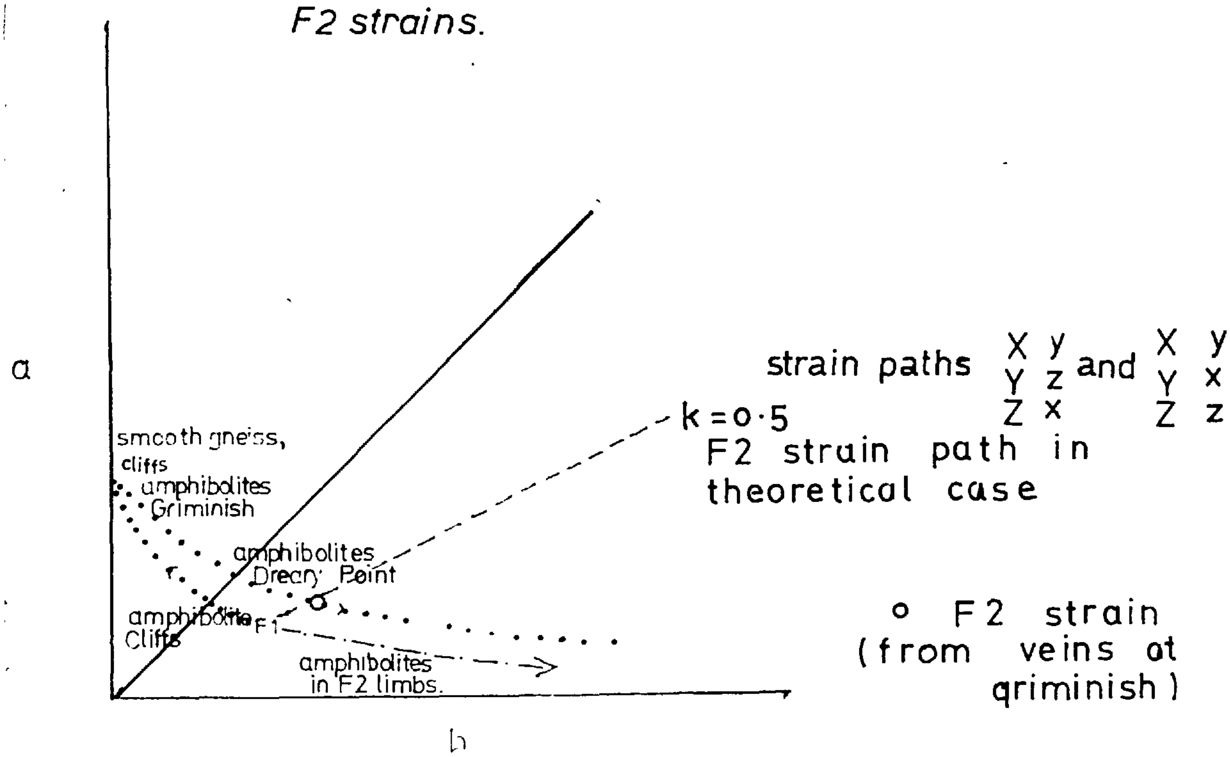


Fig 10 Strain superimposition at Griminish

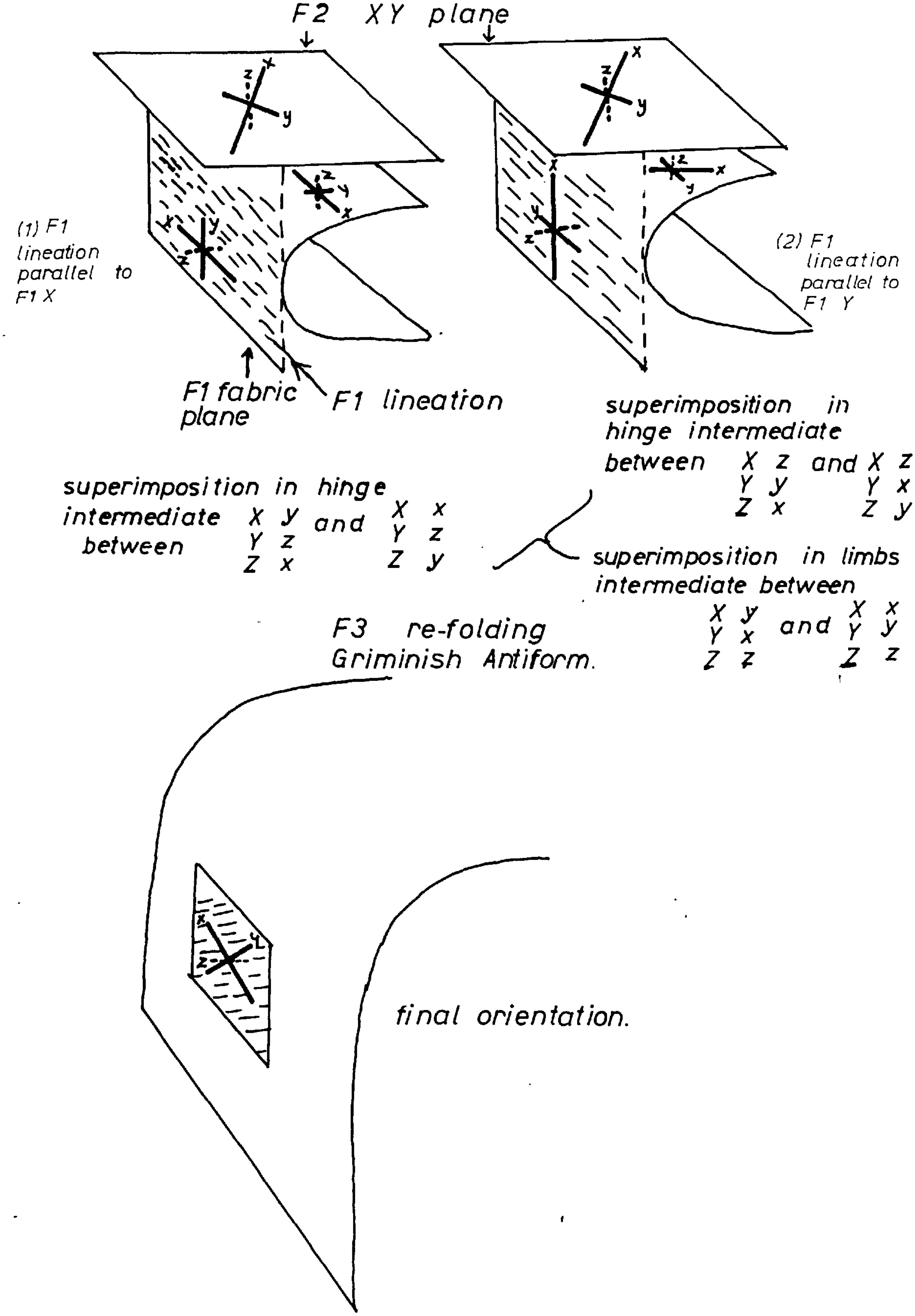


illustration for the concluding discussion of this chapter, for the present it serves to show how the natural strain paths may have been much more complex than any shown on fig 4, and that in reality superimposition of the F1 and F2 strains in F2 fold hinges at Griminish probably took place in an orientation either intermediate between

X	y		X	x
Y	z	and	Y	z
Z	x		Z	y

or between

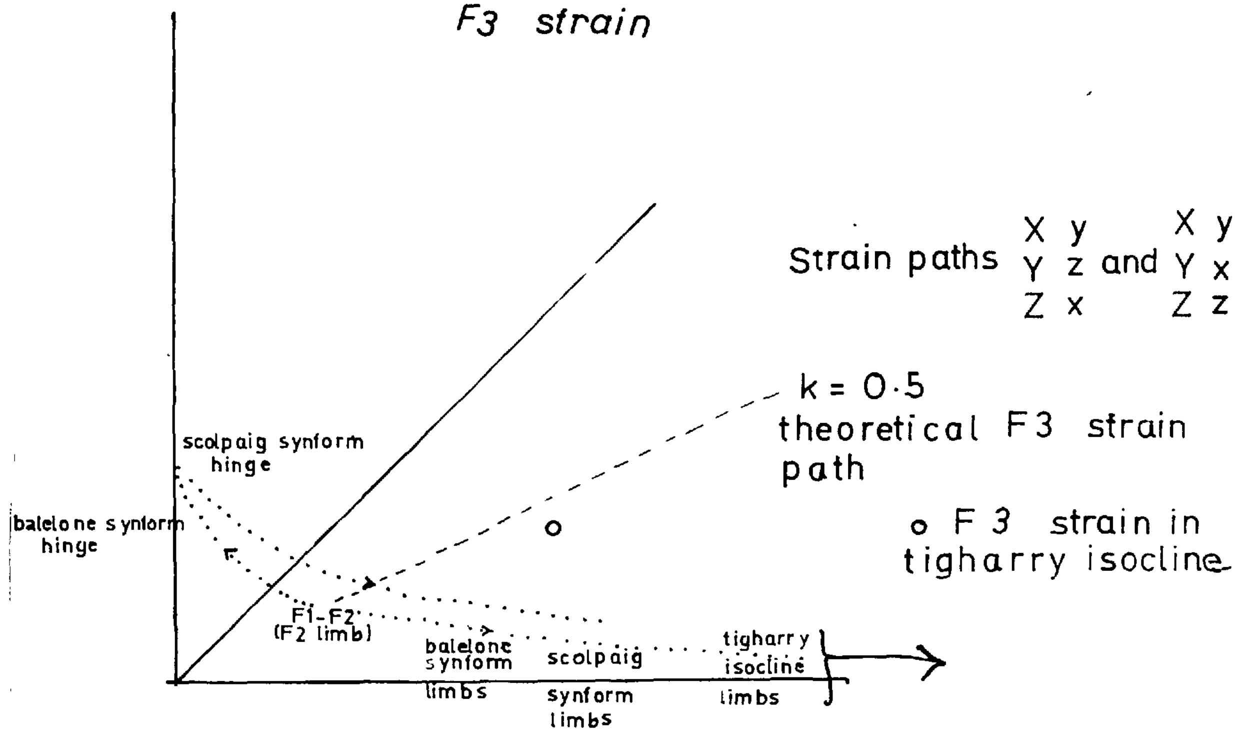
X	z		X	z
Y	y	and	Y	x
Z	x		Z	y

6:12:5 <u>The effects of F3 strain</u>

The superimposition of a third finite strain (that of the F3 deformation) with its XY plane normal to the F2 XY plane (see Chapter 1) also affected the fabrics of the North Uist coast.

F3 strain could have been superimposed either on F2 limb zones with their modified F1 PL fabrics, or on F2 hinge zones where fabrics in amphibolite bands were more linear. Oblate F3 strain superimposed on an earlier oblate fabric would cause the finite strain to take paths similar to those described above for the interaction of F1 and F2 strains; oblate F3 strain superimposed on an earlier prolate fabrics would produce strain paths such as those in Fig.7. On the north Uist coast, however, there are no examples of the significant F3 straining of an F2 fold zone except, perhaps in the extreme south-west part of the Scolpaig Synform (see

Fig 11 Effects of superimposition of F3 strain



Chapter 1). All the intense F3 deformation takes place on earlier limb zones, and amphibolite fabrics often become more linear in F3 fold hinges. This can be seen in the core of the Balelone Synform, and much more clearly in the large biotitic amphibolite in the hinge of the Scolpaig Synform where fabrics are LL. Amphibolites in the limbs of both folds are PL. These increases in linearity could result from any strain path produced by superimposition of strains whose XY planes were at right angles, and we cannot isolate any particular strain path for the fold hinges with any degree of validity. The strain path shown in Fig. 11 is once again that produced by superimposition in the orientation

X	y
y	z
Z	x

The Tigharry Isocline is an area of very intense F3 strain and the orientation of deformed veins there has given a strain ellipsoid with ratios 18:6:1, k=0.4. The isoclinal nature of this fold means that the whole of the gneiss succession on the narrow strip of exposure across the structure is effectively in the fold limbs. Both the "smooth" gneiss and the amphibolites carry well developed PL fabrics of F1 - F2 age(the Tigharry Isocline is situated on an F2 limb zone) which are overgrown by post kinematic garnets (see Chapter 5). The planar fabric is flattened around the garnets by the F3 strain which presumably had little effect on the symmetry of the fabric because of its superimposition in either the orientation

X	x
Y	y
Z	z

or the orientation:-

X	y
Y	x
Z	z

The obliquity of the F3 principal strains with the linear component of the fabric (see below) suggests, in fact, that the real orientation of F3 strain in the Tigharry Isocline was intermediate between these two.

The fact that the linear component of the fabric in the Tigharry Isocline (and at Griminish) is oblique to any of the principal strains constructed from vein orientation in those areas is an important point which now receives detailed consideration. It is evident that no new mineral lineation records the X axis of the F3 strain in the Tigharry Isocline as none records the X axis of the F2 ellipsoid at Griminish. Nor is there in either area any evidence of the deformation of older linear elements, or of any rotation of these towards the new X directions. These apparent anomalies might be explained by the suggestion that recrystallization or rotation of minerals takes place more slowly than the changes of orientation of the principal strains, but a more realistic explanation makes up the concluding discussion of this chapter.

CONCLUDING DISCUSSION: THE PROBLEMS OF MINERAL LINEATIONS, CO-AXIAL DEFORMATION AND CONSTRICTIONAL STRAIN.

One of the basic premises of the foregoing study was that planar fabrics record the XY plane of the strain ellipsoid of the deformation which produced them. There is abundant evidence for this assumption, and its validity is enhanced by almost every study of deformed objects in rocks (e.g. Cloos, 1947). The significance of the orientation of mineral lineations, however, has not been established with the same degree of certainty. Flinn (1956, 1965) and others have assumed that mineral lineations always occupy the X direction of the strain which produced them, but the frequent parallelism of fold axes and mineral lineations in many basement areas casts some doubt on the validity of this assumption in basement rocks.

Excluding mineral lineations for the moment, there seem to be two main types of linear features - those which directly record an extension direction in the rocks (e.g. stretched fossils and the long axes of deformed ooids), and those which are essentially intersection phenomena. "Intersection lineations" can be produced by the intersection of two material planes or, as in the case of fold axes, by the intersection of a strain plane on a pre-existing material surface.

As we saw earlier, co-axial deformation does not necessarily require the co-axial superimposition of strain ellipsoids, but it does require the still rather special case that the XY plane of the superimposed strain contains the linear elements of the surface on which it is superimposed. These linear elements are then parallel to the line of intersection of the two planes, about which folding of the original surface takes place. Thus, one possible explanation of parallel mineral lineations and fold axes is that the two structures are of different ages - the lineations are older and did not develop in response to the same strain conditions as the folding. Their parallelism is simply a function of co-axial deformation.

Often, however, there is good evidence that coeval mineral lineations and fold axes are parallel, as, for example in the Cliffs of the North Uist coast, where linearity increases in F2 fold hinges. If these lineations are taken as records of the X direction of the strain which produced them the implication is that fold axes often form parallel to the direction of principal extension of the rocks. Since fold axes occupy no specific orientation within the strain ellipsoid but merely form in the original rock layering, we must assume that either the principal strains were added in the rock layering, or that rotation of linear elements towards the X direction is commonly complete, a situation only possible under infinite strain. Most of the strain determinations made in the present work (all of them except that in the Hougharry Antiform of the North Uist coast) indicate that new principal strains are not always added parallel to the rock layering, for, in all except this case, the principal strains are oblique to the linear elements of the rock. Furthermore, if fold axes parallel to X directions were common, we might expect to see the long axes of boudins normal to fold axes, or perhaps some other sign of physical stretching in the axial direction. In the present area at least, such features are never seen. For the moment, therefore, let us assume that mineral lineations do not necessarily record an X direction and examine the problem from its first principles.

The feature common to both mineral lineations and fold axes is that they are both contained in the rock layering. Well documented accounts of crystals growing through well developed rock layering are, with the exception of descriptions of post kinematic garbenschiefer, rare in the literature, and in the North Uist-Sound of Harris area the phenomenon is unknown. This suggests that layering has as much, if not more, influence on the development of mineral fabrics as do the strain conditions. If the strain which creates the co-eval axes and lineations is oblate, then some degree of extension will occur in all directions within the XY plane. It may be that mineral lineations develop in a "compromise" position with respect to the extension direction while still lying within the rock layering, i.e.

they may grow along the line of intersection between the XY plane and the layering.

The necessity for a planar surface in which mineral lineations can develop might explain the different significance of the mineral lineations in basement terrains on the one hand, and slate belt areas on the other. In slates, deformed objects commonly show that the mineral lineation known as "grain" truly records the principal extension direction. It is able to occupy this orientation because a new penetrative planar fabric (the cleavage) is developed parallel to the XY plane of the strain ellipsoid, and the control of any old layering is lost. The orientation of the new cleavage is such that the lineation can grow both in a surface and parallel to X. In gneisses, new penetrative planar fabrics with this orientation are not often developed, and the old layering (the gneissic banding)maintains its control on mineral growth. Thus, apparent "axial extension" is confined to gneissic (deep seated) environments, whereas in deformed rocks high in the crust the angle between fold axes and mineral lineations is exceedingly variable and commonly very large.

To summarise, then, the factors governing the orientation of a mineral lineation are:

i) The crystals must lie in the rock layering.

ii) The crystals must lie in the extension field of some strain ellipsoid i.e. mineral lineations grow within the layering as close as possible to the maximum extension direction.

These concepts immediately make it necessary to examine the effects of constrictional strain in the same circumstances. The extension field of a prolate ellipsoid is restricted and if crystals are to grow in it their elongation must lie close to the X direction. If layered rocks are subject to a phase of constrictional deformation so that X lies close to or in the layering, folds will develop on axes close to X and (theoretically rotate towards X as the strain progresses. Axial planes should fan about X; mineral lineations should be parallel to the fold axes. Alternatively, if the layering is at a high angle to X it should fold in all directions and the distribution of axes and axial planes should be random. This situation is one which might cause minerals to grow actively through rock layers, and the rare documented examples of the phenomenon (e.g. Clough in Deach et al, 1907) might be due to it, though I know of no published description of associated random structures. If, however, a finite constrictional strain is produced by the superimposition of oblate strain ellipsoids, the lack of random structures is not surprising. With the XY planes of the superimposed ellipsoids at right angles producing a prolate ellipsoid with its X axis at a high angle to the layering we would have two sets of folds with axes at right angles to each other (and dome and basin type interference). If on the other hand the strains were superimposed so that X of the finite prolate ellipsoid lay close to the layering we should see co-axial interference with mineral lineations parallel to fold axes. Since these are the commonest structures to be associated with LL fabrics we have a further line of evidence supporting superimposition of oblate strains as the most important cause of finite constriction in rocks.

CHAPTER 7

TWO EXAMPLES OF THE USES OF STRAIN ELLIPSOIDS DETERMINED FROM THE ORIENTATION OF DEFORMED VEINS IN STRUCTURAL ANALYSIS.

7:1 INTRODUCTION

This chapter contains two detailed studies of the use of the strain ellipsoids determined from the orientation of deformed veins (see Chapter 4) as aids in the interpretation of complex geological phenomena. The information is used in the way recommended in chapter 4, i.e. confidence is placed in the predicted orientations of the principal strains while strains ratios are used cautiously and in a comparative manner. The topics under discussion in this chapter are firstly deformed lineations in the Sound of Harris steep zone, and secondly, the Tigharry Isocline, the major strain inhomogeneity of the North Uist coast. Sheet 1 in the appendix contains much relevant data.

7:2 THE DEFORMED LINEATIONS OF THE SOUND OF HARRIS STEEP ZONE

7:2:1 Statement of the Problem.

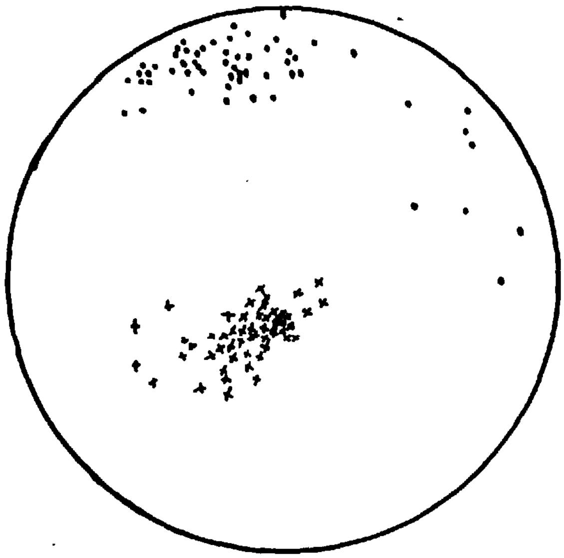
As we have seen in chapter 2, deformation in the Sound of Harris is not always co-axial. There is often a small (about 10°) angular discordance between the F1 and F2 lineations, and both these structures have locally suffered a very profound later deformation of either late F2 or F3 age. This section attempts to analyse this later deformation, and to that end the F1 and F2 linear elements (fold axes, felsic rods, intersection and mineral lineations) are grouped together as "early linear elements", or "early lineations".

In all the islands of the Sound of Harris except Boreray the orientation of these early linear elements varies in the plane of the foliation. On the scale of the individual islands the lineations fall on a partial or complete stereographic girdle about a fairly constant cluster of poles to foliation (Fig.1). Examined more closely, however, the manner of this variation is seen to be systematic (Figs. 2, 3 and 4). In western Pabbay as far east as the F2 fold zone FZ1 (area A on Fig.5) the plunge of the early lineations varies between gentle to the SE and moderate to the NW. In western Berneray it is very gentle to the NW, while in the central part of that island, on strike with western Pabbay, it is gentle to the SE.

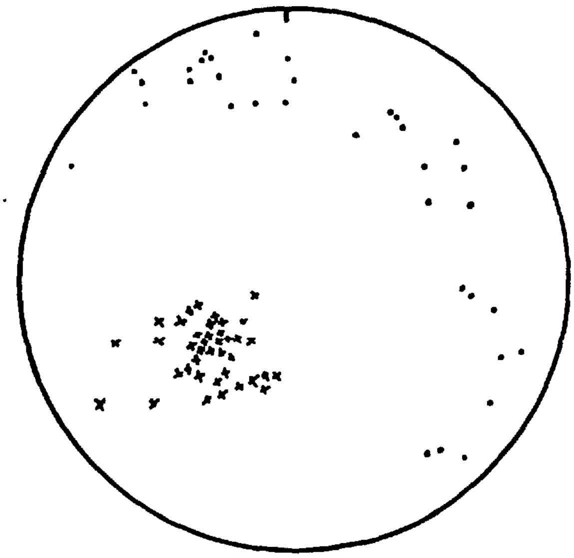
The fold zone FZ1 in Pabbay (B (i) on Fig.5) is developed in amphibolites and "pink and blue" gneisses in which the felsic veins are thick and close together (see Chapter 4). Here the early lineations plunge very steeply to both the NW and the SE. In the apparent continuation of this fold zone in south east Pabbay and Berneray, where the thickness of the felsic veins in the "pink and blue" gneisses is much less, the early lineations are strongly deformed in the foliation planes on a very small scale (plate 1).

East of this fold zone there is, in both Pabbay and Berneray, an area of generally gentle north-westerly plunges, and here exposure in Berneray ceases. In Pabbay, the major F2 fold zone FZ2 (D on Fig.5) has steeply

Fig 1 (also see Sheet 1 in the pocket)
Deformed early linear elements.

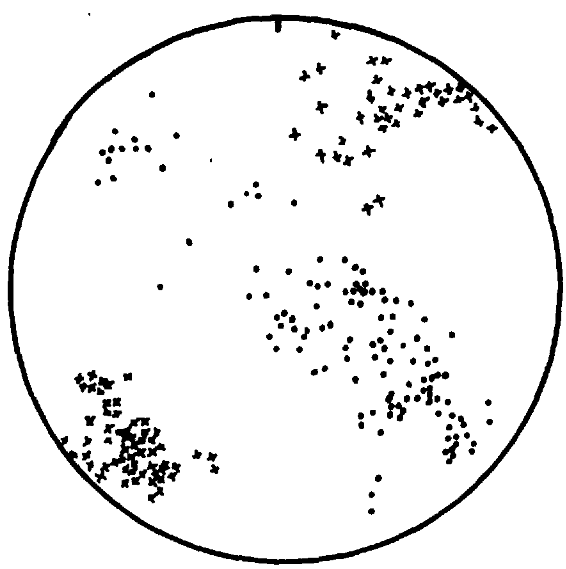


KILLEGRAY



ENSAY

Dots - early linear elements
Crosses above - poles to foliation
below - poles to axial planes of F2 minor folds.



PABBAY

Lineations are undeformed in most of Bemeray

Fig 2 Early linear elements

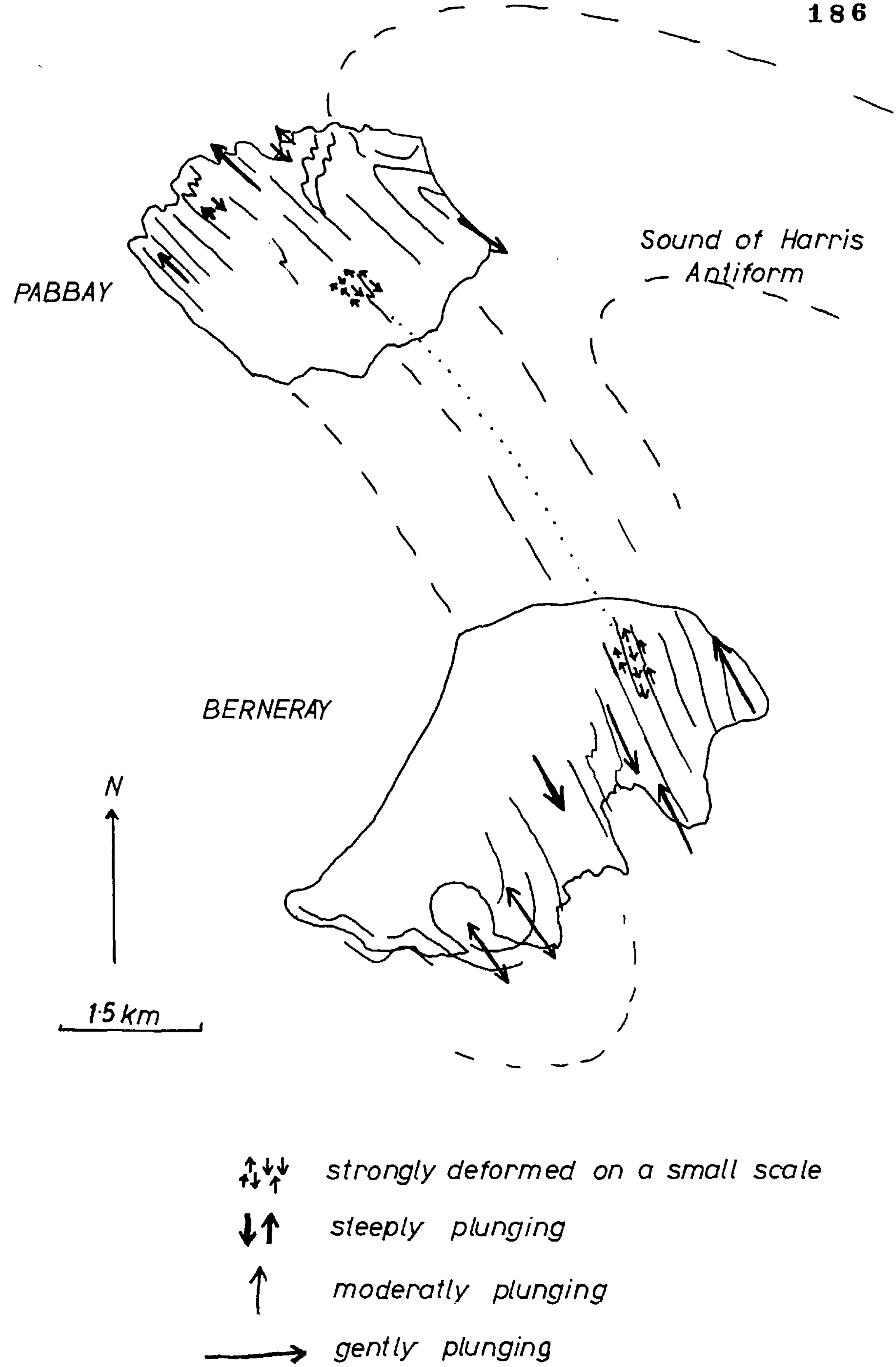


Fig 3 Schematic diagram of deformed lineations in Pabbay and Berneray

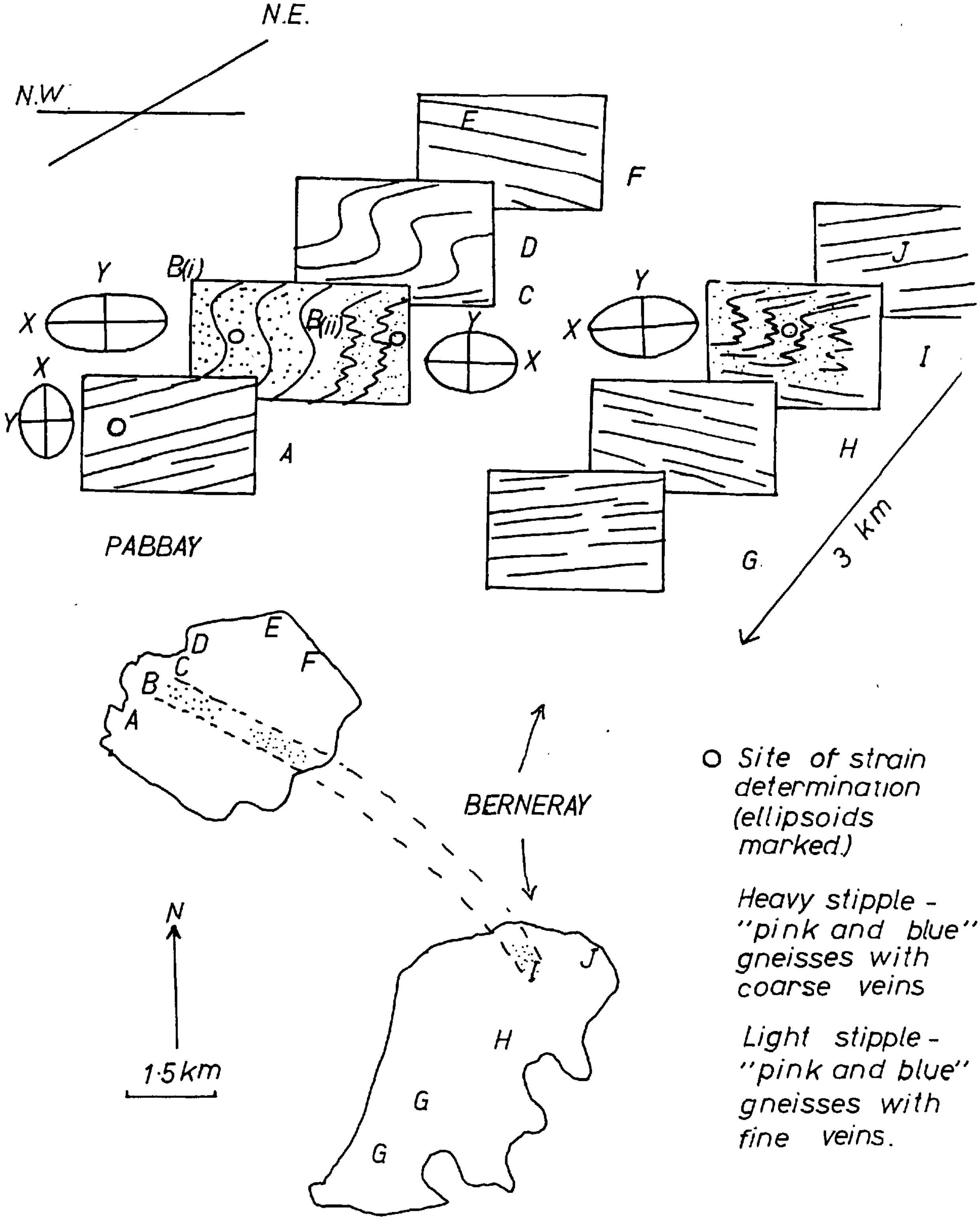
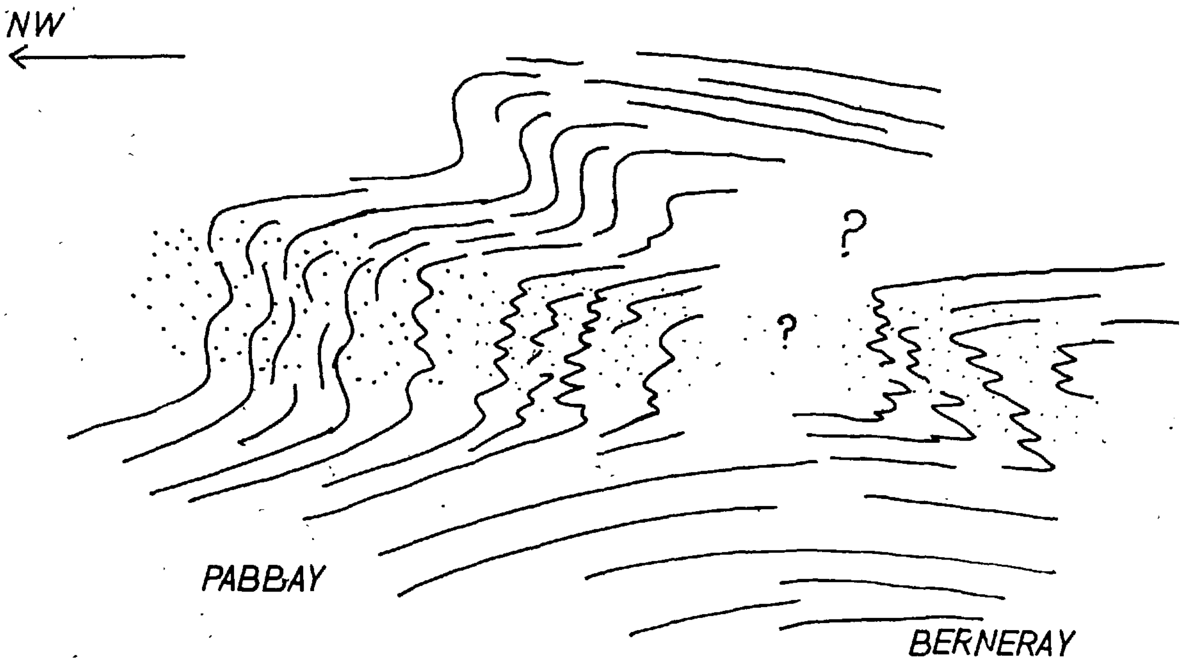
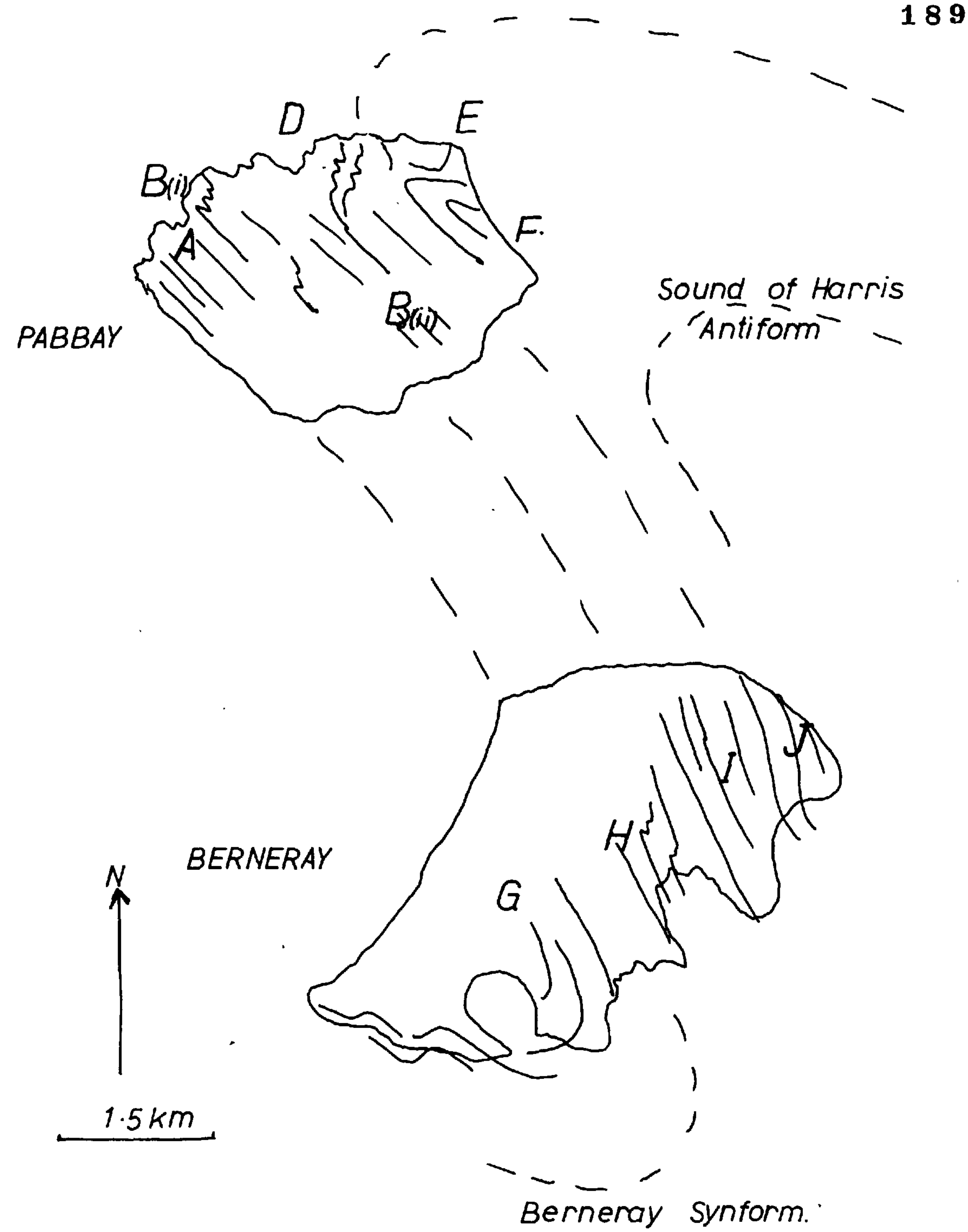


Fig 4 *Highly schematic diagram of deformed early lineations, (looking on an aggregate of foliation surfaces)*



"Pink and blue" gneisses stippled (coarsely banded - coarse stipple, finely banded, fine stipple).

Fig 5 Locality map



Lineations deformed in the plane of foliation,
Bruist, Berneray.

plunging early lineations while further east, in the Pabbay deformation low (see Chapter 2) and adjacent areas there is a gentle or moderate plunge to the SE.

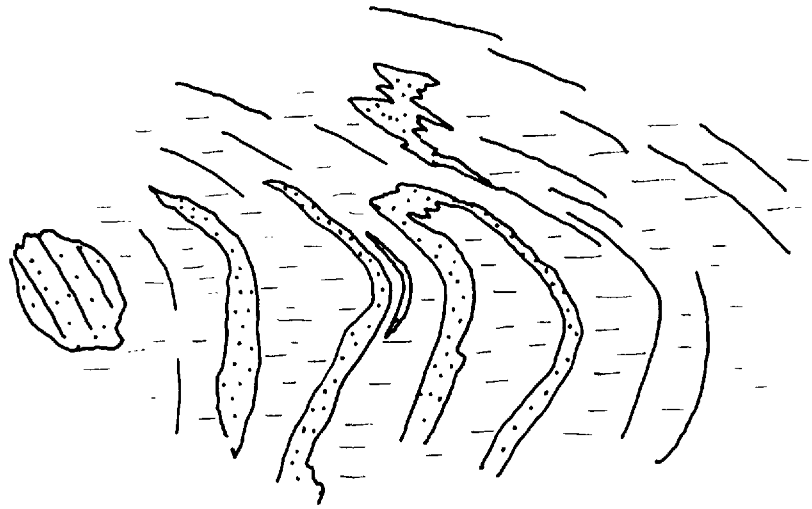
Unlike Pabbay, Berneray possesses no large areas of steeply plunging early lineations - they either plunge gently or moderately one way or the other, or are deformed on a small scale. Figs. 2 and 3 summarize all this information.

Cutting the deformed early lineations is an almost horizontal alignment of hornblende and biotite (Fig.6). It is developed only where the early lineations are steeply plunging or deformed on a small scale. It lies parallel to the "axial traces" of early lineations deformed on a small scale, and may well be co-eval with the deformation of these structures (Fig.6).

7:2:2 <u>Strain Determinations in Pabbay and Berneray</u>

The strain determinations relevant to this discussion were made from the orientation of deformed veins in four localities. Two of these are in N.W. Pabbay, one in the fold zone FZ1, and one just to the west of this. Another is in S.E. Pabbay, and another in Berneray, near Bruist. These localities appear on appendix sheet 1 as NW Pabbay (1), NW Pabbay (2), SE Pabbay, and Berneray respectively, and correspond to localities A, B (i), B(ii), and I on Figs. 3 and 5. In all these areas the rocks are "pink and blue" gneisses, the felsic veins being thicker and closer together in the localities of N.W. Pabbay than in the other two areas. In all localities except NW Pabbay (1) the early linear elements are obviously deformed.

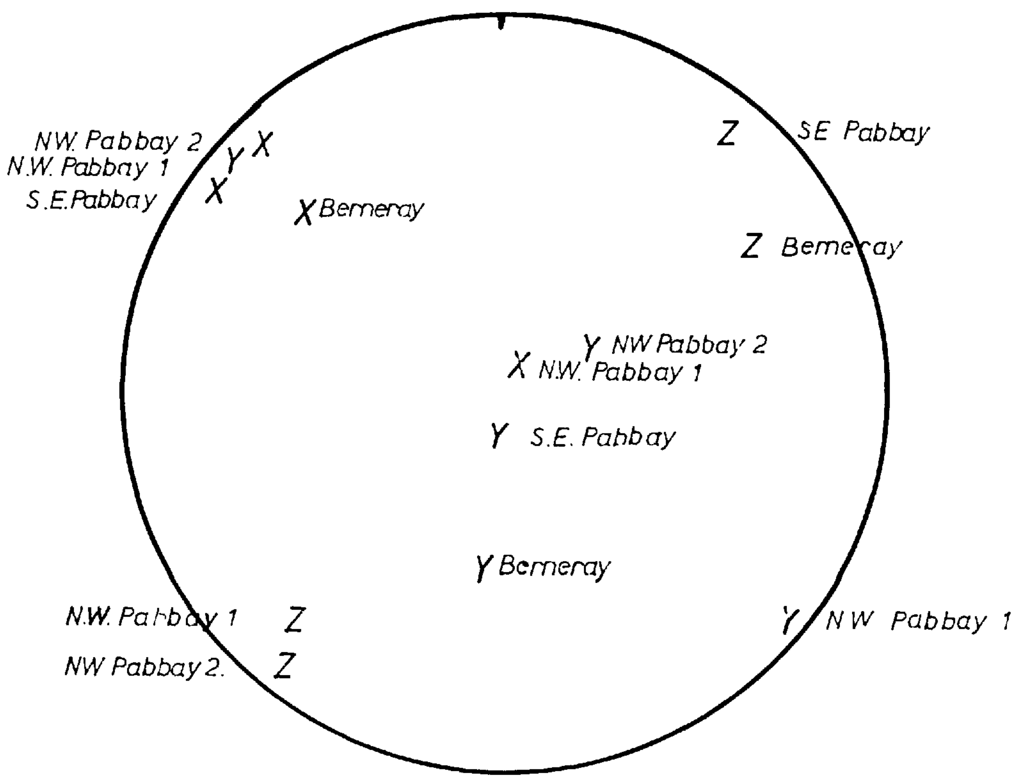
Fig 6 Deformed early lineations (including felsic rods (stippled)) cut by late mineral lineation, Bruist, Berneray, looking on the foliation surface.



10cms

traced from plate 1.

Fig 7a Orientation of principal strains, Sound of Harris, as determined from the orientation of deformed veins



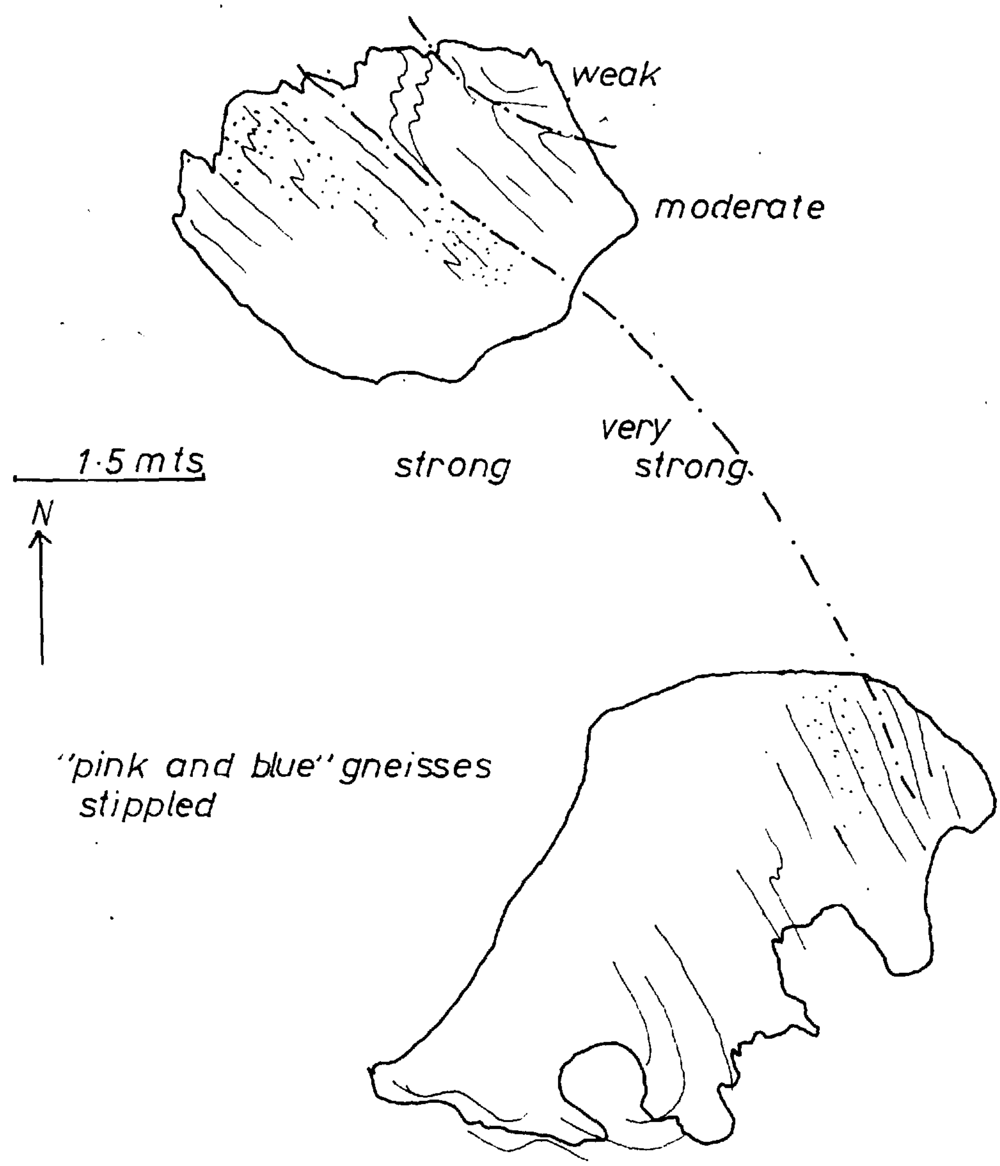
All the folds in the deformed veins from which strain ellipsoids were constructed are of F2 age, and thus at first sight the strain ellipsoids are apparently dominantly records of the F2 strain. The closeness of the veins in N.W. Pabbay has produced harmonic folding in that area, and the computed strain ratios are therefore considered to be unduly, probably ludicrously, low (see Chapter 4). The same criticism applies to both sets of data from N.W. Pabbay, however, so that comparisons between the two sets may be reasonably valid. Such a comparison shows that the strain ratios at N.W. Pabbay (2) (i.e. within the F2 fold zone FZ1) are significantly higher than those of N.W. Pabbay (1), and, although the axes of the two strain ellipsoids have similar orientations in each locality (one axis nearly horizontal NW-SE, one nearly horizontal NE-SW, and one nearly vertical) the relative values of the axes are interchanged. The Y axis of the strain ellipsoid for N.W. Pabbay (1) becomes the X axis of the strain ellipsoid for N.W. Pabbay (2), and vice-verca; Z remains the same in each. The k value of the ellipsoids in N.W. Pabbay is more or less the same (k= 0.5).

Much higher strain ratios occur in S.E. Pabbay and Berneray, and the k value of the ellipsoids here is lower than it is in either of the N.W. Pabbay localities. The orientation of the principal strains is more or less the same as in N.W. Pabbay (2), i.e. X plunges gently to the NW, Y is steep, and Z plunges gently or (in Berneray) moderately to the N.E. This information is a summary of the most likely orientations of the various principal strains and of the most realistic (favoured) strain ratios presented on appendix sheet 1. It is tabulated on Fig.7.

Fig 7b Comparison of (favoured or average) strain ratios

N.W. Pabbay 1	-	2 : 1·5 : 1	k = 0·6
N.W. Pabbay 2	-	4 : 2·5 : 1	k = 0·5
S.E. Pabbay	-	7 : 5 : 1	k = 0·2
Berneray	-	12 : 6·5 : 1	k = 0·3

Fig 8 Subjective assesment of deformation

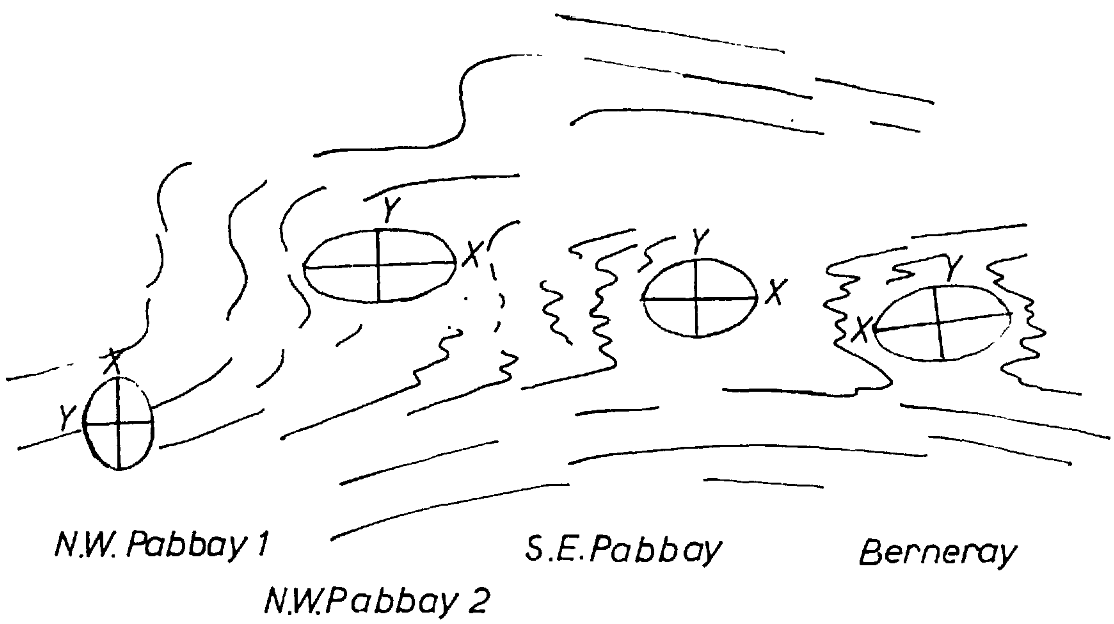


It must be remembered that if, as it would seem from the dominance of F2 folds in the deformed veins (Chapter 4), the strain ellipsoids primarily record the F2 strain, the present orientations of the ellipsoid axes are, because of re-orientation by large scale F3 folding, different from their original orientations. The original orientation of the axial surfaces of the F2 folds (and therefore the XY plane of the F2 strain ellipsoid) can be approximately reconstructed from the enveloping surface of the major F3 folds, by rotation of the early structures about the axes of the late folds (see Chapter 2). Since the horizontal axes of the constructed strain ellipsoids is more or less parallel to the axes of the major F3 folds (see later), rotation of the other two axes about this one until the XY plane of the ellipsoid has a N.E. - S.W. strike gives the original orientation of the ellipsoids.

7:2:3 Correlation of the data from strain determinations with the pattern of deformed lineations.

The variations in orientation, symmetry and magnitude of the strain in these various localities may have some relationship with the pattern of deformed lineations. The strain ellipsoid with the smallest strain ratios and a horizontal Y axis (Pabbay NW1) is associated with regular, gently plunging early linear elements. A strain ellipsoid with a lower k value, greater strain ratios, and with the horizontally oriented principal strain now the X axis instead of the Y axis, is associated with generally steeply plunging linear elements deformed on a large scale in rocks with thick felsic veins (Pabbay NW2). In the same F2 fold zone, in S.E. Pabbay and in Berneray, nearly uniaxial oblate strain ellipsoids with the greatest strain ratios and again with nearly horizontal

Fig 9 Schematic view of the relationships of the constructed strain ellipsoids with the deformed early lineations. (also see fig 3)



X axes are associated with lineations deformed on a small scale in rocks with thin felsic veins. These relationships are summarized diagramatically on Fig.9.

The association of thin veins with lineations deformed on a small scale, and of larger scale deformed lineations in gneisses with thicker veins is probably a direct function of these different lithologies; the fold axes and rodding structures developed in the thin competent veins are thinner and weaker, and therefore the scale of their deformation is smaller than that of the larger and more resistant axes and rods developed in the thicker veins. It seems that just as the wavelength of folds is determined by the thickness of competent layers involved in the folding, so the scale of the arcuate patterns of lineations deformed "in plane" is in part, a function of the coarseness of those lineations.

The lack of thick competent bands in S.E. Pabbay and Berneray may also partly account for the higher strain ratios recorded there; the "pink and blue" gneisses as a whole are, in these areas, more incompetent than the rocks of NW Pabbay and were therefore more highly strained.

Having established these relationships we must move into the more speculative sphere of a possible explanation of the deformed lineations of the Sound of Harris steep zone. The reader may recall that early linear elements in Ensay and Killegray, in the other limb of the Sound of Harris Antiform, are also deformed "in plane" in a similar manner to those of the steep zone. A separate analysis of these structures is not given, however, since an explanation of their origin can be found in the hypotheses set out below.

Initially the most significant facts are as follows (i) Where the early lineations are apparently undeformed the nearly horizontal principal strain is the Y axis of the strain ellipsoid determined from vein orientation, and where they are deformed the horizontal principal strain is an X axis; (ii) Two "phases" of deformation can be recognised, one which produced the F2 folds and rodding structures, and one which deformed them. This later deformation was presumably a kind of inhomogeneous slip in plane (perhaps analogous to the mechanisms postulated for the formation of similar folds), and probably caused by a late flattening type of deformation. Such a deformation would also have affected the fold profiles and the orientation of the veins from which the strain ellipsoids were constructed, i.e. the strain ellipsoids determined in the Sound of Harris steep zone must be the products of both these deformation increments. (iii) In S.E. Pabbay and Berneray the observable movement direction of the small scale deformed lineations lies in the foliation planes nearly parallel to the X directions of the strain ellipsoids constructed at these localities. Often this direction is marked by a late mineral lineation (see above).

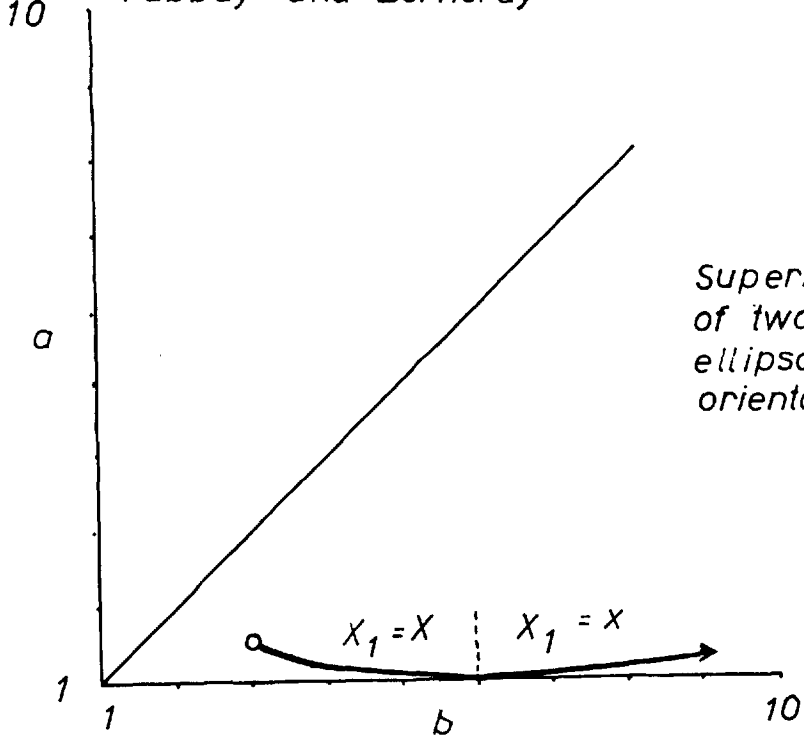
Evidence of two superimposed increments of deformation suggests the possibility of explaining the deformed lineation and the changing values of the horizontal principal strain in terms of the simplified principles of strain superimposition described in chapter 6. The validity of using these principles in the instance is indicated by the constant orientation of the strain ellipsoids from all the localities. From the data given above it is logical to assume that the deformation of the lineation is related to the change of value of the horizontal principal strain, and it may be that this change in value was the result of the superimposition of the two strain increments. The basic geometry of these deformed lineations, (the fact that they are deformed in planes which are not folded) suggests that the Z axes and XY planes of the two superimposed strains were parallel. If the interchanging of the values of the principal strains between N.W. Pabbay (1) and the other localities is due to the superimposition of two deformation increments, the relative orientations of the two incremental strain ellipsoids must have been

$$\begin{matrix} X & y \\ Y & x \\ Z & z \end{matrix}$$

For all except uniaxial oblate ellipsoids this causes the finite strain to move along a path towards $k=0$, reach it where the ratios of the second strain equal those of the first, and be deflected from it where the ratios of the second strain exceed those of the first. Along this path the intermediate axis of the first strain ellipsoid (Y) becomes the longest axis of the finite strain ellipsoid (X_1) as soon as the strain ratios of the second ellipsoid (x, y, z) exceed those of the first (X,Y,Z) (i.e. as soon as the path is deflected from $k = 0$). Similarly X of the first ellipsoid becomes Y_1 of the finite strain ellipsoid (Fig.10).

The deformation of the early lineations of the Sound of Harris can be regarded as inhomogeneous deformation (differential flow in the foliation planes) in response to the x direction of the second strain increment or (the same direction) the X_1 axis of the finite strain, and may be envisaged as a series of halts along the strain path

Fig10. Strain superimposition which may have caused the variations in strain values recorded in Pabbay and Berneray



Superimposition of two k = 0·5 ellipsoids in the orientation

$$\begin{array}{cc} X & y \\ Y & x \\ Z & z \end{array},$$

Fig10c

N.W. Pabbay 1 → N.W. Pabbay 2 (by superimposition of a second strain k = 4·5 in the orientation $\begin{array}{cc} X & y \\ Y & x \\ Z & z \end{array}$)

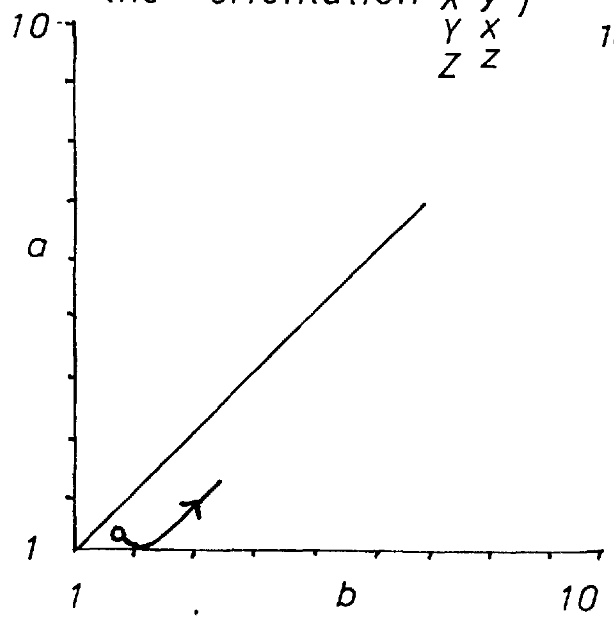
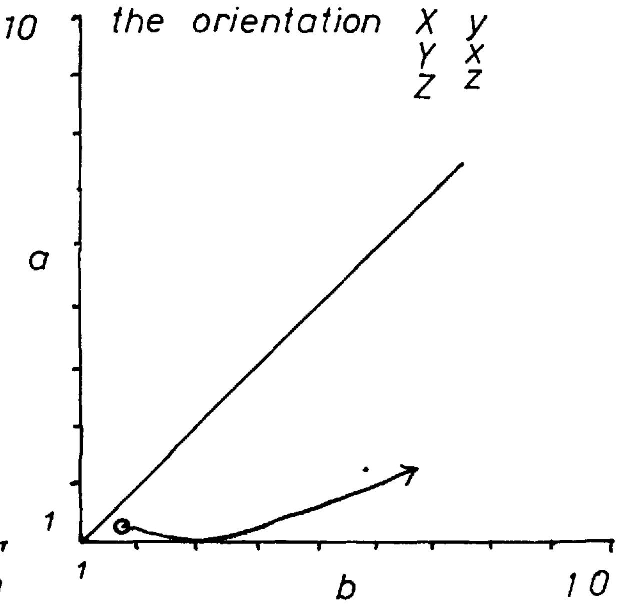


Fig 10 b

N.W.Pabbay 1 → S.E.Pabbay and Berneray (by superimposition of a second strain k = 0·8 in the orientation $\begin{array}{cc} X & y \\ Y & x \\ Z & z \end{array}$



o is the ellipsoid constructed at N.W. Pabbay 1

X x

Y y

Z z

All the qualitative data can be accommodated into this scheme. The first strain (X, Y, Z) produced the minor folds, F2 rodding structures and mineral lineations, and may now be represented by the strain ellipsoid constructed at N.W. Pabbay (1). The second strain increment increased the ratios of the finite strain, changed its k value, caused the axes of the finite strain ellipsoid to interchange their values and deformed the earlier linear elements.

When we try to fit quantitative data from strain determinations into this scheme, however, we are faced with serious anomalies. If the strain ellipsoid for N.W. Pabbay (1) is taken to represent the unmodified first strain increment the ellipsoids constructed for S.E. Pabbay and Berneray could be derived by the strain path described above by the superimposition of a second strain increment with oblate symmetry (Fig.10b), but the ellipsoid for N.W. Pabbay (2) could not; it would only be reached along this path if the second strain increment was prolate (Fig.10c), and evidence of constrictional deformation cannot be found in these rocks. We must therefore conclude that either

(i) The strain ellipsoid for N.W. Pabbay(1) does not represent the unmodified first strain increment

or (ii) The strain measurements for either N.W. Pabbay (1) or N.W. Pabbay (2) or both are erroneous, and that the error is not constant.

(iii) All the strain measurements are incorrect

or (iv) The suggested explanation is incorrect.

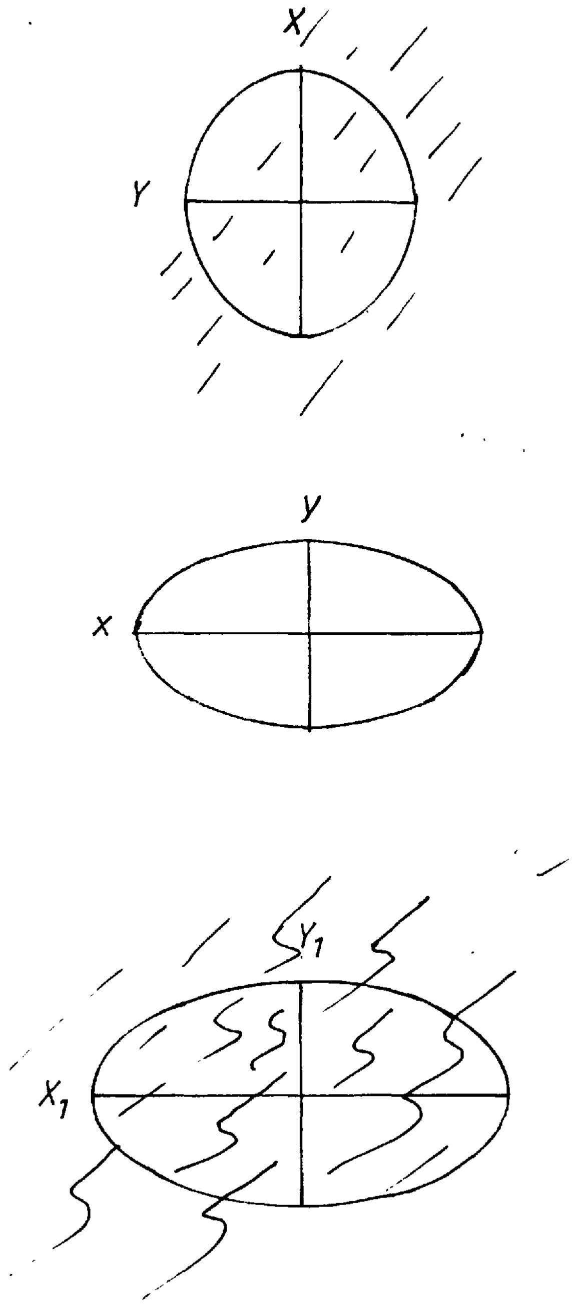
It is my personal opinion that the deformed lineations of the Sound of Harris should be explained in these terms and that the calculated strain ratios, especially those of N.W. Pabbay, are suspect. Considerably more confidence can be placed in the predicted orientations and in the interchanged values of the principal strains, and also, probably, in the changes of k value of the ellipsoids. The field associations of two deformations, one which created a suite of linear structures and one which deformed them and caused the development of a new lineation parallel to the movement direction of the deformed lineations still hold, and the association of the deformation of the lineation with a change in the value of the horizontal principal strain produced by co-axial superimposition of strains is probably correct.

Now, therefore, we revert to a purely qualitative discussion to establish the age of the second strain increment which caused the deformation of the early linear elements and examine its importance in the regional structure and structural history of the Sound of Harris.

7:2:4 The age of the deformation of the early linear elements: a regional synthesis

To have deformed in the manner described (Figs. 3, 4, and 9 and plate 1), the early lineations must originally have lain at a significant angle to the x direction of the second strain increment (Fig.12). Since the superimposition of the strain increments was co-axial, the lineations must also have lain at a significant angle to the axes of the first incremental strain ellipsoid which caused their development. This strain ellipsoid (the F2 strain ellipsoid-most of the important deformed lineations are of F2 age) was probably added to an original surface so that its XY plane lay normal to the surface and its Z axis was contained in the surface. The F2 fold axes and mineral lineations developed parallel to each other along the line of intersection of the original surface with the XY plane of the ellipsoid, within its extension field (see the conclusions of Chapter 6) (Fig.13a). The original surface (the foliation after the F1 deformation) probably had a N.W.-S.E. strike and a steep dip. The XY plane of the F2 strain ellipsoid probably had a north-easterly strike and a nearly horizontal dip when it was added to this surface, and, since F2 folding is tight in the Sound of Harris steep zone, the foliation was also brought into this orientation. Our problem is to decide whether the deformation of the lineation occurred while the foliation still lay in this orientation (i.e. that the second strain is late F2 in age, Fig.13) or whether it occurred after the rocks had been brought to their present steeply dipping NW-SE orientation in the limb of the F3 Sound of Harris

Fig 12

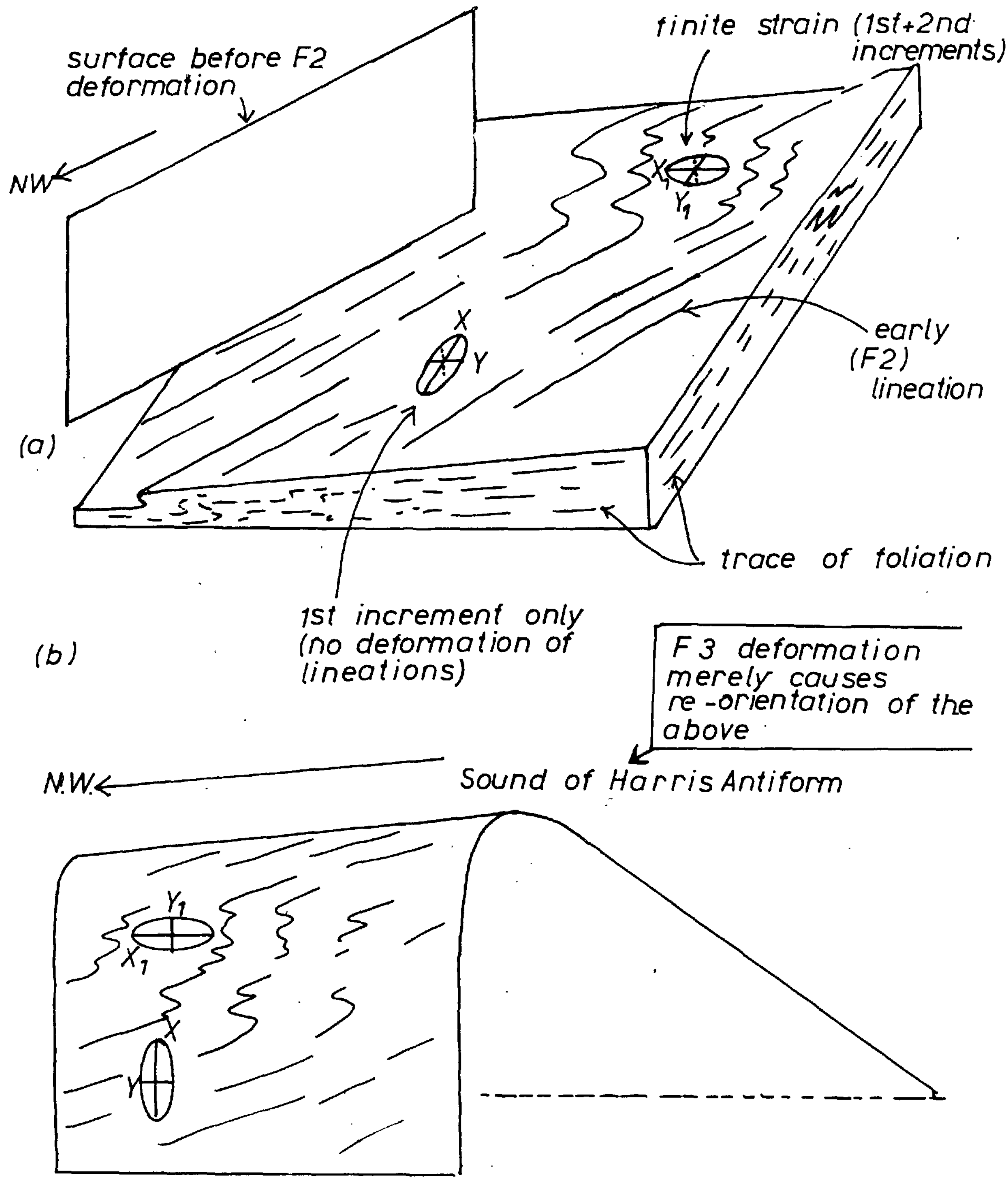


1st increment
and lineation

2nd increment

finite strain
deformation of
lineation.

Fig 13 A late F2 age for the deformation of the early linear elements - stylized block diagram.



Antiform (i.e. the second strain increment and the deformation of the lineations are F3 phenomena). The first alternative implies that F3 deformation in the Sound of Harris steep zone is negligible and that the rocks were merely re-oriented; the second implies that F3 deformation is very important. We saw in Chapter 2 that apart from some very open folds there is no direct evidence of post-F2 deformation in the Sound of Harris steep zone except for the problematical deformed lineations with which we are now concerned, yet in the other limb of the Berneray Synform F3 minor folds are abundant and well developed. We may now discover whether this apparent difference in the strain state of the two limbs is a valid one.

The idea that the deformation of the lineation is F2 in age and that the deformed elements were merely re-oriented by F3 has one serious disadvantage; the interchaning of the principal strains hardly seems compatible within the generally held concept of a "deformation phase". If, on the other hand an F3 age is postulated for the second strain increment we see that the manner of its superimposition

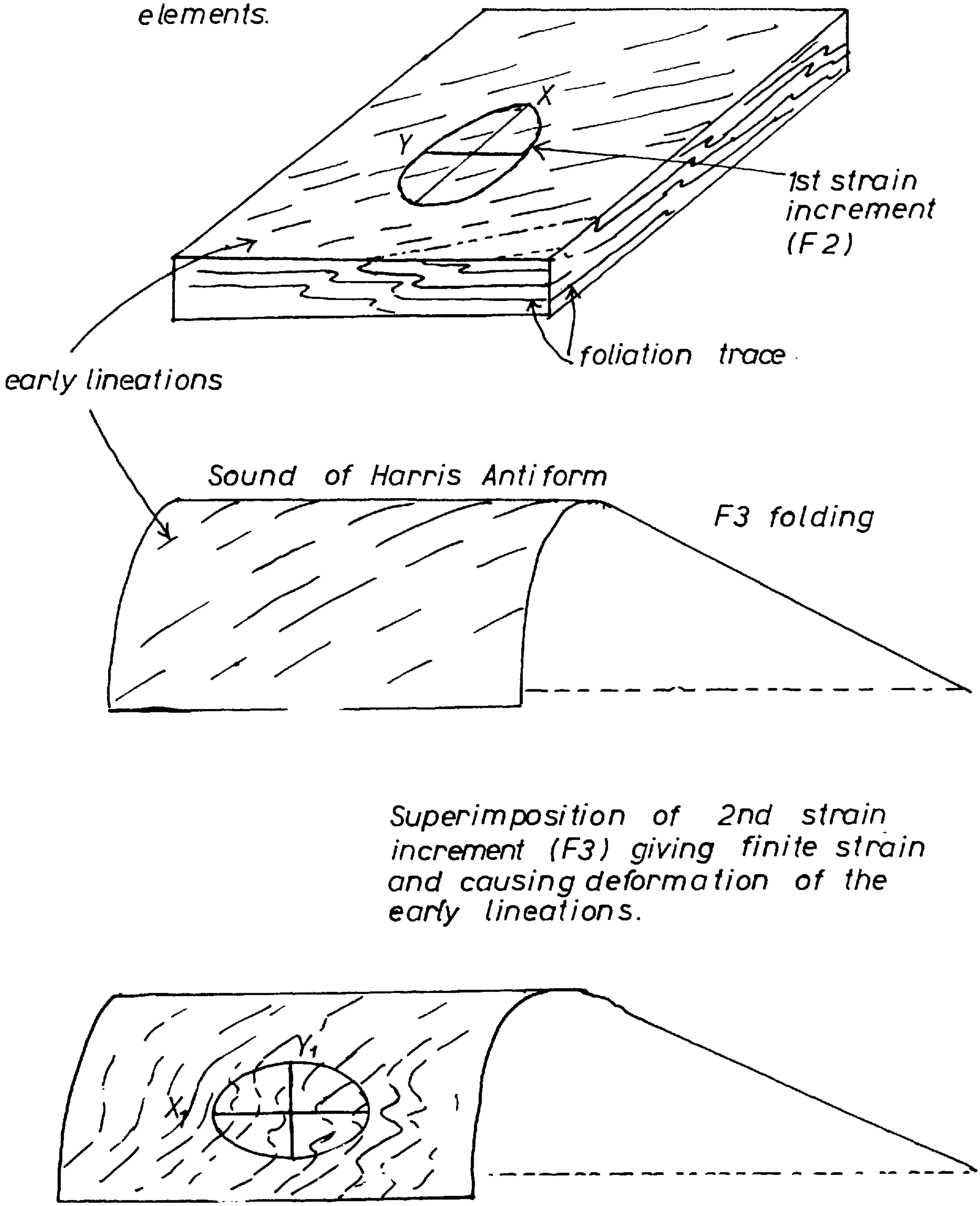
X x
Y y
Z z

is just what we should expect from F3 strain in the limb of an F3 fold (see Chapter 6).

Before discussing the implications of an F3 age for the lineation deformation, it is worth reminding the reader that the projected axis of the Sound of Harris Antiform, the hinge of the Berneray Synform; the movement direction of the small scale deformed lineations, the late lineation, and the horizontal axes (X in three cases, Y in the fourth) of the constructed strain ellipsoids are all effectively parallel. At their most extreme positions they all lie in a cone of less than 15° in diameter. Acceptance of an F3 age for the deformation of the lineation, therefore implies active F3, stretching parallel to the axis of the Sound of Harris Antiform i.e. the second strain increment was added to the NE-SW trending foliation pile so that its X direction lay in the foliation surfaces parallel to the Y axis of the F2 strain, and folding took place about this direction. The early linear elements, lying oblique to this direction were initially folded over the fold hinge and later deformed in response to a direction parallel to the fold hinge (see Fig.14). There is no structure visible in the rocks which records any differential movement normal to the fold hinges, and if the Sound of Harris Antiform were a normal fold we would have to invent complicated schemes to integrate the deformation of the lineation into a reasonable folding mechanism, or postulate massive rotation of the axes of Berneray Synform and Sound of Harris Antiform.

None of this is necessary, however, if we remember that the Sound of Harris Antiform is essentially a succession of metasediment derived gneisses plastered around the resistant block of the N.E. Pabbay deformation low. The deformation of the early lineations can now be looked upon as a contact strain effect - the product of very intense deformation at the margin of the resistant mass. This might explain the absence of strongly deformed lineations in the south-westerly part of the Sound of Harris steep zone and in the neighbourhood of the Berneray Synform, areas far from the resistant block.

Fig 14 Block diagrams showing an F3 age for the deformation of the early linear elements.



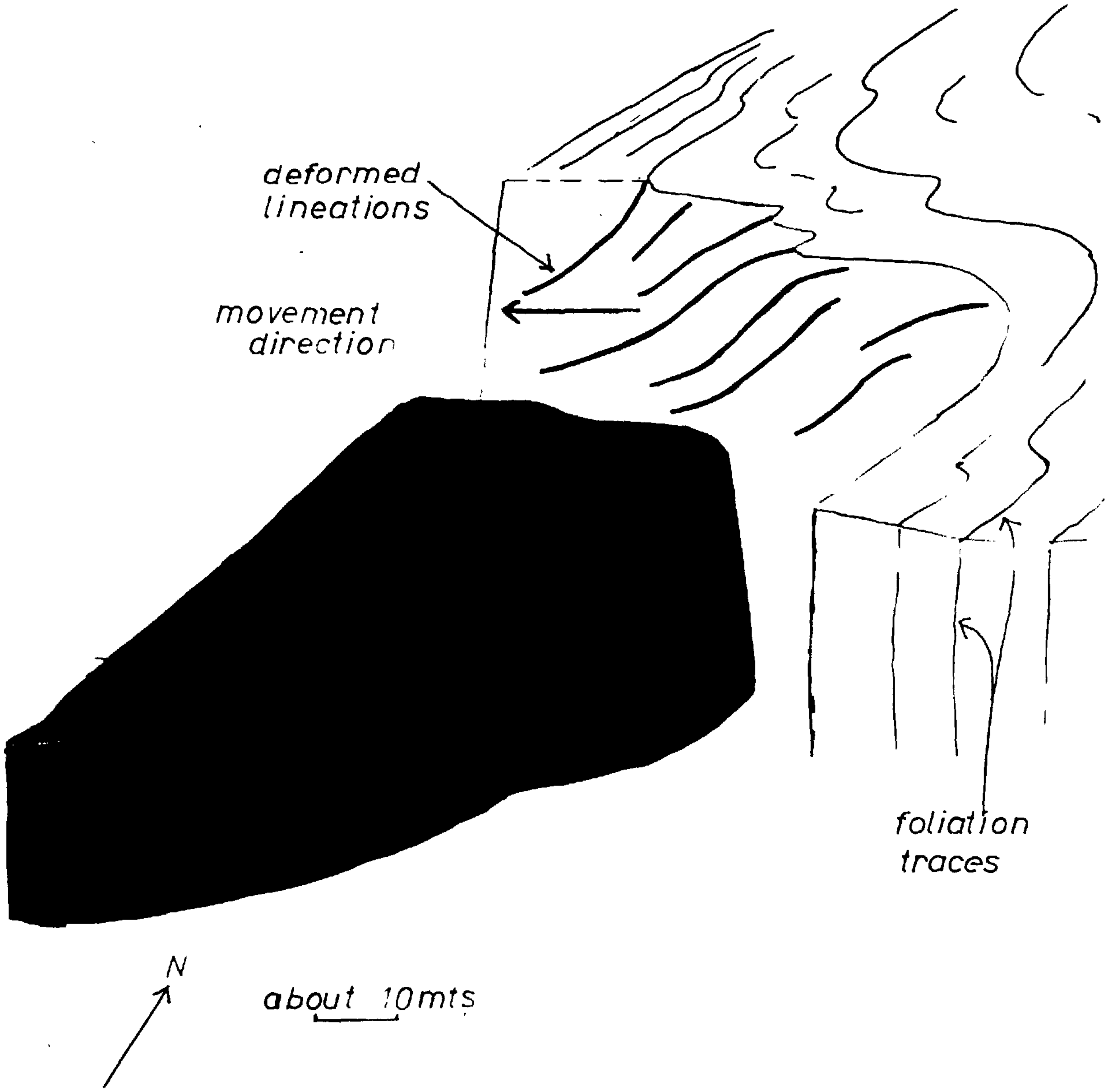
Commonly on a small scale in the North Uist area we see pre-existing lineations in acid gneiss deformed when those gneisses are flattened around boudinaged amphibolite masses. In some positions around these boudinaged masses deformation of the lineation takes place in the plane of the foliation, presumably by differential movement between the resistant material of the boudin and the more easily deformed gneiss, and the deformation persists for some distance into the acid gneiss (Fig.15).

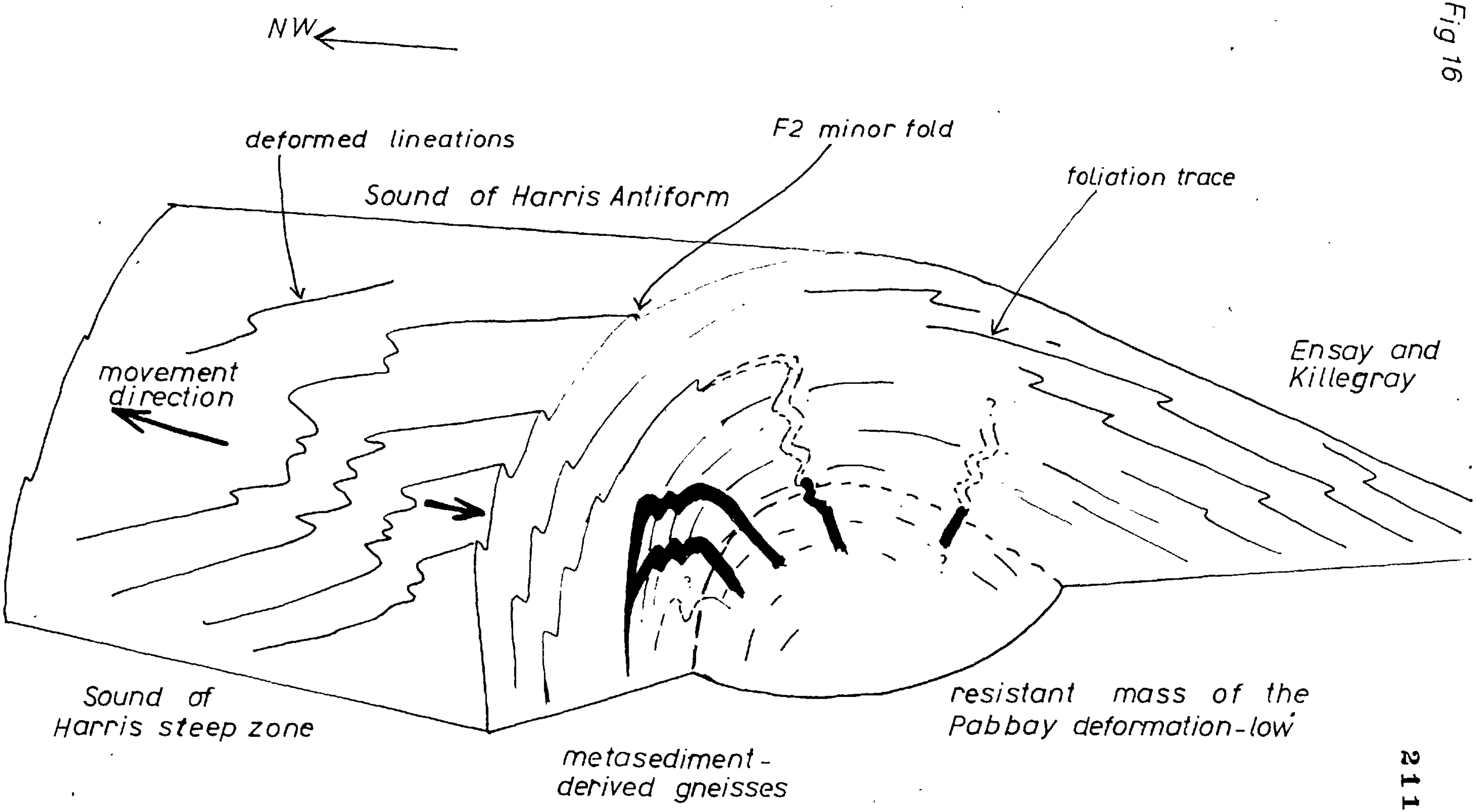
If we invoke this theory the problems of integrating the deformed lineations with the development of the major F3 folds disappears, the new horizontal X direction is due to flow of the ductile rocks around the resistant block. (see Fig.16). It is interesting to note that the lithologically identical rocks of the Langavat belt of Sound of Harris also carry lineations deformed in the plane of the foliation (Myers 1968, Chapter 8 of this thesis), and that the Langavat rocks also lie against a resistant rock mass, in this case a large igneous body (the "Tonalite" of Dearnley, 1963).

7:2:5 Concluding remarks

Even though there are few structures of obvious F3 age in the Sound of Harris steep zone, F3 deformation there is considerable and makes up the second of two strain increments which can be recognised in a study of the relationship between the deformed early lineations and the strain ellipsoids constructed from the orientation of deformed veins.

Fig 15 Exploded block diagram showing lineations deformed "in plane" along the contact of a boudinaged mass of Scourie-dyke amphibolite north of Scolpaig on the North Uist coast





Highly schematic block diagram illustrating an F3 age for the deformed lineations.

Fig 16

THE TIGHARRY ISOCLINE

7:3 A MAJOR STRAIN INHOMOGENEITY ON THE NORTH UIST COAST

7:3:1 Introductory statement

The analysis of the Tigharry Isocline as a second example of the use of strain ellipsoids constructed from deformed veins provides a means of reiterating two other important themes of this thesis, namely the importance of axial surface fanning and fold order, and the inhomogeneous nature of deformation in the North Uist area.

7:3:2 Partial recapitulation: Description of the Tigharry Isocline and its associated minor structures.

The Tigharry Isocline appears in the field as a zone of exceedingly regular, planar rocks with a very consistent orientation (Foliation dips 80+$^{\circ}$ towards 230°-240°; mineral lineations, rods, minor fold axes and the long axes of boudins are all parallel, plunging 35° towards 315°-325°). It lies between the second order Hoglan Synform in the south-west and the Balelone Synform, of similar age and size in the north east. The domain of planar rocks is bounded in the south-west by the Hoglan Amphibolite, a composite Scourie-Dyke amphibolite some 200 metres thick, and in the north-east by the Hosta Amphibolite of similar thickness and character (see Chapter 1 and maps 1, 2, and 6). In the centre of the outcrop there is a narrow belt of metasediments (including quartzites, semi-pelites and iron rich bands) interbanded with ultrabasic rocks and Scourie-Dyke amphibolites.

The vertical rocks making up the Tigharry isocline lie within the domain of the North Uist coast Synform, lying along, and parallel to the axial surface of that structure.

7:3:3 Evidence of existence of the Tigharry Isocline

The field evidence which demands the interpretation of the belt of planar rocks between Hoglan and Hosta as a major isoclinal fold has not yet been fully presented. The Tigharry Isocline has been previously accepted as fact rather than interpretation because much of the interpretation would have been out of context at an earlier stage of this thesis.

The evidence is:-

(i) The senses of the asymmetrical third order F3 folds flanking the planar rocks are opposed and indicate a major antiformal closure between them.

(ii) The Hoglan and Hosta amphibolites are uniquely thick for Scourie-Dyke amphibolites on the North Uist coast, but are themselves almost equal in size. They are probably different outcrops along the same band, which may continue further south as the amphibolite contained in the hinge of the Hougharry Antiform. (See maps 1, 2 and 6 and Chapter 1).

(iii) The belt of planar rocks lies between two second order F3 synforms whose axial surfaces are inclined inwards.

(iv) Throughout the North Uist area there are examples of small scale open synforms which contain central isoclinal antiforms. Rock layering is thickened far more in the antiformal hinges than it is in the synformal hinges, and where a single layer is subject to this kind of unequal thickening (e.g. in the South Hoglan fold, see Chapter 1) the isocline can be seen as a local area of intense strain (see also Chapter 3).

The Tigharry Isocline probably represents a "shoot-through" structure in the middle of the North Uist Coast Synform.

The closure of the Tigharry Isocline is not seen, though it may lie within the central belt of metasediments where there are suggestions of lithological duplication. The isoclinal geometry of the structure means that the bulk of the exposed planar rocks lie in its limbs.

7:3:4 The Minor structures

(A) The Fabric

The bulk of the acid gneiss in the Tigharry Isocline is "smooth gneiss" which is both strongly lineated and carries a well developed "flaggy" planar fabric with which Scourie-Dyke amphibolite units are strongly interbanded on all scales. Both the acid gneiss and the amphibolite are perfect PL tectonites. Garnet is common in many of the amphibolite bands. Two distinct sizes of garnet crystals are apparent - large crystals up to 2 cms in diameter, and small crystals averaging 1 mm. All overgrow the foliation to some extent (i.e. they cut across oriented hornblende crystals), but there is always some degree of flattening of the foliation around the garnets, and many of them have retrogressed rims. The larger garnets contain cracks which are consistently normal to the foliation.

(B) Minor folds

Minor folds occur sporadically in felsic veins and pegmatite bands throughout the outcrop of the Tigharry Isocline, and the foliation of the Hosta Amphibolite is strongly crenulated in places. The folded felsic veins are often strongly flattened buckles, though true ptygmatically folded (unflattened) veins occur, and sometimes highly modified elasticas shapes are seen (Fig.17).

(C) Boudinage structures

Boudinage is an extremely common phenomenon within the domain of the Tigharry Isocline. It is much more important here than anywhere else on the North Uist coast, and since lithological combinations are fairly uniform over much of the coast its intense local development may indicate either an unusually high strain state in the Tigharry Isocline, or that the strain rate was high, or both.

7:3:5 Strain Determination in the Tigharry Isocline

The well developed fabric, the strongly flattened minor folds and the abundance of boudins all indicate that the rocks of the Tigharry Isocline are highly strained. The similarity of these structures throughout the domain of the isocline and the large scale form of the structure suggest that it may be considered as a domain in which the strain is effectively homogeneous on a fairly large scale, and therefore it was a suitable place to attempt strain ellipsoid construction from the orientation of deformed veins.

Two areas were found in which there were sufficient numbers of veins, one in the Hoglan Amphibolite, the other in the Hosta Amphibolite. Data from each area were examined (and are presented, see appendix sheet 1 nos. 1, 2 and 3) both separately and together. Differences between the two sets of data probably have no geological significance, and are probably due to errors in measurement and inadequacy of data (see Chapter 4). The data are insufficient numerically, and more seriously, the boundary of the extension field is

Fig 17 Buckled and flattened pegmatite band in the Hoglan amphibolite

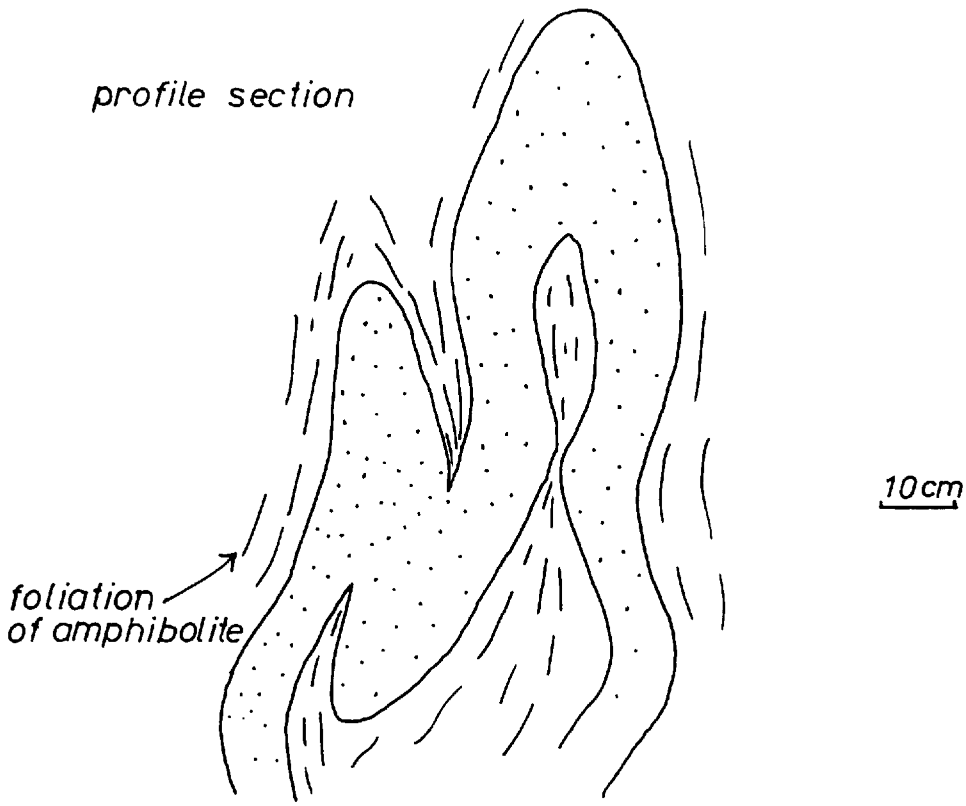
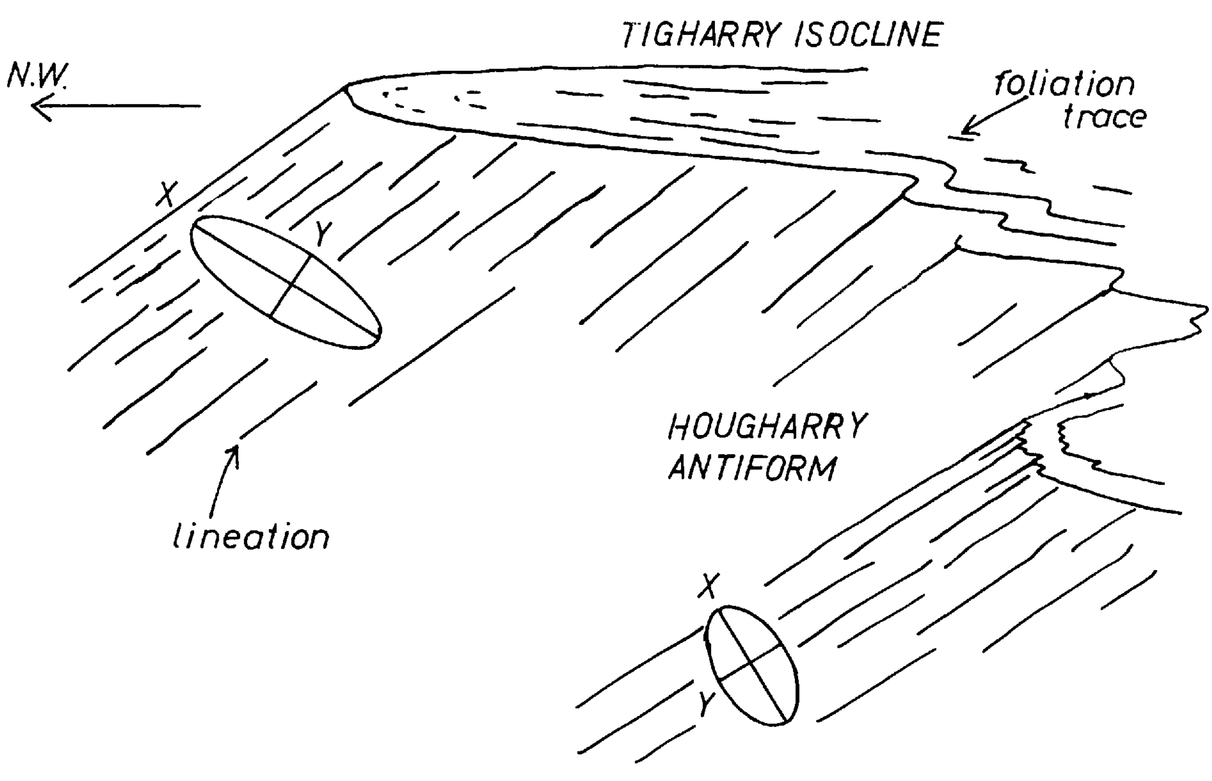


Fig 19 Schematic block diagram showing orientations of constructed strain ellipsoids in relation to the linear fabric.



imperfectly expressed (the initial orientation of the veins may not have been random). Nevertheless the results (average strain ratios 18:6:1) are perhaps the most realistic of all the strain determinations made in the North Uist-Sound of Harris area. The average orientations of the principal strains are:-

X 34° towards 165°
Y 56° towards 315°
Z 14° towards 065°

i.e. the XY plane of the strain ellipsoid for the Tigharry Isocline is parallel to the foliation (and axial surface) of the structure. The linear elements are not parallel to any principal strain, and this feature is discussed later in the chapter.

7:3:6 <u>The meaning of the strain ellipsoids</u>

The field relations set out above indicate that the Tigharry Isocline is an F3 structure. The P.L. fabric of the Hosta Amphibolite is deformed by minor folds which are most probably parasitic on the major isocline, and one of the strain determinations was made here from veins involved in these folds. Thus the strain ellipsoid in the Hosta Amphibolite appears dominantly to record F3 strain. In the Hoglan Amphibolite, however, the measurements of vein orientation were made on folded pegmatite bands and felsic veins which though apparently parasitic on the isocline (and thus of F3 age) carry the P-L fabric of the amphibolite as an axial plane structure (see Fig. I8). There are two explanations for this apparent inconsistency:

(a) The folds in the pegmatite bands, and the P-L fabric of the amphibolite are of Pre-F3 age, and were re-folded (bodily re-oriented) by the high order isocline. (Fig.18a)

(b) The pegmatite bands were originally discordant to the F1-F2 foliation in the amphibolite and, during the formation of the Tigharry isocline, buckled on a lower wavelength than the major structure. (Fig.18b).

If the first explanation is correct, F2 strain must make up a considerable proportion of strain recorded in the Hoglan amphibolite, in fact the constructed strain ellipsoid might be a record of the (final) finite strain produced by superimposition of F3 strain on F1/F2 structures in the limbs of the Tigharry Isocline. The similar values of the strain determinations in both the Hoglan and the Hosta amphibolites (Tigharry Isocline S.W. and Tigharry Isocline N.E. on appendix sheet 1), however, suggest that if this is true for one strain determination it is also true of the other.

Superimposition of F3 strain in the limbs of the Tigharry Isocline probably occurred in an orientation intermediate between

X x
Y y
Z z

and

X y
Y x
Z z

Doubts whether the strain ellipsoids record the F3 or the final finite strain are quantitatively significant. If the 18:6:1 value is the F3 component only, and F2 strain was, for instance, of the order of 6:3:1 (the value of F2 strain at Griminish), the finite strain in the limbs of the Tigharry Isocline would have ratios somewhere between 108:18:1 (k=0.6)

Fig 18 (a)

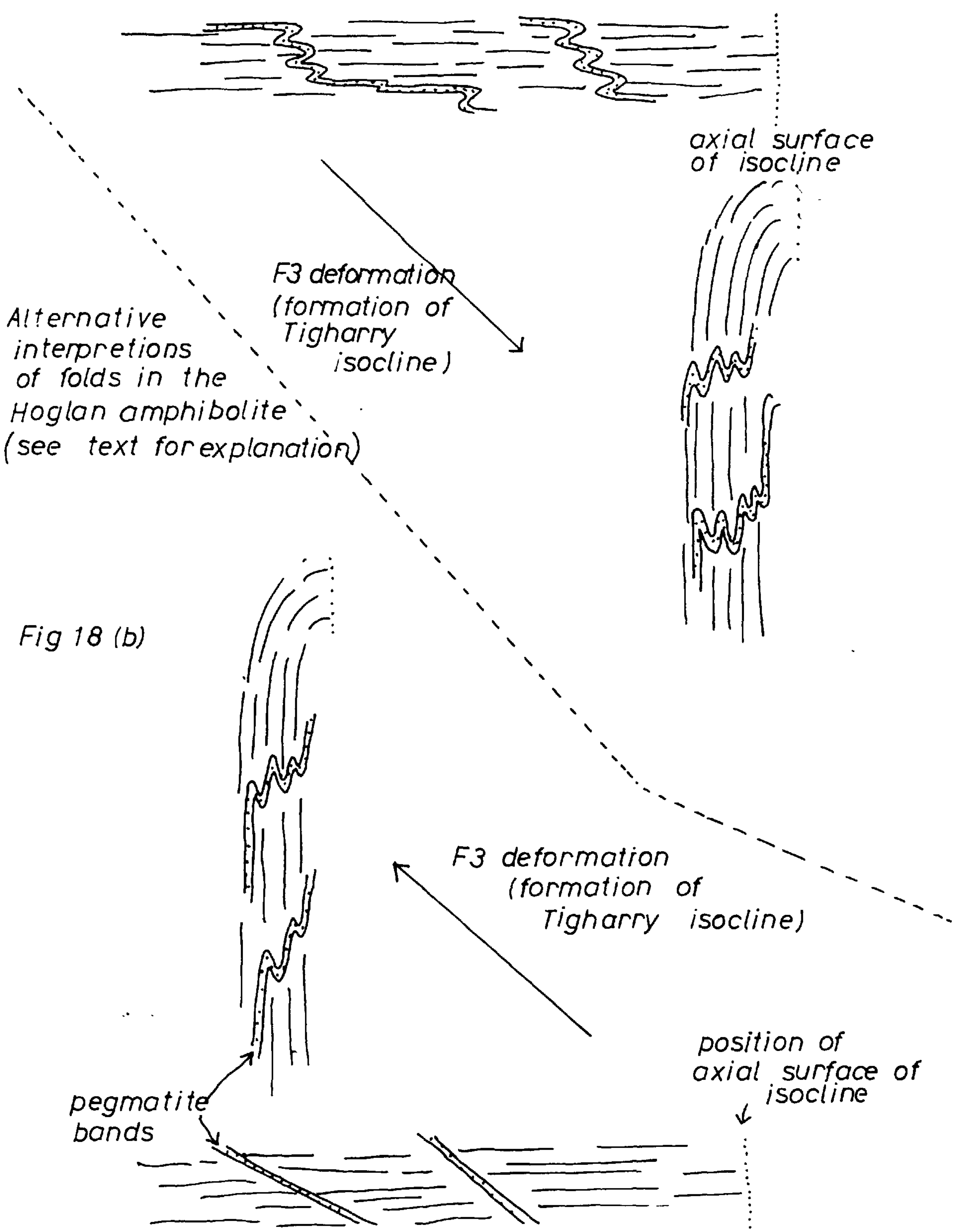


Fig 18 (b)

and 54:36:1 (k=0.1), alternatively the ratios of the final finite strain might be only 18:6:1. One wonders, however, just how significant this very large numerical difference is in terms of the physical appearance of the rock - would structures produced by an 18:6:1 strain differ markedly from those produced by an 108:18:1 strain? Folds would probably have similar geometry in either case, and both strains are great enough to effectively eliminate any original obliquity between rock layers. To recall a remark made in Chapter 4, in realms of very high strain we almost lose the value of quantitative structural geology for its own sake.

7:3:7 The relationship of the fabric to the strain ellipsoid.

This topic was examined in Chapter 6 as part of a discussion of the fabric history of the whole North Uist coast. Some points, however, need to be re-stated in this chapter. The planar part of the P.L. fabric of the Tigharry Isocline is obviously pre-F3 in age because it is folded by F3 minor folds. Although the age of the linear component of the fabric cannot be established with the same certainty (see below, and also the conclusions of Chapter 6), it too is probably mostly F1/F2, rather than F3 in age. If the strain ellipsoid recorded from the veins is essentially an F3 phenomenon, the co-incident symmetry of the strain and the fabric (the one tri-axial oblate, the other P.L.) is of no genetic significance. If the ellipsoid records the final finite strain, however, the co-incidence is significant.

As we have seen, the axes of the strain ellipsoid constructed from the orientation of deformed veins are not parallel to the linear structures of the Tigharry Isocline. A reconstruction (from the enveloping surfaces of the major F3 folds) of the orientation of foliation before F3 deformation on the North Uist coast, suggest that the XY plane of the F3 strain was added at a very high angle to an original surface which possessed a N.E. - S.W. strike and a north-westerly dip. The principal F3 strains could not have been added in the original surface, but the XY plane of the F3 strain must have contained the pre-existing lineations lying in the original surface, i.e. the two surfaces intersected along the old lineation. Folding took place about this direction, thus the F3 fold axes are parallel to the F1/F2 lineation.

The relative orientations of the X direction of the F3 strain and the F1/F2 lineation are such that "in plane" deformation of the early lineations comparable to that in the Sound of Harris might be expected. The fact that linear elements are not deformed suggests that differential slip in the foliation planes did not occur i.e. that flow between the foliation planes during or after fold formation was homogeneous. This being the case, we might expect homogeneous rotation of the lineations towards the new X direction, but this is an academic point; homogeneous rotation of the linear elements probably could not occur without rotation of the whole foliation pile (crystals probably cannot rotate through surfaces) and such rotation cannot be recognised because no structures are produced by it. If such rotation did occur it is obvious from Fig.19 that it was "frozen" at a very early stage. From the available evidence it is probably best to envisage the formation of the Tigharry Isocline as upward flow of rock, initially as a fold, caused by very intense local flattening.

Once the fold limbs had rotated near to the F3 XY plane, superimposition of oblate F3 strain on the F1/F2 PL fabric of those limbs was in an orientation such that the symmetry of the fabric was unchanged.

The amount of recrystallization which accompanied the development of the Tigharry Isocline is a difficult fact to establish. The post F2 garnets are probably partially retrogressed co-evally with the F3 strain (see Chapter 5), but if syn-tectonic recrystallization of hornblende occurred, the isoclinal geometry of the fold would have necessitated that most of this recrystallization took place in the planes of the F1/F2 fabric. Any preferred linear orientation would presumably have developed along the line of intersection of the XY plane of the F3 strain and the original surface - the same direction as the F1/F2 lineation, (see the conclusions of Chapter 6).

An interesting point is that much of the hornblende has a bluish colour in thin section. The only other occurrences of bluish hornblende in the present area are also in rocks which have obviously been very intensely strained, e.g. rocks from pinched-in F2 synforms in the Sound of Harris and from mylonite zones in South Harris itself. There might conceivably be a connection between the colour and the strain state, though whether the mineral recrystallized with a blue colour or merely turned blue under intense strain, one cannot tell.

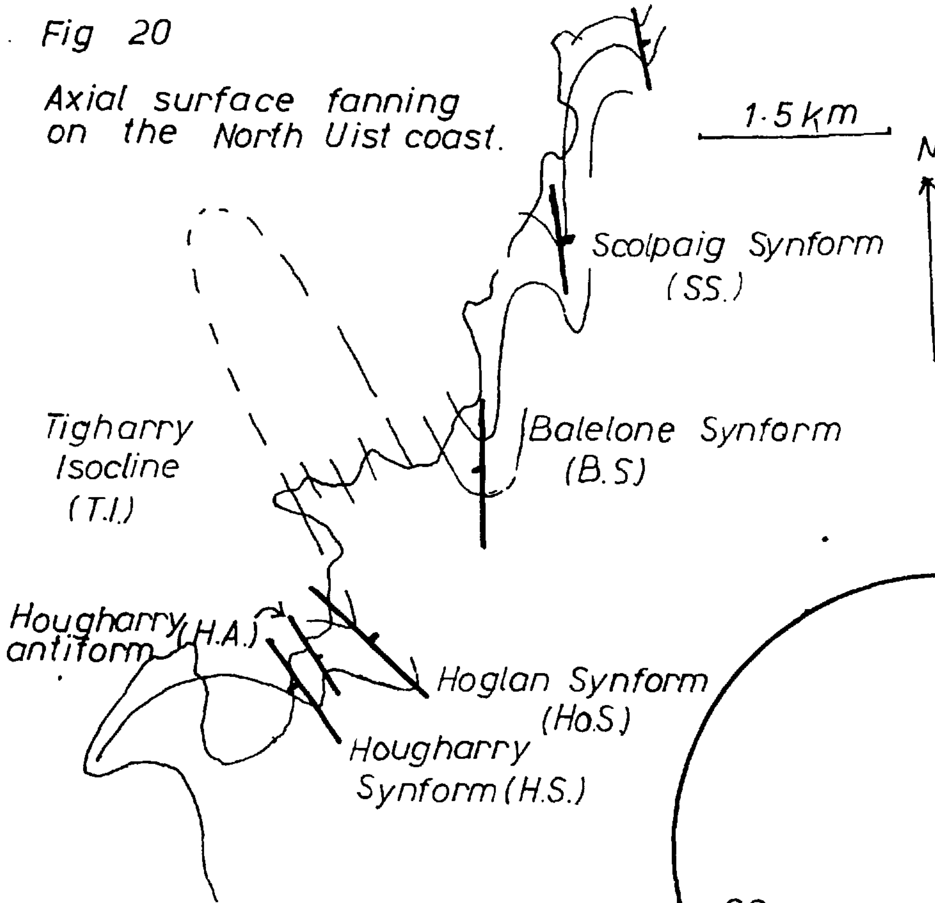
7:3:8 <u>Axial surface fanning and fold order</u>

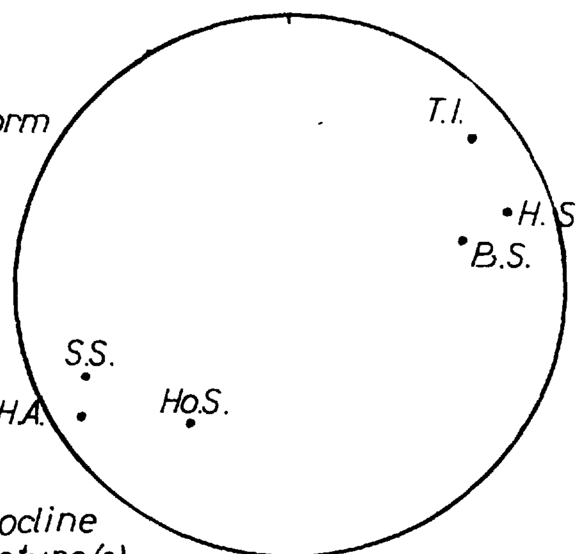
The axial surfaces of the Hoglan and Balelone Synforms are inclined downwards towards each other, and the 81° angle between them is almost exactly bisected by the foliation (i.e. the axial surface) of the Tigharry Isocline. The axial surfaces of the Hougharry and Scolpaig synforms, on the other hand, are inclined upwards towards each other (Fig.20). The fanning of the second order F3 folds on the North Uist coast is complex, and not a simple fan about the first order North Uist coast Synform. The convergence of the axial surfaces of the Hougharry and Scolpaig folds may represent the remnants of just such a simple fan which has been modified by the development of the Tigharry Isocline, or the pattern may be interpreted as the fanning of minor folds about two larger scale synforms separated by a tighter antiform.

A synform containing a central isoclinal antiform may be regarded as a group of three folds which could be considered complementary except for the fact that the axial surfaces of the synforms fan towards the antiform. This fanning was the criteria used in Chapter 1 to erect a system of fold ordering by means of which regional description proceeded, and in chapter 1 the Balelone and Hoglan synforms were called second order, and the Tigharry Isocline first order (the "complement" of the North Uist coast Synform). Attempted explanations of the structural development of the North Uist coast involving with this basic premise, however, break down for one reason or another (e.g. one demands the rotation of planes through the XY plane of the F3 strain ellipsoid for the Tigharry Isocline) and we are forced to take another line of reasoning. This is simply that if a major synform contains a central isoclinal antiform the two synformal closures which flank the antiform have axial surfaces which are inclined towards one another at an angle which depends

Fig 20

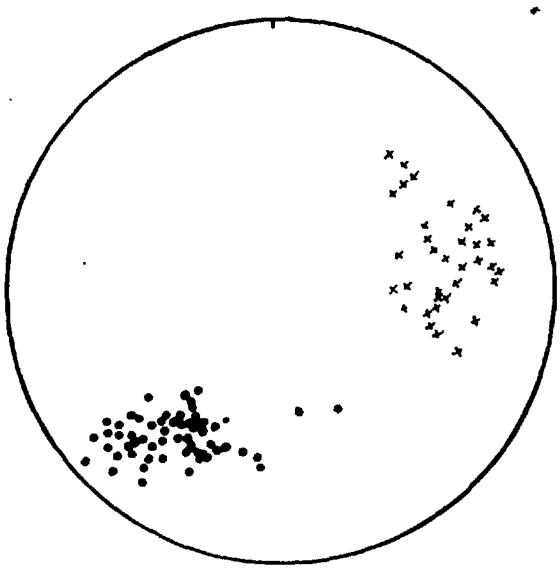
Axial surface fanning on the North Uist coast.

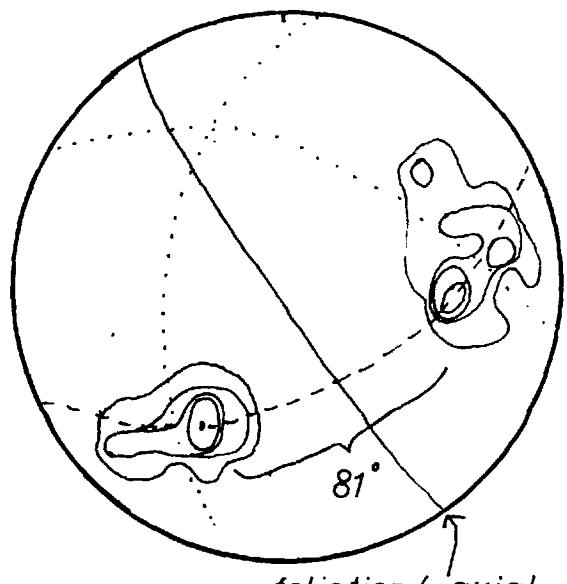




above and right : 1st and 2nd order folds.

below: third order folds north-east of the Tigharry Isocline (x) and south-west of that structure (•) poles to axial planes (contoured on right)





foliation (=axial plane) of Tigharry Isocline upon the interlimb angle of the major synform. The fanning is not produced by re-orientation by higher order re-folding, it is a simple geometrical necessity. Applied to the North Uist coast this idea means that the Balelone and Hoglan Synforms are the dual synformal components of the North Uist coast Synform, therefore they must now be classed as first order structures essentially complementary with the Tigharry Isocline, and the truly second order folds of Hougharry, Scolpaig and Griminish must be thought of as parasitic on the Hoglan and Balelone folds.

An amended scheme of fold ordering on the North Uist coast is as follows:

Order 1(A)	North Uist coast Synform and its antiformal complements to the north and south
Order 1(B)	Tigharry Isocline Hoglan Synform Balelone Synform
Order 2	Hougharry Synform Hougharry Antiform Cliffs Antiform Scolpaig Synform Griminish Antiform Griminish Synform

7:3:10 <u>Concluding remarks and further speculation</u>

The localization of very intense strain in the Tigharry Isocline is indicated by the following:

(i) A comparison of the strain measurements made inside the isocline with those made outside it (i.e. in the neighbouring Hougharry Antiform).

(ii) The widespread development of F3 boudinage within the Tigharry Isocline and a virtual absence of it elsewhere on the coast.

(iii) A comparison of values of two dimensional shortening along folded layers from within the isocline with those outside it.

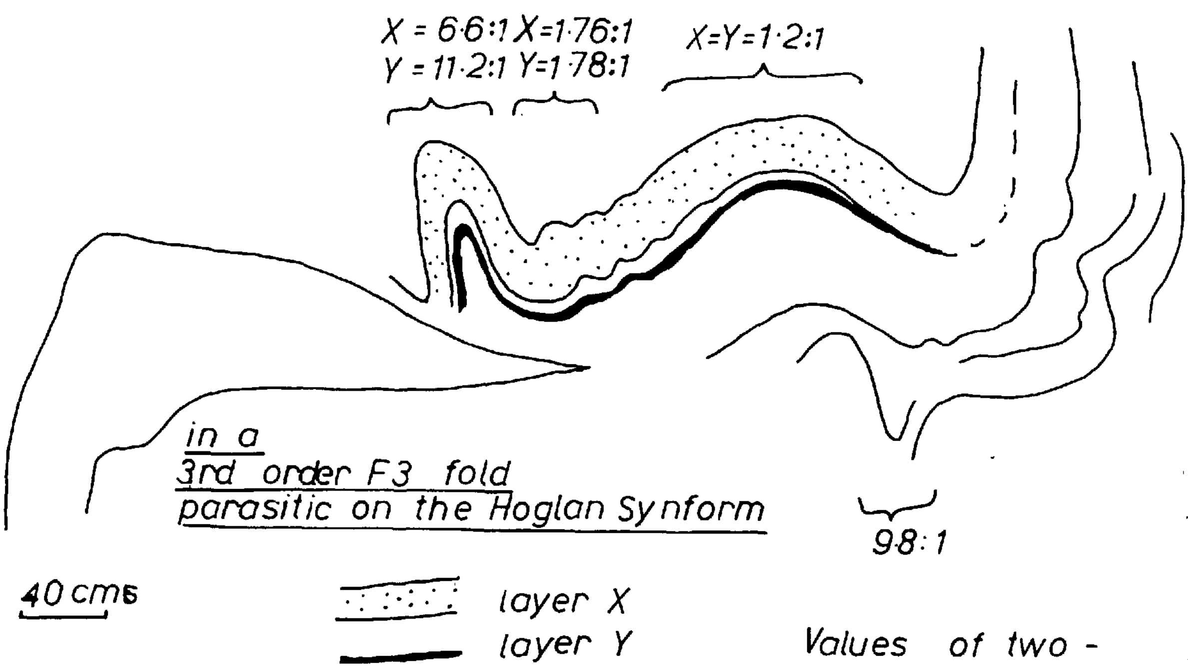
(iv) The isoclinal geometry of the structure.

(v) The visual regularity of the rocks within the isocline and the perfect development of their tectonite fabrics.

The localization of the intense strain is also suggested by the high angle still existing between the axial surfaces of the Hoglan and Balelone folds. Extensive homogeneous deformation outside the isocline would presumably have reduced this angle considerably. The reason for the localization of the strain is implicit in any proposed mechanism of "shoot through" formation (see chapter 3).

These inequalities in the strain state of the rocks over the North Uist coast are a large scale and rather extreme example of the kind of inhomogeneous deformation which characterizes the Lewisian rocks from North Uist to South Harris. The Tigharry Isocline is a homogeneous domain within the North Uist Coast synform, the synform itself being a structure in which inhomogeneity is apparent on all possible scales (the largest scale inhomogeneity being between the Tigharry Isocline and its neighbouring structures). The two concluding figures (Figs. 22 & 23) show inhomogeneous strain on two distinct scales, the one across a third order F3 synform from the south-west limb of the Tigharry Isocline, the other across the North Uist coast itself.

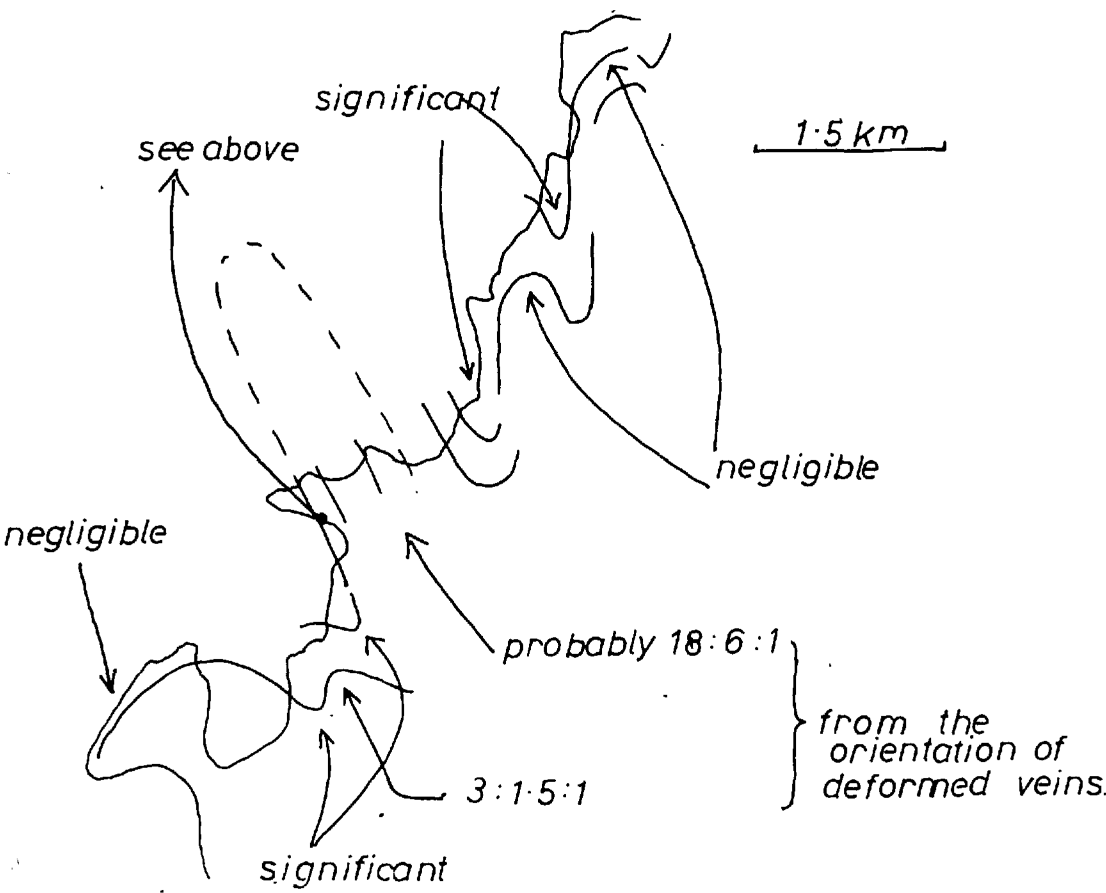
Figs 22 & 23 Inhomogeneous deformation on two different scales.



in a 3rd order F3 fold parasitic on the Hoglan Synform

Values of two-dimensional shortening given.

F3 Strain on the North Uist coast



7:4 THE CONCLUSION OF CHAPTER 7

Strain ellipsoids constructed from deformed veins have been useful in attempting to solve problems of the regional structural geology. In the Sound of Harris they helped the analysis of deformed lineations, and through their use in this context we were able to discuss in a more informed manner the age of the lineations, and whether F3 deformation is in fact important in the steep limb of the Sound of Harris Antiform where its presence might otherwise be unsuspected. In the Tigharry Isocline strain measurements helped demonstrate a profound strain inhomogeneity which was already strongly suggested by other structures. In both areas the use of strain ellipsoids raised important problems (especially those of particular relationships of fabric and strain) about which we would otherwise have been completely unaware.

CHAPTER 8

SOUTH HARRIS

8:1 INTRODUCTION

8:1:1 The scope of the work and its limitations

This chapter does not stand with the other regional studies recorded in this thesis because it is based on only about a fifth as much field work. It draws rather far-reaching conclusions from less substantial evidence than do the other two studies, and it uses Dearnley's (1963) lithological map as a basis for some of the discussion of the regional structure. My own systematic work in South Harris was restricted to the mapping of the Toe Head peninsula and of the west coast as far north as Borve Lodge, though numerous excursions have been made elsewhere. The area studied in detail is a useful one because it provides a more or less continuous coastal cross-section through all the main rock units in South Harris and, since inland exposure in the igneous and metasedimentary rocks is mostly mediocre, provides the best ground for an assessment of the structure of South Harris as a whole. Two well-exposed inland areas which, through lack of time, did not receive sufficient attention are the Renish peninsula and the magnificent exposures of anorthosite on Roneval (a hill near Rodil). Further investigation of these areas may provide evidence which might necessitate revision of some of the conclusions presented in this chapter.

8:1:2 Previous work

Compared with the rest of the Outer Isles, South Harris has been an area of geological indulgence. Jehu and Craig (1927) noted most of the important lithological groups and Davidson (1943) has described the petrology at great length. Myers, in a doctoral thesis in the present Imperial College series (1968), has discussed the migmatic acid gneisses in the northern part of South Harris, and my own area overlaps with his in the coastal exposures of the Langavat metasediments near Borve Lodge. The most important regional study of the southern part of South Harris is Dearnley's (1963), and a discussion of this work appears later in this chapter.

This chapter consists of two main parts, one a description of the geology of southern South Harris with emphasis on aspects of the structure; the other a suggested interpretation of the structure and structural history of the area. The two parts are separated by an outline of Dearnley's (1963) conclusions, and brief comments are made on the discrepancies between that work and the present one.

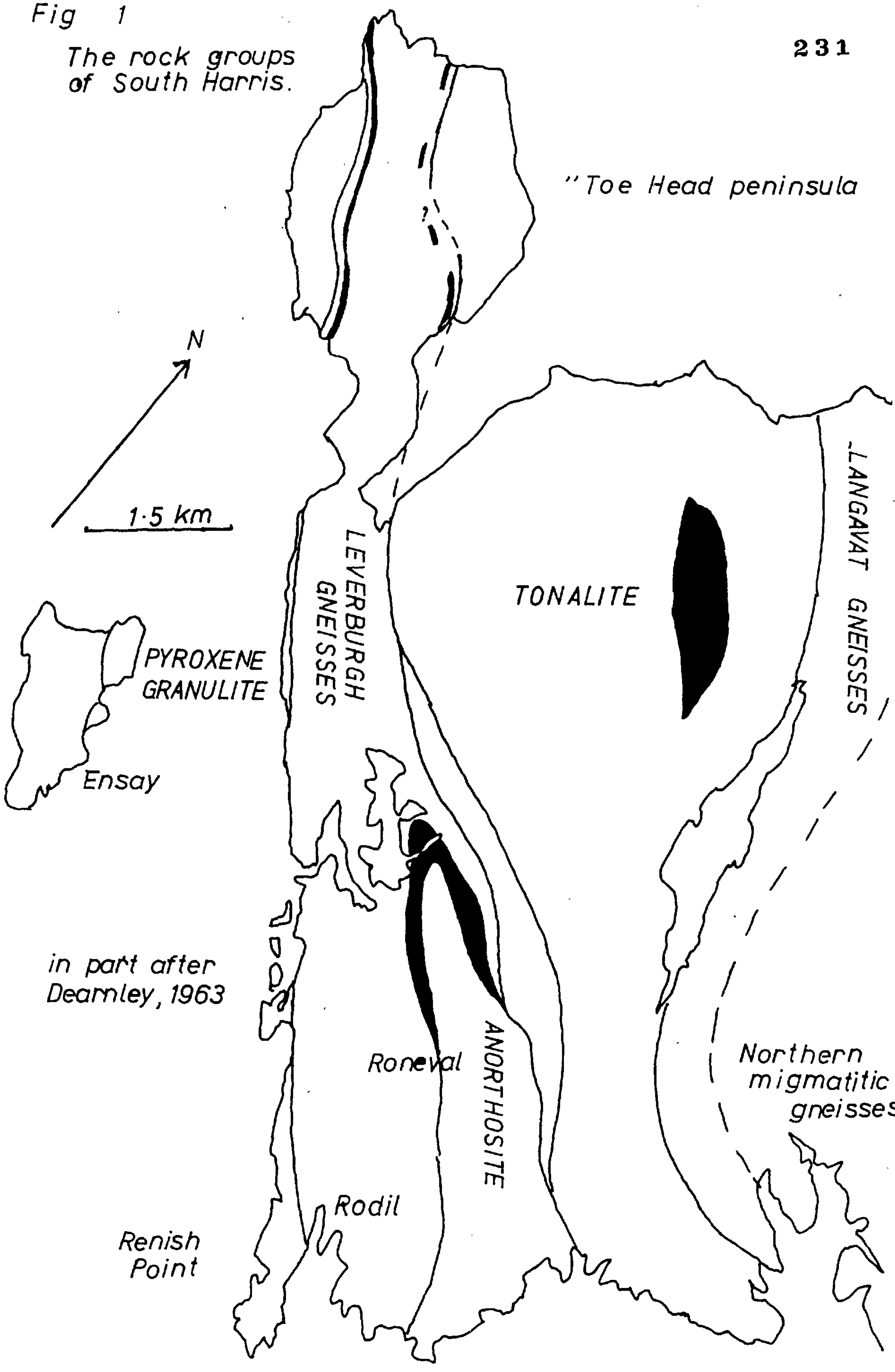
8:2 THE ROCKS OF SOUTH HARRIS

8:2:1 Introduction

South Harris contains three major acid gneiss units:- the northern "migmatic" gneisses (normal acid gneisses containing much Laxfordian intrusive granite), the Langavat Gneisses (including metasediments), the Leverburgh Gneisses (dominantly metasedimentary), and three large masses of metamorphosed, igneous rock - a tonalite, an anorthosite,

Fig 1

The rock groups of South Harris.



and a pyroxene granulite - the terminology of the igneous rocks is Dearnley's. A subordinate, but still significant, rock type is a very coarse-grained garnetiferous amphibolite called by Dearnley a "metagabbro". Thick bands of this rock run parallel to the contacts of the three larger igneous masses and bands and pods of it occur within the intrusions and throughout the Leverburgh Gneisses. In the anorthosite, dykes of garnetiferous amphibolite cross cut an early banding. Amphibolite bands similar to the Scourie-Dyke amphibolites of North Uist and the Sound of Harris occur within the Langavat Gneisses and in the northern migmatites, but are absent from the Leverburgh Gneisses. Late cross-cutting Laxfordian pegmatites are widespread; they post-date all the deformation and are therefore ignored. Each of the important rock types is now examined in turn.

8:2:2 The Pyroxene Granulite

This rock is exposed in a narrow strip along most of the south-west coast of South Harris and also occurs on the north-east headlands of Ensay and on small islands offshore of Leverburgh. It presumably continues beneath the Sound of Harris between Ensay and South Harris itself. Where virtually undeformed, the Pyroxene Granulite appears in the field as a coarse-grained rock, resembling a gabbro and containing laths of ortho-pyroxene, clino-pyroxene, felspar biotite, and (on Ensay) garnet. Its originally intrusive nature is indicated by the occasional occurrence of xenoliths of acid material (plate 1). It contains lenses and bands of amphibolite and abundant thin felsic veins. The mass is elongated N.W. - S.E.; its contacts are local zones of intense deformation which converge downwards.

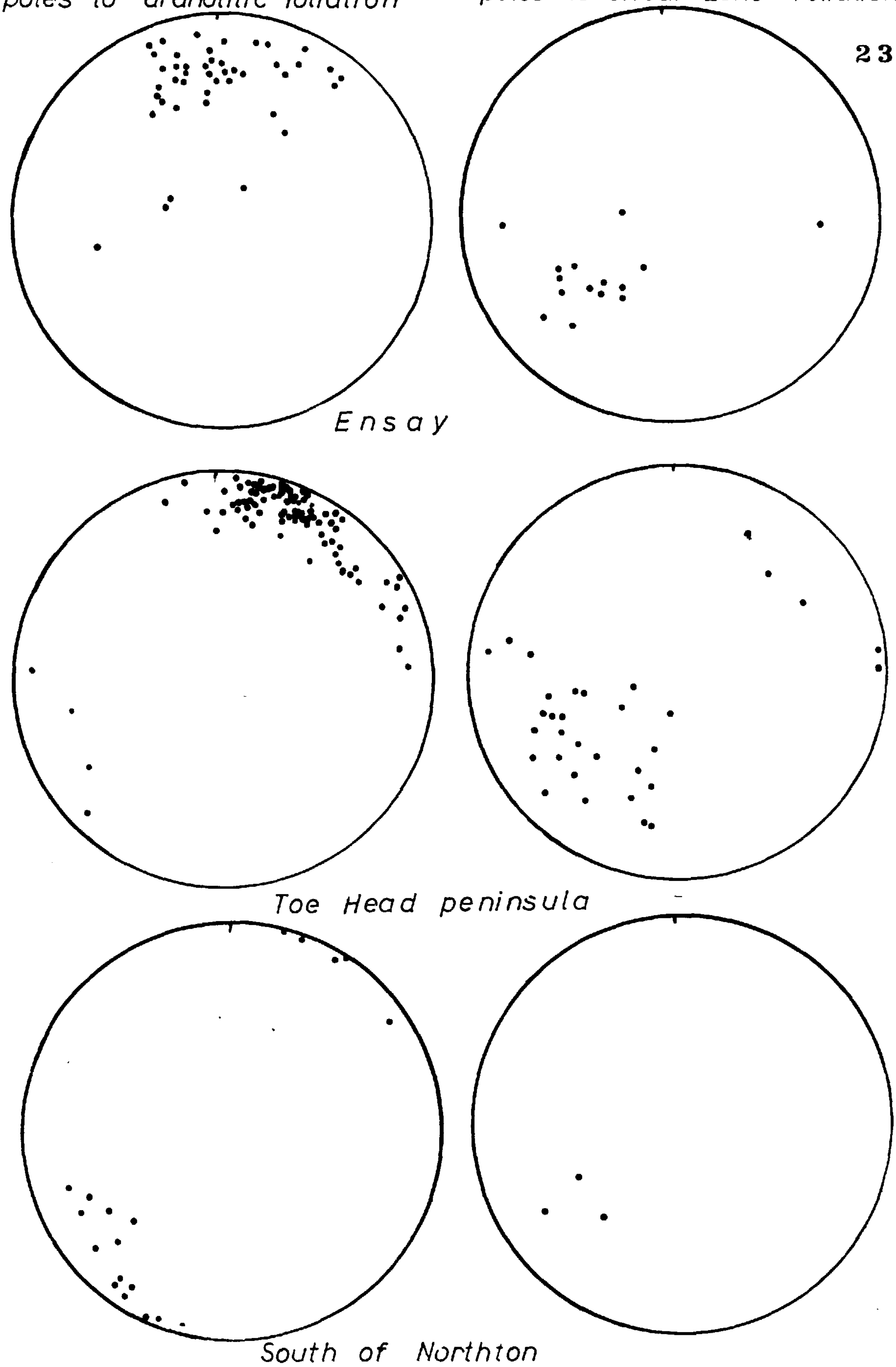
Plate 1 *Xenolith in almost undeformed Pyroxene Granulite, Ensay. (x2,approx.)*

Plate 2 *Shear zone in Pyroxene Granulite, Ensay.*

10cms

Fig 2

poles to "Granulitic foliation" poles to shear zone foliation.



Structures within the Pyroxene Granulite (Fig.2).

There is evidence of two phases of deformation within the Pyroxene Granulite. An early penetrative planar fabric, called here the "Granulitic Foliation", is cut by shear zones in which a new planar fabric - the Shear Zone Foliation - is developed. The "Granulitic Foliation" is axial planar to the folds in the felsic veins, and in one example, on Ensay, to folds in an amphibolite band. Amphibolites lying close to the plane of the "Granulitic Foliation" are often boudinaged.

The orientation of the "Granulitic Foliation" is modified as it swings in and out of the shear zones, but its general trend on Ensay is E. - W., while on the Toe Head peninsula and south east of Northton it is N.W. - S.E.. Everywhere it is nearly vertically dipping. On Ensay it is strongly oblique to the foliation of the acid gneiss and to the contact zone of the body; on the Toe Head peninsula it is usually oblique to these features whereas south of Northton it is more or less parallel to them.

The shear zones range in size from the order of a few centimetres to about ten metres in width, and the larger they are, the more complex are their internal structures (i.e. the more they depart from the simple shear model). Within the shear zones the development of the new foliation is accompanied by grain size reduction and complete recrystallization in the amphibolite facies (c.f. the shear zones of Castell Odair described in chapters 5 and 6), and the felsic veins are broken down into felspathic augen (plate 2).

On Ensay the zones dip 40°- 50° to the north-east, on the Toe Head peninsula, though the orientation is less regular, the dip is generally steeper, but in the same direction. The senses of movement in the zones are different in different areas. On Ensay the sense is always sinistral, on Toe Head it is always dextral, at Renish Point it is sinistral once more. In the area south of Northton shear zones are not particularly well developed. A very large shear-zone with complex internal structures, including local re-folding relationships, lies near the southern contact of the Pyroxene Granulite on Ensay. The orientation of the shear zones on Ensay is parallel to the orientation of the foliation of the acid gneiss there, the shear zones on Toe Head are more or less parallel to the foliation of the Leverburgh Gneisses. In both areas they are parallel to the sides of the Pyroxene Granulite.

Petrography of the Pyroxene Granulite

In thin section the least modified parts of the Pyroxene Granulite contain abundant, large, often euhedral laths of orthopyroxene and clinopyroxene and rarer large laths of plagioclase set in a fine-grained groundmass of quartz and felspar. The edges of the large crystals, though showing slight signs of alteration, are sharp and well defined. All the plagioclases, and some of the pyroxenes, show strain effects, e.g. the plagioclases have a slightly sigmoidal shape and sigmoidal twins and also show undulose extinction, (c.f. Dearnley, 1963). Some of the plagioclase laths are zoned. Together with biotite the large crystals are orientated in a rough planar fabric - the "Granulitic Foliation". In the area south of Northton, where the "Granulitic Foliation" is almost parallel to the foliation of the Leverburgh Gneisses, the fabric is much better developed and most of the pyroxene is made over to amphibole.

In the shear zones the original rock type is completely and profoundly altered. No trace of the large crystals remains and the rock in the shear zones is a fine grained, highly tectonized amphibolite.

8:2:3 The Leverburgh Gneisses

North of the Pyroxene Granulite, between that body and the Tonalite mass to the north, lies the north-westerly trending belt of the Leverburgh Gneisses. The bulk of these rocks are obviously metasediments and have been described in detail by Davidson and Dearnley (op cit). In this section I will discuss only a part of the belt, namely that in the Toe Head and Northton areas, where the Leverburgh Gneisses form a banded psammatic succession steeply dipping to the north-east. Minor folds are few, there are some near isoclinal "intra-folial" folds on the south coast of Northton and some on Toe Head developed in an amphibolite band which might conceivably represent a Scourie-Dyke amphibolite. More open, and probably later, asymmetric folds occur throughout the Leverburgh belt; their axial surfaces dip south westwards. The large scale changes in the strike of the metasediments on Toe Head which can be seen on map 3 have the same sense as these smaller late structures, and more detailed description of these structures is given later.

Throughout the succession there are ill defined zones in which the rocks lose their obvious metasedimentary aspect and begin to resemble mylonites. The "mylonite-zones" are parallel to the foliation of the sediments, sometimes folds and new planar fabrics are developed within them in the same way as they are in the largest shear zones within the Pyroxene Granulite. The largest "mylonite zone" forms the contact between the Pyroxene Granulite and the Leverburgh Gneisses and is well seen near the temple on the south coast of the Toe Head peninsula. South of Northton, where the "Granulite foliation" is highly amphibolised and the two rocks are sometimes difficult to tell apart.

Within the metasediments are pods (boudins) and bands of garnetiferous amphibolite, and larger pod-like bodies of Pyroxene Granulite (e.g. south of Northton and on Toe Head, see map 3). The foliation within these pods is almost always oblique to the foliation of the metasediments and is sigmoidally deformed (Fig. 3). The sense of movement indicated by this curving of the foliation within the detached masses is dominantly, but no consistently, dextral. These rotated blocks of igneous (?) rock provide, after the foliation itself, the most important structural elements of the Leverburgh Gneisses.

Some metres from the contact of the Leverburgh Gneisses with the Pyroxene Granulite, within and parallel to the foliation of the gneisses themselves, is a thick (up to 50 mts) band of garnetiferous amphibolite which follows the contact all the way across the Toe Head peninsula. The exposure is imperfect, but the band appears to be boudinaged along its length. It is internally foliated and the orientation of this internal foliation is consistently oblique to the general trend of the band (and to the trend of the foliation of the gneiss) by as much as 80°. The sense of this obliquity is always the same, the internal foliation runs E. - W., the band itself N.W. - S.E. A similar band of garnetiferous amphibolite follows the contact of the Tonalite mass, but it is very poorly exposed.

FIG 3 A sheared block from the Leverburgh Gneisses

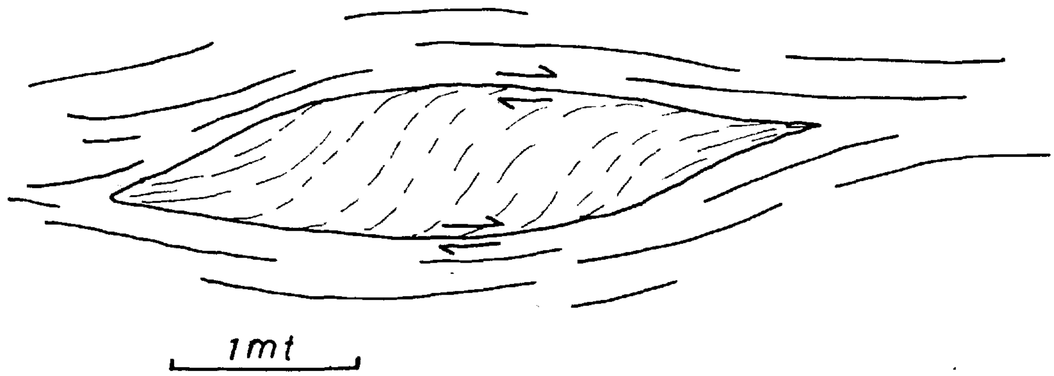
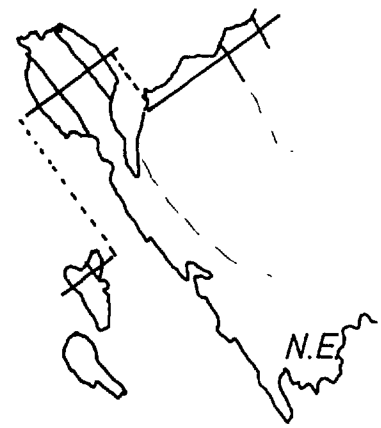
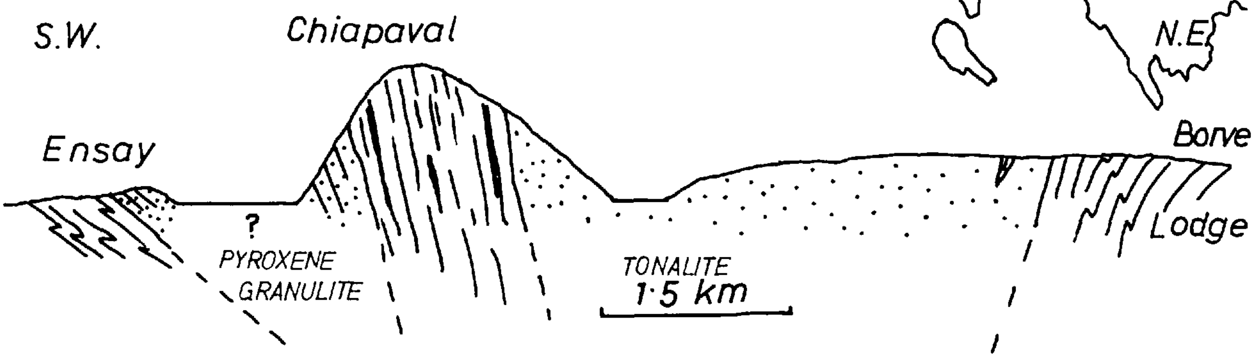
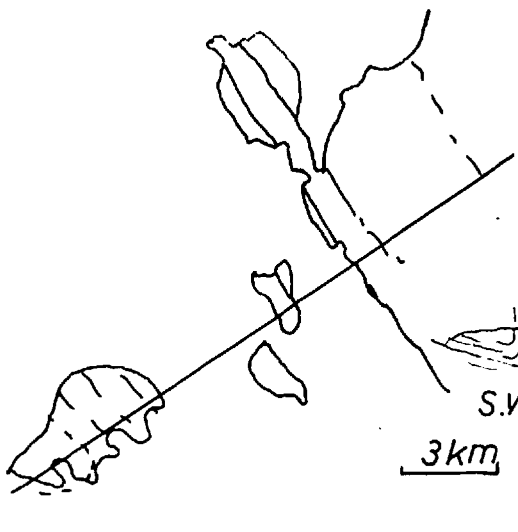


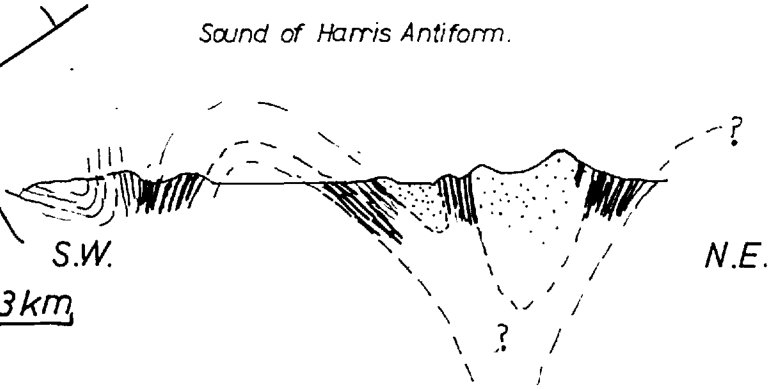
Fig 4 Sketch sections across parts of South Harris (inset maps show lines of section).





South Harris and the Sound of Harris: metasediment-derived rocks heavily shaded.





Gneisses lying between the amphibolite and the Tonalite are usually strongly mylonized.

Petrography of the Leverburgh Gneisses

Leverburgh metasediments unaffected by "mylonization" carry varied mineral assemblages which are fully discussed by Davidson and Dearnley (op cit). In the Northton area the rocks often carry fresh porphyroblasts of biotite, kyanite and garnet. These minerals overgrow the foliation, though it is also flattened about them to some extent and the garnets usually have retrogressed rims. Cracks in the garnet crystals are consistently oblique to the foliation. The textural relationships of the porphyroblasts indicate that all three minerals are stable together. Traced into the "mylonite" zones, however, the rocks undergo significant mineralogical changes, biotite remains stable but the kyanite and garnet are both retrogressed, the latter to pseudomorphic clots of amphibole and chlorite, the former, apparently, to felspar. There is a reduction in the general grain size of the rock and, especially near the contact with the Pyroxene Granulite, biotite locally becomes aligned in a new planar fabric parallel to the axial planes of the minor folds which occur in this zone. Small, new garnets locally grow in the "mylonite" zones.

8:2:4 The Tonalite

Like the Pyroxene Granulite, the Tonalite possesses an early "Granulitic Foliation" which is cut by shear zones in which a new foliation is developed. On the Toe Head peninsula and east of Northton this "Granulitic Foliation" is vertically dipping and trends NE.-SW. On the west coast of Harris, near Scarasta, the "Granulitic Foliation" cannot be seen and the fabric of the Tonalite is isotropic outside the shear zones. Within a few metres of the northern contact the "Granulitic Foliation" reappears, though it is mostly overprinted by the enormously thick zone of shearing which forms the contact in this area. Inland exposure on the Tonalite is very poor indeed. It seems that the "Granulitic Foliation" curves out of its NE-SW. trend in the northern part of the mass, i.e. it is deflected dextrally into the contact shear zone, but this observation is based on scanty evidence.

The Tonalite contains numerous amphibolite bands (dykes?) which bear the same relationship to the "Granulitic Foliation" as do the amphibolite bands within the Pyroxene Granulite to the "Granulitic Foliation" there i.e. they were in existence before the foliation developed and they are deformed. It should be noted, however, that their deformed state is not always obvious at first sight since they are not always folded nor are they boudinaged, presumably this is because they were incompetent with respect to the tonalite.

The Tonalite carries a very large number of shear-zones which are in every way comparable with the shear-zones of the Pyroxene Granulite except that their sense is, with one or two minor exceptions, consistently dextral all over the western half of the mass (I have not examined the eastern half); also their orientation is rather more irregular than that of their counterparts in the Pyroxene Granulite. The northern and southern contacts of the tonalite are large and complex shear-zones, and both show re-folding locally within them. Within the belt of sheared rocks near the northern contact of the mass is an inclusion of metasediments some tens of metres wide on the coast and persistent for some distance inland. Dearnley called these rocks the "Bay Stengie series".

Near the two contacts of the Tonalite are bands of brecciated rock. The origin of this material is obscure, it may be either tectonic or igneous and is now strongly deformed. On the north-east coast of the Toe Head peninsula there is some evidence that the Tonalite may have been a composite intrusion, for xenoliths of fine, dark "tonalite" occur (and are deformed) within coarser, lighter tonalite.

Large garnets(up to 2 cms in diameter) are exceedingly abundant in many parts of the Tonalite. On the Toe Head peninsula they overgrow the Granulitic Foliation, but unlike the garnets in the metasediments, the cracks within individual crystals are usually parallel to the Granulitic foliation (the cracks in the garnets in the Leverburgh metasediments are consistently normal to the foliation in those rocks). Though obviously later than the Granulitic foliation, the garnets in the Tonalite are completely retrogressed in the shear-zones, i.e. they are older than the shear-zones.

8:2:5 The Langavat Gneisses

The Langavat gneisses form a NW-SE. trending belt lying to the north of the Tonalite mass, between it and the migmatitic gneisses which occupy the northern part of South Harris. They are regular fine-grained, flaggy biotitic or hornblendic gneisses and contain horizons of pelite, quartzite and marble as well as amphibolite. Except that they contain a higher proportion of obvious metasediment, they are identical with the fine grained biotitic rocks of the Sound of Harris (see Chapter 2).

The Langavat belt is some 1,000 metres wide in its coastal exposures on the west coast of South Harris, and my detailed examination of it is restricted to these exposures. The northern boundary of the belt is gradational over a short distance, the characteristic fine grained flaggy gneisses giving way to more granitic acid gneiss with abundant granite sheets in the neighbourhood of Borve Lodge (see Myers, 1968). The southern contact of the belt is not exposed on the coast.

Structure of the Langavat gneisses

The Langavat gneisses dip south-westwards, the dip gradually steepening towards the south west. Minor folds are common and most folded layers approach similar geometry in profile (i.e. they are strongly flattened). The age relationships of the minor folds of the Langavat Belt are a problem. One conspicuous set has more or less constant "z" profiles throughout the coastal exposure and a gneissic cleavage commonly runs parallel to the axial surfaces of the folds. Planar fabrics in amphibolites (probably Scourie-Dyke amphibolites) are folded by these structures. There are, however, other folds, usually fairly open, and usually with an opposite sense of asymmetry to the "z" folds described above (i.e. their axial surfaces dip southeastwards more gently than the general trend of the banding, instead of more steeply than the banding). These opposed senses do not seem to belong to minor folds of a single generation around a major structure, the two sets of folds appear to be distinct, and of different ages.

The linear elements of the Langavat Gneisses are deformed in the plane of the foliation, and plotted on a stereogram they describe a partial great circle about poles to foliation (c.f. Myers, 1968). Small scale deformed lineations visible on the scale of single outcrops are not usual; as in north-west Pabbay, the curves of the lineations are fairly large scale phenomena.

8:2:6 The Anorthosite

I have made no systematic study of this rock mass, and the absence of one is the most serious omission from this work. The following notes are the result of two brief visits.

(i) The body is elongated NW-SE.

(ii) There is an internal banding which is often coarse and may possibly be original, and this is cut by dyke-like bodies of amphibolite. In the centre of the body these dykes are apparently undeformed.

8:3 DEARNLEY'S WORK

The terminology is Dearnley's as far as possible. The conclusions of Dearnley's 1963 paper are as follows.

(i) Three distinct periods of folding and metamorphism can be recognised in South Harris.

(ii) The first period of granulite facies metamorphism is seen in the pyroxene granulites and is correlated with the Scourian of the Scottish mainland. Relic traces of pre-Laxfordian metamorphism and folding can be seen in the Leverburgh paragneisses.

(iii) The South Harris igneous complex (i.e. the tonalite, anorthosite and garnetiferous amphibolites) is later than, and not co-magmatic with, the pyroxene granulites.

(iv) The South Harris igneous complex represents a non-orogenic igneous series of chemically closely related rock types essentially contemporaneous with the Scourie-Dykes of the mainland and their representative in the Hebrides.

(v) After the intrusion of this suite a second period of granulite facies metamorphism, associated with folding along NW-SE axes (the Early Laxfordian) affected the whole area. The structure of the Langavat belt is antiformal, plunging to the south-east. The north westerly plunge of structures in the igneous complex and the Leverburgh belt may have been inherited from the older Scourian structural pattern.

(vi) A third period of retrograde metamorphism and NW-SE folding (the Late Laxfordian) then followed; it was associated with the formation of the injection complex (called the Northern Migmatitic Gneisses in this thesis) and the development of zones of retrograde metamorphism in the igneous complex and the Langavat paragneisses. (Apparently there is a mis-print in this statement, for earlier in the paper the Late Laxfordian folds are described as NE-SW trending; they are called "cross folds" and drawn on the map in this orientation).

The interpretations presented in this thesis differ from the second, third and fifth of Dearnley's conclusions, and from the sixth if my supposition of a printing error is correct. The nature of the differences is described below.

<u>Dearnley's conclusion</u> (ii)(concerning the relict Scourian features).

It is contended here that the large crystals of pyroxene and plagioclase in the Pyroxene Granulite are original igneous phenocrysts rather than metamorphic porphyroblasts. The texture of the rock is thoroughly porphyritic, the large crystals are large euhedral or subhedral laths showing signs of straining related to the first phase of deformation affecting the rock mass, and some of the plagioclases appear to be zoned. These seem to me igneous rather than metamorphic features. I have seen no evidence of relict pre-Laxfordian structures or mineral assemblages in the Leverburgh para-gneisses. These rocks have suffered very profound Lax-fordian metamorphism and deformation.

Dearnley's conclusion(iii) (that the Pyroxene Granulite is different from the other metamorphosed igneous rocks of South Harris).

This conclusion is based on geochemical evidence (which is not disputed here) and on apparently different relation-ships between the igneous bodies and the amphibolite bands within them. These different relationships are not recognised in this thesis. The amphibolite bands in the Tonalite on the one hand and the Pyroxene Granulite on the other seem to bear similar relationships to the granulitic foliation in both bodies, namely that they are sometimes boudinaged or folded so that the foliation is axial planar to the folded bands. Commonly, however, they are merely oblique to the foliation and neither boudinaged nor folded; perhaps they were incompetent with respect to the igneous masses. Both sorts of relationship occur in both the Tonalite and the Pyroxene Granulite, if anything the last is more common in the Tonalite.

As a digression, it may be noted that if these garnet-iferous amphibolites are equivalent to the Scourie-Dyke amphibolites, then the granulitic foliation in the two igneous bodies is Laxfordian in age. They do not quite resemble the Scourie-Dyke amphibolites elsewhere in the Hebrides, however, and proof of their age is lacking. They, and the "Granulitic foliation" of both the Tonalite and the Pyroxene Granulite might be Scourian. In this thesis, however, the amphibolite bands are taken to represent Scourie-Dykes, and all the igneous rocks are assumed to belong to the same general phase of igneous activity as the Scourie-Dyke suite, probably slightly early in an intrusive sequence, (c.f. Dearnley, op cit).

Dearnley's conclusion(v) (An early Laxfordian Granulitic facies metamorphism).

Discrepancies here are very minor. I do not envisage a phase of regional granulitic facies metamorphism, but rather local recrystalization in the granulitic facies brought about by local deficiencies of water which may be related to variations of rock type. The matter is discussed more fully in an appendix to this chapter.

Dearnley's conclusion(vi)(Late Laxfordian cross folds).

The changes in strike of the Langavat belt are not regarded here as folds in the true sense, but as accommodations of the paragneisses to shape variations of the Tonalite. Minor folds with a N.E. - S.W. orientation do not occur.

General points

Many features given emphasis in this thesis are not mentioned in Dearnley's paper. Most have a structural bias, e.g. the mode of deformation of the early (Granulitic) foliation of the igneous rocks; the existence of shear zones; and the obliquity between the internal foliation of the igneous bodies and the foliation of the paragneisses. Some of Dearnley's structural observations have not been reiterated in this work, e.g. the large scale isoclinal folds which Dearnley postulated in the Leverburgh and Langavat (c.f. Myers, 1968) belts have not been recognised.

8:4 INTERPRETATION OF THE STRUCTURE OF SOUTH HARRIS IN RELATION TO THAT OF NORTH UIST AND THE SOUND OF HARRIS.

8:4:1 General Structural pattern

The major structure of the southern part of South Harris is synformal (Fig.4). The gneisses of Ensay and Killegray in the gently dipping limb of the Sound of Harris Antiform dip north-eastwards beneath the Pyroxene Granulite, while the Langavat gneisses dip south-westwards beneath the Tonalite. The Berneray Synform and Sound of Harris Antiform are asymmetrical structures whose sense is compatible with a parasitic relationship on a larger scale synformal structure in South Harris. The structure may not be a synform in the true sense, however, the synformal geometry may be brought about by the accommodation of the Langavat and Sound of Harris gneisses around the margins of the igneous masses. In either case it may well be that the Langavat and Sound of Harris rocks are parts of one belt continuous in depth beneath the igneous bodies, a possibility suggested by the remarkable litholog-ical similarity of the two groups of rocks (see Chapter 2).

The elongation of the igneous masses, the orientation of the shear zones within them and the trend of the synformal structure itself are all similar to the trend of the F3 structures of North Uist and the Sound of Harris, and this evidence, together with the congruent relationship of the major F3 folds of the Sound of Harris suggests an F3 age for all these structures.

It was shown in chapter 2 that the F2 folds of the Sound of Harris are refolded across the two major F3 folds in that area, and that their sense of asymmetry remains consistantly 'Z'. The earlier of the two sets of folds in the Langavat belt have a similar style and asymmetry to these F2 folds of the Sound of Harris. We have already seen that the Langavat folds fold an earlier foliation in amphibolite bands which may well be Scourie-Dyke amphibolites, and thus they may also be of F2 age. It may be that a suite of F2 folds in the paragneisses is re-folded beneath the South Harris synformal structure.

The major igneous bodies of South Harris and the Laverburgh gneisses lie within the main synformal domain, and the deformation state of each of these rock groups is now examined in turn.

8:4:2 <u>The Pyroxene Granulite</u>

In each of the separated fragments of exposure of the Pyroxene Granulite the orientation of the "Granulitic Coliation" and the senses of movement in the shear zones are different. Between the Toe Head peninsula and Ensay an antiformal culmination of the "Granulitic Foliation" is indicated by the opposed senses of the minor folds produced in the "Granulitic Foliation" as it enters and leaves the shear zones. This structure, together with the different senses of movement in the shear zones of Ensay and Toe Head indicates a north-westerly movement of the Pyroxene Granulite relative to the acid gneisses on either side. On Renish point (the southernmost corner of South Harris) the trend of the "Granulitic Foliation" and the predominant sense of shearing suggests a relative south-easterly movement of the Pyroxene Granulite mass in that area. One possible interpretation of the complete pattern of deformed "Granulitic Foliation" is that it defines a series of concentric convex curves facing north-westwards in the N.W., and perhaps also south-eastwards in the S.E. The shear zones seem to form an integral part of this pattern, lying on the flanks of the concentric ellipses of foliation. The senses of movement of the shear zones, and the general pattern of the deformed foliation (if the interpretation above is correct) suggest that the Pyroxene Granulite has suffered a flattening deformation and has extended in at least two directions. It was noted above that this deformation is probably equivalent to the F3 deformation of North Uist and the Sound of Harris.

The pattern of the deformed "Granulitic Foliation" suggests that before this F3 deformation it lay in a more or less N.E.-S.W. orientation, similar to the postulated original trend of the F2 structures in North Uist and the Sound of Harris. It therefore seems likely that the Granulitic foliation is of F2 age. This foliation probably formed mainly by the mechanical re-orientation of the pyroxene and plagioclase phenocrysts, though there is local evidence (e.g. corrosion of the pyroxenes) that it was associated with incipient amphibolite facies metamorphism. In the area south of Northton where the "Granulitic Foliation" is parallel to the sides of the body it is a thoroughly amphibolized fabric, but this might have been brought about by F3 deformation in that area (see below).

The structural history of the Pyroxene Granulite might be interpreted as follows. An early phase of deformation, probably equivalent to the F2 deformation of North Uist and the Sound of Harris produced a N.E. - S.W. trending foliation throughout the mass. A second phase of deformation, probably the F3 deformation of the areas further south deformed this foliation and probably produced a profound change in the shape of the body. The mass was flattened and extended and the foliation perhaps deformed into a series of concentric ellipses (ellipsoids in three dimensions?) whose dimensions progressively increased outwards from the centre in the directions in which the body was elongated. Initially this deformation may have been accomplished in a truly homogeneous manner, but eventually plastic failure occurred and displacements were taken up in discrete shear zones where new foliations were developed in response to locally very intense strains. Perhaps this more homogeneous deformation occurred as the strength of the pyroxene granulite was exceeded. Where the early foliation had been brought parallel to the sides of the body, however, (e.g. South of Northton) shear zone development was restricted and movement taken up along the old foliation planes.

8:4:3 <u>The Tonalite</u>

The northern and southern boundaries of the Tonalite, though enormously complicated, are essentially dextral shear zones. The internal "Granulitic Foliation" is sigmoidally deformed, though its "S" form is extremely asymmetrical; the curve of the N.E. -S.W. trending foliation into the shear zone at the southern contact being very sharp and taking place over a metre or so, while the curve into the northern shear zone contact is much more gradual. The shear zone at the northern contact is much broader than its counterpart in the south, and we may associate the asymmetry of the sigmoidal internal foliation with more intense deformation at the northern (i.e. the external) contact of the mass.

The shape of the Tonalite mass is almost identical with the shape of the detached blocks of the Leverburgh belt; like them it thins to a "tail" at its end where the internal foliation is parallel to the foliation of the gneisses of the matrix. Like the Pyroxene Granulite the contaots of the Tonalite appear to converge in depth, but unlike that body the sigmoidal form of the deformed "Granulitic foliation" suggests an important differential shear component on either side of the body, a feature also indicated by the consistent sense of the shear zones throughout the intrusion.

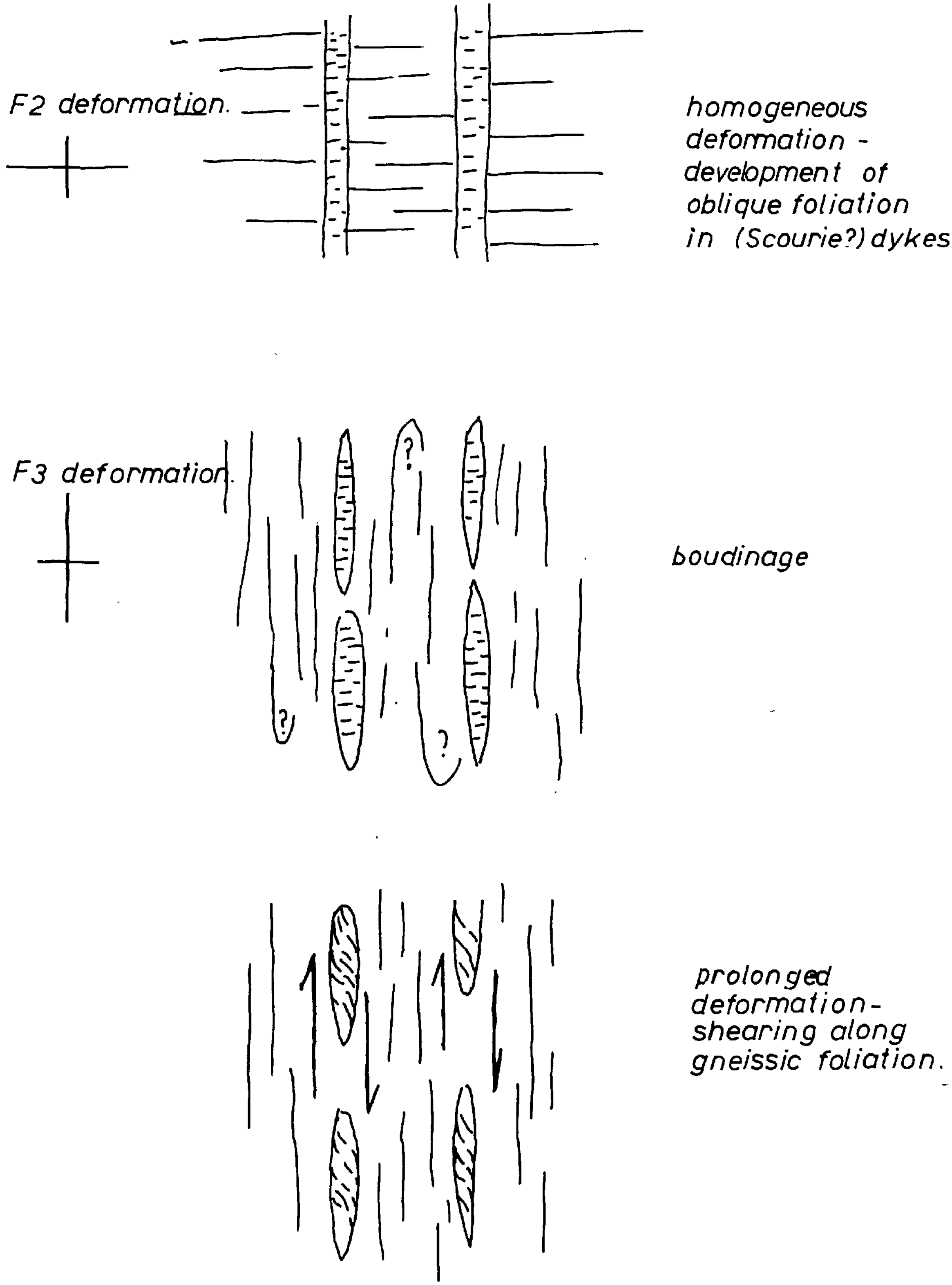
As in the Pyroxene Granulite the shear zones trend in broadly the same direction as the F3 structures of North Uist and the Sound of Harris, while the trend of more or less unmodified "Granulitic foliation" is N.E.-S.W., the same as the original trend of F2 structures further south.

8:4:4 The Leverburgh Gneisses

Since folds are relatively unimportant, the key to the structure of the Leverburgh belt lies in the boudinaged and apparently rotated blocks of garnetiferous amphibolite within it, and a hypothesis of the origin of these structures is now given. Structures such as those shown in Fig.3 might be interpreted as simple rotated blocks whose internal foliation has been moved significantly from its original orientation. The existence of more or less continuous concordant bands of this garnetiferous amphibolite whose internal foliation is consistently oblique to the foliation in the acid gneisses, however, makes it seem likely that the detached blocks represent the boudinaged variants of these obliquely foliated bands. The orientation of the internal foliation of both the blocks and the bands is reasonably consistent and very close to the orientation of the "Granulitic Foliation" of the Pyroxene Granulite on Toe Head, one would not expect this consistency if a suite of blocks of different shapes and sizes were individually rotated within the matrix.

The development of oblique internal foliations in members of a gneissic succession is not an easy phenomenon to explain; one reasonable possibility is shown in Fig.5 where a discordant band behaves in an incompetent manner within a deforming matrix. It would appear that before a foliation developed within them, the amphibolite bands possessed a N.W.-S E. trend (parallel to the original trend of the Scourie-Dyke amphibolites in North Uist and the Sound of Harris). On deformation they developed a foliation almost normal to their length and trending originally in an E. - W. or N.E. - S.W. direction (the F2 direction), though the bands themselves, perhaps by virtue of their relative incompetence did not fold, but were merely homogeneously thickened. F3 deformation brought the foliation of the Leverburgh Gneisses into a N.W.-S.E. orientation (see later in this chapter), but the orientation of the amphibolites was such that it remained relatively unchanged. F3 flattening caused boudinage of the amphibolite bands (unfortunately implying a change in the relative competence of metasediment and amphibolite with time) and differential movement in the foliation planes of the gneiss caused shearing around the margins of the amphibolite blocks. The internal foliations were sigmoidally deformed, dextral shearing causing the orientation to change from

Fig 5 Possible origin of sheared blocks



N.E. - S.W. to E. - W. or W.N.W. - E.S.E., while sinistral shearing rotated internal foliations nearer to N.-S.

The occurrence of both dextral and sinistral movement senses around these deformed blocks is puzzling; I can see no clearly defined pattern in it. It may be that the shearing movements in the Leverburgh Gneisses were of an inhomogeneous nature on small scale (the large scale movements are apparently more easily understood - see later).

8:4:5 Interpretation of the Structural History of Southern South Harris.

The problems involved in the interpretation of the Leverburgh Gneisses occur again when the structural evolution of the southern South Harris area as a whole is considered.

Initially it would seem logical to interpret the igneous masses with internal foliations oblique to those of the acid gneiss matrix as blocks which have rotated within that matrix. Gay has shown, however, that bodies capable of free rotation within a matrix rotate at different rates depending on their shape, and since in Harris the internal foliations of igneous bodies with widely differing shapes can be parallel to each other, an explanation of the structures in these terms is impossible. Ramsay (1962a) has suggested that during a flattening deformation foliation may be rotated around spherical objects which themselves remain fixed and do not rotate, and here seems to lie the answer to the problem of the major structure of South Harris. The following hypothesis of the structural history of the area is coined in these terms.

At the end of the phase of deformation which produced the "Granulitic Foliation" in the igneous rocks (probably the F2 deformation) the foliation of the igneous and the sedimentary rocks may have been nearly parallel. Metamorphism continued after deformation had ceased, and porphyroblasts of biotite, kyanite garnet and other minerals grew across the F2 fabric of the Leverburgh Gneisses. In the Tonalite a similar post kinematic metamorphism occurred, and garnets grew across the "Granulitic Foliation" of that body. Dearnley records similar occurrences in the Pyroxene Granulite.

The F3 deformation which followed this static metamorphism produced the structural architecture of South Harris as we now see it. In the early part of this phase-re-orientation of the metasedimentary rocks occurred around the igneous bodies, perhaps as the South Harris synformal structure developed. The igneous rocks probably did not change their orientation during this massive rotation of the paragneisses, and may therefore have been either spherical or elongated N.W.-S.E. before the F3 deformation commenced. Perfect sphericity is doubtful in igneous bodies of this scale, especially since they had suffered a previous deformation, and it is logical to assume that they were originally elongated in the direction of the Scourie-Dyke swarm (NW-SE) and that this basic shape persisted through the F2 deformation.

The internal cracks in the post F2 garnets might record the rotation of the paragneisses in much the same way as do the Granulitic foliations of the igneous rocks.

The cracks are mostly alligned NE-SW and are parallel to foliation in the Tonalite (perhaps their orientation was influenced by the orientation of the fabric which they overgrew), but normal to the foliation of the Leverburgh Gneisses. They also may have remained static during the rotation of the gneisses about them.

Intense regional flattening followed this rotation of the paragneisses into the major structure and the "Granulitic Foliation" of the igneous rocks was deformed, that of the Pyroxene Granulite during flattening and extension of the body, that of the Tonalite by differential movement of the gneisses on either side. Amphibolite bands in the Leverburgh Gneisses were strongly boudinaged. Differential movement within the Leverburgh belt, probably induced by the shape changes of the igneous bodies on either side, was taken up along the foliation planes and mylonite zones developed where this displacement was particularly marked, e.g. along the margins of the newly formed (and doubtless still developing) boudins, and along the contacts of the igneous masses. Thus the shear zones in the igneous masses and the mylonite zones of the Leverburgh Gneisses are broadly complementary structures. Amphibolite facies metamorphism accompanied these movements, though metamorphic reactions were only completed in regions where the strain was particularly intense (e.g. garnets show retrogressed rims throughout the Leverburgh Gneisses, but are only completely altered in the mylonite zones such as that along the contact with the Pyroxene Granulite). The flattening of the foliation around the post F2 porphyroblasts may have occurred at this time.

It seems that the role of the Leverburgh Gneisses during this deformation was a passive one as they accommodated themselves to the shape changes in the igneous bodies. Evidence for this statement is apparent from the form of the late asymmetrical folds which may be seen in various places in the Leverburgh Gneisses, and a description of these structures follows.

On the Toe Head peninsula the swing of strike of the "Granulitic Foliation" of the Pyroxene Granulite as it enters and leaves the shear zones are duplicated on a larger scale. In the southern part of the outcrop the Granulitic Foliation trends WNW-ESE, curving to NW-SE in the central part, then back to WNW-ESE in the north near Liuri. In the south of the peninsula at least, the contact of the Pyroxene Granulite mass curves in sympathy with the foliation, and the change in orientation persists across the strike of the Leverburgh Gneisses and perhaps also affects the contact of the Tonalite (exposure is inadequate here). The axial trace of this structure and of smaller scale folds congruent with it runs NNE-SSW. A possibly complementary warp is seen near Toe Head itself, where the Leverburgh Gneisses gradually curve from NW-SE nearer to WNW-ESE. The sense of these structures indicates that the north-eastern part of the Leverburgh belt moved relatively to the north-west against the Tonalite, while the south-western part of the belt moved relatively to the south-east against the Pyroxene Granulite. This large scale movement pattern cannot be related to any fold in the Leverburgh belt, but is directly related to the relative movements of the two large igneous bodies. The small scale shearing movements around amphibolite pods within the Leverburgh Gneisses do not fit in with this general picture, and further examination of these structures is required. As noted earlier small scale shearing movements within the Leverburgh Gneisses may have been inhomogeneous.

A final problem of South Harris geology is the different modes of deformation of the Tonalite and the Pyroxene Granulite, the former showing the effects of differential shear, the latter the effects of flattening. It may be simply that the Pyroxene Granulite is more strongly deformed. The sigmoidally deformed internal foliation of the Tonalite might, if it were strongly flattened, resemble the more symmetrical pattern of the deformed "Granulitic Foliation" of the Pyroxene Granulite. This would not, however, explain the consistency of sense of the shear zones in the Tonalite and the opposed senses of movement of the shear zones in the Pyroxene Granulite, and the real answer may be that the two intrusions were originally of different shapes.

The structural history of South Harris is summarized on Table 1, and diagramatically on Fig. 6.

Fig 6 Highly schematic representation of the deformation history of South Harris

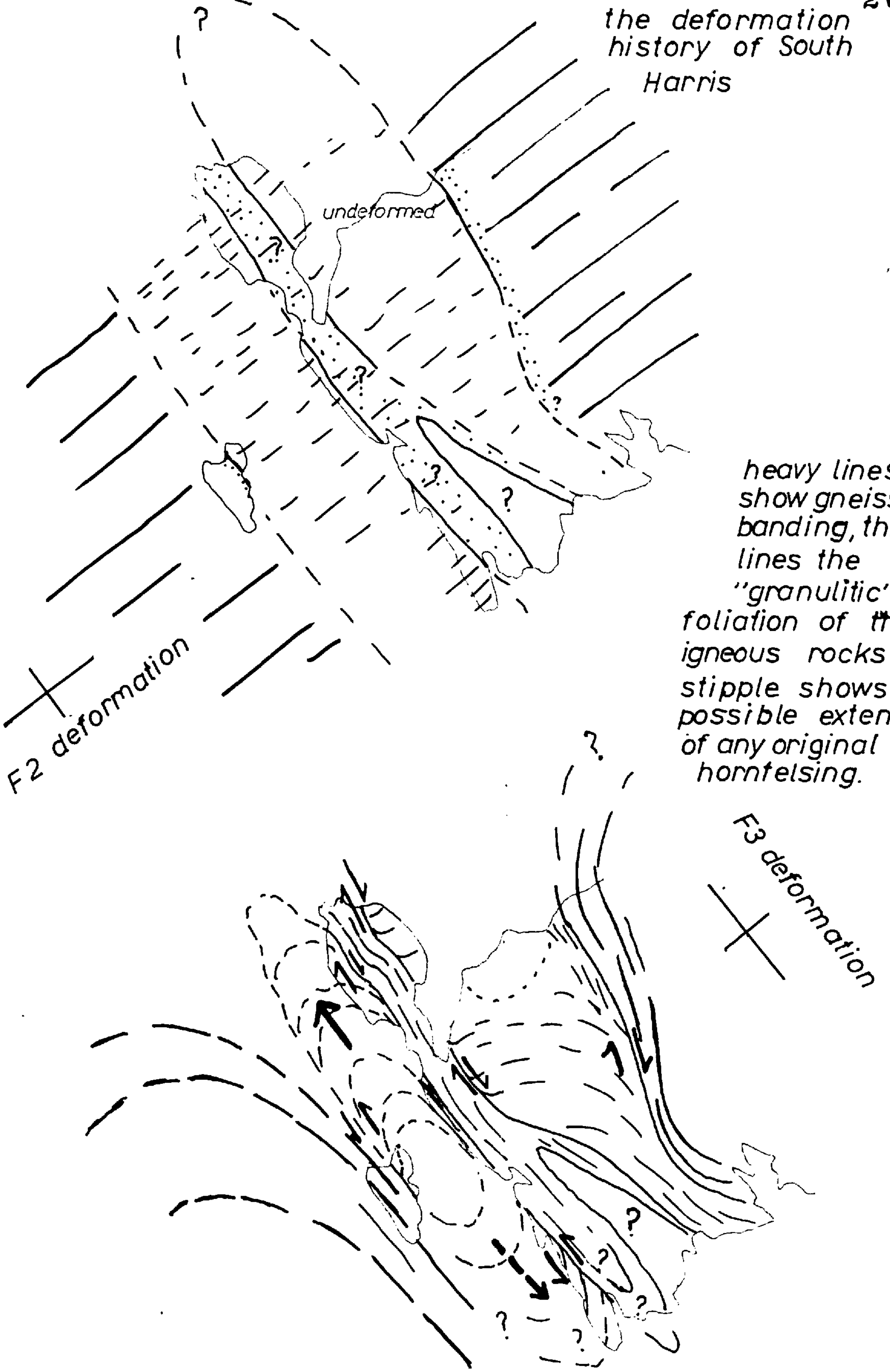


TABLE I

post F3	Brittle structures, faults, joints, some pseudo-tachylite.
?	Growth of new garnets in mylonized Leverburgh Gneisses.
	Regional flattening.
	Deformation of the igneous bodies development of shear-zones and mylonite zones.
F3	Boudinage of amphibolites in the Leverburgh Gneisses.
	Initiation of South Harris synformal zone and consequent rotation of the paragneisses around the igneous bodies, mylonization; amphibolite facies metamorphism.
	Growth of post-kinematic porphyroblasts.
F2	Development of fabrics with N.E.-S.W. strike in the igneous rocks. Scourie dykes are cross foliated. Folds in the Langavat and Sound of Harris Gneisses with "Z" asymmetry.
F1	The only evidence of F1 deformation is the foliation in possible Scourie-Dyke amphibolites in the Langavat belt.
	Intrusion of igneous rocks and Scourie-Dykes.
	Banded Scourian paragneisses.

8:5 APPENDIX

8:5:1 The Significance of the Paragneisses

Metasediments are better developed in South Harris than anywhere else in the Hebrides. They are probably ultimately of Scourian age but the reason for their sometimes remarkable state of preservation is very much a matter of debate (see Coward et al 1969). There is in many parts of the Hebrides an association between metasediments and igneous bodies, and nowhere is this more apparent than in South Harris where the largest igneous bodies are associated with the largest masses of metasediment in the Hebrides. It is possible that rocks adjacent to igneous intrusions suffered an early contact metamorphism which may have rendered them more resistant to later gneissification. Coward et al (op cit) and Myers (1968) have stressed that most of the gneissification of the metasediments of the Hebrides is of Scourian age, and there is good evidence for the idea in many localities. In South Harris, however, there is evidence to suggest the post-Scourian age of the igneous rocks and perhaps the importance of Laxfordian metamorphism in the formation of acid gneiss from metasediment has been underestimated.

My personal views on the South Harris paragneisses are as follows:-

(i) The Sound of Harris, Langavat, and Leverburgh Gneisses formed, in Scourian times, one important elongated belt of metasediments of various lithological types but distinct from the regional acid gneiss.

(ii) The large igneous bodies of South Harris were intruded into this metasedimentary belt and the metasediments in South Harris were hornfelsed, those lying between the two largest igneous bodies (now the Leverburgh Gneisses) being most seriously affected, those lying further away (now the Sound of Harris Gneisses) not affected at all.

(iii) Hornfelsing may anneal rock, and probably thereby reduce its permeability. Mineralogical changes brought about by means of pore water during Laxfordian metamorphism would be slower in the hornfelsed rocks, and the degree of alteration of the metasediments therefore would be dependent on the extent of the hornfelsing, thus true metasediments are rarer in the Sound of Harris than they are in the Langavat belt, otherwise the "metasediment derived" biotitic gneisses in the two areas are identical. The most abundant well preserved metasediments are in the Leverburgh Gneisses lying between two igneous bodies, where the early contact metamorphism was presumably the most intense.

8:5:2 <u>The Granulite Facies in South Harris</u>

Obviously the remarks made above concerning contact metamorphism, permeability, and the preservation of metasediments are also likely to be of significance in a consideration of the state of regional metamorphism of the South Harris rocks. Although I have not observed it personally both Dearnley and Davidson record orthopyroxene in some of the Leverburgh gneisses, and Dearnley records that some of the pyroxene in the Pyroxene Granulite crystallised later than the deformation which produced the "Granulitic

Foliation" (post F2). These observations, especially the former, suggest that at least one metamorphism of South Harris belongs in the Granulitic facies of the present scheme of classification of metamorphic rocks, and this conclusion is reached by both Dearnley and Davidson.

We need not, however, envisage a phase of regional Granulitic Facies metamorphism. An attractive hypothesis for the origin of these high grade mineral assemblages follows Buddington (1963) who suggests that igneous masses metamorphosed in a partly open system at temperatures equivalent to those of the amphibolite facies will, in their marginal areas where water has permeated, be metamorphosed in the amphibolite facies, with pyroxene crystallizing only when all the available water has been used up. In the central parts, however, where water has not permeated, any recrystallization will be in the granulite facies (see also chapter 5). Acid gneisses with high permeability would, under the above conditions, be metamorphosed in the amphibolite facies. Rocks adjacent to the igneous bodies, annealed by earlier contact metamorphism would be relatively impervious, however, and once the water already contained within them had been used up in recrystallization of biotite and amphibole, they, like the igneous rocks would begin to recrystallize in the granulitic facies. Evidence of such incipient granulite facies metamorphism is seen in the Leverburgh Gneisses where the presence of a garnet-kyanite assemblage (the post-F2, pre-F3 assemblage) probably indicates the presence of high confining pressures at this time. Here, then, is more evidence compatible with Winkler's (1967) and Buddington's (1963) suggestions that the crystallization of ortho-pyroxene at temperatures equivalent to those prevailing in the amphibolite facies metamorphism necessitates that P_s is greatly in excess of P_{H_2O}.

CONCLUDING DISCUSSION

Structural mapping of the North Uist coast and the Sound of Harris revealed that the Lewisian gneisses of the area have been subjected to three important phases of ductile deformation after their intrusion by a suite of basic dykes. The form of the deformed dykes and the nature of their tectonite fabrics greatly aided the interpretation of the structure and structural history of the complex as a whole. The structural history is essentially one of "re-working" of a pre-existing basement complex.

Usually, deformation was too intense to enable any study of the pre-dyke history of the complex to be made; indeed we usually see concordant amphibolites rather than cross-cutting dykes. The coexistence of areas where discordant relationships between dykes and acid-gneiss banding are still preserved and areas where the two structural elements are completely concordant indicates a very important feature of the deformation state of the area as a whole, namely the overall inhomogeneity of that deformation state.

Once the importance of inhomogeneous deformation was appreciated, it became necessary to examine the situation more closely, and if possible from a quantitative standpoint. Much of this thesis has been devoted, directly or indirectly, to just such an examination. The difficulties of applying quantitative techniques to basement rocks are considerable, however, and the strain determinations made in this thesis do not have the accuracy which we have come to expect from quantitative structural geology in orogenic terrains. Nevertheless, they are important if they only serve to emphasise the inhomogeneous state of deformation in the area, for in a final assessment the most important facets of this work are the identification of strain variations, the astonishingly short distances over which these variations occur, and their relationships with the lithology and metamorphic state of the rocks. Before proceeding with this discussion it is worth summarizing some of the examples of strain variations which have been mentioned during the course of the thesis.

(1) Shear-zones

Shear-zones are the most extreme examples of inhomogeneous deformation. XZ ratios of up to 400:1 were obtained by Ramsay and Graham from the centre of a shear zone cutting completely undeformed metagabbro (in press). This remarkable variation takes place over two or three centimetres, and is associated with a profound variation in the state of recrystallization. Shear zones probably develop under special conditions, presumably in rocks incapable of deforming in a more homogeneous ductile manner. Water content may be a critical factor of these special conditions.

(2) Metasediments and high regional deformation

Strain ratios of 7:5:1 were obtained from metasediment-derived rocks in Pabbay, while 500 metres away quartzo-felspathic gneisses show only slight effects of Laxfordian deformation. A similar situation exists in South Harris where strongly deformed metasediments lie adjacent to less deformed masses of metamorphosed igneous rook.

(3) Intense strain at the contacts of resistant masses.

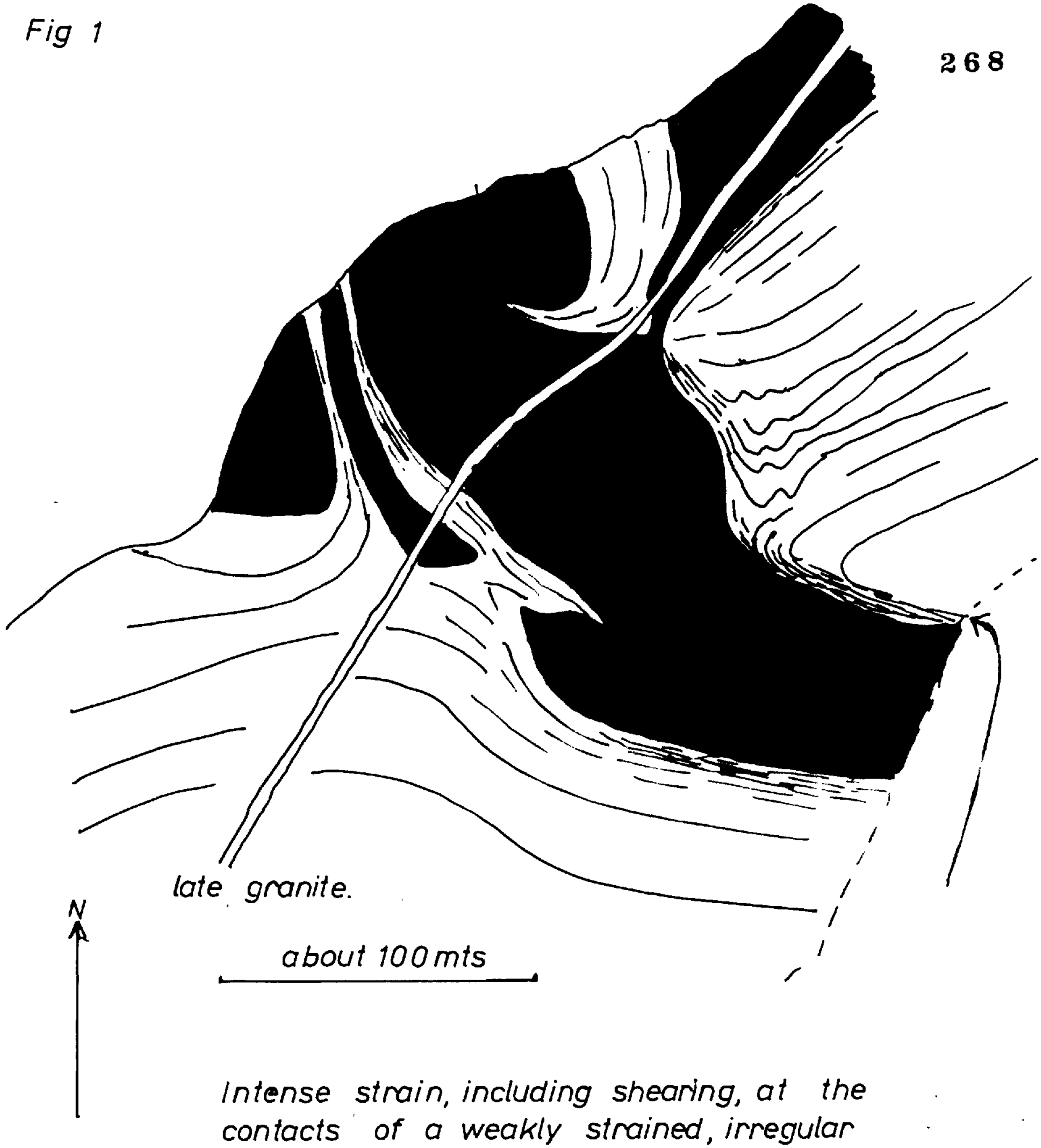
The strong deformation of the metasedimentary belts mentioned above may be in part due to "contact strain" effects as well as to their distinctive lithology. Smaller scale examples of "contact strain" phenomena can be seen at the margins of the Castell Odair basic mass and also (more dramatically) at the margins of scarcely-deformed Scourie dyke amphibolites at Udal in North Uist, where very intense shearing occurs locally along the lithological discontinuity between acid gneiss and amphibolite, (Fig.1).

(4) "Shoot through" structures

These are a special type of strain inhomogeneity. The largest one recognised in the North Uist-Sound of Harris area is the Tigharry Isocline. Strain ratios of 18:6:1 were recorded within the domain of the Tigharry Isocline, while much lower values are evident elsewhere on the North Uist coast (e.g. 3:1:5:1 in the Hougharry Antiform).

Strain inhomogeneities are apparent in the structures of all three deformation phases in the North Uist-Sound of Harris area; those of the third deformation (F3) are usually the most obvious because the lack of later deformation makes the F3 structures easiest to interpret, though variations of F2 strain are well seen, for example, in the Sound of Harris. The inhomogeneities occur on a variety of scales, the largest scale being that of the major F3 structures. The major F3 synforms are relatively tight and are areas of significant F3 deformation, while the complementary major antiforms are broad and show little evidence of F3 deformation. This pattern applies all over the southern Hebrides (see Coward, 1969, Francis 1969,

Fig 1



Intense strain, including shearing, at the contacts of a weakly strained, irregular mass of Scourie dyke amphibolite in the low deformation area at Udal, North Uist. (part of a detailed map)

black - Scourie dyke amphibolite.
lined - acid gneiss banding. spacing is an indication of intensity of shearing.

and Coward et al, in preparation). In the North Uist-Sound of Harris area it is exemplified by significant F3 deformation in the North Uist coast Synform, the Berneray Synform and the South Harris synformal zone, and by a lack of F3 deformation in the antiformal domain of northern North Uist and, more dramatically, in the Sound of Harris Antiform.

The structural pattern in each of the synformal zones is characteristic. The South Harris synformal zone is much tighter than the North Uist coast Synform, the rocks within it are dominantly planar and linear elements are frequently deformed "in plane". Boudinage is a common feature, and the few visible folds are enormously flattened. Conversely a large-scale impression of the North Uist coast is essentially one of linearity, for the rocks are extensively folded on variable axial surfaces about constant axes and F3 boudinage is a rare phenomenon. In a flight of geological fancy we might equate these different large scale impressions with different positions on the strain path which was found appropriate to explain the fabric variations of the North Uist coast, namely superimposition of an oblate regional F3 strain on a surface representing an F1-F2 strain state in an orientation

$$\begin{matrix} X & y \\ Y & z \\ Z & x \end{matrix}$$

The North Uist coast area would represent the halting of the path of the finite strain in the constrictional field of the Flinn diagram, the Leverburgh belt (the central part of the South Harris synformal zone) would lie further along the path in the flattening field.

At the end of this thesis it is appropriate to make use of some of its dominant themes to explain the cuspate pattern and the associated strain variations of the southern Hebridean major structure. Two explanations have already been outlined (Chapter 3), one that the structure may represent a deformed interface analogous with a basement-cover relationship, the other that it is a result of an upward flow of material during the F3 deformation. Francis, in his thesis, presents avariation of the basement-cover hypothesis, envisaging folding of the "cover" and contemporaneous faulting of an ancient basement.

In the Sound of Harris and also in parts of South Uist (Coward, 1969) it is apparent that F3 antiformal cores contain rocks that are not only unaffected by F3 strain but also show very little evidence of any earlier Laxfordian deformation, i.e. the F3 antiforms formed around earlier "deformation lows". In the Sound of Harris Antiform it is clear that the absence of early strain is coincident with a lithological change, the highly deformed rock probably being of metasedimentary origin, the relatively undeformed material being more "normal" acid gneiss of uncertain origin. Similarly the Tonalite and anorthosite masses of South Harris, though lying within a zone of very intense deformation, have cores which are undeformed, save in restricted areas (shear-zones). On a much smaller scale, for example in the basic mass of Castell Odair on the North Uist coast, we see no evidence of penetrative deformation in the centre of the body, though marginally it is strongly deformed and the acid gneiss outside it is plastered about it (effectively folded around it). Bodies of rock which act as resistant masses during deformation are common on all scales throughout the North Uist-Sound of Harris area and seem to occur throughout the Lewisian outcrop, indeed, we may look upon the central Scourian zone of the mainland of Scotland as a very large resistant block which escaped regional Laxfordian deformation. The major structure of the Southern Hebrides may represent the modifications of easily deformed rock about bodies which are more resistant. Ductile gneisses deformed around ellipsoidal resistant masses would appear at some levels of erosion as broad antiformal domains, material "pinched in" between the masses would have synformal geometry at similar levels, and would be very strongly deformed.

Intense deformation in the synforms would then become analogous to intense "contact strain" seen at the margins of resistant bodies (e.g. the Leverburgh and Langavat rocks of South Harris, the gneisses of the Sound of Harris steep zone and the margins of the Castell Odair basic body).

The physical properties which allow one rock type to deform easily yet an adjacent different rock type to remain virtually unaffected are probably complex, and study of them belongs primarily to the field of experimental structural geology. From simple observation, however, we are able to recognise one of them, and may well be the most important. I have tried to emphasise during this thesis the intimate association of water, strain and recrystallization in the rocks of the North Uist-Sound of Harris area. Anhydrous or slightly hydrated rocks either remain largely undeformed, or suffer failure in discrete shear zones; thoroughly hydrated rocks suffer penetrative deformation. Reiterating another conclusion made earlier in this work, it is likely that permeability is a basic control on whether or not a rock deforms in a ductile manner. Igneous bodies may lie as resistant masses within acid gneiss because they are less permeable than the gneiss, biotitic metasedimentary material is more strongly deformed than biotite free acid gneiss.

It is hardly going too far to suggest that if any mode of deformation is characteristic of basement areas in general (one may find examples of it in Scandinavia, Africa and Greenland, as well as everywhere in the Lewisian) then it is the deformation of metasediment derived gneisses around resistant blocks of igneous rock, ancient basement or "primeval crust". The undeformed material occupies domal areas, the genisses deformed around it may be tightly folded into pinched-in synforms between the domes, otherwise they are plastered around them. The different deformation states are due chiefly to differences in lithology, and change over the same negligible distances as changes in rock type occur.

These structures are not necessarily mantled gneiss domes; they are not gravity controlled in the sense that they formed by upward flow of light material, quite the reverse, probably, in the Hebridean examples. At higher levels this sort of structure may find a parallel in such phenomena as the external massifs of the Alps or the Pre-Cambrian Padarn ridge in the Palaeozoic fold belt of North Wales. The Padarn ridge analogy is a good one since the slates adjacent to it are probably the most strongly deformed rocks in North Wales. The Cambrian slate belt might be interpreted as an area of high deformation lying in the centre of the geosynclinal trough, but it may simply represent an increase of regional deformation at the margin of the resistant block. We might look on the Cambrian slate belt as a high level version of the Langavat belt of South Harris.

My final comments concern the very approach to basement geology. With the exception of the North-west Highland memoir (Peach et al, 1907) most of the significant research into basement rocks has occurred in the last twenty years, a period which has also seen the development of radiometric dating and the emergence of a branch of structural geology concerned with unravelling polyphase deformation in orogenic terrains. Sutton and Watsons' study of the Lewisian emphasised the existence of two orogenies. Their conclusion was based on field relationships and was substantiated by radiometric dating.

Since Sutton and Watson the geochronological bias of basement geology has increased, both in terms of radiometric dating and as a result of a consuming desire of some field geologists to discover (at all costs?) enough phases of deformation and metamorphism to cover at least one page of type script.

Sutton and Watsons' concepts of orogeny in the interpretation of basement rocks have also been enlarged upon by recent workers. Sutton and Watson used the term to describe two periods of polyphase deformation and metamorphism separated by a long time interval and an anorogenic phase of dyke intrusion, and in such a context the use of the word "orogeny" might be valid. Since 1951, however the word has unfortunately appeared in descriptions of a single fold phase or of a phase of recrystallization which happens to register a particular radiometric date. This practise is surely a rather dubious one.

The question whether orogenic belts of Caledonian or Alpine type may be recognised in basement terrains has been a controversial subject for at least twenty years and has had continual affirmations (e.g. Holmes 1948) and denials (e.g. Macgregor, 1951, Brock 1959, Gill 1951). My own opinions are on balance those of the latter authors. Admittedly, the Lewisian outcrop of Scotland is geographically restricted, but there is little evidence within it of a Laxfordian orogenic belt, and no single Laxfordian trend.

We may have placed too much emphasis on the chronological aspects of basement geology and too little on its spatial aspects. Useful progress might be made if we dispensed with the rather inflexible, uniformitarian concept of orogeny and reduced our geochronological bias. Metamorphic variations within basements and the structural patterns of basement rocks are often direct functions of variations in rock type. Once we except this, we have a very general and adaptable framework in which special studies such as geochronology and geochemistry find a valuable place.

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Map 1
The North Uist coast
foliation
125 mts
1:5000
N
late fracture
pegmatite
Tertiary(?) dyke
Scourie dyke amphibolite
partially amphibolized
totally amphibolized
metagabbro of the Scourie dyke suite.
anthophyllite (?) gneiss
metasediment
feint stipple
beaches
Castell Odair
GRIMINISH SYNFORM
"Dreary point" (Blaaskie)
GRIMINISH ANTIFORM.
Griminish point
Sloc Roe
Scolpaig bay
SCOLPAIG SYNFORM.
"South Scolpaig"
CLIFFS ANTIFORM
The Cliffs
Natural arch.
Vorlish.
BALELONE SYNFORM.
Raikinish.
Hosta beach
HOSTA AMPHIBOLITE
TIGHARRY ISOCLINE
HOGLAN AMPHIBOLITE.
Tigharry.
Hoglan bay
HOGLAN SYNFORM
SOUTH HOGLAN FOLD
HOUGHARRY ANTIFORM
HOUGHARRY SYNFORM
Hougharry
"Hougharry bay"
Aird an Runaird
R.H.Graham. Ph.d. 1969.

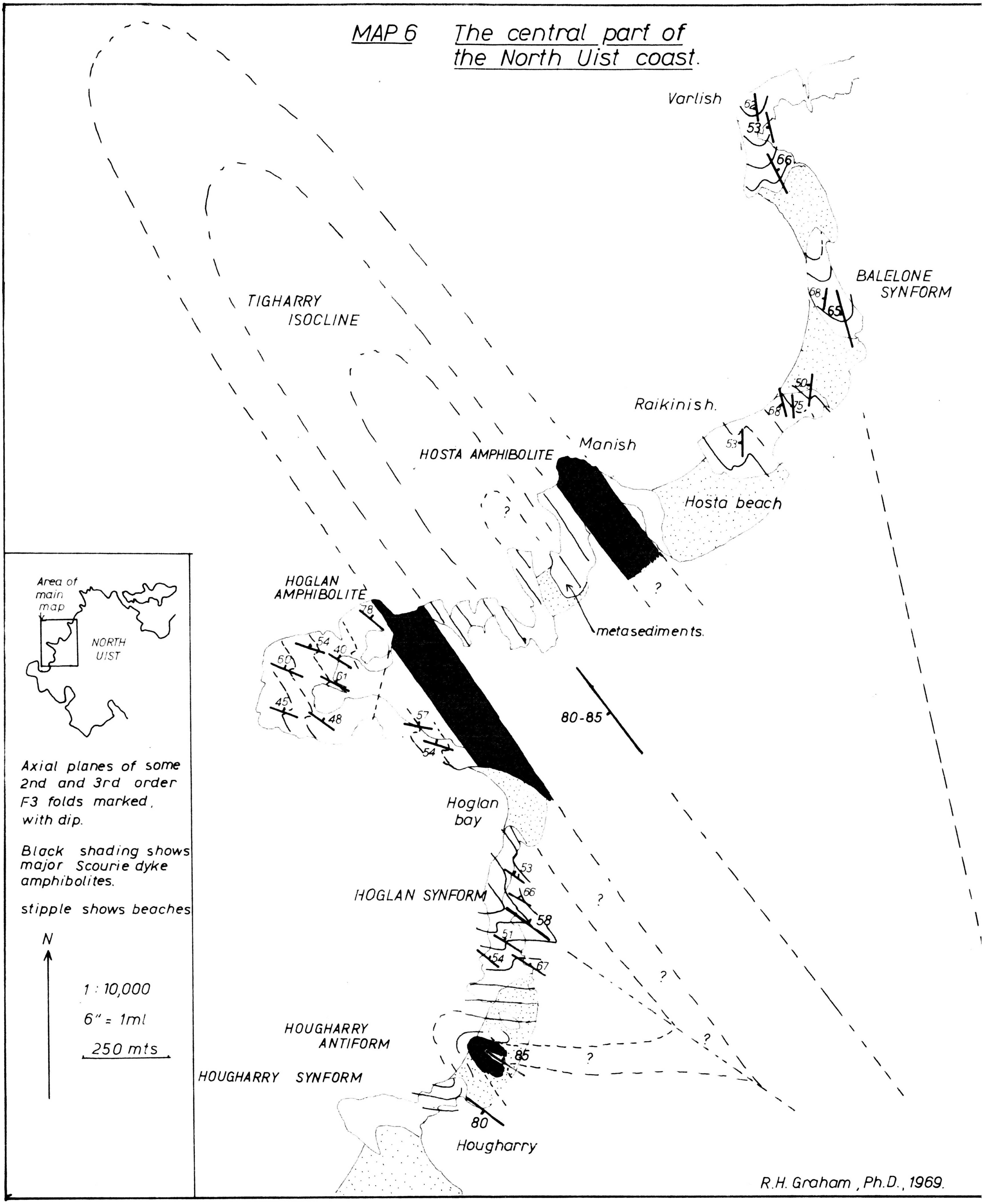
Map 2


North Griminish
Griminish Synform
Castell Odair
"Dreary Point"
Griminish Antiform
Griminish Point
Sloc Roe
Scolpaig Bay
Scolpaig Synform
"South Scolpaig"
The Cliffs
Cliffs Antiform
Natural arch
Varlish
Balelone Synform
Raikinish
Hosta Beach
Manish
Tigharry Isocline
Hoglan Beach
Hoglan Synform
Hougharry Antiform
Hougharry Synform
Hougharry
"Hougharry Bay"
Aird an Runaird
Structural map of the North Uist Coast.
foliation.
Axial trace of F2 fold.
Axial trace of F3 fold.
Boundary between "rough" gneiss and "smooth" gneiss
Area of a strain determination
PL
P>L
L>P
LL
PP
Fabric data.
NORTH
Area of main map
North Uist
Scale about 12" = 1 mile (1:5,000)
0 metres. 250
R.H. Graham. Ph.D 1969.

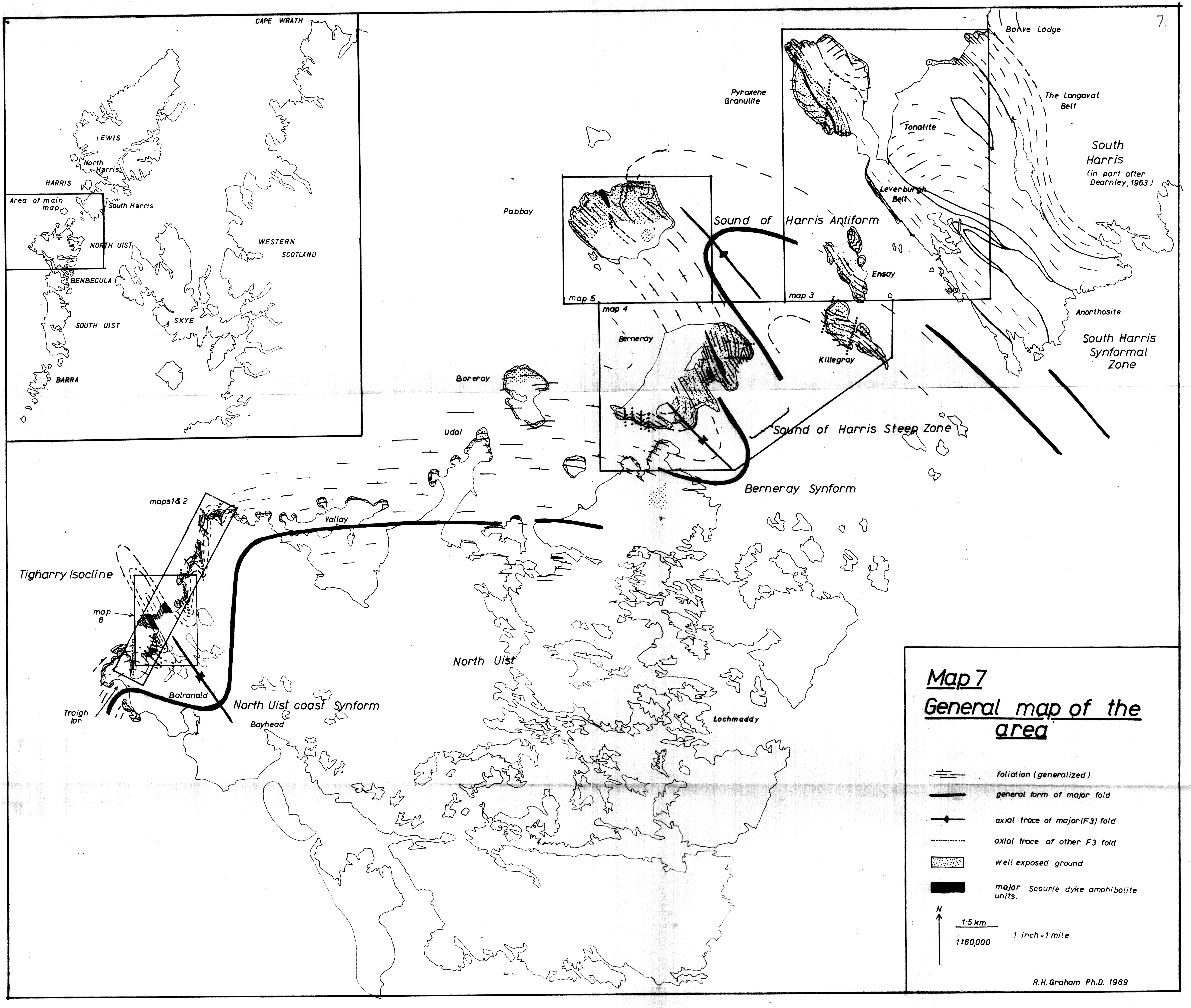
Map 3 Ensay and part of South Harris.
foliation in acid gneiss and metasediment
"granulitic foliation" in igneous bodies
shear zone in igneous bodies
Scourie dyke amphibolites in acid gneiss and metasediment.
late fracture
pegmatite and Tertiary (?) dyke
deformed breccia
garnetiferous amphibolite
Pyroxene Granulite
Tonalite
ultrabasic
N
250 mts
1 10,000
6"=1 mile
R.H.Graham Ph.D. 1969
TOE HEAD
Liuri
"toe head peninsula"
Chaipaval
Temple
Scaraster
BORVE LODGE
metasediment (Bay Stengie)
Northton
Greaval
Leverburgh
ENSAY
metasediment.
57°51'
7°8'


Map 4
Berneray and Killegray
foliation
late fracture
F3 axial trace
Scourie dyke amphibolite
metagabbro
ultrabasic
pegmatite
boundary of area of strain determination.
N
250 mts
1:10,000
6"=1mile
R.H.Graham PhD. 1969
Killegray
SOUND OF HARRIS ANTIFORM
Sand Hill
Bruist
Loch Bruist
"Sand Hill Beach"
Borve Hill
Berneray Granite
School house
Borve
Loch Borve
Rovoisinish
Berneray
BERNERAY SYNFORM
North Uist ("newton peninsular")

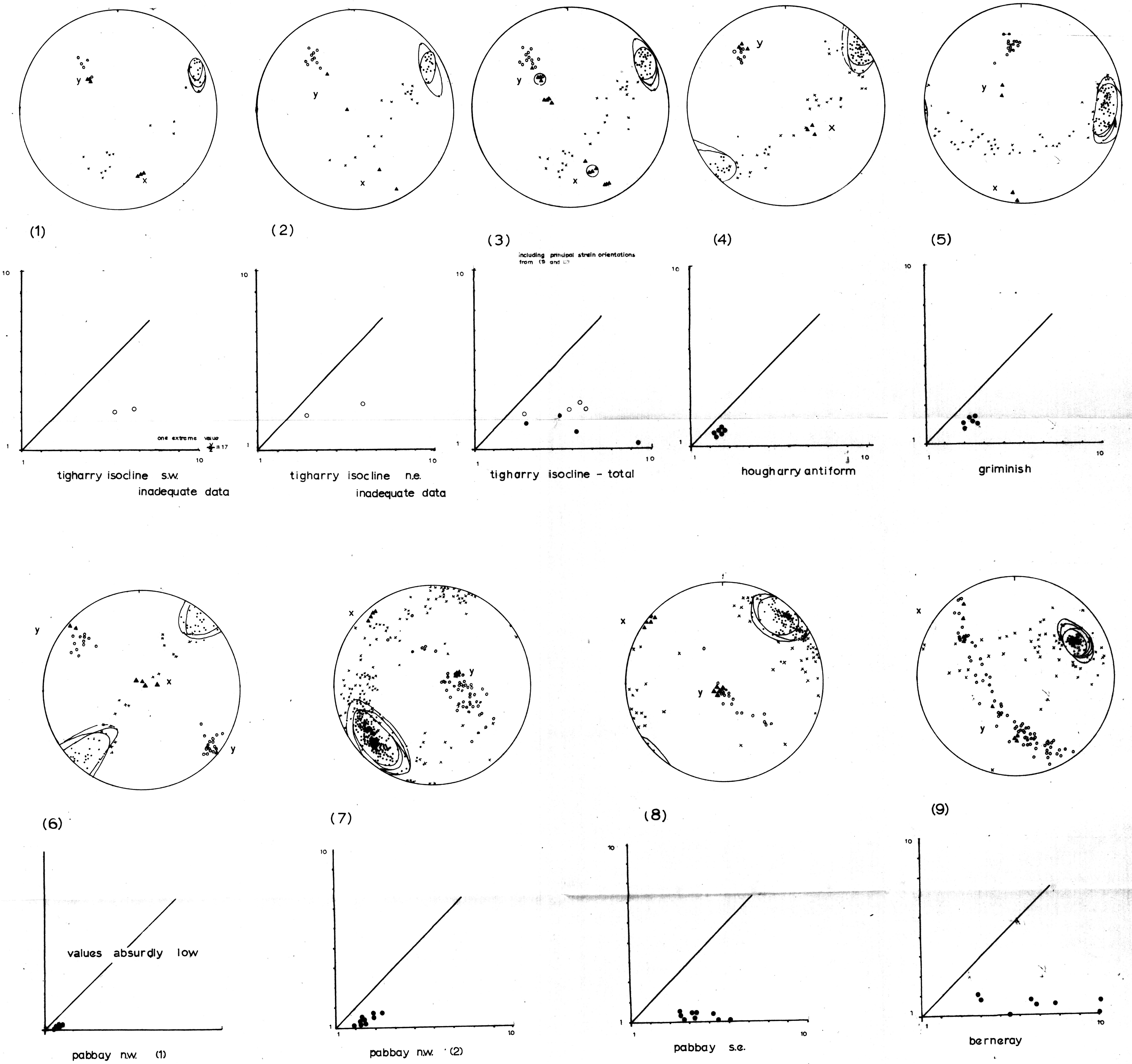
Map 5
Pabbay
Foliation (gneissic banding)
F2 gneissic cleavage
Late fractures.
boundary of area of strain determination
F3 axial trace
Scourie dyke amphibolite
Ultrabasic rock
Pegmatite.
Stipple - beaches.
N
Scale 1:10,000
250mts
P1
P2
P3
P4
FZ1
FZ2
Scarsdale point
Alarip bay
R.H.Graham. Ph.D. 1969.

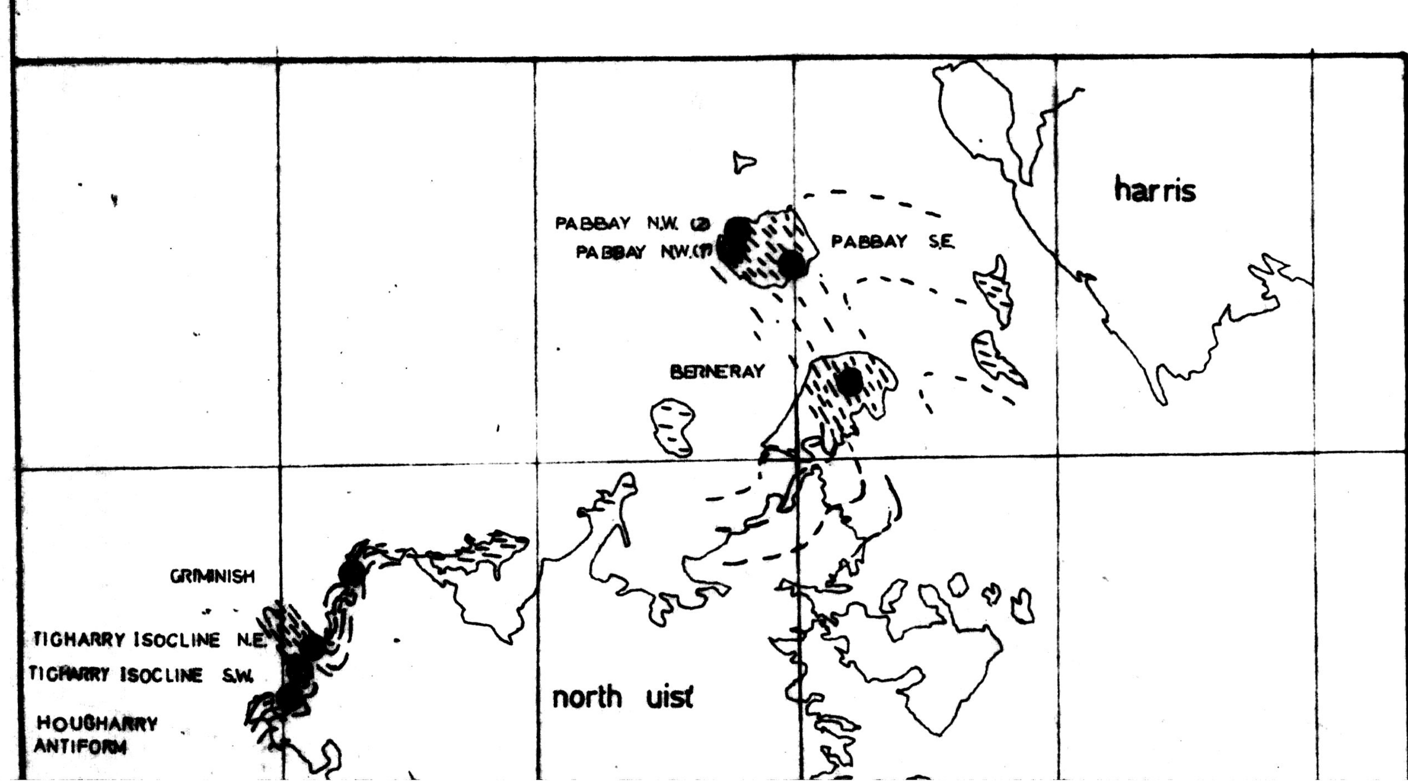


MAP 6 The central part of the North Uist coast.


CAPE WRATH
LEWIS
North Harris
HARRIS
Area of main map
South Harris
NORTH UIST
WESTERN SCOTLAND
BENBECULA
SKYE
SOUTH UIST
BARRA
Borve Lodge
Pyroxene Granulite
The Langavat Belt
Tonalite
South Harris
(in part after Dearnley, 1963)
Leverburgh Belt
Pabbay
Sound of Harris Antiform
Ensay
map 5
map 3
map 4
Berneray
Killegray
Anorthosite
South Harris Synformal Zone
Boreray
Sound of Harris Steep Zone
Udal
Berneray Synform
maps1& 2
Vallay
Tigharry Isocline
map 6
North Uist
Balranald
North Uist coast Synform
Traigh Iar
Bayhead
Lochmaddy
Map 7
General map of the area
foliation (generalized)
general form of major fold
axial trace of major (F3) fold
axial trace of other F3 fold
well exposed ground
major Scourie dyke amphibolite units.
N
1·5 km
1:60,000
1 inch = 1 mile
R.H. Graham Ph.D. 1969

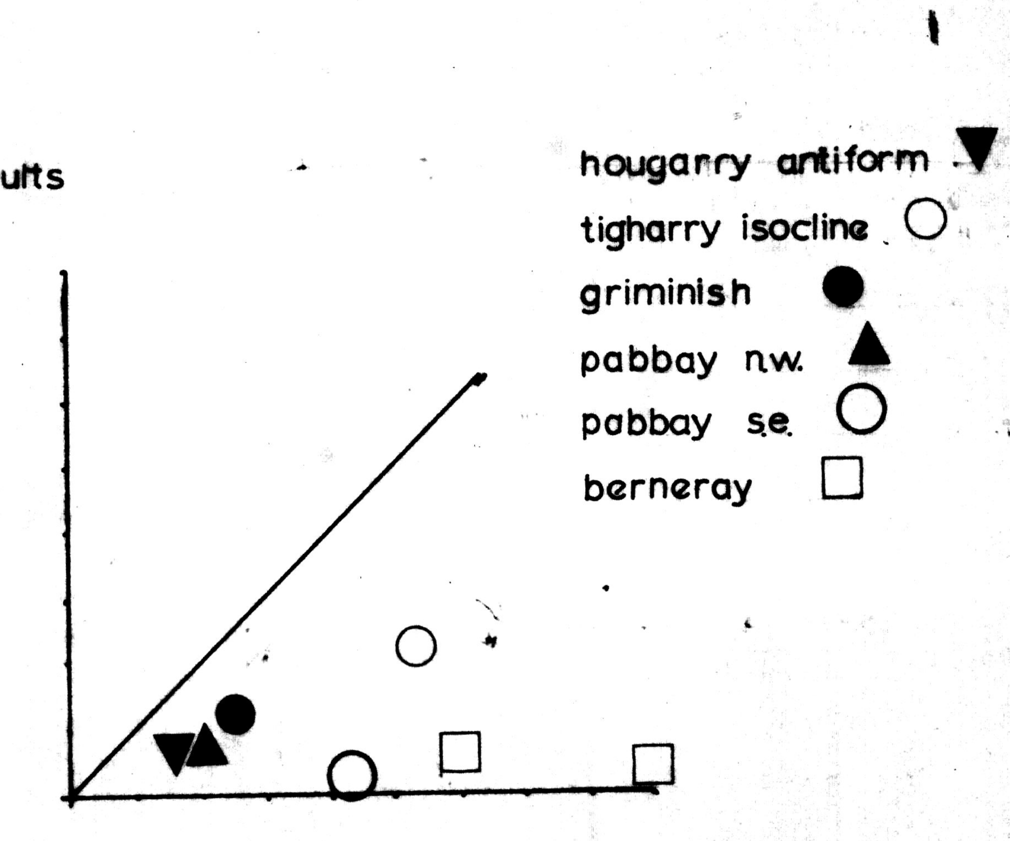
Sheet 1 Strain determinations from the orientation of deformed veins.





· poles to extended veins
× poles to inflexion surfaces of folded veins
∘ fold axes and lineations
▲ possible positions of X and Y axes of strain ellipsoid
(▲) favoured positions of the above
(Z lies in the centre of the extension field projection)

Favoured results



R.H.Graham Phd 1969