

THE RELATION BETWEEN METAMORPHIC REACTIONS AND
TECTONIC STRUCTURE AT SCOURIE, SUTHERLAND: A
STUDY OF REGENERATION.

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SUMMARY.

Many geological problems are still unsolved in the much studied and classic Scourie - Laxford region of the Lewisian basement. A small part of this region has been mapped in detail by the present writer. The results of previous work are not reiterated, but instead new data arising from field work are presented. The distinction between Scourian and Laxfordian structures and associated mineral fabrics is emphasised. The Scourian pyroxene granulites, in the area studied, have been affected by a late Scourian deformation during which the rocks were hydrated, and by a Laxfordian deformation during which the rocks were further hydrated and significant quantities of biotite appeared. Petrological studies have enabled the two metasomatic events to be closely defined - the late Scourian event involved addition of only water, while the Laxfordian event involved addition of water, potash, barium, and possibly other minor elements. Analyses of both Scourian and Laxfordian minerals are presented. Chemical reactions are presented which describe the degeneration of pyroxene to hornblende and of hornblende to biotite. The stability of muscovite in Scourian granite sheets, and the presence of kyanite, staurolite and corundum in an aluminous gneiss group are investigated. The presence of these minerals in Laxfordian shear zones permits the PT conditions during deformation and metamorphism to be deduced. Whole rock analyses show that assemblages in shear zones are oxidised relative to assemblages in adjacent rocks. The amount of water necessary to cause this oxidation is calculated, and this provides the key to the significance of the Laxfordian shear zones. The regional shape change resulting from the shear zones is very small, and it seems that the shear zones are more significant as channels for migration of

moderate quantities of fluids. The absence of Laxfordian deformation in areas where there is no Laxfordian metasomatism is analysed in terms of the thermodynamics of irreversible processes. It is shown that the process of deformation is not a spontaneous process considered alone, but is coupled with processes of fluid diffusion and chemical reaction, these latter processes occurring spontaneously. The fluid source is considered to be the Laxfordian granite complex at Laxford and a model, based on a stress function analysis, is built up to show how the fluid diffused out of this region and how instabilities leading to the generation of shear zones in the Scourie block were initiated.

CONTENTS.

LIST OF FIGURES	6
LIST OF TABLES	8
ACKNOWLEDGEMENTS	10
CHAPTER 1 INTRODUCTION	11
1.1 Scope of the present work	12
1.2 Lewisian terminology and chronology	12
1.3 Summary of past work in the region	15
1.4 Summary of work by Beach, Coward and Graham	17
1.5 Distribution of the rock types at Scourie and the foliations developed in them	21
1.6 A note on magnetic anomalies	22
1.7 A note on format	24
CHAPTER 2 A CONSEQUENCE OF FIELD MAPPING	25
2.1 The Scourian granulites	26
2.2 Amphibolisation of the granulites	27
2.3 The Scourie dyke suite	30
2.4 Major structure	36
2.5 The Laxfordian shear zones	47
CHAPTER 3 MINERAL FABRICS	57
3.1 Optic fabrics	58
3.2 Shape anisotropy	62
3.3 Redistribution of material	66
3.4 Some comments on the deformation of polycrystalline aggregates	67
CHAPTER 4 AN INTRODUCTION TO THE MINERALOGICAL STUDIES	71
4.1 Sampling	71
4.2 Petrography	72
4.3 Analytical techniques	78
4.4 Thermodynamics	81
4.5 Tables and abbreviations	81

CHAPTER 5	REACTIONS INVOLVING PYROXENE	84
5.1	Petrography	84
5.2	The reaction pyroxene → hornblende	89
5.3	Simplified pyroxene-hornblende equilibria	95
5.4	Minor reactions associated with amphibolisation	105
5.5	Hornblende pods	108
5.6	The isochemistry of amphibolisation	111
CHAPTER 6	REACTIONS INVOLVING BIOTITE	112
6.1	The transformation of hornblende to biotite	117
6.2	The Carmichael evaluation	118
6.3	Stability of muscovite	120
6.4	Biotised hornblende pods	121
6.5	The GK gneiss	121
6.6	Chemical analyses from the biotised rocks	129
6.7	Shear zones cutting dolerite dykes	134
6.8	Discussion of the metasomatism	137
6.9	Conclusion	140
CHAPTER 7	REACTIONS INVOLVING OXIDATION	141
7.1	Oxidation during amphibolisation	141
7.2	Oxidation in shear zones	142
7.3	Mineral assemblages in the Tarbet Steep Belt	147
7.4	A note on opaque minerals	150
CHAPTER 8	A DISCUSSION OF THE MINERAL ANALYSES	152
8.1	A note towards the definition of a meta-morphic system	152
8.2	The relation between rock oxidation and mineral chemistry	154
8.3	The significance of potassium and titanium in hornblende and biotite	155
8.4	A statistical study of hornblende compositions	162

CHAPTER 9	THE RELATIONS BETWEEN METAMORPHIC REACTIONS AND TECTONIC STRUCTURE	166
9.1	The concept of fugacity: fluid diffusion in rocks	167
9.2	Amphibolisation as a metasomatic event	171
9.3	Biotisation as a metasomatic event	173
9.4	Work done and heat produced in a shear zone	174
9.5	The importance of oxidation in shear zones	177
9.6	An instability leading to the generation of shear zones	180
9.7	Entropy production and irreversible processes	183
CHAPTER 10	A REGIONAL MODEL	191
10.1	Introduction	191
10.2	Stress function analysis	195
10.3	The source and motion of the fluid phase	199
10.4	End	202
	PERSPECTIVE	203
	REFERENCES	204
COVER POCKET	a) Map of area between Scourie and Tarbet b) Map of Natural Arch Dyke	

LIST OF FIGURES.

1.1	Map of part of Sutherland	16
1.2	Map of the Scourie-Laxford region	18
1.3	Section from Ben Auskaird to Rhiconich	20
1.4	Magnetic anomalies in the Scourie district	23
2.1	Map of the Mullach Dyke	32
2.2	A) Lineation L2 in the Natural Arch Dyke B) Dyke-gneiss relations in the Tarbet Steep Belt	33
2.3	Coast section across Tarbet antiforms	37
2.4	A) Lineation L2 in the Tarbet Steep Belt B) Laxfordian deformation of steep belt	41
2.5	Deformation of Tarbet Steep Belt: flattening	44
2.6	Deformation of Tarbet Steep Belt: simple shear	45
2.7	Deformation of L1 in a shear zone	50
2.8	Poll Eorna shear zone	51
2.9	Section across Sithean Mor	52
3.1	Felspar fabric in shear zone	59
3.2	Plagioclase optic fabrics	60
3.3	Log Rf/ ϕ plots for hornblendes	65
5.1	Geoscan traverses across hornblende	93
5.2	Geoscan traverse across hornblende	94
5.3	Oxidation during formation of hornblende	100
5.4	Stability curves of amphiboles	101
6.1	Untitled	116
6.2	Relationship of high alumina minerals in sample 53	125
6.3	GK gneiss whole rock analyses, a) K2O vs Al2O3 b) K2O vs TiO2	135
6.4	GK gneiss whole rock analyses, K2O vs O.R.	136
8.1	Biotite analyses, TiO2 vs mg	156
8.2	Hornblende analyses, TiO2 vs mg	157
8.3	Coexisting hornblende and biotite, TiO2 vs mg	158
8.4	Hornblende analyses, TiO2 vs SiO2	160

8.5	Hornblende analyses, K ₂ O vs TiO ₂	161
10.1	Scourie- Laxford section, showing principal stress trajectories	193
10.2	Orientation of principal stresses during the main Laxfordian deformation	194
10.3	Theoretically derived stress trajectories	198

LIST OF TABLES.

4.1	Localities and brief hand specimen descriptions of samples mentioned in text.	73
4.2	Summary of mineral assemblages found in the different areas of the Scourie-Laxford transition.	76
5.1	Scourian assemblages: modal analyses.	87
5.2	Amphibolised Scourian rocks: mineral analyses.	88
5.3	Hornblende pods: modal analyses.	106
5.4	Hornblende pods: mineral analyses.	107
6.1	Shear zones and adjacent assemblages: modal analyses.	114
6.2	GK gneiss: modal analyses.	122
6.3	Biotised gneiss: mineral analyses.	130
6.4	GK gneiss: mineral analyses.	131
6.5	Barium and manganese analyses of selected minerals.	132
6.6	GK gneiss: whole rock analyses and CIPW norms.	133
7.1	Oxidised assemblages: mineral analyses.	143
7.2	Modal analyses of oxidised shear zone.	144
7.3	Tarbet suite: modal analyses.	148
7.4	Tarbet suite: mineral analyses.	149
8.1	Statistical data for hornblende analyses.	164

LIST OF PLATES.

- | | | |
|-----|---|-----|
| 2.1 | A) Unmodified Scourian granulite facies mineral banding | 28 |
| | B) Linear fabric in hinge zone of asymmetric Scourian fold | |
| 2.2 | Laxfordian shear zones near Scourie | 48 |
| 3.1 | Fabric developed in shear zone crossing the GK gneiss | 64 |
| 5.1 | A) Partially amphibolised granulite: clinopyroxene rimmed by hornblende | 86 |
| | B) Amphibolised granulite: hornblende corona with quartz free rim | |
| 6.1 | Hornblende - biotite fabric in Laxfordian shear zone | 113 |

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CHAPTER 1.

INTRODUCTION.

During the late nineteenth century the Geological Survey of Great Britain carried out extensive mapping of the rocks in the North West Highlands of Scotland. This resulted in the publication of the now classic memoir (Peach et al 1907). The quality of this work is perhaps indicated by the virtual absence of further research into the Fundamental Complex, as it was then named, until Sutton and Watson published the results of a more detailed survey. They had studied the areas around Torridon (Ross and Cromarty) and Scourie (Sutherland) and their paper (Sutton and Watson 1951) created a more general interest in the geological problems of the Lewisian basement.

Sutton and Watson established a generalisation that later work has confirmed. They followed the Geological Survey in showing that it was possible to differentiate between an older and a younger part of the basement. Dolerite dykes cutting the Scourian were unaltered, whereas their equivalent in the Laxfordian were metamorphosed. Field observations led Sutton and Watson to conclude that the Laxfordian event was essentially a remetamorphism of the Scourian rocks, and from this they suggested that regeneration of the original basement complex could be observed.

Though the idea of regeneration has been challenged since then, it is evident that it can still be applied to many parts of the Lewisian. However few attempts have been made to examine in detail the effects of the metamorphic and tectonic processes involved in regeneration since Teall published some fundamental conclusions and asked some very pertinent questions (Teall 1885). Teall showed that a foliated amphibolite was derived from an igneous dolerite, the extent of the amphibolite defining a shear zone in the dolerite dyke (this dyke being the Scourie Dyke at Poll Eorna, Scourie). The close association of mineral

transformations and tectonic deformation was thus demonstrated.

There is a curious dichotomy apparent in the method of approach to field observations between Teall (1885) and Bowes and Khoury (1965). Figure 1 of Teall and figure 1 of Bowes and Khoury both in fact refer to the development of amphibolite by shearing of a dolerite, but the observations of the latter appear to be governed by interpretation. It is this reversal of scientific method that makes a summary of much of the previous research in the Scourie district difficult.

1.1 Scope of the present work.

In a second paper (1961) Sutton and Watson described the nature and distribution of shear zones of Laxfordian age within otherwise little modified Scourian rocks at Scourie. The shear zones were interpreted as being coeval with the strong Laxfordian metamorphism and deformation seen at Laxford Bridge.

Following the detailed analysis of shear zones elsewhere made by Ramsay and Graham (1970), it was decided that since shear zones were recognised as common products of deformation in basement rocks, further research into the nature of shear zones was necessary. Papers in preparation by Graham present the results of research into the shear zones of South Harris in the Outer Hebrides, while this thesis is a report on research into the shear zones found near Scourie.

1.2 Lewisian terminology and chronology.

Terminology and chronological classification are two aspects of the same problem, that of relating tectonic structures and metamorphic events to a time scale, thus permitting correlations to be made between apparently dissimilar areas. Misleading ideas arise when field geologists correlate between areas on the basis of similarity (of fold style, dykes, etc.), and when isotope geologists collect samples with an incomplete knowledge of their structural relationships. A recent paper by Park (1970), though it does not solve all the problems pertaining to

Lewisian terminology, clarifies their nature and gives a summary of knowledge to date.

Following from the work of Giletti et al (1961), many local and detailed isotopic studies have been made, a number concerning the Lewisian in Sutherland (Evans 1965, Evans and Tarney 1964, Lambert and Holland 1970).

The meaning of the three most prominent terms in use - Scourian, Inverian and Laxfordian - has been extended by some to denote an orogeny in each case. The terms were originally used to describe particular tectonic and metamorphic associations that could be defined in the field. It is only since the attempt to apply these terms to areas other than those where they were initially defined that confusion has arisen. It would thus seem more logical now to redefine these terms as periods of time in which separate tectonic and metamorphic events can be placed as more isotopic data become available.

There has also been a tendency to define these terms (Scourian, etc) on the basis of the metamorphic facies that is seen in the type areas. Thus the Scourian is said to be granulite facies and the Laxfordian is migmatite and amphibolite facies. Only when the terms are defined as strict periods of time, with isotopic events fixed in these time spans, will it be possible to discuss different areas of Lewisian in a reasonable way and attempt to correlate variations in metamorphism etc. between these areas.

Terminology adopted in this work.

The sequence and range of isotopic age dates obtained from Lewisian rocks is now well known (see the recent summary by Park 1970). There is still little agreement concerning the names that should be used to describe the known events or the interpretation that can be placed on these names. It amounts to an expressed bias to use any but the division into Scourian and Laxfordian in the most general sense.

The reasons for rejecting terminologies such as that suggested by Park (1970) are as follows:

a) The sequence of events in the Scourie area is still not known with any certainty.

b) In spite of the attempts of some workers to impose a regularity on the structural and metamorphic sequences on the mainland and the Outer Hebrides, the similarities between the sequence at Scourie and in other areas are becoming fewer.

c) In the Scourie area there are at least three different hornblende fabrics developed, one main age of biotite fabrics, and a number of pegmatites and granites of different ages. Coward (personal communication) has observed that axial surface hornblende fabrics at Badcall (near Scourie) are cross cut by pegmatites that have been dated at 2600my (Lambert, personal communication). Early hornblende fabrics such as this have not been generally recognised.

d) There is no continuity between Scourie and Badcall - an area of pyroxene granulite separates the two areas where amphibolised and folded rocks are seen, and correlation over just a few kilometres is very difficult.

Until it has been decided which of the metamorphic and structural events are the most significant in the history of the Lewisian, it is very difficult to choose a terminology that is objective and meaningful. For example, the term 'Badcallian' introduced by Park (1970) is certainly an objective and non genetic term (though it refers to a granulite facies metamorphism) because of its derivation from an obscure and previously little published place name; it becomes a little meaningless when it is realised that many equally objective terms could have been introduced, such as Geisgeilian, Duartmoreian, Newtonian, Drumbegian, Neddian - or even perhaps Scourian. Undoubtedly one of the most firmly established events on the mainland is the age of intrusion of the Scourie dyke suite. The terms Scourian and Laxfordian, as used by Sutton and Watson (1951), are considered to be entirely adequate and without confusion. The words 'Inverian', 'Badcallian' etc. have not attained the generality in meaning that 'Scourian' and 'Laxfordian' have and so cannot be used without the specific and implied meanings that they carry. For this reason they have not been used in this thesis.

1.3 Summary of past work in the region.

The area studied in this thesis is small and it is necessary at the outset to outline the regional context of this area. Fig.1.1 gives a synthesis of the major tectonic structures that have been mapped in the Lewisian of Sutherland. Sources of information for this compilation were Evans (1963), Dash (1969), Holland (1966) and Beach et al (in preparation). The location of the area studied by the writer is also shown on this figure.

The north west trending line through Laxford is significant - it divides this region into two parts about which the following generalisations can be made:

a) to the north of this line the rocks are granitic and pegmatitic; they record large finite Laxfordian strains and give Laxfordian isotopic ages;

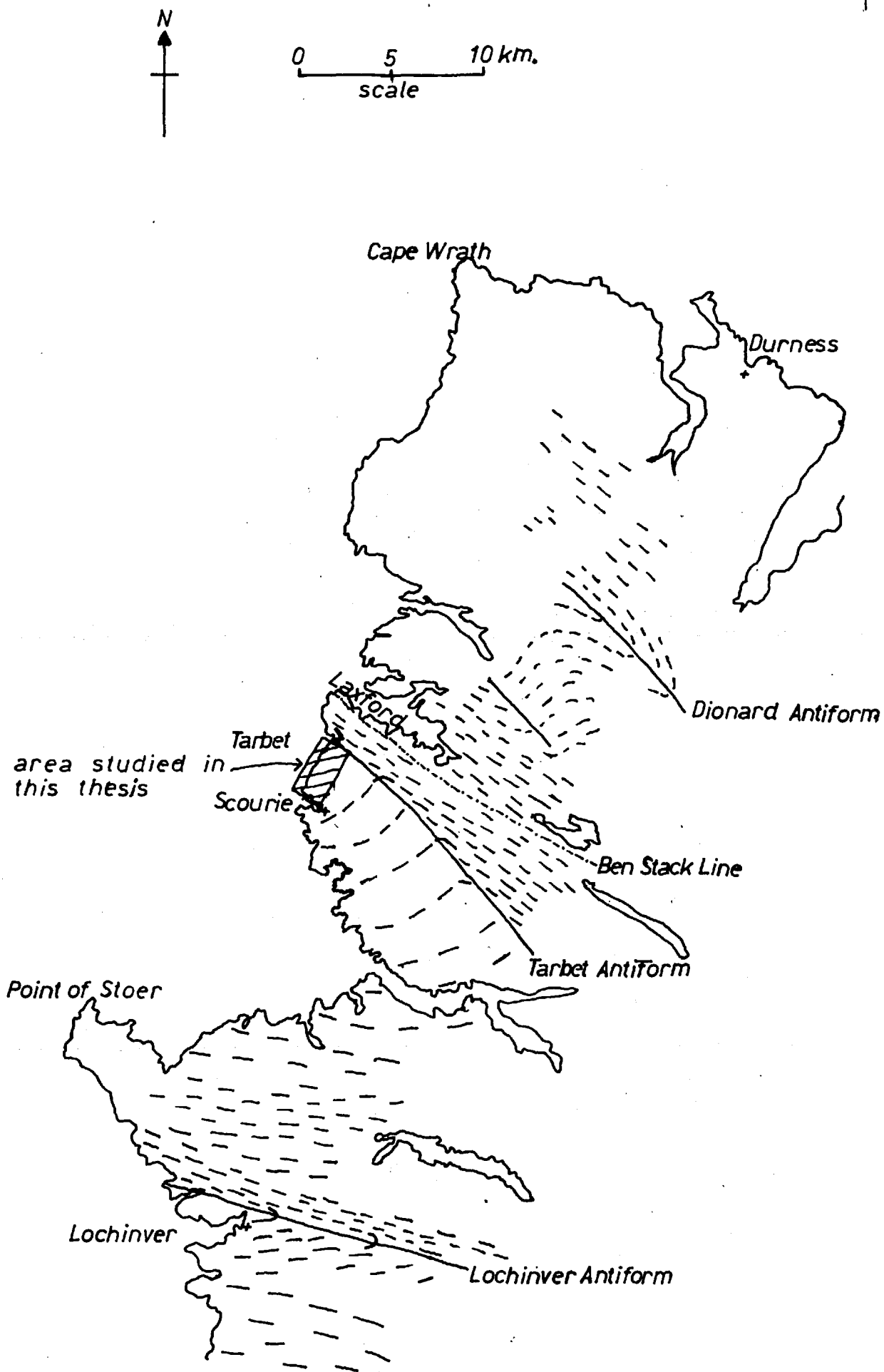
b) to the south of this line the rocks are non granitic, and remnants, ranging in size from a few millimetres to many kilometres, of Scourian granulite facies assemblages can be seen.

The transition zone in the Scourie - Laxford region is thus of fundamental importance, and was recognised by Peach et al (1907).

Since Evans (1963) suggested that the Lochinver Antiform was of late Scourian age, the real age of the geometrically similar structure in the Scourie-Laxford region has been in dispute. One of the principal results of remapping in the latter area has been the recognition of certain tectonic structures, regarded as Laxfordian by Sutton and Watson (1951, 1961), that formed prior to dyke intrusion. The details of these structures are presented in the next chapter. A series of Scourian folds have been mapped in the transition zone by Beach et al (see next section) and it has been found that the strong Laxfordian deformation is concentrated in the northern part of this fold belt. The most southerly of these major folds lies near the northern boundary of the area studied in this thesis. This fold, referred to as the Tarbet Antiform, is geometrically asymmetric, and the outcrop of the steep northern limb, referred to as the Tarbet Steep Belt, corresponds closely to the extent of the Claisfearn Zone of Sutton and Watson (1951).

MAP OF PART OF SUTHERLAND

showing schematically the major tectonic structures and foliations in the Lewisian rocks



The Ben Stack line, shown on fig.1.1, is based on the work of Holland (1966) and marks a geochemical boundary defined by him in terms of bulk rock chemistry. It thus forms the southern limit of the maximum development of the Laxfordian granites, and as such was recognised by Peach et al (1907). The tectonic significance of this line is difficult to judge. Peach et al report it as being regionally transgressive to the foliation (1907, p126). This will be referred to again in the next section.

1.4 Summary of work by Beach, Coward and Graham.

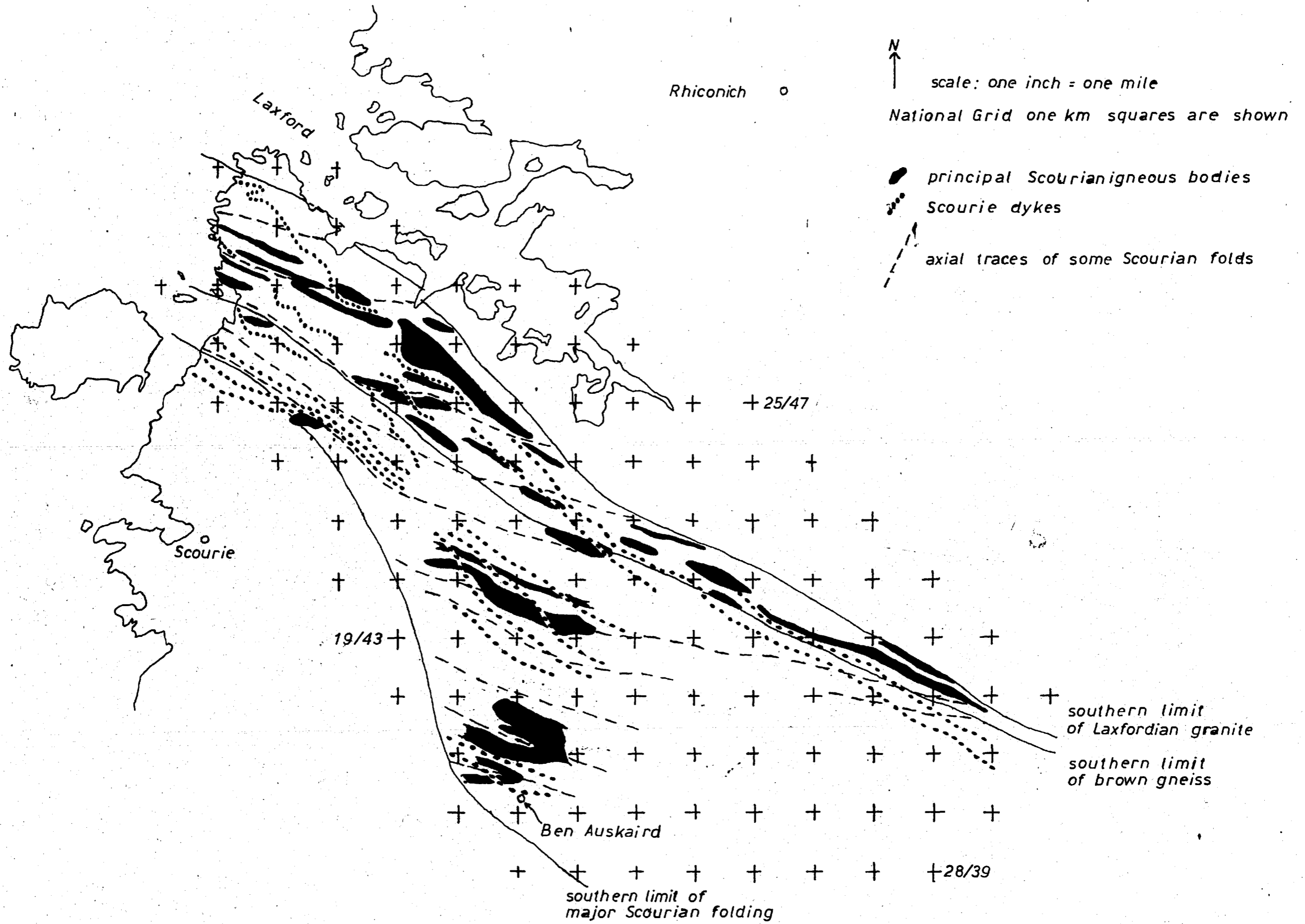
No systematic mapping of the tectonic structures in the whole of the Scourie - Laxford region has been carried out since the work of Clough (in Peach et al 1907). A number of workers have expressed, but not published, ideas concerning the structure of this region, but without a structural map to refer to it is difficult to judge which of these ideas is correct. In an attempt to remedy, at least partially, this state of affairs, Beach, Coward and Graham recently mapped part of the Scourie - Laxford region on the scale of 6 inches to one mile. The results of this work will be presented in a separate paper (Beach et al, in preparation).

Fig.1.2 is a simplified and reduced version of this new map. A brief summary of this work is now presented.

There are a large number of igneous bodies present in the area mapped, a far greater number than are found to the south of this area. These igneous bodies have been mapped as lithologic units, and the following assemblages have been recorded:

- a) microcline-plagioclase-quartz-muscovite-(biotite)
- b) plagioclase-quartz-muscovite-biotite-(hornblende)
- c) plagioclase-quartz-biotite-hornblende
- d) microcline-plagioclase-hornblende-diopside-sphene
- e) plagioclase-hornblende-pyroxene
- f) plagioclase-pyroxene-garnet
- g) pyroxene-garnet-olivine
- h) anthophyllite-talc
- i) plagioclase-quartz-(biotite)
- j) garnet-quartz

SUMMARY OF GEOLOGY IN SCOURIE-LAXFORD AREA



The greatest concentration of these igneous bodies is in the Foindle Zone of Sutton and Watson (1951) and its south-eastward extension. Frequently in this area gneissic assemblages consisting of garnet-biotite-plagioclase-quartz-sulphides (the brown gneisses of Sutton and Watson 1951; the distinctive brown weathering of these rocks is a result of the decomposition of metallic sulphide phases) are closely associated with the igneous bodies. This association has been described in the Lewisian of the Outer Hebrides by Coward et al (1969) and the brown gneisses there were interpreted as relicts of metasedimentary rocks because of their similarity with the undoubted metasediments of South Harris. In the Scourie area it is possible to draw a line marking the southern limit of the outcrop of the brown gneisses (fig.1.2). So far rocks of this type have not been recognised north of Laxford.

The igneous bodies undoubtedly pre-date the granulite facies metamorphism ending at 2600my. since they contain granulite facies mineral assemblages (though these are frequently altered to more hydrous assemblages) and the mineral banding in these bodies is continuous with that developed in the more acid gneisses during the granulite facies metamorphism.

By mapping the different igneous bodies, the brown gneisses and the minor fold vergence, a number of major folds have been defined. The axes of these folds trend ESE, and a few are shown on fig.1.2. These folds post-date the granulite facies metamorphism and are cross cut by members of the Scourie dyke suite; they are thus Scourian in age. On the coast the most southerly of these folds lies just south of Tarbet (GR.165488), but in going inland in a south easterly direction to Ben Auskaird (GR.210402) successive folds are seen to develop in an en echelon fashion (see fig.1.2).

Laxfordian deformation has been superposed on the Scourian folds. On the northern slopes of Ben Auskaird the axial surface fabrics of these folds are vertical. The dykes in this region are slightly deformed but this Laxfordian deformation was evidently not strong enough here to overprint the mineral fabrics in the gneisses to the extent that new Laxfordian fabrics were developed. North of Ben Auskaird there is evidence

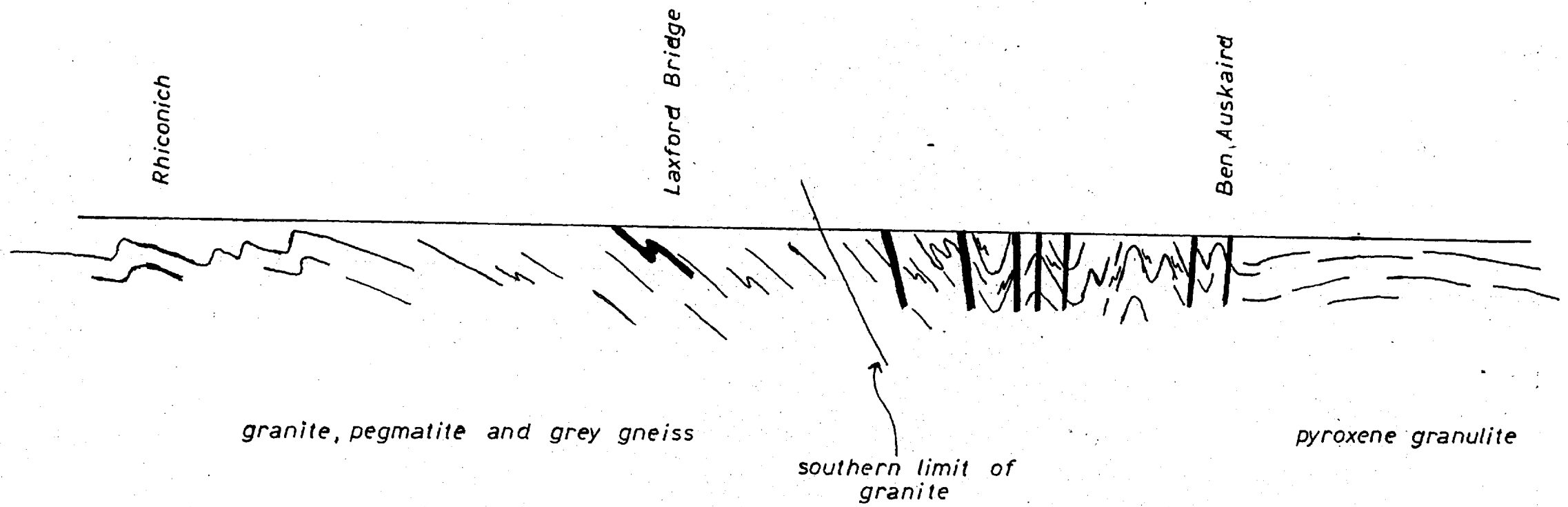
that the Scourian folds have been flattened. The folds become isoclinal and the axial surfaces dip to the SSW, becoming progressively less steep towards the north. The fold axes have been deformed in the plane of the foliation, and variations of up to 90° in the pitch of the fold axes have been mapped. It is only about one mile south of Badnabay (GR.220466) that new Laxfordian mineral fabrics become prominent in the gneisses. This contrasts with the coastal section where Laxfordian fabrics can be seen right up to the southern limit of development of Scourian folds.

The line marking the southern limit of Laxfordian granite is shown in fig.1.2. The transition from granitic to non-granitic rocks is very rapid and it is evident that the line marking this transition cuts across the axes of the Scourian folds. It is in this region that the Laxfordian flattening deformation mentioned above reaches a maximum. The Scourian fold axes now plunge consistently to the south east (though the angle of plunge varies between 15° and 40°) and they are parallel to the Laxfordian mineral lineation (for reasons presented later this mineral lineation is interpreted as a finite stretching direction). Northwards into the granitic rocks evidence of a Scourian phase of folding is obscured, partly because of the intense recrystallisation here, and partly because of the large quantity of pegmatitic and granitic material present in these rocks. However, it is here that the first evidence of a second phase of Laxfordian deformation is seen. The planar fabrics in the Scourie dykes are folded, and a set of minor folds (with a vergence consistent with their being on the southerly limb of an antiform) can be seen in the gneisses. This deformation increases in intensity northwards to Rhiconich (GR.255522), where the earlier foliation is almost flat lying and is folded into a number of asymmetric folds with short steeply dipping and northerly facing limbs. A few miles north of Rhiconich, Dash (1970) has mapped the Strath Dionnard Antiform (see fig.1.1) and this records the effect of a third phase of Laxfordian deformation. This antiform is an open fold with gently dipping limbs.

FIG. 1.3

Section from Ben Auskaird to Rhiconich

scale: one inch = one mile



Thus, four phases of deformation have been recorded (one Scourian, three Laxfordian), the interesting features of this sequence being that the southern limit of each successive phase of deformation occupies a more northerly position than the preceding phase, and that the different phases of folding are virtually coaxial.

In conclusion fig.1.3 shows a simplified cross section of the area between Ben Auskaird and Rhiconich.

1.5 Distribution of the rock types at Scourie and the foliations developed in them.

The gneisses in the Scourie area everywhere vary in composition from acid to ultrabasic (see Peach et al 1907, chapters 8,9; Sutton and Watson, 1951). However, it was recognised early in the course of this work that a general tri-partite division could be made - a division into pyroxene, hornblende and biotite bearing rocks respectively.

Pyroxene bearing assemblages generally belong to the early Scourian granulite facies rocks. These rocks are excellently exposed in the area between Scourie and Badcall, but northwards from Scourie to Tarbet much of the original pyroxene has been replaced by or recrystallised to hornblende. It will be seen later that the appearance of hornblende reflects an important event, but one which is localised in space. In the area northward from Scourie unaltered granulite facies rocks are restricted in occurrence to small irregular patches that escaped amphibolisation. Thus a north west line through Scourie Bay marks broadly the southern limit of the major development of hornblende in this area.

Biotite may be found in many rock types at Scourie, but the third division set up above of biotite bearing rocks includes only assemblages in which biotite is an important phase in terms of both rock chemistry and fabric. Biotite is then essentially restricted to rocks that gained their present day characteristics during the Laxfordian period.

Mapping of lithologic units within gneissic basement terrains is generally difficult due to the lack of clear

definition and persistence of such units. It is more important to map foliation and to distinguish foliations of different ages. Peach et al (1907) and Sutton and Watson (1951) distinguished two different foliations in the Scourie area, the mineral banding of the granulite facies rocks and the mineral foliation in rocks of Laxfordian age. This distinction has to be modified slightly in the light of the work by Beach et al. An important axial surface fabric developed with the Scourian folds and during the Laxfordian this fabric increased in intensity. Thus the second foliation of Peach et al (1907) is made up of two distinct components.

In the area between Tarbet and Scourie, and also at Badcall, hornblende fabrics are associated with Scourian folds, but since these structures are nowhere continuously developed, correlation northwards is difficult. These fabrics are discussed in detail in the next chapter.

1.6 A note on magnetic anomalies.

Part of the original one inch aeromagnetic survey of Great Britain (carried out by the IGS.) is reproduced in fig.1.4, which shows the magnetic anomalies in the area around Scourie.

It is evident from a study of the magnetic anomalies in the area between Lochinver and Scourie, and occurring in the regions where Lewisian rocks are exposed, that the most pronounced positive anomalies correspond very closely with the known outcrops of Scourian granulites. Thus the area of Scourie Mor (south of Scourie Bay) has an anomaly on the regional map of +400 gammas. There is a very steep magnetic gradient northwards across Scourie Bay until near Tarbet there is a magnetic low of -250 gammas on the regional survey. The fall off in magnetism of the rocks is a result of the granulites in this area being amphibolised (Bernius, personal communication). Thus the magnetic gradient across Scourie Bay coincides with the southern limit of major development of hornblende in once granulite facies rocks.

The work of Evans (1963) in the Lochinver region suggested that amphibolisation of the granulites commenced in the steep limb of the Lochinver Antiform and spread outwards, in particular

taken from IGS one inch survey
contours 50 gamma
scale: 1 inch = 1 mile



southwards, from there. There is an anomaly of -400 gammas corresponding to the outcrop of the steep limb of the Lochinver Antiform. It will be seen in the next chapter that there is a geometrically similar structure in the Tarbet area. Again, the magnetic anomaly of -250 gammas corresponds very closely with the outcrop of the steep limb of the Tarbet Antiform. It is suggested that the correspondence is real, and that the magnetic lows mentioned were produced at the same time as the formation of the major folds and amphibolisation of the granulite

Northwards from Tarbet there is a steady gradient up to the magnetic highs of the Strath Dionnard. The rocks between Laxford and Durness are all Laxfordian in age, and Holland (personal communication) has suggested that the rising magnetic gradient towards Strath Dionnard indicates that the Scourian granulite facies rocks are rising nearer the surface in this region. At present there seems to be no firm basis for this supposition.

1.7 A note on format.

Samples collected during the course of field work are referred to throughout the text and tables by numbers assigned them in the field. Not all the samples are in the Imperial College collection, but those that are can be easily located by using the field number quoted.

The use of abbreviations has been restricted to commonly accepted contractions of units of measure, metric measures being preferred. Mineral names have been shortened only in the tables.

Stereograms are presented as lower ^mhemisphere, equal area projections.

CHAPTER 2.

A CONSEQUENCE OF FIELD MAPPING.

A presentation of essential field observations and some preliminary discussions in this chapter will provide a framework for the detailed petrologic studies which follow. A large amount of field data has already been published for the area in which the present work has been carried out. The reader is referred to Peach et al (1907) and Sutton and Watson (1951, 1961) for many details. The main points which need to be discussed are the development of hornblende and biotite mineral fabrics from originally pyroxene bearing rocks, the spatial extent of the structures with which these fabrics are associated, and the age relations of the different fabrics and structures.

The structural analysis of basement terrains is generally more problematic than similar work in younger orogenic belts. The main problem encountered in the work in Sutherland was a lack of markers which defined geometry sufficiently precisely for detailed analysis to be carried out. It is for this reason that the techniques of Ramsay (1967) are virtually absent from the work presented here.

In a terrain where the finite structure developed so heterogeneously it is difficult to adhere to a strictly chronological sequence in presentation. This account is divided into the following sections:

- 2.1 The Scourian granulites.
- 2.2 Amphibolisation of the granulites.
- 2.3 The Scourie dyke suite.
- 2.4 Major structure.
- 2.5 The Laxfordian shear zones.

Sutton and Watson (1951) divided the area between Scourie and Laxford into a number of structural and metamorphic zones, related to the effects of Laxfordian reworking. Field work described in this chapter is based mainly on the Scourie Zone (in which the effects of Laxfordian deformation are localised in extent) and in part of the Claisfearn Zone.

A structural map of the region to be discussed is to be found in the cover pocket. Place names referred to in the text can be located on the map.

2.1 The Scourian granulites.

The Scourian granulite facies rocks are the oldest components so far recognised in this part of the Lewisian complex. Only a small amount of pyroxene granulite remains in the portion of the Scourie Zone studied here, but it is sufficiently widespread to enable the first processes of alteration to be frequently observed. The granulites have a number of characteristic features which are considered to have been established prior to the events which are the main topic of discussion in this chapter. The later structural and metamorphic effects are over-printed on those established during the formation of the granulites. These characteristics may be briefly summarised as follows:

- a) the granulites have a flat lying mineral banding, the only observable perturbations of which are occasional small relict fold hinges and pods lying between mineral bands;
- b) there is a general absence of planar and linear mineral fabrics defined by pyroxene (weak lineations parallel to the relict fold hinges may sometimes be seen); coarse quartz schlieren are fairly frequently developed and this lineation is also parallel to the relict fold hinges;
- c) there is a large range in rock composition in the granulite suite, and these contrasting assemblages are simultaneously involved in later events.

2.2 Amphibolisation of the granulites.

The area to the south of Scourie Bay consists dominantly of pyroxene granulite. However, there is a rapid transition (cf. note on magnetic anomalies) across the Bay to mostly amphibolised rocks on the north side. The process involved is one of hydration of the pyroxene bearing rocks and will be discussed in detail later. At present it is sufficient to note the association of hydration and heterogeneous deformation as outlined in the next paragraph.

A). Observed structures.

The Scourian mineral banding is folded into a number of large asymmetric folds consisting of long flat limbs and short steep limbs. It is only by virtue of the latter that the structures are recognised, since the flat limbs often consist of unaltered or little altered pyroxene granulite and conform to the general structural style of the granulites. The fold axes of these structures plunge from 20-30°, their orientation changing from SW on Sithean Mor to W in the area south of Tarbet. The steep limbs are invariably amphibolised, and the granulite mineral banding shows only a little thinning. These limbs vary from 10m to 500m in width.

In some areas (eg. Sithean Mor) minor folds parasitic on the larger structures are common. These folds have wavelengths and amplitudes up to 1.5m and the pyroxenes have been amphibolised throughout the structures.

B). Fabrics related to structure.

Mineral fabrics varying from LS to L (Flinn 1965) are associated with amphibolisation. The large asymmetric folds have certain fabric elements associated with them, and in particular their steep limbs have an LS fabric. The lineation, denoted L1, is a coarse mineral rodding defined by aggregates of hornblende and sometimes of quartz, and is parallel to the



A) Unmodified Scourian granulite facies banding



B) Linear fabric in hinge zone of asymmetric Scourian fold

fold axes. The hinge zones of the larger of these folds have a linear L fabric; in them the mineral banding is broken and folded into a coarse rodding lineation (see plate 2.1) that is again parallel to the axes of the folds. In the smaller asymmetric folds the mineral banding is continuous through the hinge zones. The planar element of the fabric is developed parallel to the axial surface of the folds: in the asymmetric folds this is very nearly parallel to the steep limb, whereas in the minor folds it more obviously cross cuts the mineral banding. In the flat limbs of asymmetric folds which have suffered amphibolisation a variety of fabrics are seen. Where the banding is not folded, pyroxenes are merely pseudomorphed by hornblende and the granular texture is preserved. The axial surface fabric of the minor folds has a variable dip and nowhere completely overprints the mineral banding. A lineation L1 of hornblende aggregates lies parallel to the fold axes. An extreme development of this lineation is seen where originally dispersed granulite pyroxene grains alter and collect into hornblende segregations. The banding is destroyed and a perfectly linear fabric is produced. The process of segregation is seen to take place over a distance of a few centimetres and the resulting pods measure up to 3cm across and 20cm in length.

C). Age of structures and fabrics.

All the fold structures described above are found to be truncated by the Scourie dykes. The fabrics which are closely associated with the structures were also formed prior to dyke intrusion.

It is evident that amphibolisation and deformation were associated phenomena, and that a system of LS fabrics was produced which reflect variations in deformation in the rock. There is no evidence that the structures and fabrics so far described formed during more than one phase of deformation.

South of Tarbet the lineation L1 is seen to be folded by structures shown to be Scourian by Beach et al (in preparation)

Thus there are two ages of Scourian folding in this area. It is impossible to tell if there was further hydration of the complex coeval with the second Scourian folds or whether amphibolisation was restricted to the first Scourian fold phase.

Some of the first fold structures mentioned above may be adequately described as shear zones from the point of view that the steep limbs of asymmetric folds are approximately linear zones of amphibolised and deformed gneiss, while the adjacent flat limbs are undeformed pyroxene granulite. Other fold structures belonging to the same suite do not record this special type of strain distribution (see Ramsay and Graham 1970 for detailed strain analyses of shear zones), and must be considered as folds in the more general sense. There is a complete gradation between the two types of structures.

A number of samples of pods of hornblende were collected from a number of different structures belonging to the first Scourian fold phase, and from other rocks recording the effects of amphibolisation. It is hoped that these rocks will be dated, using the potassium-argon method, in the near future and provide some knowledge of the true age of amphibolisation.

2.3 The Scourie dyke suite.

The Scourie dyke suite has already received much attention (Peach et al 1907, Sutton and Watson 1951, 1961) and for this reason reiteration of general field relations is unnecessary. A discussion of dyke chronology is presented by Park (1970). The general outcome of this is that the dykes represent a unique time marker.

The usefulness of the dykes in attempting to unravel the various deformations in a particular region can never be overemphasised. For this reason, the topic discussed in this section is some detailed mapping of deformed dykes in the immediate area of the Tarbet Antiform. The following dykes are described, their locations being shown on the map:

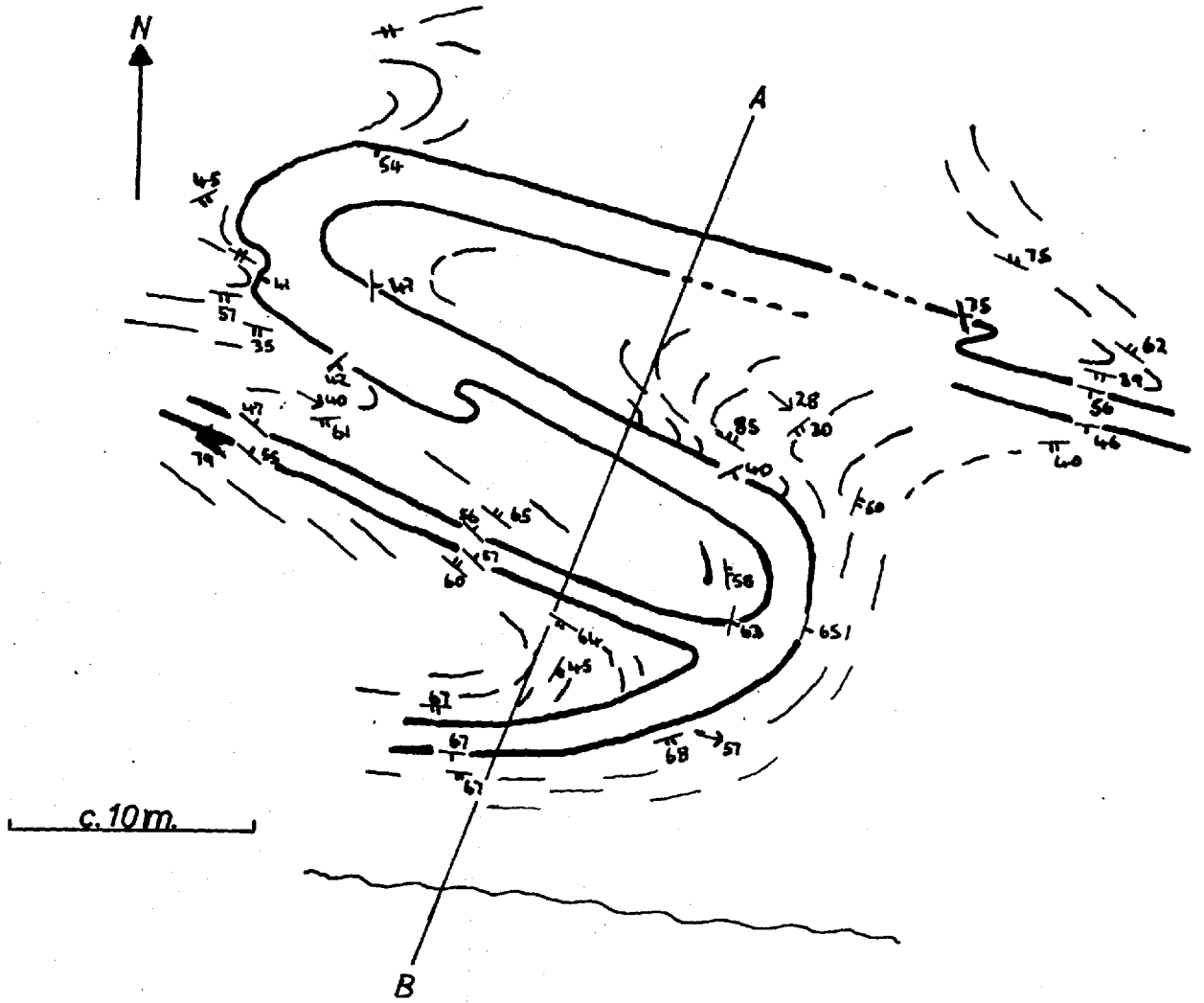
- a) Natural Arch Dyke,
- b) Mullach Dyke,
- c) Tarbet Dyke.

A). Natural Arch Dyke.

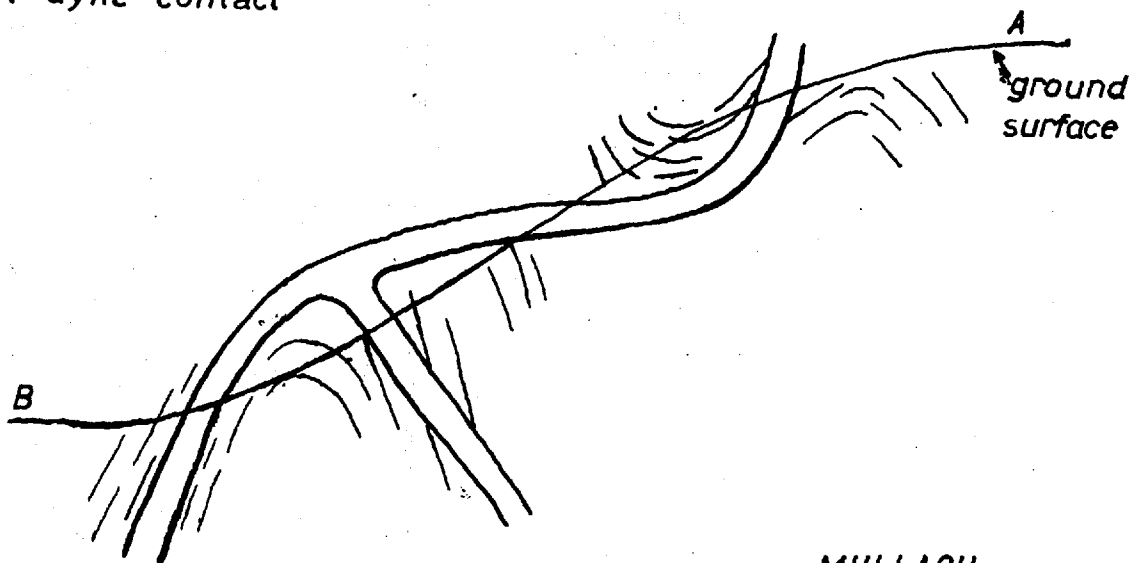
This is the best exposed dyke and a detailed map is presented in map 2 (cover pocket). Mapping of the mineral fabric in the dyke is purely qualitative.

It is evident that the dyke has been both thinned and folded and it is possible to show that these effects are results of separate phases of deformation. This is most obvious from inspection of the main antiformal hinge zone of the dyke (see map). Here the dyke has been thinned (though not actually broken) and after extension, this zone has been compressed and the cusp structure formed. Boudinage of an igneous dyke indicates that during this deformation the dyke was more competent than the surrounding gneiss. At a later stage cusps formed, and it is apparent that the gneiss was more competent than the amphibolite in the cusp. This phenomenon of changing relative competency with a change from dolerite to amphibolite has already been described by Coward (1969) and Francis (1969) in the Outer Hebrides.

Distinct fabrics were formed during heterogeneous thinning and during folding of the dyke. In the first phase of deformation (thinning) a planar fabric was formed in an amphibolite skin to mainly undeformed cores of dolerite. This planar fabric is parallel to the contact of the dyke. During folding more complex fabrics were formed. The dyke is crossed by a number of zones of high strain which are approximately parallel to the fold axial surface. These zones are analogous to the shear zones described by Ramsay and Graham (1970), except that the boundary conditions specified in that paper are not completely fulfilled here. For this reason the strains developed in these zones cannot be analysed in terms of simple shear. However, the zones are bounded by undeformed dolerite and where the direction in which



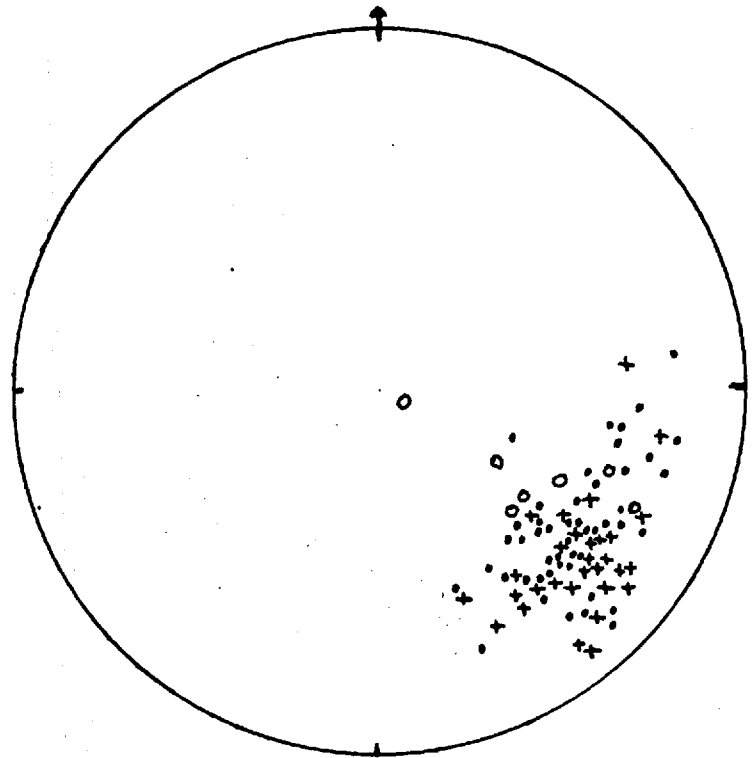
- ⊥ gneiss foliation
- gneiss lineation
- ⊥ dyke contact



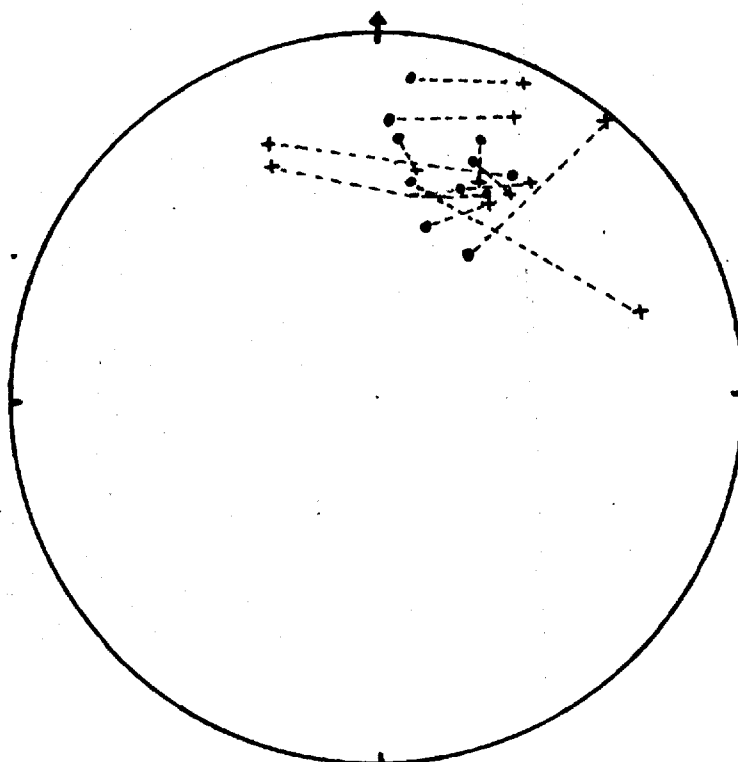
MULLACH
DYKE.

A. Lineation L2 in Natural Arch Dyke.

- L2 in dyke.
- + L2 in adjacent gneiss.
- Fold axis defined by dyke contact.



B. Dyke-gneiss relationships in the Tarbet Steep Belt.



- pole to dyke contact.
- + pole to gneiss foliation at dyke contact.

the foliation curves at the margin of the zone can be seen, a sense of displacement may be defined. The mineral fabric developed in these zones is an LS fabric and by analogy with the structures analysed by Ramsay and Graham (1970) the lineation is considered to define the direction of maximum finite extension. Fig.2.2A shows a plot of the lineations from the dyke.

The fabrics formed during folding are often superposed on the earlier fabric. It can be seen in the field that strong linear fabrics result from this superposition.

The brevity in extent of outcrop of this dyke severely hinders any interpretation of the original relation of the dyke and the gneisses. Two points are clear from the existing exposure:

- a) the southerly fold limb shows a lesser development of fabric than the northerly fold limb, and
- b) the thinned parts of the southerly fold limb did not form axes of folding in the second phase of deformation in a manner similar to the formation of the main fold hinge (see map). It would appear that the southerly fold limb has been modified little after initial extension.

The foliation of the gneisses is parallel to the axial surface of the fold defined by the dyke, and the lineation in the gneisses is continuous with the lineation in the dyke. The latter has already been interpreted as a finite stretching direction, and this interpretation must be extended to the gneissic lineation.

An interpretation of the fold either in terms of single layer buckling or in terms of heterogeneous shear is difficult. Finally it must be pointed out that few other dykes in this region have been folded like the Natural Arch Dyke.

B). Mullach Dyke.

This dyke differs from the Natural Arch Dyke in that it lies on the south, rather than the north, of the Tarbet antiform. Fig.2.1 presents an outcrop map of the dyke. The outcrop lies

on a steep slope and there is no topographic control. The map is presented as it would be seen looking directly down on it and for this reason the fold shape is accentuated.

Two observations made here need emphasising:

a) the dyke has a steeply dipping branch which consists of undeformed dolerite (both centre and margins), and this branch cuts through steeply dipping foliation;

b) both the main dyke and the gneissic foliation are folded.

The first observation indicates that the original mineral banding of the gneiss had been folded in part into a steep position prior to dyke intrusion, in a similar manner to that already described in section 2.2. The folding of the dyke is a result of Laxfordian deformation. The geometry that is now seen is the pattern that results when two discordant surfaces (ie. the dyke and the gneissic foliation) are folded together (Ramsay 1967, ch.9). However, with an apparently variable dip of the foliation prior to folding, plots of the dihedral angle across a fold profile are difficult to interpret. Modification of this geometry by superposition of additional flattening strains (Ramsay 1967, p411) would result in alternate dyke fold limbs dipping more and less steeply than the gneissic foliation.

c). The Tarbet Dyke.

This dyke is situated in the Tarbet Steep Belt and it has been mapped straight on to the areal map (cover pocket). It differs from most other dykes in its greater thickness and in the predominance of gabbroic (rather than doleritic) textures.

The important observations are:

a) Within the dyke, strains are heterogeneous, high strains being developed mainly in zones which again approximate to the shear zones described by Ramsay and Graham (1970). Large portions of the dyke remain undeformed. The lineation within the strain zones has the same plunge as and is continuous with the

lineation in adjacent gneisses (cf section 2.3A and fig.2.2A and 2.4).

b) The dyke has a wave-like outcrop, the parts of the dyke striking southerly being the least deformed, whilst those parts striking south east contain a greater number of the shear zones.

c) The dyke contact cross cuts the gneissic foliation. In general the strike of the contact lies in a clockwise direction from the strike of the foliation.

D). Other dykes.



A number of other dykes have been mapped in the Tarbet Steep Belt. The usual features observed are a dyke up to one or two metres wide, with a foliation developed throughout, cross cutting the pre-existing gneissic foliation. The relation between the dyke contact and the gneissic foliation is shown in fig.2.2B. In all cases the contact lies in a clockwise direction from the foliation, and in all but two cases the contact dips more steeply than the foliation. This is an important relation that will be referred to later.

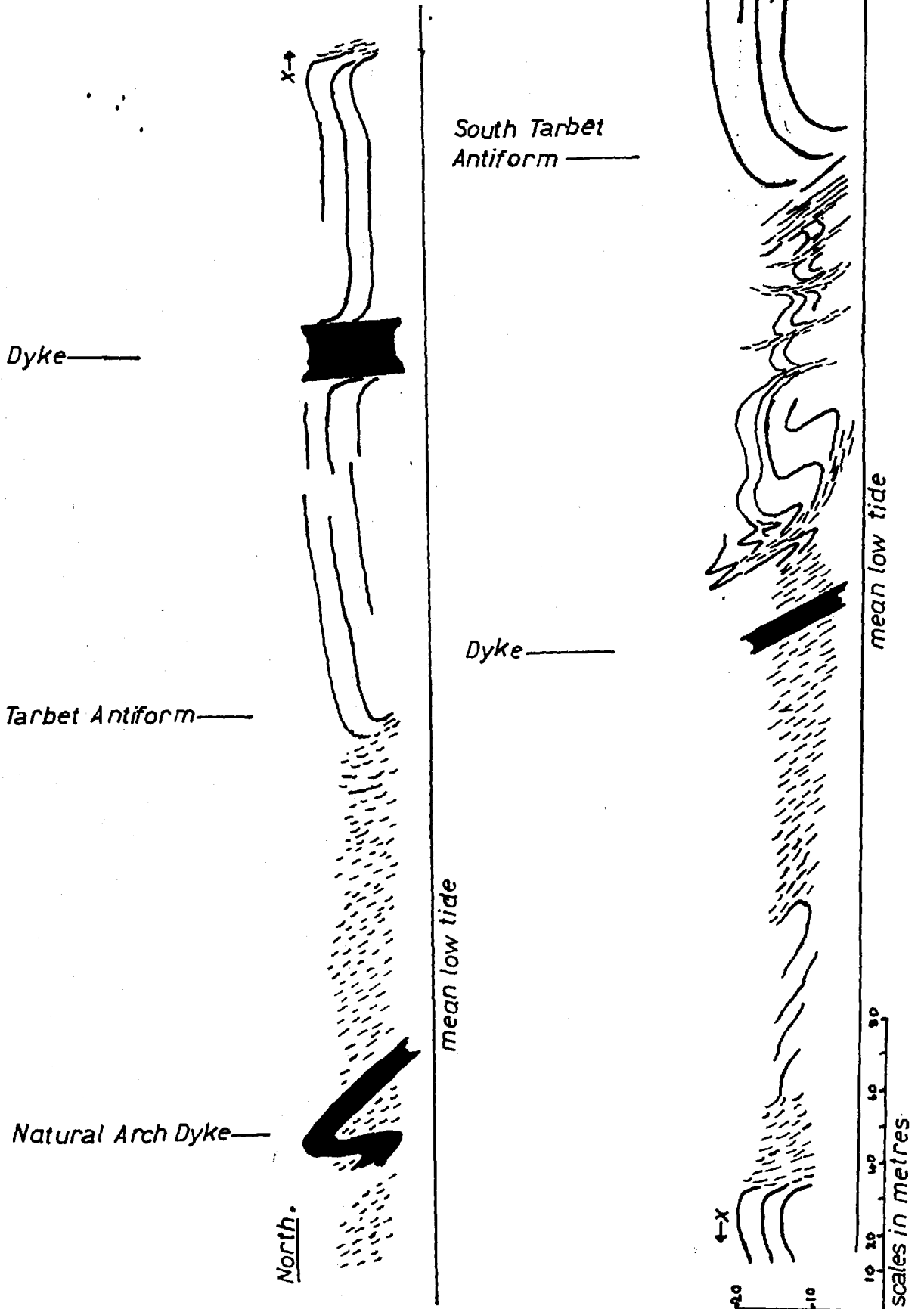
2.4 Major structure.

The presence of a major fold structure in the Scourie-Laxford region has been proposed by both Lambert and Holland (unpublished). Lambert details a structure similar to the antiform dated as late Scourian by Evans (1963, 1965) at Lochinver (personal communication) and very different from that described by Sutton and Watson (1951, 1961).

The recent work of Beach et al (in preparation) has shown beyond doubt the existence of a series of folds of Scourian age (cf, section 1.4), though these do not correspond at all closely to the early structures envisaged by Lambert. This section will present details of some work carried out by the writer in the area around Tarbet, with particular reference to the nature of the strains resulting from Laxfordian deformation, and to the

COASTAL SECTION ACROSS TARBET ANTIFORMS

 Scourian and modified Scourian fabric
 Laxfordian SL fabric



nature of the transition between the area of low and the area of high Laxfordian deformation.

From detailed mapping along the coast a continuous section across strike has been compiled. Part of this section is reproduced in fig.2.3, and its location is indicated on the map (cover pocket).

The flat lying Scourian banding and modified banding is overturned across two antiformal fold hinges, the main Tarbet antiform, and the subsidiary South Tarbet antiform, to the south. Northwards from the Tarbet antiform is a tract of gneiss one kilometre wide which forms the overturned limb of the Tarbet antiform and is named the Tarbet Steep Belt. The characteristic features of this belt are the steep, south westerly dipping foliation, the presence of strong LS mineral fabrics, and the almost complete absence of remnants of minerals and textures that were formed in the early Scourian (granulite facies) metamorphism.

A). Fabrics and minor structures in the Tarbet Steep Belt.

The Scourian mineral banding can be traced over both the South Tarbet and the Tarbet antiforms. The South Tarbet antiform, and the folds immediately north of it (see fig.2.3), fold the lineation L1. This feature cannot be seen at the Tarbet antiform because of the poor exposure in this area.

Within the Tarbet Steep Belt it is still possible to recognise modified Scourian banding. Layers of hornblende replacing pyroxene can be seen, and in basic bodies, remnants of granulite facies assemblages can be found. It is also possible to trace Scourian granite sheets (in the unmodified Scourian these sheets are conformable with the main mineral layering) as continuous layers in the Tarbet Steep Belt. North of Tarbet village it becomes more difficult to recognise modified Scourian banding, except in the basic and ultrabasic bodies (cf. Sutton and Watson 1961, fig.4) because of the intense recrystallisation in this area.

It is not possible to separate Laxfordian and Scourian mineral fabrics within the Tarbet Steep Belt, as can be done in some areas (cf. section 1.4). This is because the Laxfordian deformation and recrystallisation is sufficiently strong to overprint earlier fabrics. However, it is known (Beach et al) that the Scourian folds did have axial surface mineral fabrics.

The mineral fabrics dominant in the Tarbet Steep Belt are LS fabrics, with components S₂⁰ (this is the second foliation of Peach et al 1907) and L₂. This LS fabric is also present in the dykes, thus showing that its present form is a result of Laxfordian deformation and recrystallisation.

It is a feature of the more competent bodies (ie. Scourie dykes and early basic igneous bodies) that Laxfordian deformation resulted in a heterogeneous strain distribution within these bodies. In particular, zones of high strain are frequently seen. In both dykes and basic bodies these shear zones have a sense of downthrow to the south (cf. Sutton and Watson 1961 fig.4). Beach et al (in preparation) have shown that this statement can be generalised to the whole region affected by Laxfordian strains. This phenomenon indicates that the shear zone structures are related not to the fold forming phase of (Scourian) deformation, but to a phase of deformation superposed on the folds. This is apparent immediately in the case of the dykes since they are themselves younger than the folds, but it must also be true for the basic bodies.

By analogy with the analysis of Ramsay and Graham (1970) the lineation within these shear zones is interpreted as recording the finite stretching direction. This lineation is continuous from dyke (or basic body) into gneiss, and thus the lineation L₂ in the gneisses is also interpreted as recording the finite Laxfordian stretching direction. Minor folds are not common in the Tarbet Steep Belt, but invariably the minor fold axes are parallel to L₂. This is consistent with the interpretations made above, since the axes of Scourian folds will be rotated into the finite stretching direction after a moderate Laxfordian deformation.

At any one outcrop the lineation has a constant plunge, but this plunge varies from outcrop to outcrop. Fig.2.4A is a plot of the lineation L2 in the Tarbet Steep Belt. There is a pattern in the variation of direction and amount of plunge of the lineation which suggests that the finite strain considered across the strike of the foliation is not homogeneous. This is most clearly seen by reference to the wave-like form of the Tarbet Dyke (see map). The lineation is steeper in plunge in the gneiss tracts along strike from the parts of the dyke striking nearly southerly (ie. the least deformed parts of the dyke, see above section 2.3C), and similarly, L2 is less steep in the gneiss tracts along strike from the parts of the dyke having a south easterly strike. The latter are interpreted as the more deformed belts. Two interpretations of this pattern are

a) L2 has been deformed in a plane subsequent to its formation, and

b) rotational strains were present during the formation of L2.

B). Deformation in the area between the Tarbet and the South Tarbet antiforms.

This is an area that is structurally heterogeneous; the following structural elements can be seen:

a) flat lying but amphibolised Scourian mineral banding,

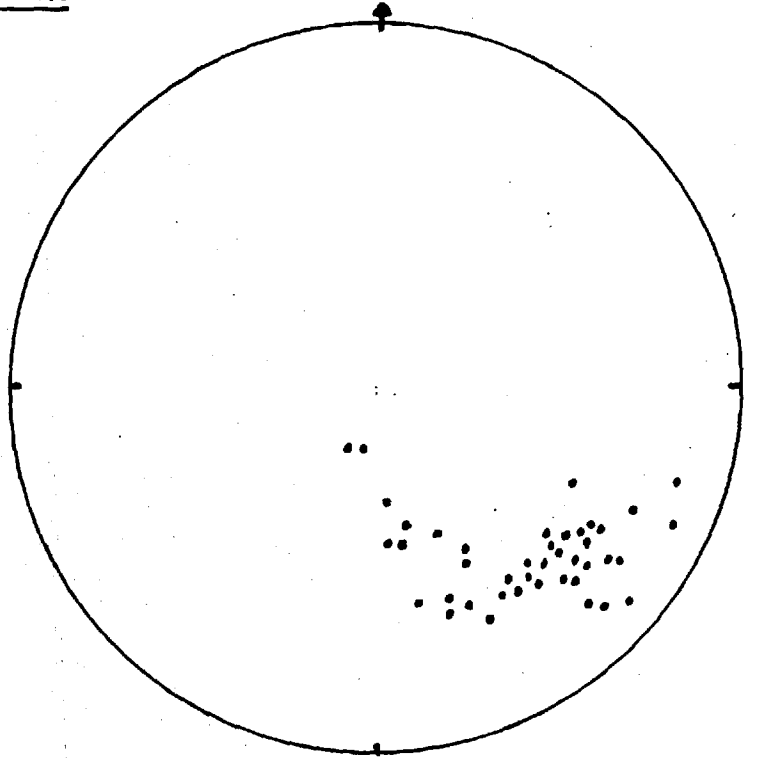
b) Scourian banding folded by Laxfordian deformation. (see above section 2.3B),

c) deformed Scourian fold structures.

The existence in this area of pre-dyke structures similar to those described earlier in section 2.2 is shown by the fact that a) vertical foliation is cut by undeformed parts of the Mullach Dyke, b) the dykes truncate the lineation L1 in the areas where the Scourian banding is still flat lying, and c) the dykes truncate axial surface hornblende fabrics.

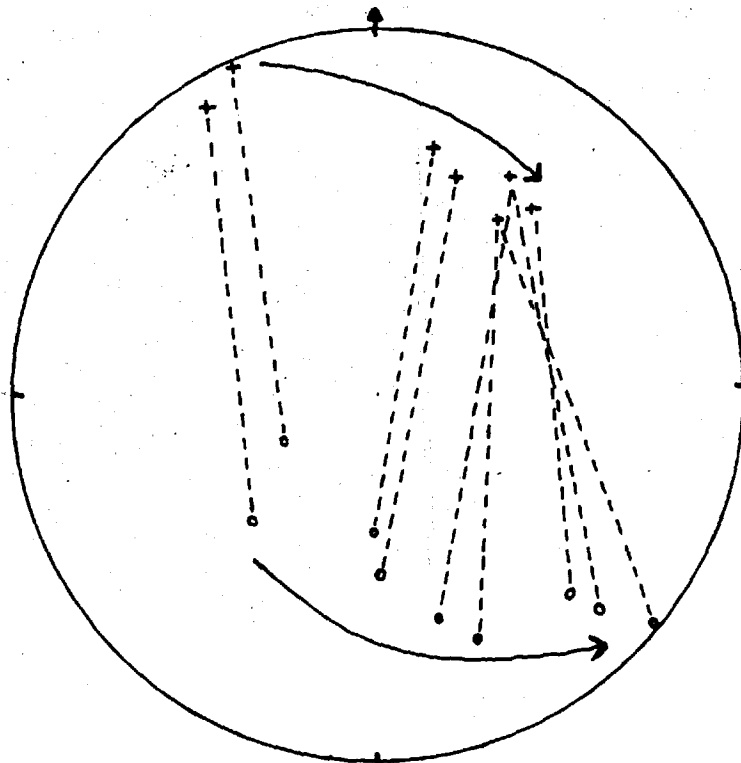
The Laxfordian modification of the earlier steep belts will be described here. The lineation L1 is developed strongly in these steep belts and there is no evidence to suggest that

A. lineation L2 in Tarbet Steep Belt



B. Laxfordian deformation of steep belt
 arrows show direction of rotation.
 tie lines join readings at one outcrop.

- + poles to gneiss foliation.
- lineation L1.
- lineation L2.



the lineation did not develop at the same time as the steep belts Laxfordian deformation of these structures is shown by:

- a) change in strike and dip of the foliation,
- b) rotation of L1 in the plane of the foliation, and
- c) growth of S2 and D2, L1 eventually being overprinted.

Fig.2.4B shows the geometry of this rotation. The rodding lineation L1 is rotated through about 150° (measured in the plane of the foliation) and is seen to be rotated through the direction of the lineation L2. The latter lineation has already been interpreted as the finite stretching direction. For the observed rotation of L1 to be compatible with this, it is necessary that an earlier deformation, pre-dating the formation of L2, resulted in substantial rotation of L1. This earlier rotation could either be pre-dyke or post-dyke in age. There are reasons to believe that both a pre-dyke and a post-dyke (but pre L2) rotation could have occurred. The evidence is as follows:

- a) it is thought that L1 formed prior to the second phase of Scourian deformation, and that it is deformed by these structures;

- b) in mapping the Natural Arch Dyke, two phases of (Laxfordian) deformation were distinguished, and it was during the second of these that the main Laxfordian LS fabrics were produced.

Whilst L1 was rotated in these deformed steep belts, the foliation also changed from a dip of $80-90^{\circ}$ to the south east to a dip of $40-50^{\circ}$ to the south south west.

C). Discussion.

The work described in this thesis was carried out prior to the recent mapping done by Beach, Coward and Graham. Initially it was necessary to deduce, from the work done in the small area around Tarbet, the age of the Tarbet antiform. This is no longer necessary because the mapping carried out by Beach et al, covering a larger region, has shown that the Tarbet antiform is one of a series of Scourian folds. In this section attention is still

focussed on the Tarbet Steep Belt as defined earlier, and an attempt is made to evaluate the nature of the dominant Laxfordian strains in this area. The arguments are based on the relationships between the Scourie dykes and the gneissic foliation. Two general models are proposed, one involving flattening strains, the other involving simple shear strains. The problems are considered in two dimensions only.

The present positions of the dykes and gneissic foliation are known. The dykes dip consistently more steeply than the foliation (this generalisation has been confirmed in the whole region mapped by Beach et al) and in the Tarbet Steep Belt the discordances seen in vertical section are about $5-10^{\circ}$. The dip of the foliation prior to Laxfordian deformation is known to be 90° in the little modified folds north of Ben Auskaire (section 1.4) and it is assumed that the pre-Laxfordian foliation at Tarbet was also vertical.

It is thus possible to construct fig.2.5, which shows a section across part of the Tarbet Steep Belt and the discordant relations given above. In the first model it is assumed that the Laxfordian strains are the sum of increments of irrotational flattening strains. In this case the changing angular relations between two lines (or planes) during deformation are dependent only on the ratio of the principal strains, and this is described by the Wettstein relation (see Ramsay 1967, p67, 129). The position of the trace of the XY plane of the strain ellipsoid has to be assumed. The position of the foliation now seen is the result of the sum of Laxfordian and earlier strains, and if the strain is to be removed with the foliation considered as a passive marker, the XY plane cannot be coincident with the plane of the foliation, since this would involve theoretically infinite strain ratios. In this example the trace of the XY plane is taken to be a few degrees less steep than the trace of the foliation (see fig.2.5).

Using the Wettstein relationship, the strain has been removed from the angular discordance shown in fig.2.5 until the trace of the foliation is vertical. In this particular example

DEFORMATION OF TARBET STEEP BELT

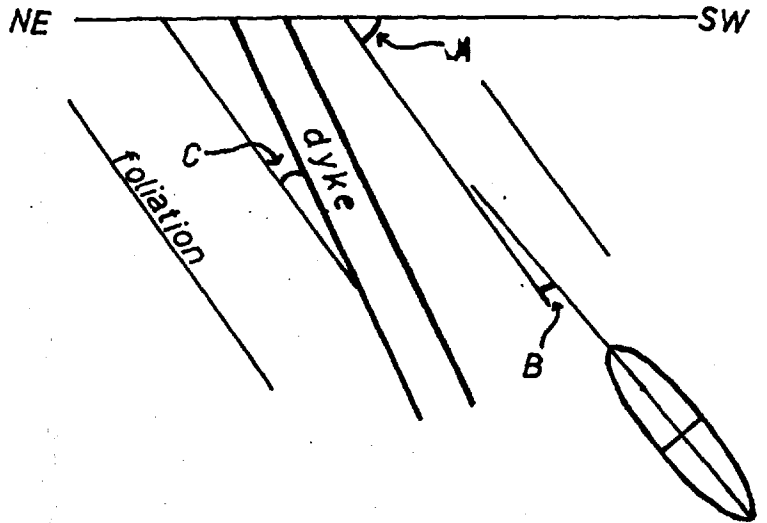
a) irrotational flattening

deformed

$A = 50-60^\circ$

$B = 5^\circ$

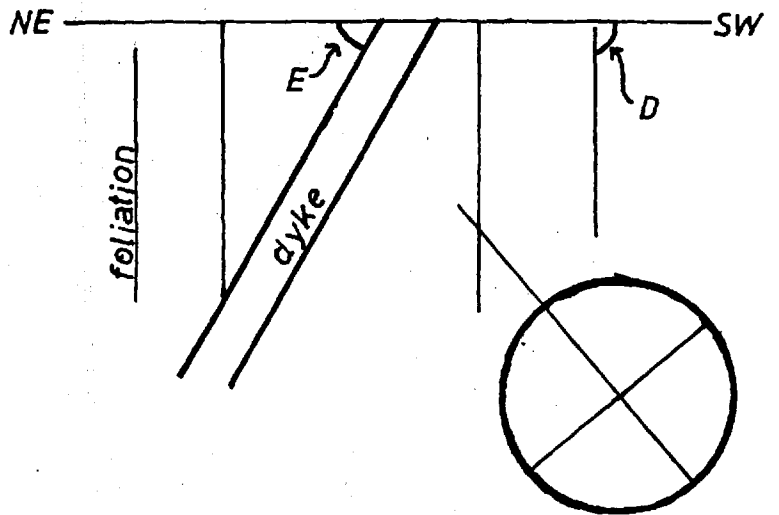
$C = 5-10^\circ$



undeformed

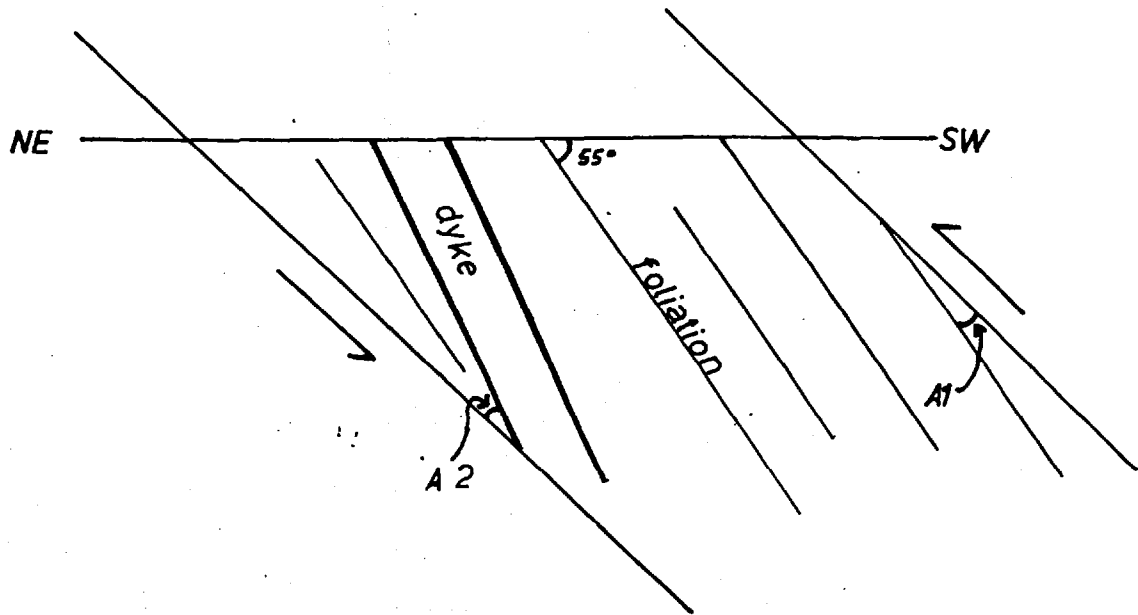
$D = c. 90^\circ$

$E = c. 60^\circ$



DEFORMATION OF TARBET STEEP BELT

b) simple shear



$$\cot A' = \cot A + \gamma$$

foliation	deformed	undeformed	$\therefore \gamma$	$\therefore 1+e1/1+e2$
	$A1'$	$A1$		
	5°	40°	10.2	30
	10°	45°	4.7	10

dyke	$A2'$	γ	$\therefore A2$	\therefore initial dip
	10°	4.7	135°	0°
	15°	10.2	170°	30° SW
	20°	4.7	150°	15° SW

the strain ratio necessary to cause this deformation is found to be $1+e_1/1+e_2 = 10/1$. The strain is then removed from the dyke in the same way, but this time using the calculated strain ratio and leaving the initial position of the dyke as unknown. It is then found that in the undeformed state the dyke dips 60° to the north east.

In spite of the simplicity of the calculation, a strain ratio of 10/1 is compatible with observed rock fabric (thinning of mineral banding) in the Tarbet Steep Belt, and the estimated initial position of the dyke is the same as the orientation of undeformed dykes in the Scourie Zone (Sutton and Watson 1961), and throughout Assynt in general.

The strain has been removed from the same angular relations as shown in fig 2.5 until the foliation is vertical on the basis that the finite strain is the sum of increments of simple shear. The equations of Ramsay (1967, section 3.10) were used in this second model. It will be appreciated from these equations that small variations in the angles of discordance affect the computed value of shear strain very much. This is a feature of simple shear type deformations. For this reason the results of three different calculations are presented in fig.2.6. Again, the initial orientation of the dyke has been computed. In each case an orientation very different from that of the undeformed dykes at Scourie is obtained and it must be concluded that rotational simple shear strains played only a minor role in the development of the Tarbet Steep Belt during the main Laxfordian deformation.

This simple analysis suggests that the main Laxfordian deformation in the Tarbet Steep Belt was accomplished by flattening amounting to a bulk strain of about 10/1, the finite XY plane of the strain ellipsoid dipping to the south west; the principal elongation direction (that is, L2) plunges about $30-40^\circ$ to the south east in this plane. This statement must remain a generalisation because of the variations in structure within the Tarbet Steep Belt that have already been discussed, and because of the assumptions made in the computation. The sharp transition from

high to low finite Laxfordian strain southward across the Tarbet antiform results from the fact that Laxfordian strains have been superposed on an earlier (Scourian) structure consisting broadly of steep foliation to the north and flat lying foliation to the south of the Tarbet antiform.

2.5 Shear zones.

A). Ideal geometry.

The Laxfordian shear zones in the Scourie region have been described by Sutton and Watson (1961), and shear zones of similar types are the subject of a paper by Ramsay and Graham (1970). From the analytical work of the latter, a number of features define the ideal geometry of a shear zone:

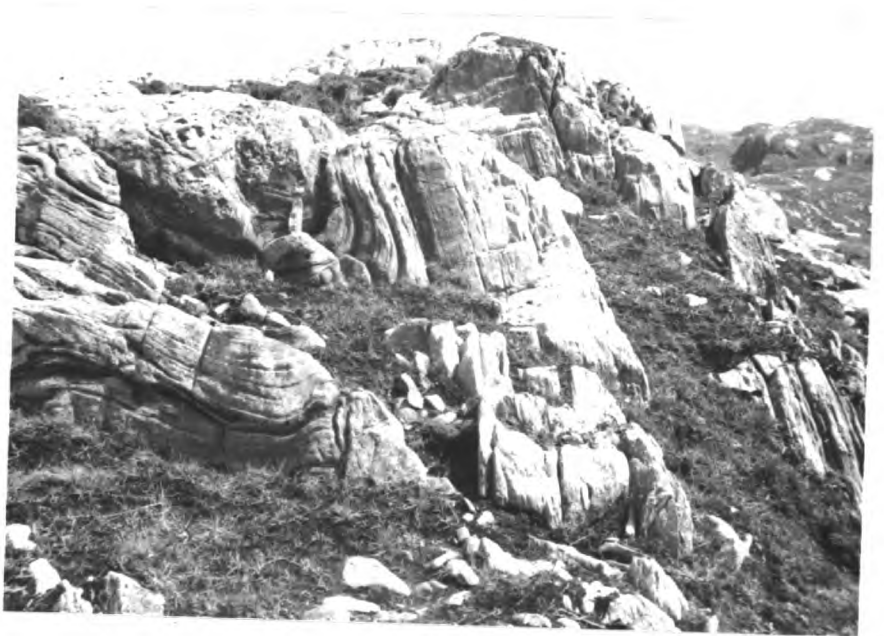
- a) the walls of the zone are parallel planar surfaces,
- b) outside the zone, the rock is undeformed during the period of formation of the zone,
- c) there must be a constancy of structure in one direction within the zone, i.e. the zone possesses certain elements of symmetry,
- d) the rock undergoes no volume change.

If these conditions are fulfilled, strains can be analysed precisely in terms of simple shear within the zone. It must be emphasised that no genetic implications are made by defining a shear zone in terms of geometry and strain distribution.

B). Departures from ideality.

In the Scourie area the commonest departures from these conditions are that the bounding surfaces of the zones are not always planar and parallel, and that the shear zones are in fact finite in length. Often volume changes can be inferred to have occurred within the shear zones. Proof of low to negligible strains in rocks flanking a shear zone is most

Laxfordian shear zones near Scourie



easily obtained when the zones cut across otherwise undeformed igneous rock. It is more difficult to demonstrate this point when the zones cut an already foliated or banded gneiss.

It is mainly because of this last point that it has not been found possible to analyse the strains in the shear zones at Scourie quantitatively using the methods described by Ramsay and Graham (1970).

C). Essential characteristics of the shear zones at Scourie.

The locations of the shear zones studied here are shown on the map (cover pocket). They are broadly the same ones as those studied by Sutton and Watson (1961). Three features are used to distinguish Laxfordian shear zones:

a) Geometry.

The shear zones rotate the Scourian mineral banding to a vertical or steeply dipping position, and the zones are geometrically monoclinical. They are essentially parallel sided and finite in length, their dimensions being up to 2m in width and 200m in length. The lineation L1 is folded into the shear zones. However, the angular divergence of the monocline fold axis and L1 external to the shear zone is often small, and the locus of the deformed lineation cannot be clearly defined and used in the way described by Ramsay (1967, ch.8). In a few cases, L1 lies initially at a high angle to the shear zone. The locus of one such deformed lineation is shown in fig.2.7. Interpretation of this and other loci is difficult, and since no additional information was gained from the loci, discussion is omitted.

b) Mineralogy and fabric.

The rocks within the shear zones show a thorough recrystallisation, both new fabrics and new minerals appearing. Biotite (a new phase) and hornblende (generally a recrystallised phase) are the most conspicuous minerals in the shear zones, and the fabrics result from the orientation of these minerals. The fabric developed in the shear zones is denoted LS, the planar

DEFORMATION OF L1 IN SHEAR ZONE.

FIG 2.7

L2 assumed to be stretching direction
in shear zone.

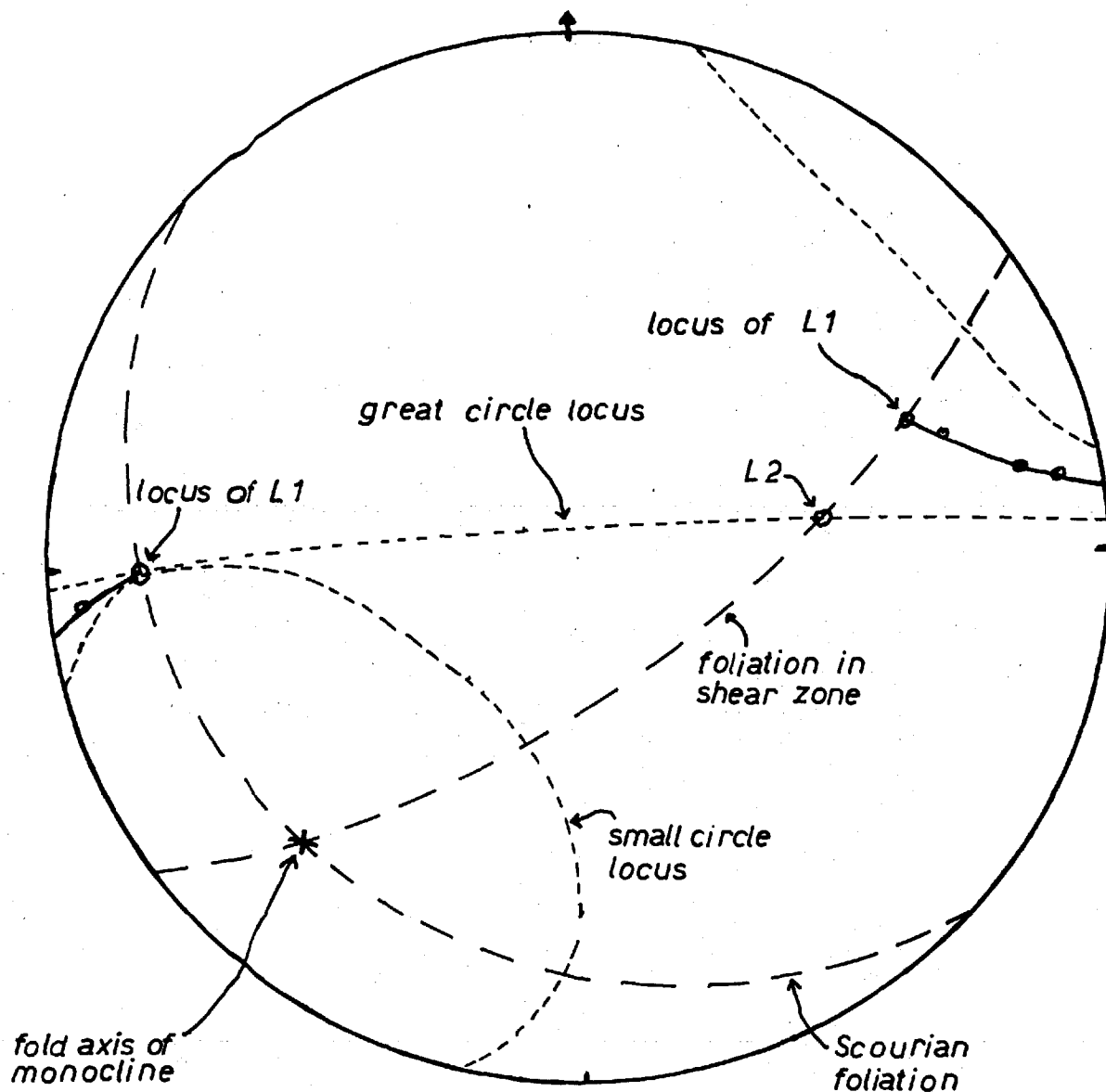
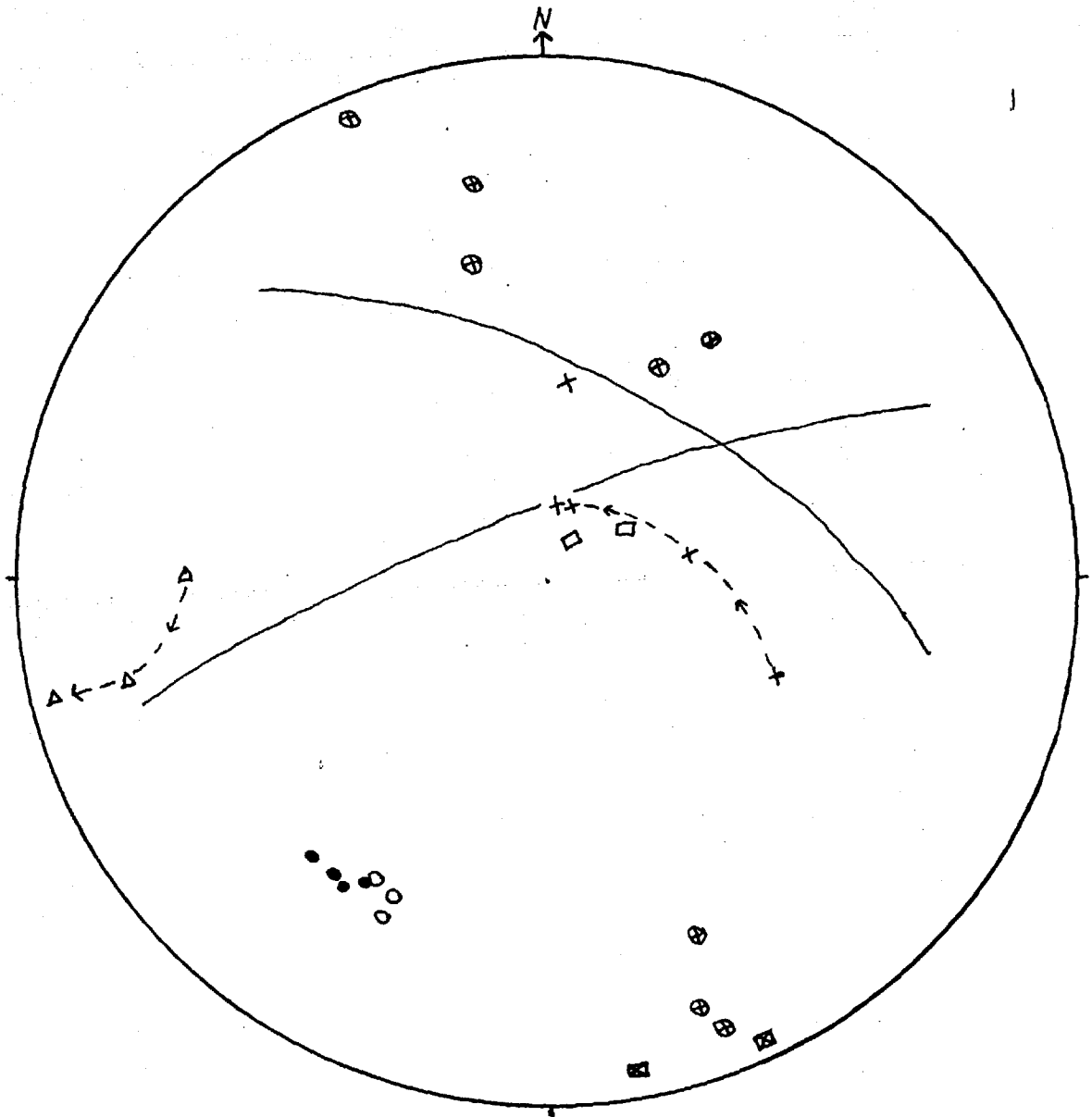


FIG 2.8

POLL EORNA

data from shear zone crossing the Scourie Dyke



pole to dyke contact

- east of shear
- west of shear

⊕ pole to gneiss foliation

△ L1 in gneiss

+ L2 in gneiss

⊕ pole to dolerite foliation

□ L2 in dolerite

Section across Sithean Mor

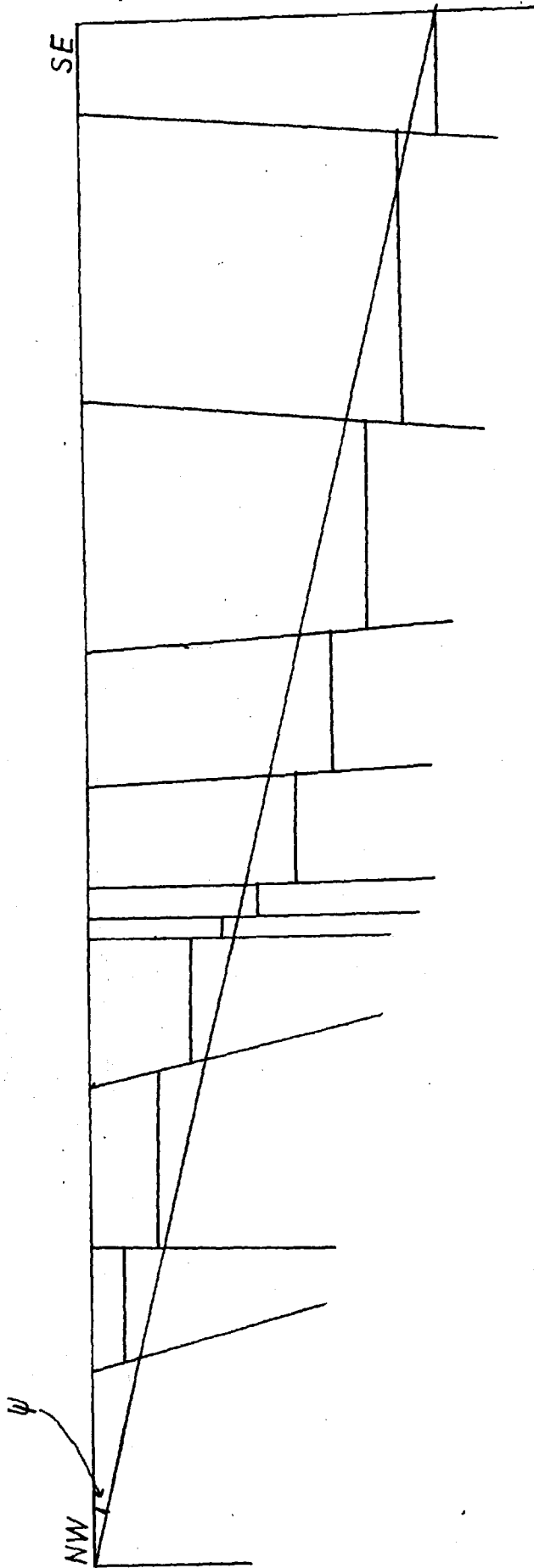
showing hypothetical displacements on shear zones

scale: 1 : 3517

$$\psi = 13^\circ$$

$$\tan \psi = 0.23$$

$$\frac{1+e1}{1+e2} = 1.5/1 \text{ approx.}$$



element of the fabric lying parallel to the original Scourian banding in the main part of the zone. Very rarely the new planar fabric is seen to cross cut the older banding in the margins of a shear zone.

c) Age.

Shear zones provide the only evidence of ductile deformation (in the area in which they occur) after the intrusion of the Scourie dykes and before the pre-Torridonian faulting. The shear zones displace the dykes, horizontal displacements often being only one or two metres. The small scale of the displacements results from the fact that the direction of inferred movement in the shear zones (cf. section 2.4A and see later) is very nearly parallel to the dyke walls. The displacements have not been shown on the map.

D). Displacements on the shear zones.

Because of the small angle between dyke wall and movement direction in the shear zone, any calculations of the amount of displacement on a shear zone are very sensitive to small errors in this measured angle. Only in one example was it considered that an accurate calculation of this displacement could be made.

The shear zone described by Teall (1885), from the Scourie dyke at Poll Eorna, can be traced from the dyke into the adjacent gneisses. The dyke at this point is about 25m wide. The shear zone within the dyke is about 1m wide and is bounded by undeformed dolerite. It can be traced northward into the gneisses as a linear zone of approximately constant width. However, when it is traced southward into the gneisses it increases in width to about 10m. This is partly due to interference of the shear crossing the dyke with an earlier shear along the southern margin of the dyke. For this reason only the northern margin of the dyke is considered, where no earlier shear zone is present.

The data collected at this locality is conveniently shown on a stereogram (fig.2.8). The horizontal displacement

of the dyke contact, measured in the plane of the shear zone foliation, is 11.3m. The position of the foliation in the shear zone records the position of the XY plane of the finite strain ellipsoid in that zone (see Ramsay and Graham 1970). It is not necessary for the finite strain ratio to be very large before the foliation becomes nearly parallel to the margins of the shear zone. This is easily appreciated from the graphs presented by Ramsay (1967, p86). Thus it can often be assumed that the finite elongation direction in the centre of the shear zone is parallel to the actual movement direction. This has been done in this example.

It is a simple geometrical calculation to compute the real displacement on the shear zone from the measured data. It is found to be 22m. for this example. It is unfortunate that this procedure could not be applied to a number of shear zones in the area.

A series of shear zones have been mapped across Sithean Mor (see map) which all have roughly the same width as the one at Poll Eorna. In order that some idea of the bulk strains resulting from displacements on shear zones in this region may be gained it has been assumed that each shear zone has a displacement of 20m. The movement directions are all slightly different for this set of shear zones (see map) and so a vertical section has been constructed. This is shown in fig.2.9. The most interesting feature of this construction is that the calculated bulk regional strain resulting from the narrow zones of shear is very small.- a shear strain of 0.23 gammas in a vertical direction. The implication of this is that in this area deformation by formation of shear zones was not a means of producing a large finite regional shape change. This contrasts with a region such as South Harris where deformation by shear zones has resulted in a large finite regional shape change (Graham personal communication). What is the significance of the shear zones at Scourie if regional deformation during their formation was relatively insignificant? This question, and related topics, are discussed in detail in chapters 9 and 10.

E). Discussion of field relations.

The general field observations are very much in agreement with those of Sutton and Watson (1961, p92-96), and for this reason it is unnecessary to give details here. However, a few points will be discussed.

There seems no doubt that the shear zones form a conjugate set with preferential development of one half (those with "down-throw" to the south east). The direction and angle of plunge of the line of intersection of the two shear planes is important if an interpretation of the stress orientation initiating shearing is to be made. The axial surface of the monoclinial fold geometry of the shear zones is very nearly parallel to the foliation within the shear zones, and so it is sufficient to plot the poles to this foliation in conjunction with the sense of asymmetry. This has been done, taking small areas at a time. The general result, though not precise, indicates a plunge of 20° to the south west of the line of intersection. The difference between this and the figure obtained by Sutton and Watson (70° to the south west) must be accounted for. Two factors contribute to this difference. First, Sutton and Watson included in their assessment structures that are now known to be Scourian in age (ie. the steep belts described in section 2.2 here). Second, the orientation of some shear zones is governed by these earlier structures in that the shear zones develop parallel to the earlier steep foliation. In the assessment made here, both these categories were excluded. The angle between the planes of shear in the conjugate set was found to be about $20-30^{\circ}$ (≈ 20).

In agreement with Sutton and Watson (1961) it is found that there is no reason to suppose that the majority of shear zones formed at more than one distinct time. Strictly, this idea needs slight modification. Along dyke margins the gneissic banding is often deflected into a vertical position (Peach et al 1907, plate XXI). From the point of view of linear distribution of strains and development of LS mineral fabrics, these structures must be grouped with the shear zones. However,

these structures contain no biotite and are deformed by shear zones which contain biotite. The importance of this mineralogical distinction will become apparent later. For the present it is sufficient to distinguish an earlier set of shear zones along dyke margins. The dykes are affected by these shear zones and this leads to a characteristic pattern often seen in the dykes of this region - the margins of the dyke are foliated, while the central portions are undeformed. This relationship, produced by heterogeneous development of strains, has been cited as evidence for multiple dyke intrusion by Bowes and Khoury (1965).

There is no evidence of multiple ductile deformation within the shear zones as proposed by Sutton and Watson (1961). Flinn (1965) has suggested that mineral reorientation during deformation is governed by the shape of the crystal and by the shape of the strain ellipsoid. The shear zones described by Ramsay and Graham (1970) show equal development of linear and planar fabrics (LS) which is compatible with a ~~constant~~ constant volume, plane strain deformation ellipsoid. In a similar way, most of the shear zones described here have LS fabrics, and though there is no way of proving it, it seems very likely that the strains are predominantly of the simple shear type. By analogy with Ramsay and Graham (1970), the lineation is interpreted as the finite elongation direction within the shear zone. Seen in this light, the shear zones are the result of one phase of deformation.

The interpretation of shear zones forming a conjugate set implies certain assumptions have been made regarding the genesis of the shear zones. However, the elaboration of a mechanistic model for the origin of the shear zones is delayed until later in the thesis.

CHAPTER 3.

MINERAL FABRICS.

Many studies have been made of the relationships between the different mineral grains in metamorphic rocks. Work by Kretz (1966, 1969), Flinn (1969) and Vernon (1968, 1970) has aimed at defining the criteria that may be used to judge whether mineral grains are in equilibrium with each other. Other studies (eg. Stauffer 1970, Beach 1969) have considered that information concerning the mechanism of deformation of a particular rock can be gained from microscopic examination of the mineral grains in that rock.

In most metamorphic rocks recrystallisation masks any evidence that individual grains may have been deformed (eg. by glide planes), and the recrystallised grains now seen are judged to be related to late- or post-tectonic annealing. The process of recrystallisation is little understood and the controlling factors of such a process in silicate rocks are far from clear. It seems doubtful if studies of optic orientation of minerals in such rocks could give information on, for example, the stress history during deformation.

In this chapter brief consideration will be given to three aspects of mineral fabrics. These are a) optic fabrics, b) dimensional orientation of minerals, and c) redistribution of material during deformation and metamorphism.

One of the problems that hinders interpretation of metamorphic fabrics is that end products, and not processes, are observed. From a study of shear zones such as those referred to in chapter 2 it is possible to observe the initial and final fabrics, and inferences as to the processes linking both states may be more easily made.

3.1 Optic fabrics.

An initial study of optic fabrics in shear zones has been made by Graham (in Ramsay and Graham 1970); studies of quartz orientation in relation to variations in finite strain suggested that there was a connection between mineral fabric and finite strain. With this in mind quantitative work has been carried out on feldspar and quartz fabrics from shear zones at Scourie.

The initial study was based on a dolerite from Castell Odair in North Uist (Ramsay and Graham 1970) and similar rock from a dyke at Scourie (sample 2). The undeformed igneous rocks have a sub-ophitic texture of randomly oriented plagioclase crystals. In the shear zone (the boundary conditions are those of simple shear, see Ramsay and Graham 1970) the dolerite is converted to an amphibolite (cf. Teall 1885). The sub-ophitic texture is flattened but preserved in outline, and each original plagioclase crystal has been recrystallised to an aggregate of 20-30 grains, the self boundaries of which are polygonal.

The orientation of all three optic axes (ie. X,Y,Z) of the plagioclase has been measured for the two shear zones mentioned. It was found that for both samples the X and Z axes show a near random distribution whereas the Y axis shows a broad girdle distribution. Fig.3.1 shows one of these girdles of Y axes. This procedure has been repeated for plagioclase in samples collected from shear zones cutting the granulites at Scourie. The result, a girdle of Y axes and randomly distributed X and Z axes, is the same as before except that the girdle is much broader in the sheared granulites.

Two factors are suggested that may control or influence the development of such a fabric:

a) Host control.

When a large crystal recrystallises into smaller grains it is possible (see for example Hobbs 1969) that the host grain lattice controls the orientation of developing nuclei. The essential initial difference between the plagioclase in the dolerite and in the granulite is that in the dolerite the

FELSPAR FABRIC IN SHEAR ZONE.

plot of plagioclase optic Y axes.
sample no sheared dolerite

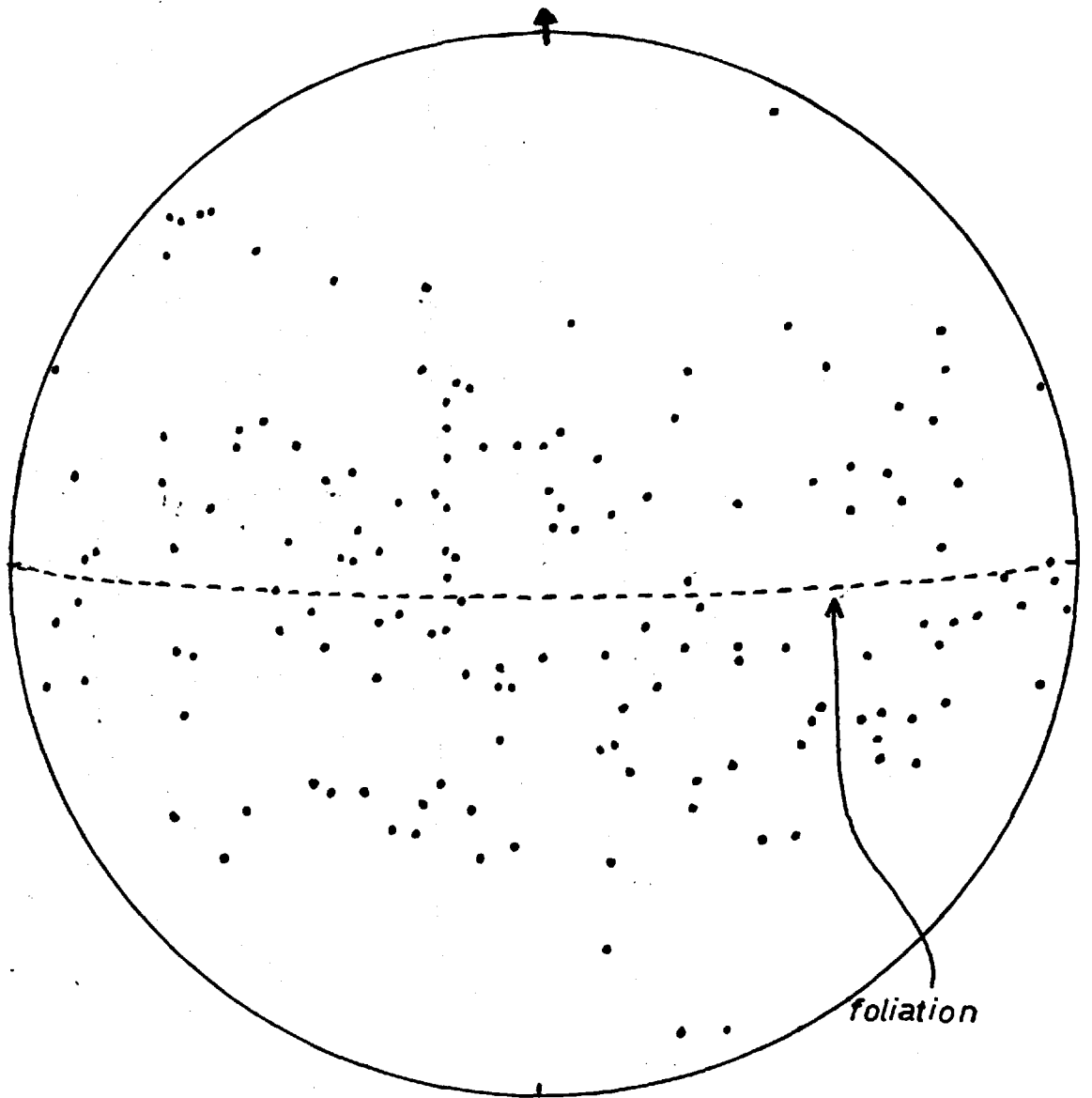
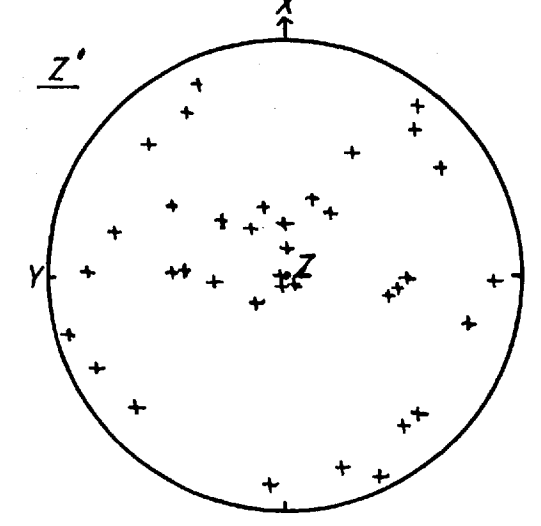
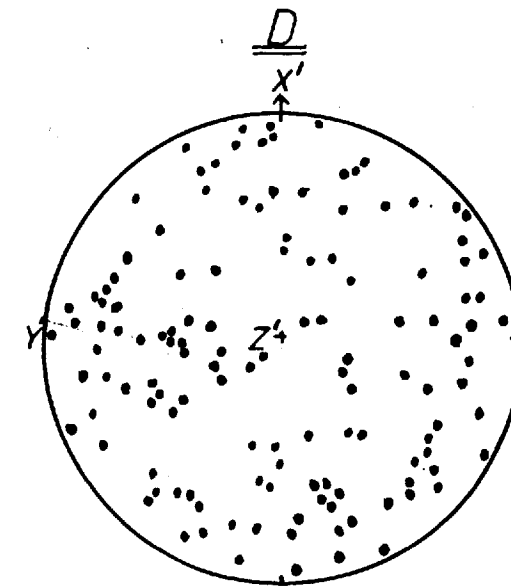
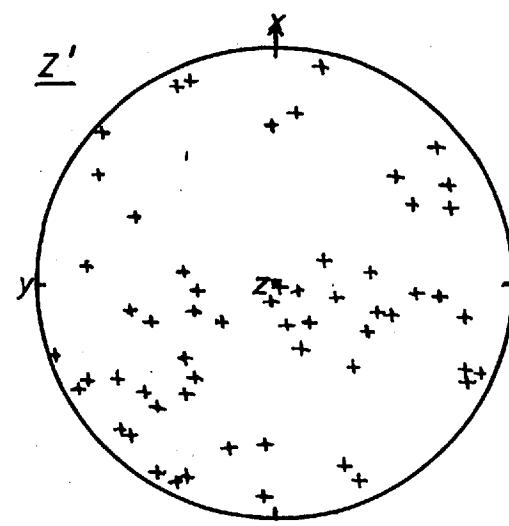
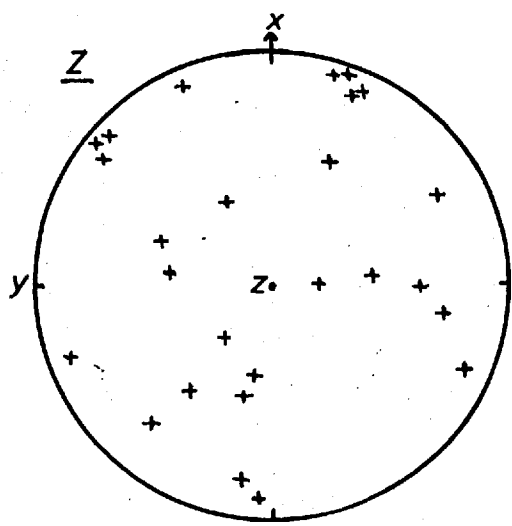
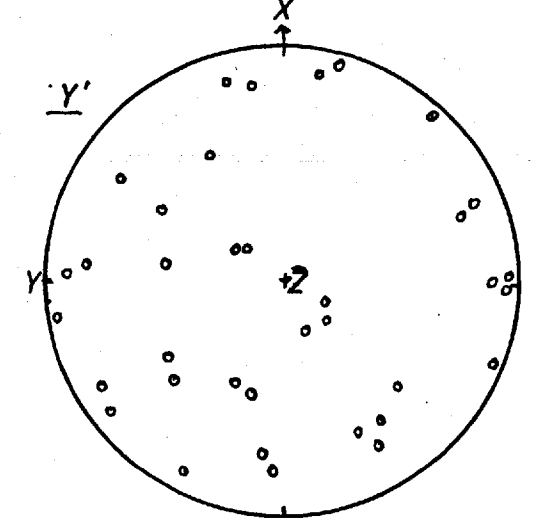
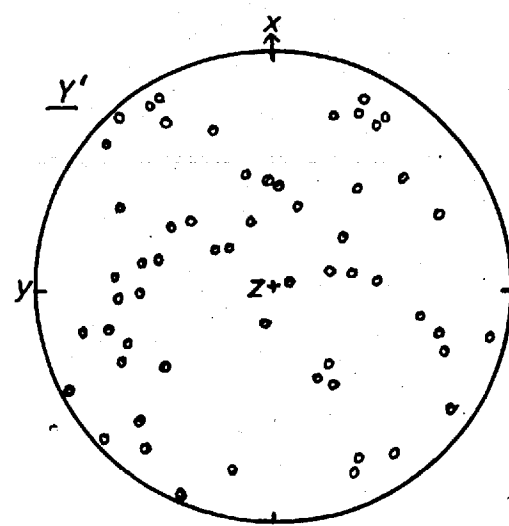
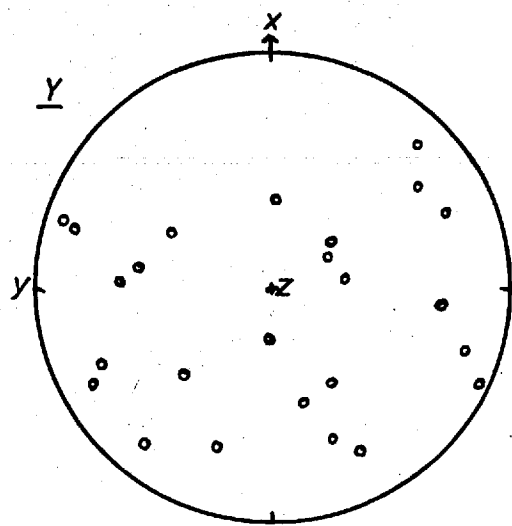
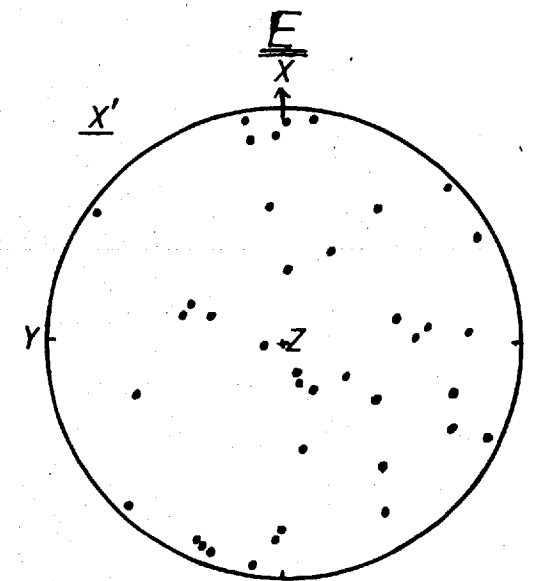
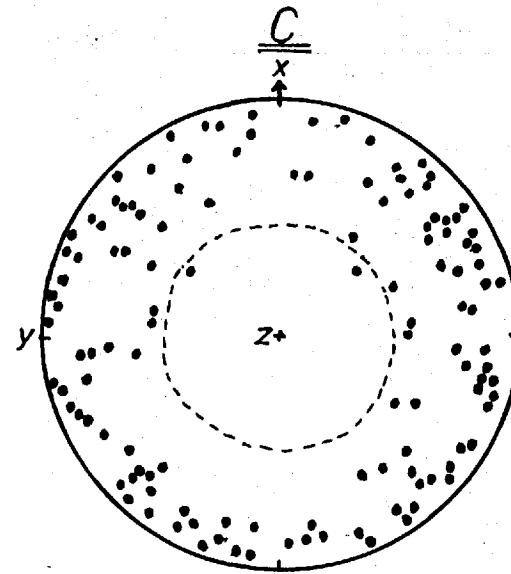
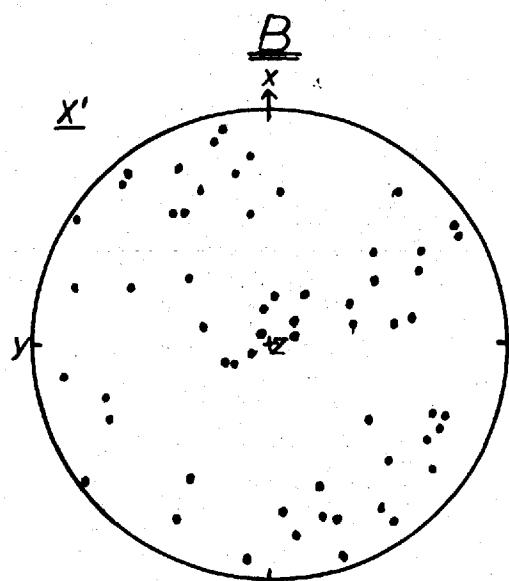
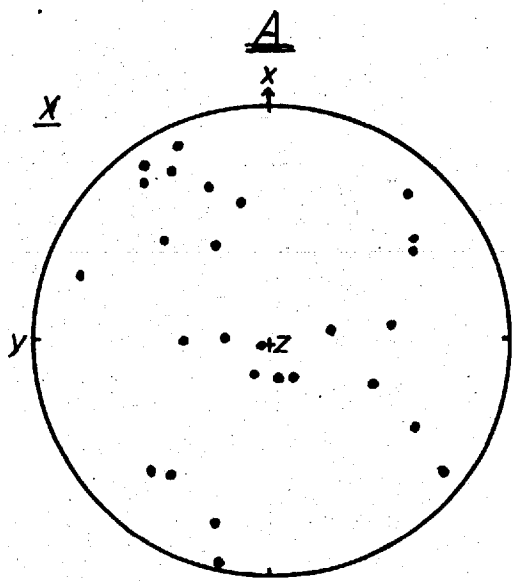


FIG 3.2

PLAGIOCLASE OPTIC FABRICS
see text for explanation



crystals had a high shape anisotropy, whereas in the granulite the crystals were approximately equidimensional. Thus the plagioclase in the dolerite will undergo a marked rotation during the first increments of strain and a preferred shape (and hence optic) orientation will be initiated. The plagioclase in the granulite will undergo relatively little rotation during strain. If the orientation of the recrystallised grains is controlled by the host lattice orientation, then stronger preferred orientations are to be expected in the sheared dolerite.

In an attempt to define this effect, a further quantitative study of undeformed dolerite (sample 2) was made. It is found that though the rock is undeformed, the large plagioclase crystals contain small and often discrete recrystallised grains. The study involved measurement of the optic orientation of these discrete grains and of their host crystals. The orientation of the grain boundaries of the discrete grains was also measured. The data were plotted in a number of ways, as follows and as shown in fig.3.2:

- A) The host crystal optic axes (X,Y,Z) are plotted in a cartesian framework (x,y,z).
- B) The optic axes of the recrystallised grains (X',Y',Z') are plotted in a cartesian framework (x,y,z).
- C) The boundaries of the new grains against host crystals are plotted in a cartesian framework (x,y,z).
- D) The axes of each new grain (X',Y',Z') have been rotated to a fixed position, and the orientation of the grain boundaries is plotted in this framework.
- E) The axes of the host crystals (X,Y,Z) have been rotated to a fixed position, and the axes of the new grains (X',Y',Z') are plotted in this framework.

Far more data are needed before any firm conclusions can be made concerning the control of host crystals on new grain orientation. From the limited data presented there is no apparent relation between host crystal and new grain orientation.

b) Surface anisotropy (Spry 1969, table 7, p148).

The most abundant phase in the shear zones being considered is hornblende. In this mineral shape anisotropy is a reflection of lattice structure and of surface anisotropy. Hornblende crystals become oriented during deformation because of their shape anisotropy, and with hornblende and plagioclase growing in close proximity, the surface anisotropy of the hornblende could affect the orientation of the nucleating plagioclase. The magnitude of this effect must depend on the number of hornblende-plagioclase boundaries.

It is suggested by the present writer that the quartz fabric diagrams presented by Ramsay and Graham (1970, fig.17) show the control exerted by the hornblende crystals on the orientation of the quartz grains. In these altered dolerites quartz occurs as small grains contained within individual hornblende grains. In any one hornblende crystal it is likely that the orientation of the smaller quartz grains will be very much influenced by the orientation of the hornblende. It is because of the fact that the hornblende crystals adopt a preferred orientation themselves (because of their shape anisotropy) after a certain amount of deformation that the quartz grains within these hornblendes also adopt a preferred orientation. If this hypothesis is correct, then the changing quartz fabrics in these shear zones (see Ramsay and Graham 1970) do reflect indirectly the changing finite strain states across the shear zone.

Quartz fabric diagrams from a shear zone cutting a granitic gneiss at Scourie (sample 18) show a random orientation of the quartz optic axes. This rock consists dominantly of quartz and feldspar, both shape and surface anisotropies being small.

3.2 Shape anisotropy.

The dimensional orientation of minerals in metamorphic rocks is an important aspect of mineral fabrics. Alignment of minerals in planar and linear fabrics is partly a reflection of the amount of strain and type of strain in a rock (Flinn

1965). When a rock is deformed crystals with a shape anisotropy will rotate: they can be considered as planes and lines within a strain ellipsoid (Ramsay 1967, p162). Coward (1969) and Graham (1969) expanded the idea that obliquely superposed strains could account for changes in mineral fabrics in areas of polyphase deformation, and applied this concept to the Laxfordian rocks in the Outer Hebrides.

In addition to external rotation of crystals by strain, there may be internal rotation of a crystal (eg. by gliding, see Carter and Raleigh 1969). In most metamorphic rocks this latter process is masked by recrystallisation.

It is commonly observed that minerals such as hornblende and biotite recrystallise to produce planar and linear fabrics. It is less readily appreciated that recrystallisation can be itself a mechanism of strain in a rock. The principle of least work is equivalent to saying that when rocks deform, they do so by the easiest possible process. Experience in metallurgy has shown that cataclasis becomes less important as a mechanism of deformation at high temperatures. If recrystallisation, a temperature controlled process, is the only process operative in a rock under deviatoric stress, then material will be redistributed to accomplish maximum strain with minimum work. In general where recrystallisation is merely dominant, this process only contributes to the total strain.

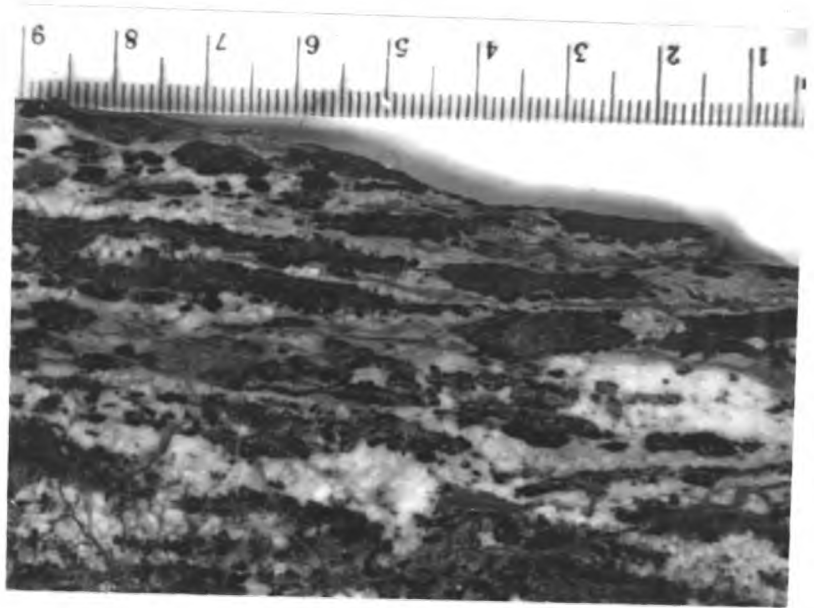
Two examples of a strain resulting from recrystallisation will be described.

Plate 3.1 shows a specimen from a shear zone crossing Scourian gneiss (sample 53). The undeformed equivalent of this sample contains spherical garnets (for details see chapter 6). In the shear zone these garnets have been altered to aggregates of biotite. The biotite has not formed spherical aggregates, but aggregates which in two dimensions are seen to be elliptical. The growth of biotite in a preferred direction has formed a crude foliation in the rock, and has accomplished a shape change relative to the undeformed rock.

Sample 22 was collected near Sithean Mor and is from a minor fold of Scourian age (see section 2.2A,B). Hornblende

PLATE 3.1

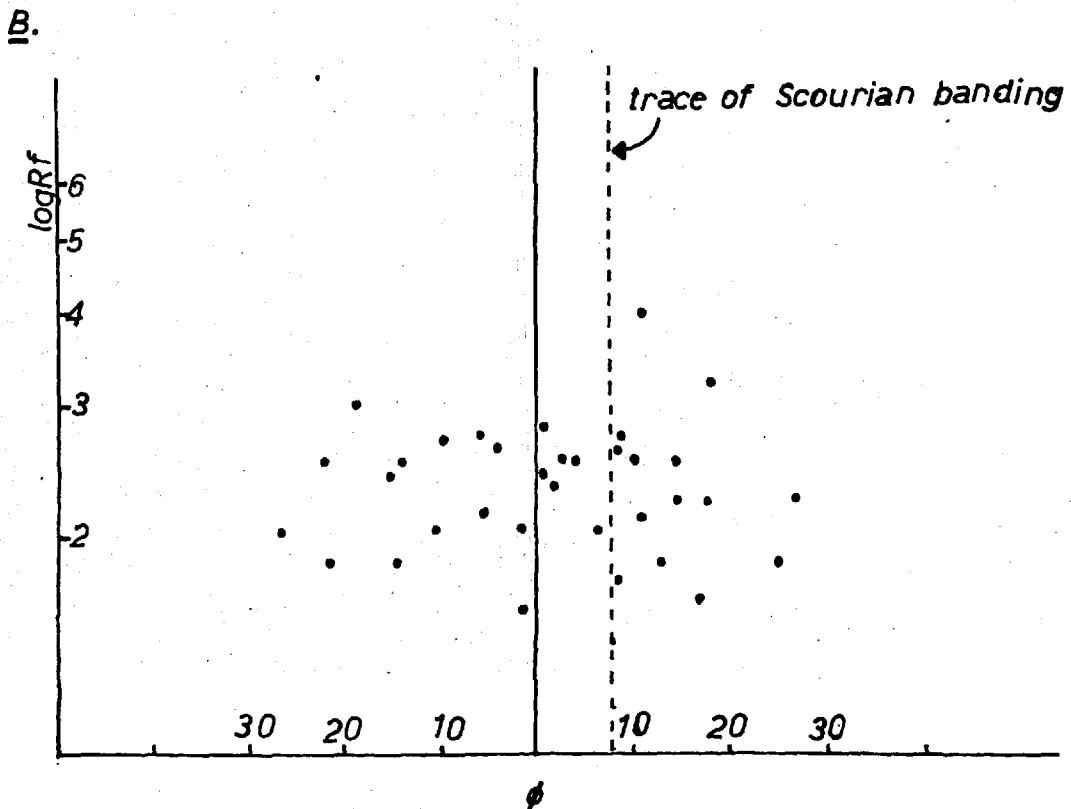
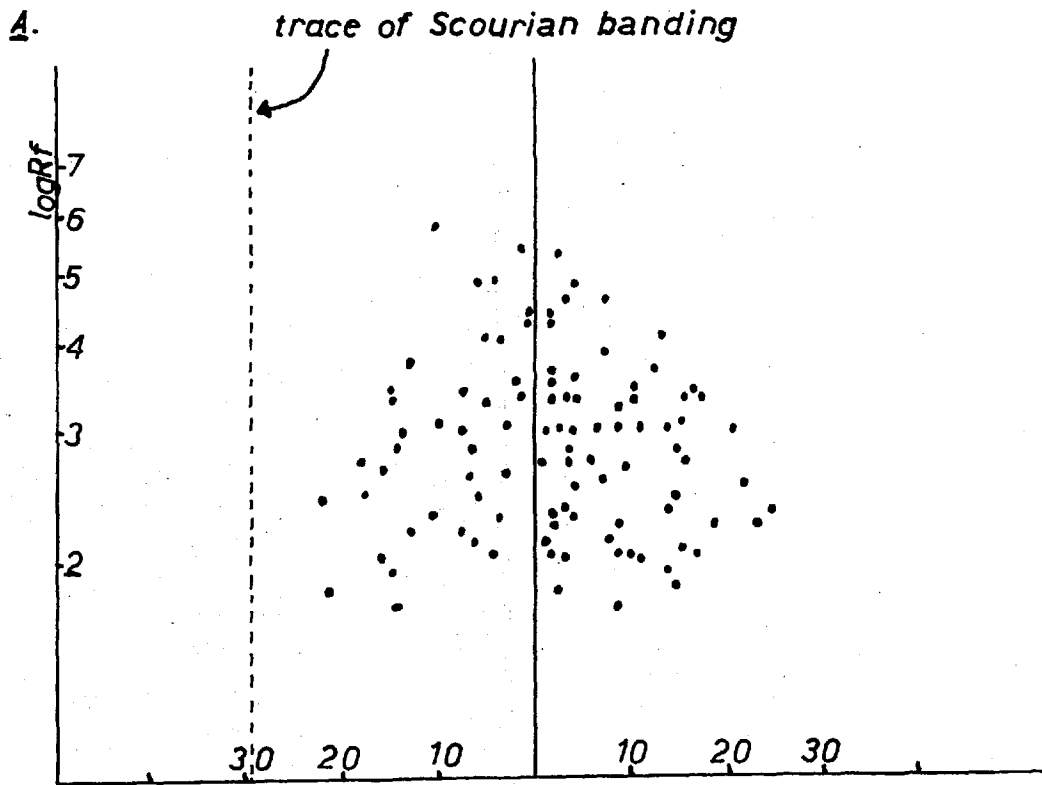
Fabric developed in shear zone crossing the GK gneiss



amphibolite sample 22 from limb of minor fold

A. XZ plane normal to fold axis

B. XY plane parallel to fold axial surface



aggregates pseudomorph the earlier granulite facies pyroxenes in many rocks in this area. In the fold sampled the hornblende aggregates are ellipsoidal in shape. Orthogonal faces were cut in the sample using the axial surface and fold axis as reference directions. In each section prepared the hornblende aggregates were then treated as ellipses and the orientation and dimensions of the major and minor axes measured. It was only possible to do this analysis for two faces because on the third face the elliptical shape of the aggregates was not defined clearly enough to permit accurate measurement. The results are plotted in fig.3.3 which shows that on this R_f/ϕ plot the points form a symmetrical cluster (see Dunnet 1969). There is no obvious reason why the results presented here should not be interpreted in the same way as described by Dunnet (1969). Thus the strain ratio of the measured aggregates is approximately $1+e_1/1+e_2/1+e_3 = 3/1.5/1$, and the initial shape factor is approximately $2/1.5/1$.

The principle of these examples is that if recrystallisation occurs in a deforming rock then it will contribute to the strain in that rock. It is not implied that recrystallisation necessitates strain, for example post-tectonic annealing will not in general alter the shape of the rock.

Recrystallisation is dependent mainly on temperature and if a stable phase is recrystallising the rate controlling factor may be temperature. This relationship is sufficient reason for proposing recrystallisation as a mechanism of strain at high temperatures. The strain patterns produced in this way will depend on the precise mechanism by which recrystallisation takes place, a subject still in dispute.

3.3 Redistribution of material.

In general the rocks studied here suggest that migration of the mineral phases has taken place only over very small distances, with the result that processes of transposition are unimportant in the evolution of these rocks.. The following evidence is used to infer the absence of large scale mineral

migration:

a) The pyroxene in the granulite is often merely replaced by hornblende during amphibolisation, and the granulitic texture is mimetically retained. The banding is seldom destroyed during this amphibolisation.

b) In the process of segregation described in section 2.2B the minimum distance of migration of hornblende necessary to produce the observed fabric is of the order of 1-2cm.

c) Within the shear zones no new segregation banding is produced. The discontinuous banding developed in some shear zones is the result of flattening and elongation of original single crystals. In shear zones cutting sub-ophitic dolerite the strained sub-ophitic texture can still be seen. This suggests an absence of processes involving mineral migration in these rocks.

d) The Scourian granulites are characterised by an absence of major potash bearing phases. During the Laxfordian such phases appear in the shear zones and other Laxfordian assemblages. Scourian granitic rocks do contain potash feldspar, but during Laxfordian metamorphism and deformation of these rocks this potash feldspar does not migrate out of its host rock to react with adjacent assemblages. Thus though potassium in some form is very mobile during the Laxfordian metasomatic event, potash feldspar is relatively immobile.

e) Quartz veining is in general absent from the area studied.

These observations suggest that processes of solution, migration and redeposition, the results of which are frequently seen in greenschist facies rocks, were almost inoperative in the rocks studied here.

3.4 Some comments on the deformation of polycrystalline aggregates.

It is possible to recognise a number of mechanisms by which crystalline aggregates deform when subjected to a deviatoric stress system. It is much more difficult to derive constitutive equations for these mechanisms. In this discussion reference

will be made to the following types of deformation mechanism:

a) cataclasis, including gliding, glide twinning and grain boundary slip;

b) pressure solution, including grain boundary diffusion and diffusion through a fluid phase;

c) lattice diffusion - this can be considered in two ways, first, movement of a unit of given composition through a crystal of the same composition (ie. diffusion of vacancies in the opposite direction) leading to a shape change, and second, differential migration of certain ions, etc. through a lattice leading to changes in composition;

d) recrystallisation.

In general, mechanisms of deformation may be divided into a stress induced group and a temperature induced group. Each mechanism has a particular activation level, and over a range of geological conditions, different mechanisms will become dominant under different conditions.

Gliding and cataclastic deformation are stress induced, and evidence that this mode of deformation has occurred in a rock is most readily obtained from rocks deformed under dry or low grade metamorphic conditions. The brittle strength of a crystal decreases with increasing temperature, and it might be expected that deformation resulting from cataclasis would increase with temperature. However, under conditions of rising temperature, new temperature activated mechanisms begin to operate, and it is possible that the lower deviatoric stresses that these high temperature systems can sustain (see Heard 1963, Price 1971) preclude deformation by gliding. Unfortunately there is no way of telling if cataclasis has been important during the deformation of high grade metamorphic rocks because of the recrystallisation that most phases have undergone.

A little more is known about diffusion mechanisms, both from observations made on natural rocks and from experimental and theoretical work. The results of processes termed pressure solution are most frequently seen in low grade metamorphic rocks. It is not clear whether pressure solution occurs by

by diffusion through grain boundary layers of a crystal or through an intergranular fluid phase. Because of the small dimensions of intergranular spaces that are likely to have existed in metamorphic rocks, it seems probable that surface layers of the crystal must have been involved in diffusion. Experimental work in metallurgy has shown that activation energies for grain boundary diffusion are lower than for lattice diffusion (see LeClaire 1951), and that this difference is larger at lower temperatures.

Green (1970) has considered crystals under deviatoric stress and presents equations defining the chemical potential of the crystal throughout and at the boundary of these crystals. Stress-strain rate laws are then derived for grain boundary and lattice diffusion (see Green 1970); neither of these laws predicts a steady state creep. One possible reason why the results of pressure solution are more frequently encountered in unmetamorphosed or low grade metamorphic rocks may be the higher stress differences that probably existed in these rocks during deformation, since then larger gradients in chemical potential of each phase will also have existed.

Self diffusion of vacancies in metals is known to become an important mechanism of deformation above temperatures of half the melting point, and the stress-strain rate law for this process can be calculated theoretically (see McLean 1965). At high temperatures and low differential stresses vacancy diffusion must be considered as a feasible mechanism of deformation in crystals.

The term recrystallisation is used to describe processes whereby dislocations, etc. are cleared from crystals and lattice blocks become reoriented. Recrystallisation may involve nucleation and growth, or simply growth of strain free blocks, and the driving forces involved are lattice strain energy and grain boundary energy. Lattice diffusion is enhanced by the presence of many dislocations that result from deformation of a crystal, and fast rates of lattice diffusion may occur if recrystallisation takes place during deformation. The time required for a given amount of recrystallisation to occur

decreases logarithmically with increasing temperature.

Formation of new and stable grain nuclei requires that any nucleus has a low interfacial energy (ie. it has a coherent boundary with its host), whereas further growth of the nucleus is promoted by high interfacial energy (ie. nucleus-host boundaries are incoherent). If recrystallisation occurs by growth of strain free blocks then the nucleation stage is no longer necessary, and incoherent boundaries may migrate relatively quickly. The data presented in fig.3.2 shows a) that the boundaries of new grains are irrational and incoherent, and b) that there is no relation between the lattices of the host crystal and the new grains. This suggests that recrystallisation occurred by growth of strain free blocks. If nucleation had occurred then preferred orientations should be apparent in fig.3.2 because only favoured nuclei would have grown.

The textural evidence presented in the previous section suggests that grain boundary diffusion of mineral phases did not occur to any extent in the rocks studied. One very convincing example is provided in rocks consisting of pyroxene, hornblende, plagioclase and quartz. Along initially pyroxene-plagioclase boundaries these two minerals have reacted to form hornblende (this is described in chapter 5). However, along quartz-pyroxene boundaries no hornblende has been formed and it must be supposed that the plagioclase could not diffuse along the quartz-pyroxene boundary and react to form hornblende.

It is concluded from this discussion that in the rocks studied the mechanism of deformation involved temperature activated processes of recrystallisation with simultaneous short range lattice diffusion.

CHAPTER 4.

AN INTRODUCTION TO THE MINERALOGICAL STUDIES.

Two generalisations relative to the work to be discussed now can be made from the field studies. They are, firstly, that in a terrain that consisted originally of granulite facies rocks, two important periods of subsequent tectonic deformation (Scourian and Laxfordian) can be distinguished, and secondly, that each set of structures has associated with it distinctive mineral fabrics and mineral assemblages.

4.1 Sampling.

For these reasons, sampling was carried out within the framework of tectonic structure, and the chapters that follow present the results of laboratory investigations of rocks collected from a variety of structural environments. The samples may be grouped as follows:

a) Partly altered Scourian granulites, with pyroxene still present. These are used to substantiate a link between the altered and unaltered granulites. The petrochemistry of the latter has been studied in detail by Muecke (1969).

b) Samples from the steep and flat limbs of the asymmetric Scourian fold structures (see section 2.2). These rocks are dominantly hornblende bearing; sometimes the steep limbs of these folds contain significant quantities of biotite which, for reasons that will be discussed later, is attributed to later Laxfordian introduction of potash. These biotite bearing rocks can be distinguished from Laxfordian shear zones by differences in mineral fabric.

c) Samples from Laxfordian shear zones, these containing mainly biotised assemblages. With each sample from a shear zone, another sample from the immediately adjacent rock was collected for comparative purposes.

d) Samples from the Tarbet Steep Belt. This is another structurally distinct unit (see section 2.4) and consists of strongly recrystallised hornblende and biotite bearing assemblages.

e) Samples from hornblende pods. These result from the amphibolisation of mafic Scourian granulite facies pods. Some of them are also involved in the later Laxfordian shear zones. These pods were collected with very specific reference to tectonic structure, not only for petrologic study, but also for a programme of potassium-argon radioisotope dating. It is hoped that this latter work, the results of which will be published elsewhere, will help solve some of the geochronological problems in this structurally complex region.

Table 4.1 presents a list of samples referred to in the text; the sampling locality is given as an eight figure National Grid reference, the prefix NC being assumed in each case. The structural map presented with this thesis (cover pocket) shows the national grid one kilometre squares in the area mapped, enabling sample localities to be found rapidly. Table 4.1 also indicates briefly the fabric elements visible in hand specimen for each sample.

4.2 Petrography.

It is possible to define events of amphibolisation and biotisation because the parent granulite facies rocks as such are mineralogically dry and depleted in potassium (Muecke 1969, Holland and Lambert 1971). Amphibolised and biotised assemblages are readily observed to be derived from the granulites since remnants of pyroxene and pseudomorphs after pyroxene are often preserved, and this forms the basis of the notion of regeneration. The prime object of this study was to define the various mineral transformations, and to evaluate their mineralogical relevance and relation to the tectonic structure in which they occur.

TABLE 4.1.

LOCALITIES AND BRIEF HAND SPECIMEN DESCRIPTIONS OF SAMPLES MENTIONED IN TEXT.

No.	Grid ref.	Chief characteristics.
2.2	14994618	undeformed dolerite with sub-ophitic texture, from a Scourie dyke, adjacent to shear zone
2.7	14994618	sample from zone of simple shear, LS fabric
12	15034666	amphibolised granulite from large Scourian steep limb, mineral banding destroyed
17	15154683	amphibolised granulite from large Scourian steep limb, mineral banding destroyed
18	14934588	Scourian granite sheet adjacent to shear zone, coarse quartz and felspar schlieren
19	14934588	Scourian granite sheet in shear zone, L2S2
22	14814595	amphibolised granulite from Scourian minor fold, axial surface hornblende fabric
24	15244622	shear zone, weak L2S2
25	15244622	adjacent to 24, granular texture
26	15174623	amphibolised granulite, shows developing hornblende segregations
28	15174623	amphibolised granulite, coarse linear segregations of hornblende and felspar, L1
32	15234622	pyroxene granulite
38	15234624	adjacent to small shear, granular hornblende texture
41	15234624	adjacent to small shear, coarse hornblende texture
42	15834816	amphibolised granulite, well defined banding, L1
43	15834816	amphibolised granulite, modified banding, L1
44	15834821	amphibolised granulite, modified banding, L1
45	16004824	amphibolised granulite, modified banding near vertical, L1
46	16024825	steep foliation, not overturned, banding thinned, weak L1, L2
47	16074829	Scourian granite sheet in overturned foliation, strong L2S2

table 4.1 continued.

52	14994615	GK gneiss from shear zone, strong L2S2
53	14994615	GK gneiss from shear zone, strong L2S2
55	14994615	GK gneiss from shear zone, strong L2S2
56	14994612	GK gneiss, granulite facies assemblage near to above shear zone
100	16094838	from Tarbet Steep Belt, L2S2
102	16154840	from Tarbet Steep Belt, L2S2
103	16234839	from Tarbet Steep Belt, L2S2
123	15834653	shear zone, strong L2, thin banding visible
124	15834653	steep limb of Scourian fold, no L2, granular texture
126	15834653	shear zone, L2S2
127	15834653	shear zone, L2S2
140	15814732	shear zone, L2S2
141	15814732	adjacent to 140, granular and banded, L1
142	15794741	shear zone, thinned banding, L2
143	15794741	adjacent to 142, granular and banded, L1
147	15274723	steep limb of Scourian fold, thinned banding
148	15274723	steep limb of Scourian fold, granular texture
149	15274723	banded pyroxene granulite adjacent to 147/148
152	15204706	shear zone, L2S2
153	15204706	adjacent to 152, granular texture, L1
162	14374582	hornblende pod in amphibolised granulite
164	14734596	shear zone, L2S2
167	14794594	hornblende pod in amphibolised granulite
168	14784595	amphibolised granulite, pseudomorphed pyroxene aggregates
171	14724610	undeformed GK gneiss
173	14724610	undeformed GK gneiss
174	14724610	undeformed GK gneiss
175	14924609	shear zone, L2S2
177	14684618	hornblende pod in Scourian
178	14724612	hornblende pod in amphibolised granulite
179	14724612	hornblende pod in shear zone, strong L2
180	14724612	shear zone, thin banding visible, L2S2
184	14784644	margin of hornblende pod in shear zone
185	14784644	centre of 184

table 4.1 continued.

187	14874661	pyroxene bearing mafic pod
188	15014655	hornblende pod from steep limb of Scourian fold
190	15004603	GK gneiss in shear zone, L2S2
191	15074600	shear zone, thinned banding visible, L2S2
192	15074600	hornblende pod in shear zone, L2
194	15194595	GK gneiss in steep limb of Scourian fold, L1
195	15134585	shear zone, banding still visible, L2
199	16074829	from Tarbet Steep Belt, L2S2
200	15884800	amphibolised granulite adjacent to 201
201	15884800	shear zone, L2S2
202	16264845	hornblende pod in Tarbet Steep Belt, L2
220	15074625	shear zone, L2S2

TABLE 4.2.

SUMMARY OF MINERAL ASSEMBLAGES FOUND IN DIFFERENT PARTS OF THE SCOURIE-LAXFORD AREA.

1). Principal assemblages of the Scourian granulites.

ultramafic division:

cpx-opx-parg-op+plag-(ol-ap-sp-biot-hbl)

mafic division:

cpx-opx+parg+gar-plag-op-(qtz-ap-biot-hbl)

intermediate division:

plag+qtz-cpx+opx-(scap-ap-biot-hbl-cal)

acid division:

plag-qtz-opx+cpx-op-(gar-ap-cal-biot-hbl)

gar-qtz-plag-(opx)

qtz-plag-Kfsp-(cal-op)

2). Principal assemblages found in amphibolised Scourian rocks.

hbl-(op-plag-biot)

hbl-plag-op-(qtz-ap-epid-zir)

hbl-plag-qtz-op-(ap-biot)

3). Assemblages recorded from Laxfordian shear zones.

biot-hbl-plag-qtz-op-(ap-rut)

hbl-biot-cal-epid

hbl-epid-plag-qtz-haem-il

gar-biot-plag-qtz-ky-op-(str-sp-cor-musc-haem)

table 4.2 continued.

- 4). Assemblages found in the Claisfearn and Foindle zones of Sutton and Watson (1951).

plag-qtz-hbl-+biot-(op)
 hbl-epid-plag-qtz-(chl-haem)
 gar-biot-plag-qtz-sulphides
 plag-Kfsp-qtz-musc-(biot)
 plag-Kfsp-hbl-cpx-sph
 plag-pyx-gar₊ol
 plag-qtz-(biot)
 gar-qtz
 anth-talc-trem

- 5). Assemblages found at Laxford Bridge.

basic gneiss:

hbl-epid-plag₊gar-op

grey gneiss:

biot-hbl-epid-plag-qtz-op₊Kfsp

granite sheets:

plag-Kfsp-qtz-(biot-hbl-cpx-op)

Fortunately, it is unnecessary to give detailed petrographic descriptions in this thesis since the excellent works by Peach et al (1907, chapters 4,9), Sutton and Watson (1951) and Muecke (1969) deal with this topic. However table 4.2 gives for reference a list of the main assemblages found in the areas covered by this thesis.

The methods used in this investigation started with microscope inspection of thin sections, and techniques included modal analysis and use of the universal stage microscope. The statistics of modal analysis have been discussed by Chayes and Fairburn (1951), Chayes (1954) and Kalsbeek (1970), and it was in the light of these discussions that the counting interval used was the minimum possible (0.05mm) and the number of counts per assemblage was maximised (more than 3000 counts per analysis was aimed at). It was still not possible to make counts of the fine inclusions present in some of the phases. The result of the method used is that the assemblages are dealt with on the scale of a thin section, except in special cases where larger or smaller scales are used. The number of counts made in each analysis is indicated in table presenting this analysis. Plagioclase compositions, determined optically using a universal stage microscope, are also given in the tables of modal analyses.

4.3 Analytical techniques.

Microscope work was followed by a programme of qualitative and quantitative chemical analyses of selected minerals and mineral assemblages, using the Geoscan electron microprobe. This instrument provides a powerful and rapid method of analysing minerals. The construction and mode of operation of the instrument are described in papers by Long (1967) and Philibert (1970). The great advantage in using a microprobe is that single grains or parts of grains can be analysed separately, a necessity in rocks where disequilibrium assemblages may be present. In the present work, a number of grains of the same phase in each assemblage have often been analysed.

The techniques generally used in obtaining quantitative analyses are documented clearly by Philibert (1970). A number of the features of the analytical technique used in this work will be briefly mentioned.

Machine drift may be one of the largest sources of error during analysis, and it is difficult to correct adequately for this error. Thus before an analysis set was commenced, tests for drift in counting rates on standards were carried out for periods of up to one hour. If a drift of greater than 1% was encountered during an analysis, then the data were rejected. Brief tests for compositional homogeneity of the grains selected for analysis were also carried out. It is possible, using back-scattered electron and X-ray pictures, to ensure that the surfaces of the grains being analysed are indeed flat, and this is necessary if accurate results are to be obtained. Analysis was then restricted to small homogeneous and flat areas of the grains selected. Analysis was not restricted to one spot (this being only a few microns in diameter) on each grain because hydroxyl bearing silicates tend to decompose under electron bombardment. This procedure also reduces the errors that might result from the presence of minute and unseen impurities in the specimen. Counting was carried out over intervals of ten seconds, and the position of the beam was moved after two such intervals. The total number of counts accumulated for any one element in a grain is determined by many factors. Statistically, counting errors are reduced to below 2% of the measured concentration if the total number of counts collected on each grain is of the order of 10^5 . Corrections for counting dead time require that an upper limit of 10^4 counts per second be placed on the count rate. These factors must be taken into account in deciding the conditions (ie. specimen current, accelerating voltage) under which the analysis is to be carried out.

During an analysis the following sequence was used. With the Geoscan two elements can be analysed simultaneously. First, counts on the relevant standards (oxide and synthetic standards were used) were accumulated, then counts for the element pair

were accumulated for each grain or parts of grains selected. The standards were returned to and further counts accumulated, and this was followed by counting of the backgrounds of both standards and specimens. Two sets of standard counts permits linear interpolations to be made, and small amounts of drift can be corrected. This procedure is written into the correction programme. The time taken for analysis of each element pair was about half an hour. An analysis set is made up of any number of element pair analyses. It is not necessary to carry out each element pair analysis under the same operating conditions, and this permits a flexibility necessary if the statistical counting errors mentioned above are to be reduced. Care was taken to analyse the same parts of grains for all the element pairs within an analysis set.

The raw data were submitted on paper tape and run with the BM-IC-NPL correction programme (Mason et al 1969) using the IC computer facilities (CDC.6600). A turnaround time of about 24 hours between carrying out the analyses and obtaining corrected results permits easy corrections to be made to any operational errors that may have occurred. The computer programme used makes corrections for dead time, mass number, atomic absorption and fluorescence (see Sweatman and Long 1969).

It is of considerable value if there is a method of checking independently the accuracy of the analyses. For this reason selected homogeneous minerals containing all the elements being analysed were included in the standard blocks. These minerals have been analysed by wet chemical methods and by various persons using the Geoscan. An analysis of one of these minerals was included in each analysis set to act as internal standard.

A large amount of information can also be gained by using the Geoscan in a qualitative way. The two methods used are called wavelength scans and Geoscan traverses. The first method involves scanning of the spectrometer through large angular ranges, while the electron beam is focussed on one mineral grain. In this way peaks corresponding to all the elements present in the mineral are obtained on the output chart. This method is valuable in

determining the various minor elements present in different phases. The second method is essentially the reverse of this, that is, the spectrometer is set at an angle corresponding to a particular peak of the element required and the specimen is moved at a controlled speed and in a controlled direction under the electron beam. This method provides a means of detecting zoning in crystals, etc.

I would at this point like to thank Mr Paul Suddaby and Mr Malcolm Frost for the large amount of time they spent instructing and helping me in the use of the Geoscan and the data processing methods.

4.4 Thermodynamics.

In the following chapters mineral reactions have been evaluated, where possible, thermodynamically using known data. The source books for these data are Robie and Waldbaum (1968) and Burnham et al (1969), these being the most up to date catalogues available. Standard symbols are used throughout, as follows:

P	pressure, bars
V	molar volume, cm^3
T	temperature, $^{\circ}\text{C}$
S	entropy, cal/deg/mol
G	free energy of formation, cal/mol
K	equilibrium constant of formation
f	fugacity, bars
μ	chemical potential

4.5 Tables and abbreviations.

For convenience the contents of the tables presented in the following chapters are summarised below.

Table 5.1 Modal analyses of Scourian and amphibolised Scourian rocks.

- 5.2 Mineral analyses from amphibolised Scourian rocks.
- 5.3 Modal analyses of the hornblende pods.
- 5.4 Mineral analyses from the hornblende pods.

- 6.1 Modal analyses of samples from shear zones and adjacent rocks.
- 6.2 Modal analyses of the GK gneiss.
- 6.3 Mineral analyses from biotised gneiss.
- 6.4 Mineral analyses from the GK gneiss.
- 6.5 Barium and manganese analyses of selected minerals.
- 6.6 Whole rock analyses and CIPW norms of the GK gneiss.
- 7.1 Mineral analyses from oxidised assemblages.
- 7.2 Modal analyses of oxidised assemblages.
- 7.3 Modal analyses of samples from the Tarbet Steep Belt.
- 7.4 Mineral analyses from the Tarbet Steep Belt.
- 8.1 Statistical data for the hornblende analyses.

The following abbreviations of mineral names have been used:

Plag(P)	plagioclase
Pap	plagioclase antiperthite
Ab	albite
An	anorthite
Kfsp	potassium feldspar
Qtz	quartz
Gar(G)	garnet
Hbl(H)	hornblende
Parg	pargasite hornblende
Trem(T)	tremolite
Ged	gedrite
Anth	anthophyllite
Ol	olivine
Cpx	clinopyroxene
Opx	orthopyroxene
Hyp	hypersthene
Biot(B)	biotite
Musc(M)	muscovite
Ky	kyanite
Str(S)	staurolite
Sp	spinel
Cor	corundum

Epid(E)	epidote
Chl(C)	chlorite
Op	opaque
Mag	magnetite
Il	ilmenite
Haem	haematite
Ap	apatite
Rut	rutile
Sph	sphene
Zir	zircon
Cal	calcite

subscript 'c' = centre of analysed grain

subscript 'e' = edge of analysed grain

subscript 's' = sample from shear zone

subscript 'a' = sample adjacent to shear zone

CHAPTER 5.

REACTIONS INVOLVING PYROXENE.

The main concern of this chapter will be a presentation of data relating to the alteration of pyroxene in the early Scourian granulites to hornblende. This is the first major alteration process that affects the granulites in the area studied here and it has already been indicated (section 2.2) that initial development of hornblende is associated with the formation of Scourian structures.

In the first sections discussion is confined to the commonest rock types seen in the field. They are followed by a short discussion of a rock type that is insignificant in terms of its total volume percent in the whole complex, but that occurs persistantly enough to warrant attention. These rocks are referred to as hornblende pods.

5.1 Petrography.

A recent study of the Scourian granulites (Muecke 1969) showed that the commonest rock type in these granulites consists of ⁿcliopyroxene, orthopyroxene, a plagioclase of composition An₃₅ and quartz. In particular primary hornblende is absent from these rocks and potash feldspar occurs only in small amounts within the plagioclase (ie. as antiperthite). The change from pyroxene- to hornblende - bearing rocks is thus an important event in that it indicates a change from anhydrous to hydrous mineral assemblages.

Following from the field observations presented in section 2.2 it might be expected that hornblendes formed at this time would vary texturally from grains or aggregates

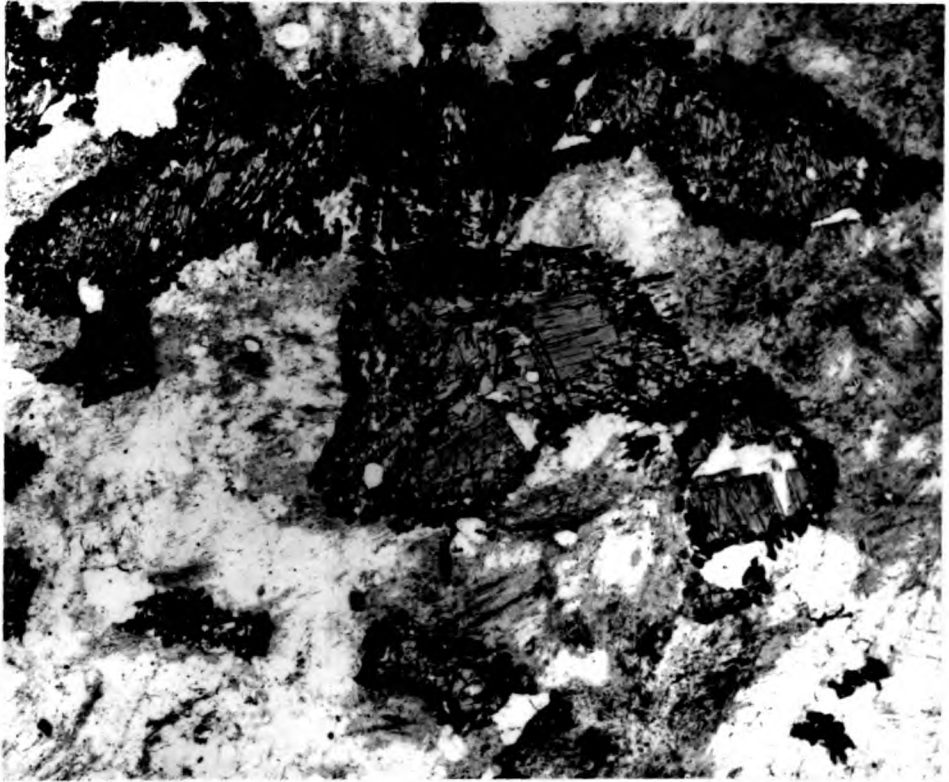
merely pseudomorphing pyroxene to completely recrystallised hornblende. This variation is seen in thin section and briefly the following textural sequence is seen with progressive recrystallisation of the assemblages:

a) Pyroxene is pseudomorphed from the margins inward (generally it is not possible to tell whether a clino- or an orthopyroxene is pseudomorphed; if relicts of pyroxene are found these are invariably a clinopyroxene) by an aggregate of small hornblende crystals. Intergrown with the hornblendes are small quartz grains; these are always concentrated at the centres and absent from the edges of the pseudomorphed pyroxene (see plate 5.1). The plagioclase is not always completely recrystallised and the antiperthite texture may still be visible. The large quartz crystals in the assemblages recrystallise, and the fine inclusions seen in these crystals in the granulite facies assemblages are no longer present.

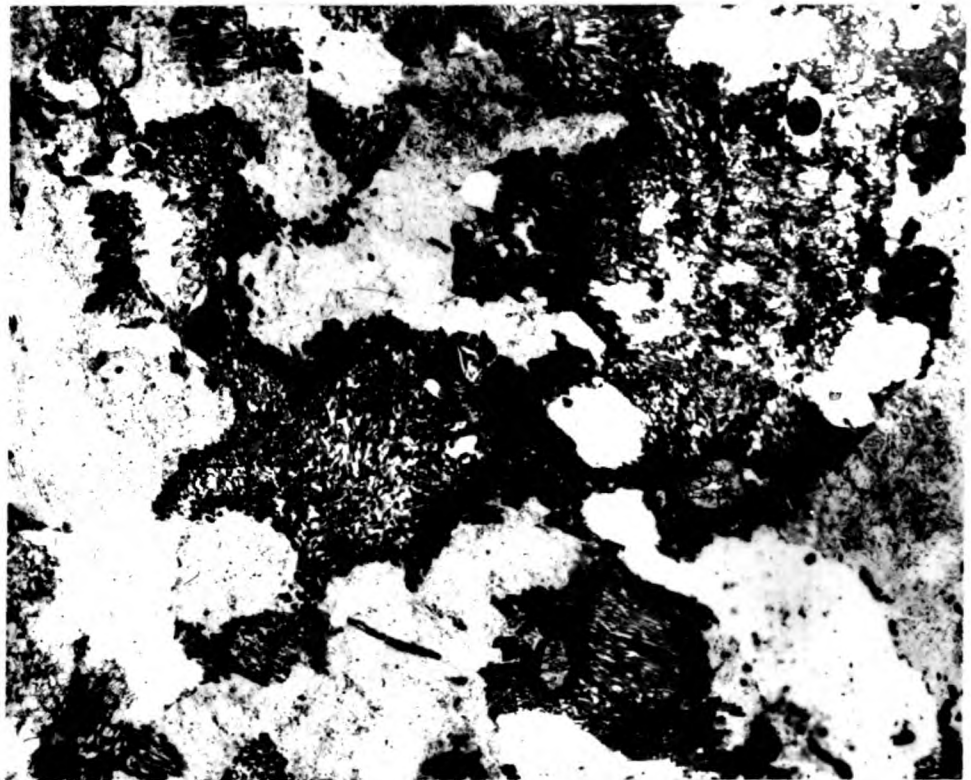
b) The hornblende begins to recrystallise into larger grains, though each pseudomorphed pyroxene is still recognisable as an aggregate of hornblendes. The intergrown quartz grains begin to coalesce but retain their uneven distribution. Any small inclusions of opaque material that may have been present in the hornblendes have either disappeared or coalesced into larger grains that generally grow exterior to the hornblende. This phase is now recognised as ilmenite.

c) The original aggregate of hornblendes recrystallise into one large crystal. The quartz grains have coalesced further but are still contained near the centre of the hornblende. The plagioclase has recrystallised into polygonal grain aggregates without antiperthite inclusions.

It is emphasised that quartz is intergrown with the hornblende and that this quartz remains within the hornblende crystal despite the continued recrystallisation of the latter. In the early aggregates of hornblende the marginal hornblende is always darker in colour than that at the centre of the aggregates; this is in part a result of the translucent nature



A) Partially amphibolised granulite: clinopyroxene rimmed by hornblende



B) Amphibolised granulite: hornblende corona with quartz free rim

TABLE 5.1.

SCOURIAN ASSEMBLAGES: MODAL ANALYSES.

	32	41	38	168	26	28	147	148	149	12	17
Plag	48.0	40.1	62.7	34.1	28.6	57.3	55.5	45.2		58.0	55.4
Pap	14.0	10.2							65.0		
Qtz	0.3	7.0	13.1		3.7	4.2	17.0	14.5	0.4	16.8	22.2
Cpx	21.4	6.8	4.7	2.4	1.8				9.4		
Hbl	13.8	34.3	12.4	65.5	61.1	31.2	18.7	36.4	19.5	16.2	10.0
Trem					tr	tr					
Ged						tr					
Biot		1.3	4.9	2.2			5.9	1.3	2.9	7.3	11.6
Epid					1.5	1.6					
Op	2.2	0.2	1.4	4.8	0.5	1.4	2.6	2.2	2.5	1.3	0.8
Rut		1.2			0.3	0.9				tr	
Ap	0.3	0.5	0.8		0.4	1.4	0.3	0.4	0.3	0.4	tr
Cal					1.5	tr					
Zir					0.6	2.1					

counts 4900 10100 10900 2900 6100 8900 3700 3500 4000 4600 3200

An% 37 36 38 33 30 31 34 37 30 30

TABLE 5.2

AMPHIBOLISED SCOURIAN ROCKS: MINERAL ANALYSES.

	SiO ₂	Al ₂ O ₃	TiO ₂	FeO	MgO	CaO	K ₂ O	total	mg
28H	42.8	11.0	0.7	16.0	10.2	11.7	0.7	93.1	39.0
168H	42.4	12.7	0.9	18.2	9.2	11.8	1.2	96.4	33.5
168H	41.0	13.4	1.2	18.8	8.6	11.5	1.4	95.9	31.3
168B	35.9	15.4	3.4	18.5	11.1	0.0	9.6	93.9	37.4
12H	41.9	12.0	1.3	16.1	10.5	11.7	1.2	94.7	39.5
12B	36.1	15.3	4.6	17.3	12.4	0.0	9.5	95.2	41.9
12B	35.7	15.3	4.7	16.9	12.4	0.0	9.6	94.6	42.2
12H	42.1	12.1	1.2	16.0	10.5	11.6	1.2	94.7	39.6
17H	42.6	12.4	1.1	17.2	10.0	11.6	0.5	95.4	36.8
17H	42.6	12.1	0.9	17.3	10.0	11.6	0.6	95.1	36.6
17H	42.6	12.3	0.9	17.3	10.3	11.5	0.6	95.5	37.3
17B	35.8	16.1	2.6	17.2	13.5	0.1	8.3	93.6	43.9
17B	36.4	16.4	2.4	16.7	13.2	0.0	8.7	93.8	44.2
200H	39.8	12.6	1.0	15.9	10.5	11.6	0.6	92.0	39.7
143H	43.8	11.6	0.9	18.0	9.8	11.2	0.9	96.2	35.2
143H	44.4	11.8	0.6	17.3	9.9	11.1	0.6	95.7	36.5
143H	44.9	11.3	0.4	15.9	11.2	10.7	0.6	95.0	41.3

of the quartz inclusions, but also reflects a compositional difference (see later).

Some modal analyses of Scourian and amphibolised Scourian rocks are presented in table 5.1, while table 5.2 lists the mineral analyses carried out on these rocks.

It will be noticed that the minerals epidote, biotite, calcite and zircon are recorded in table 5.1. Their occurrence will be discussed in section 5.4.

5.2 The reaction pyroxene \longrightarrow hornblende.

There are a number of possible ways to describe the reaction equilibria between pyroxene and hornblende. The simplest involves idealisation of the complex chemical formulae of pyroxene and hornblende; a balanced reaction may then be written using these simple formulae. This type of description is useful when it is wished to evaluate the stability field of pyroxene and the likely interaction of pyroxene with other phases.

A very different approach was adopted by Evans (1963). He analysed the mineral phases in a Scourian granulite and in an amphibolised granulite, and then recalculated the granulite facies assemblage in terms of minerals found in the amphibolised rock. Thus a reaction was evaluated in terms of mineral compositions actually encountered. This type of description is very specific and it is difficult to compare the reactions derived in this way with theoretically or experimentally evaluated reactions.

The Scourian granulites and their amphibolised equivalents can be described in terms of the phases clinopyroxene, orthopyroxene, plagioclase, hornblende, iron oxide and quartz. A univariant reaction between these phases is defined if the composition of these six phases is known in terms of five components, and other components that may be present are considered, from the point of view of the reaction, as available and mobile. This follows from consideration of the phase rule. If it is assumed that the composition of the plagioclase, iron

oxide and quartz is fixed, then knowledge of the composition of the other phases enables reaction coefficients to be calculated. The calculation involves solution of a number of linear equations and this is most easily done by using determinants. It will be realised that a particular solution (ie. a set of reaction coefficients) is only applicable to one set of mineral compositions. A family of reactions would be needed to describe a suite of rocks such as the Scourian granulites and amphibolised granulites.

One solution is given below as an example. The pyroxene compositions have been taken from Muecke (1969, sample 17, tables 4.1,4.2) and the hornblende composition is taken from table 5.2 presented here. The pyroxene formulae are based on six oxygens, while the hornblende formula is based on twenty four oxygens. The plagioclase composition used is An50 and iron oxide is calculated as magnetite.

The number of atoms in each mineral is as follows:

	Si	Al	Ti	Fe	Mg	Ca	K	Na
CP	1.86	0.26		0.2	0.72	0.9		
OP	1.9	0.21		0.54	1.23	0.07		
PL	2.5	1.5				0.5		0.5
HB	6.1	2.2	0.1	2.1	2.3	1.9	0.2	0.3

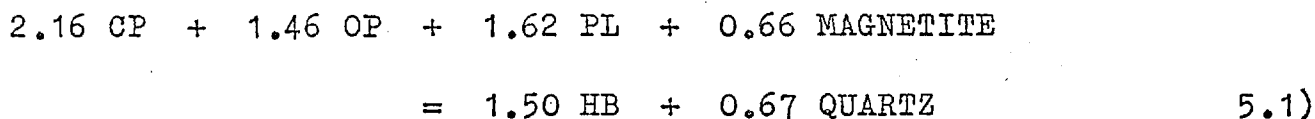
With four phases, only three components need be defined, and the following array is written:

	CP	OP	HB	PL
AL	0.26	0.21	2.2	1.5
MG	0.72	1.23	2.3	0.0
CA	0.9	0.07	1.9	0.5

Then the following sequence of determinants is evaluated

$$\begin{array}{r}
 \left| \begin{array}{ccc} 0.21 & 2.2 & 1.5 \\ 1.23 & 2.3 & 0.0 \\ 0.07 & 1.9 & 0.5 \end{array} \right|_{\text{CP}} + \left| \begin{array}{ccc} 2.2 & 1.5 & 0.26 \\ 2.3 & 0.0 & 0.72 \\ 1.9 & 0.5 & 0.9 \end{array} \right|_{\text{OP}} + \left| \begin{array}{ccc} 1.5 & 0.26 & 0.21 \\ 0.0 & 0.72 & 1.23 \\ 0.5 & 0.9 & 0.07 \end{array} \right|_{\text{HB}} \\
 \\
 + \left| \begin{array}{ccc} 0.26 & 0.21 & 2.2 \\ 0.72 & 1.23 & 2.3 \\ 0.9 & 0.07 & 1.9 \end{array} \right|_{\text{PL}} = 0
 \end{array}$$

The reaction coefficients are obtained directly and the reaction is written as:



where the mineral names substitute for the formulae given above. Thus in a rock containing 40 modal percent hornblende, 1.5% quartz will have been produced and 2.86% magnetite consumed. This is consistent with the observed petrography.

Evaluation of many such reactions for minerals of different composition would have little immediate meaning. However, it will be appreciated from table 5.2 that the hornblendes that are formed during amphibolisation of the granulites are remarkably uniform in composition. This suggests that there is some constraint within the general system of pyroxene-hornblende that results in a convergence of hornblende compositions.

Within a system consisting of two pyroxenes and plagioclase, the Fe, Mg for the hornblendes will be derived chiefly from the pyroxenes. Ca, Al will come from the plagioclase, and in addition Ca will be supplied by the clinopyroxene. Thus there must be a complex balance in the reaction between the three phases. While there is a range in the composition of the granulite facies pyroxenes (Muecke 1969), a variable of considerable importance is the ratio modal clinopyroxene/modal orthopyroxene. Modal analyses presented by Muecke (1969) show no obvious relation between this ratio and the rock type, except that the mafic granulites generally contain greater amounts of clinopyroxene relative to orthopyroxene.

It is also found (compare table 5.2 with Muecke 1969, tables 4.1,4.2) that the late Scourian hornblendes have a higher Fe/Mg ratio than the granulite pyroxenes from which they were derived. This is consistent with the observation that the modal quantity of opaque minerals is lower in the amphibolised rocks, as is shown by the following average figures:

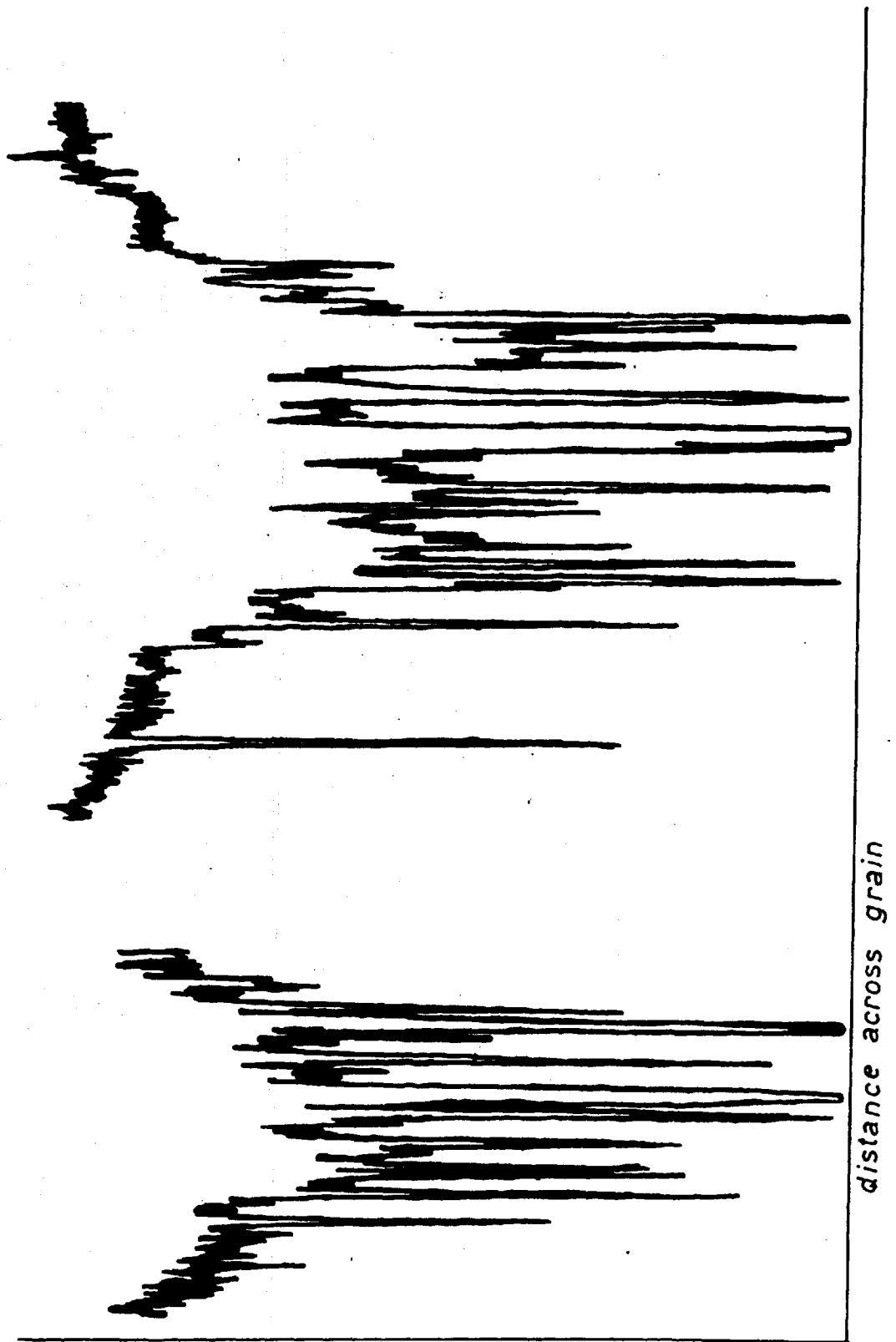
Granulites: felsic division	2.1%	} from Muecke (1969).
intermediate division	4.8%	
mafic division	4.5%	
Amphibolised granulites:		
table 5.1	1.6%	

These figures suggest that the opaque phase is consumed during amphibolisation reactions. The reaction 5.1 that was evaluated above also shows this.

Finally the textural evidence presented in section 5.1 indicates that free quartz was not a reaction product during the early stages of amphibolisation. The reaction 5.1 describes only the general balance between pyroxene and hornblende, with no regard for the development of coronas. Deviations from the general reaction permit some understanding of the reaction mechanism; this is discussed below.

Fig.5.1 presents some Geoscan traverses across aggregates of hornblende such as that shown in plate 5.1. This figure records counts per second for an AL KA1 peak as the electron beam traverses the grain. It is evident that the amount of aluminium decreases towards the centre of the hornblende and that this decrease is gradual. The sharp depressions in this trace coincide with the positions of intergrown quartz grains. Fig5.2 shows a similar traverse, this time across a hornblende aggregate that has recrystallised (stage c in section 5.1 above). Aluminium is now evenly distributed across the hornblende, while the sharp depressions indicate that quartz grains are still present near the centre of the hornblende.

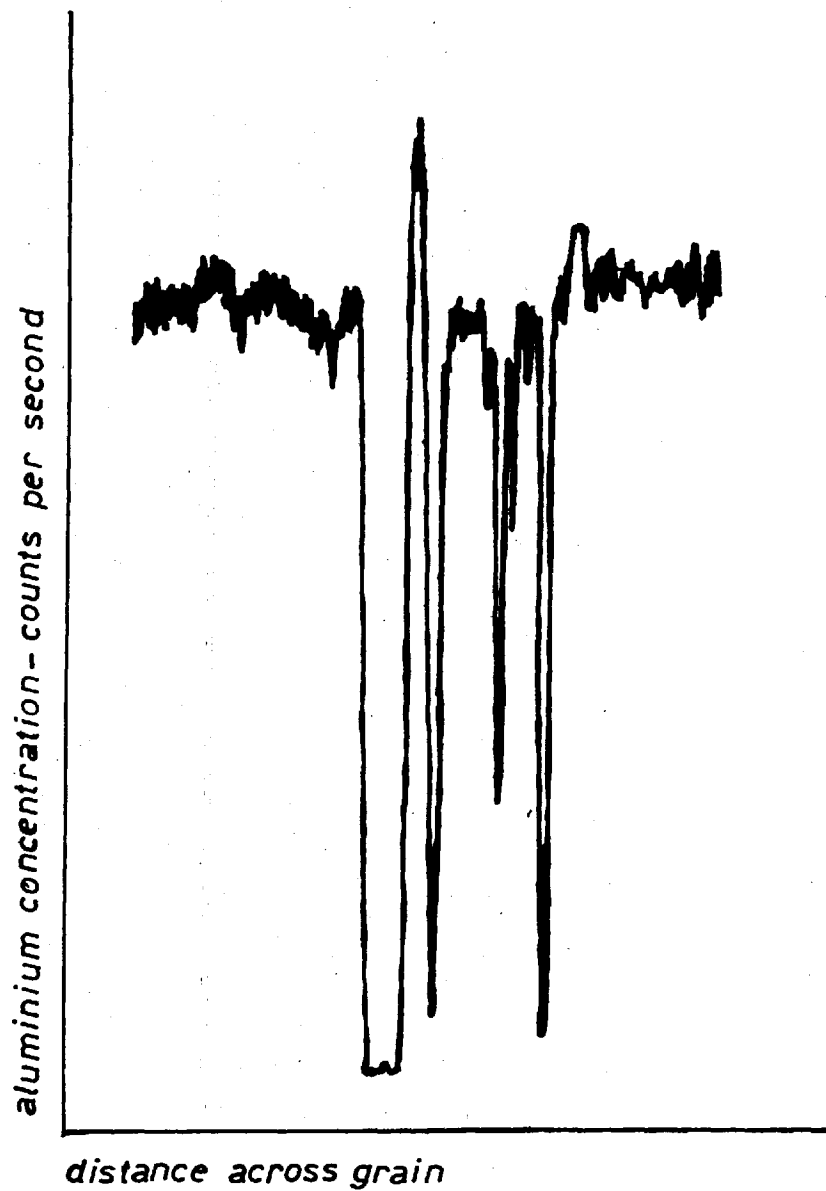
GEOSCAN TRAVERSE ACROSS HORNBLENDE
showing concentration of aluminium
Sample 168



aluminium concentration - counts per second

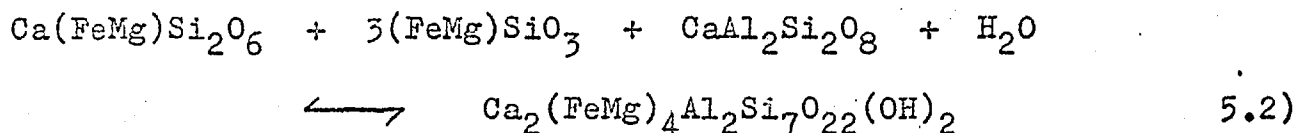
*GEOSCAN TRAVERSE ACROSS HORNBLENDE
showing concentration of aluminium
Sample 28*

FIG 5.2

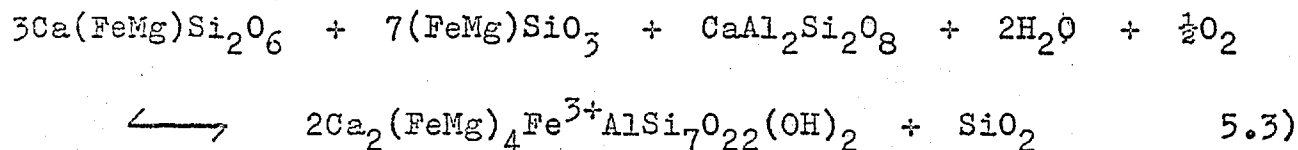


5.3 Simplified pyroxene - hornblende equilibria.

It has been possible to deduce chemical reactions in terms of simplified mineral compositions which show, in general, how the corona textures described above could have originated. These hypothetical reactions are written as follows:



$$\left. \begin{array}{l} \Delta V = -12.7 \\ \Delta S = -1.2 \end{array} \right\} \begin{array}{l} \Delta V_{\text{solids}} = +11.6 \\ P_{\text{H}_2\text{O}} = 5\text{Kb} \\ T = 900^\circ\text{C} \end{array}$$



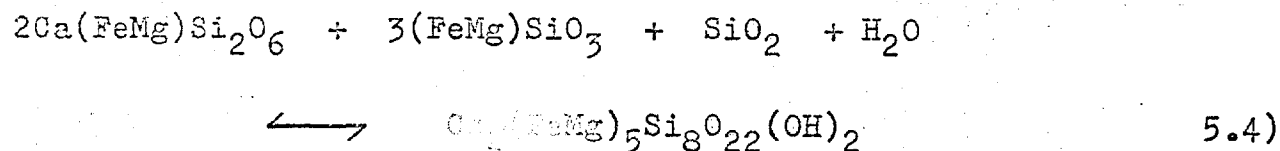
$$\left. \begin{array}{l} \Delta V = +0.6 \\ \Delta S = -3.0 \end{array} \right\} \begin{array}{l} \Delta V_{\text{solids}} = +49.1 \\ P_{\text{H}_2\text{O}} = 5\text{Kb} \\ T = 900^\circ\text{C} \end{array}$$

Initially pyroxene and plagioclase react to form a hornblende (reaction 5.2). Gradually, as the pyroxene diminishes at the expense of a growing rim of hornblende, the supply of plagioclase decreases and the reaction products become a decreasingly aluminous hornblende and an increasing amount of free quartz, as indicated by reaction 5.3. Since the pyroxenes contain significant amounts of aluminium (Muecke 1969), an aluminium free amphibole would not form. Both reactions consume more orthopyroxene than clinopyroxene, and this may explain the general absence of orthopyroxene relicts in the partially amphibolised granulites.

There are two factors which could cause reaction 5.2 to decline and cease: a) decline in the supply of anorthite molecule, b) consumption of all the orthopyroxene.

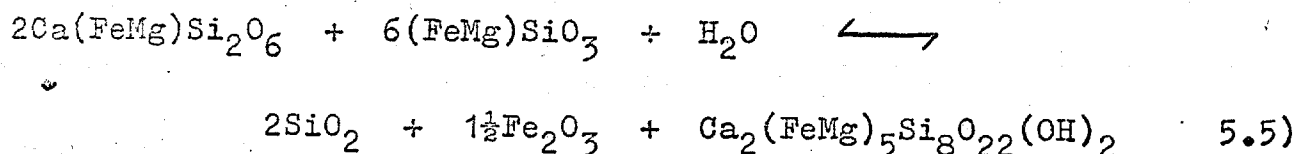
The anorthite will be derived from an andesine plagioclase; textural evidence suggests that at least initially the reactions taking place are solid-solid reactions. The anorthite molecule will have to diffuse out of the plagioclase and through an increasing amount of hornblende before it can react with the pyroxene. The width of the quartz free hornblende rims (eg. in an aggregate 3mm in diameter, these rims are about 0.2mm wide) suggest that the distance of solid diffusion of the anorthite molecule is small under the conditions at which reaction 5.2 occurred.

A decrease in the supply of anorthite necessitates the formation of a less aluminous hornblende if amphibolisation is to proceed, and this is apparently inconsistent with the simultaneous development of free quartz. Formation of tremolite, an aluminium free amphibole, can be considered as the extreme case of the reaction system. The following reaction can be written:



$$\left. \begin{array}{l} \Delta V = -0.6 \\ \Delta S = +0.1 \end{array} \right\} \begin{array}{l} \Delta V_{\text{solids}} = +23.7 \\ P_{\text{H}_2\text{O}} = 5\text{Kb} \\ T = 800^\circ\text{C} \end{array}$$

Alternatively:



$$\left. \begin{array}{l} \Delta V = +14.1 \\ \Delta S = +39.5 \end{array} \right\} \begin{array}{l} \Delta V_{\text{solids}} = +42 \\ P_{\text{H}_2\text{O}} = 5\text{Kb} \\ T = 900^\circ\text{C} \end{array}$$

There are a number of obvious reasons why neither of these reactions describes accurately the degeneration of the granulite pyroxenes being considered. Firstly, reaction 5.4 consumes quartz; secondly, reaction 5.5 produces haematite; and thirdly, as reaction 5.5 is written, the entropy change is $+39.5 \text{ cal/mol/deg}$ indicating that tremolite is the high temperature side of the reaction. Reaction 5.1 is quartz producing, as required, but here the entropy change is again positive ($+14.7 \text{ cal/mol/deg}$), and there is a charge imbalance in this reaction (excess negative charge on the right hand side). These factors are all symptomatic of the difficulty in writing realistic pyroxene-hornblende equilibria.

Reaction 5.3 has been written as a possible solution to this problem. With partial oxidation of the iron in the hornblende and substitution of aluminium in the Z sites of the hornblende, the charge imbalance is corrected and the entropy change becomes negative. The charge imbalance is greater than could be compensated for by substitution of alkalis into the A sites of the hornblende, i.e. by formation of the edenite molecule.

A preliminary examination of the structural formulae of some analysed hornblendes (table 5.2) suggests that this model may be correct in principle. A typical structural formula (hornblende-sample 28) calculated on the basis of 24 oxygens is as follows (H_2O has been added to the analysis at the level of 2wt.% oxide):

Si	6.5	}	8.0
Al	1.5		
Al	0.48	}	4.92
Ti	0.08		
Fe ²	1.02	}	4.92
Fe ³	1.03		
Mg	2.31	}	
Ca	1.9		
Na	0.08		
K	0.13		
(OH)	2.03		

Ferric iron was calculated so that there was an overall charge balance.

The situation is complicated by the fact that the pyroxenes, on the basis of six oxygens, contain 0.05-0.08 atoms of ferric iron. Also magnetite is involved in the left hand side of the reaction, as is shown by reaction 5.1.

However, if reaction 5.3 is taken as a model, it is possible to calculate the amount of oxygen necessary to cause the oxidation shown in reaction 5.3, and, on the assumption that this oxygen is derived by dissociation of pure water, to estimate the amount of water relative to rock in the system during oxidation. The calculation is carried out as follows:

$$\begin{aligned} \text{Assume } P_{\text{H}_2\text{O}} &= 7\text{Kb} \\ T &= 800^\circ\text{C} \\ P_{\text{O}_2} &= 10^{-18}\text{bars} \end{aligned}$$

$$\text{ie. } f_{\text{H}_2\text{O}} = 7847\text{bars}$$

$$K = \frac{(H_2O)}{(H_2)(O_2)^{\frac{1}{2}}} = 7.638 \cdot 10^8$$

$$\text{therefore } P_{\text{H}_2} = 1.082 \cdot 10^4 \text{ bars}$$

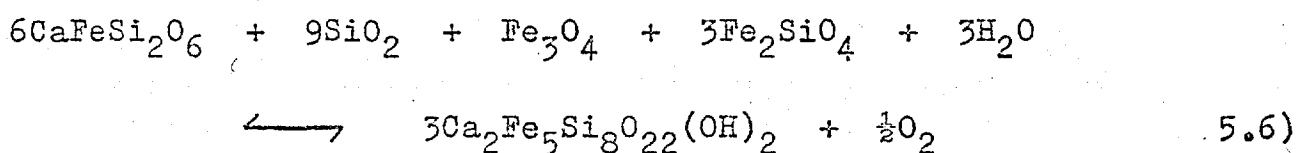
A rock containing 33 modal percent hornblende requires 1/1625 mols hydrogen by dissociation of water at 800°C, $1.082 \cdot 10^4$ bars.

Assuming that the volume of hydrogen and water are similar, $4.04 \cdot 10^{-3} \text{gH}_2\text{O}/\text{cm}^3$ rock are required to cause the necessary oxidation.

This calculation has been repeated for a different oxygen partial pressure, and at different temperatures ($P_{\text{H}_2\text{O}}$ constant). A felsic rock containing 10 modal percent hornblende was also considered. The results are shown graphically in fig.5.3.

If reaction 5.3 is a true description of the second stage of amphibolisation of the pyroxenes, then it is necessary that water, in excess of that contained in the hornblendes, was present in the system during the progress of amphibolisation. It is apparent from fig.5.3 that the amount of water required to cause oxidation at a specified value of oxygen partial pressure is particularly sensitive to variations in temperature. More detailed knowledge of the oxidation states of the rocks and minerals involved is necessary before the amount of water present can be defined accurately.

Reaction 5.4 has been investigated experimentally by Boyd (1954) for magnesium end members and this reaction might be expected to occur in the systems being considered here, since excess quartz is generally present. The stability of iron tremolite



has been investigated experimentally by Ernst (1966). The stability curves of iron tremolite (QFM buffer) and of magnesium tremolite, for $P_{\text{H}_2\text{O}} = P_{\text{total}}$, are shown in fig.5.4. The stability curve of magnesium tremolite has been extrapolated to 5Kb by calculation. It can be seen that tremolite is stable over a wide range of temperature, the breakdown temperature at a defined oxygen partial pressure being determined by the iron/magnesium ratio.

Aluminous hornblendes are generally stable to higher temperatures than tremolite. The stability curve of pargasite (Boyd 1956) in a silica deficient system, and of amphibole in natural olivine tholeiite and high alumina basalts (Yoder and Tilley 1962) are also shown on fig.5.4. The determination of Binns (1969a) is based on the stability of hornblende in natural metabasites; the difference between the curves of Binns and of Yoder and Tilley may be a result of the iron content of

OXIDATION DURING FORMATION OF HORNBLENDE

Relation between temperature and quantity of water necessary to cause a given oxidation

- A. rock containing 40% hornblende $P_{H_2O} = 7Kb, P_{O_2} = 10^{-16}$ bars
- B. rock containing 40% hornblende $P_{H_2O} = 7Kb, P_{O_2} = 10^{-18}$ bars
- C. rock containing 10% hornblende $P_{H_2O} = 7Kb, P_{O_2} = 10^{-18}$ bars

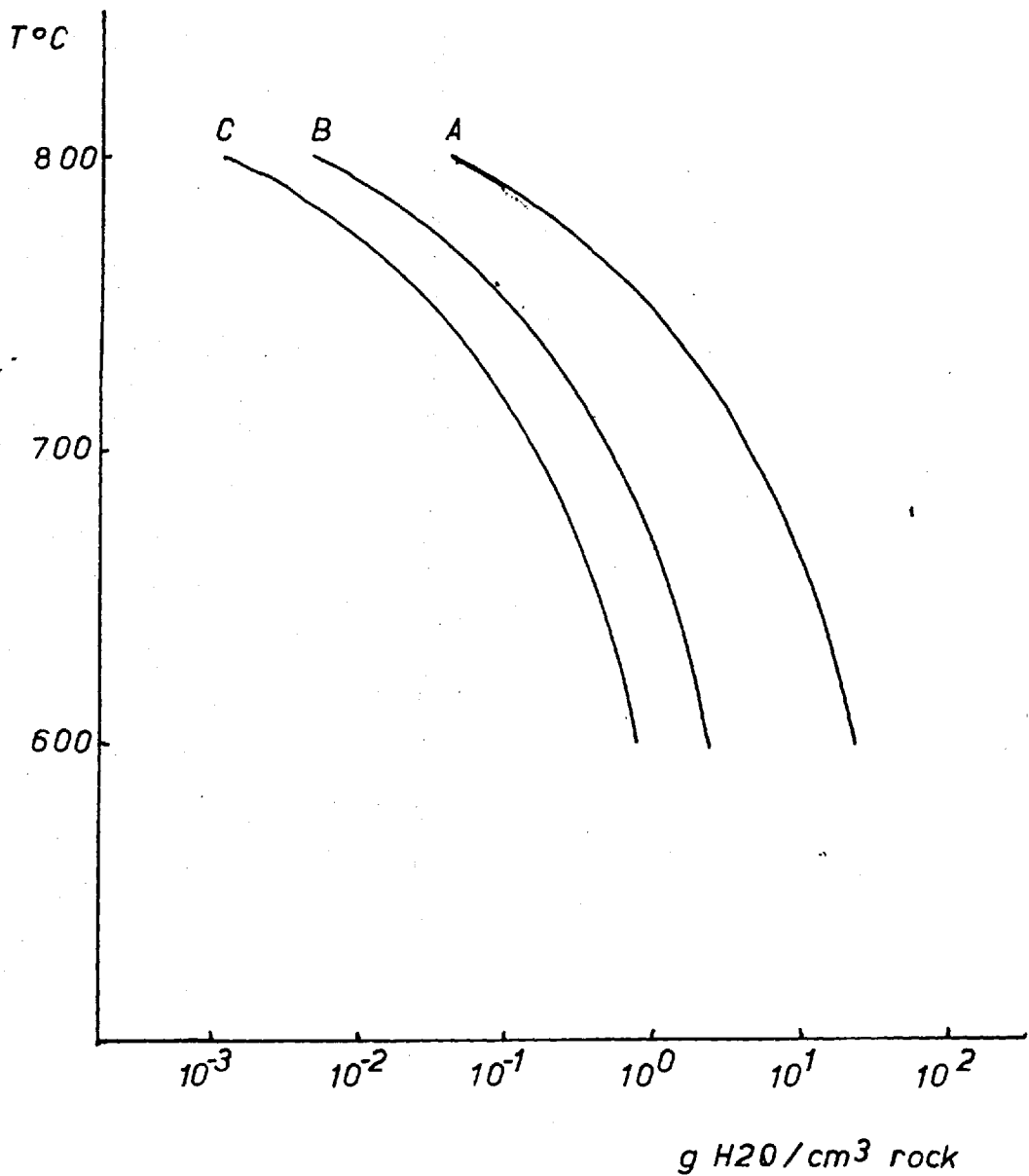
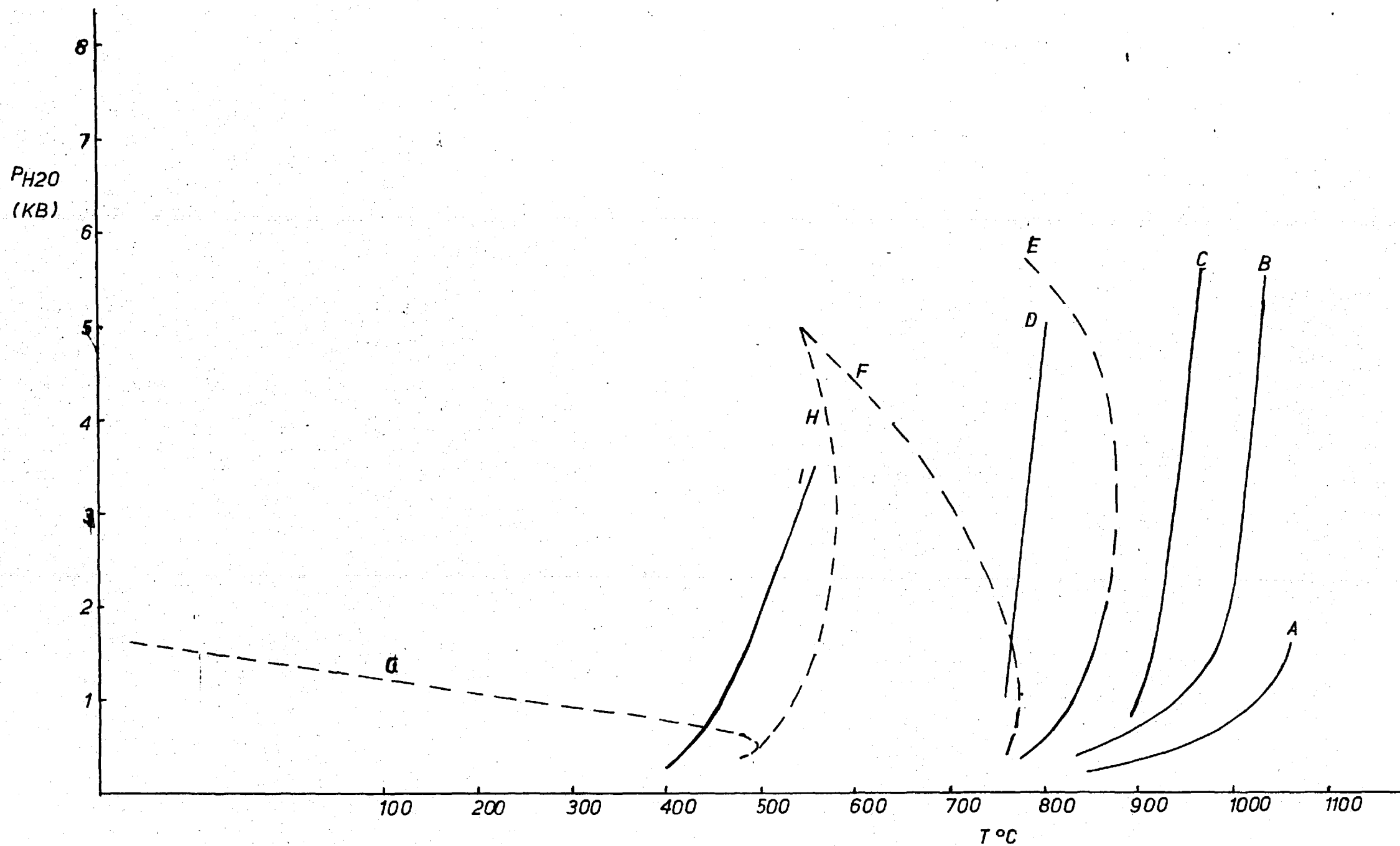


FIG 5.4

STABILITY CURVES OF AMPHIBOLES
(dashed curves have been calculated)

- A. Pargasite (Boyd)
- B. High alumina basalt (Yoder and Tilley)
- C. Olivine basalt (Yoder and Tilley)
- D. Hornblende (Binns)
- E. MG tremolite (Boyd) $P_{H_2O} = P_{TOTAL}$
- F. MG tremolite $P_{H_2O} = 0.5 P_{TOTAL}$ $\Delta S = 10$
- G. MG tremolite $P_{H_2O} = 0.5 P_{TOTAL}$ $\Delta S = 1$
- H. MG tremolite $P_{TOTAL} = P_{H_2O} + 0.5 K_b$
- I. FE tremolite (Ernst) QFM buffer



the hornblende or a feature of the experimental technique.

If at a specified temperature and pressures of water and oxygen, hornblende and tremolite coexisted stably, then the hornblende will be relatively iron rich. Thus a high temperature may prohibit the formation of a tremolite of composition consistent with the bulk composition of the rock. It is suggested that this is the reason why reactions analogous to reaction 5.4 did not occur during the amphibolisation of the pyroxenes being considered. In some of the assemblages that have been studied small amounts of tremolite are seen to coexist with hornblende (eg sample 28 - see table 7.1 for analyses), and this tremolite is interpreted, on the basis of textural evidence, as having formed during the final stages of amphibolisation. Sample 28 contains a small amount of hornblende with a very low aluminium content and an mg ratio of 66, and this coexists with an aluminous hornblende with an mg ratio of 35. Based on the assumptions that

a) the relation between the iron/magnesium ratio and the stability temperature of both tremolite and hornblende is linear,

b) for similar iron/magnesium ratios hornblende is stable to higher temperatures than tremolite (as indicated by the experimental results shown in fig.5.4) and this temperature difference is constant over the Fe/Mg range,

c) $P_{H_2O} = P_{total} = 5Kb$,

d) P_{O_2} is buffered by QFM,

e) magnetite is present,

then, using fig.5.4, the hornblende and tremolitic hornblende specified above could have formed in the temperature range 750 - 800°C. It is in this temperature range that melting of basaltic compositions begins (Yoder and Tilley 1962) providing condition c) above is maintained. The validity of this assumption is discussed in chapter 9; in general it is considered that addition of water to the granulites occurred in the region $P_{H_2O} = P_{total}$. It is thus possible that incipient melting could have occurred during amphibolisation. The perfect linear fabrics developed in hornblende-plagioclase rocks, and described

in section 2.2B, where complete segregation of hornblende and plagioclase has occurred, may represent the result of such a process.

It is important to be able to estimate the thermal stability of the pyroxene-hornblende equilibrium curves under conditions where $P_{H_2O} \neq P_{total}$, since this is relevant to both the formation and degeneration of pyroxene. The equilibrium curve for reaction 5.4 is known (Boyd 1954) under the conditions $P_{H_2O} = P_{total}$. Since volume and entropy data are available for all the phases involved in this reaction, it is possible to calculate the position of the equilibrium curves for the conditions a) $P_{H_2O} = \frac{1}{2}P_{total}$, and b) $P_{H_2O} + \frac{1}{2}Kb = P_{total}$. The former has been evaluated for entropy changes of 10 and 1 cal/mol/deg and the latter for an entropy change of 1 cal/mol/deg. The results are plotted on fig.5.4. It is obviously critical to know exactly the value of the entropy change for the reaction and errors in the initial data lead to a possible range ($\Delta S = 0-5$ cal/mol/deg) for reaction 5.4. These low values of the entropy change, and relatively large solid volume changes, lead to a low thermal stability of the equilibrium curve when $P_{H_2O} \neq P_{total}$. Reactions 5.2 and 5.3 have similarly low entropy changes and though it is not possible to plot equilibrium curves in $P_{H_2O}-T$ space for these reactions it is apparent that the stability fields of hornblende will be very much reduced under conditions where P_{H_2O} is less than P_{total} .

Not all pyroxene-hornblende equilibria are affected in this way by varying water pressures. For example, Binns (1969a) estimates his determination to be affected by only 20°C when the conditions change from $P_{H_2O} = P_{total}$ to $P_{H_2O} = \frac{1}{2}P_{total}$ (at $P_{H_2O} = 5Kb$). Binns (1969a) proposes a reaction involving one gram-molecule of hornblende, and quotes ΔV_{solids} as being $-8cm^3$. The entropy change for this reaction can then be estimated as 53 cal/mol/deg. Yoder and Tilley (1962) suggest a similar stability for their determined curves, but in this case there is no thermodynamic data to support their notion.

The data presented in this section has an important bearing on the mechanism of hydration of the granulite terrain, and can

be used to evaluate the possible quantities of heat evolved if the reactions occur metastably below equilibrium and are thus exothermic. This will become a point for discussion in chapter 9.

Finally, there are two problems concerning the formation of hornblende from pyroxene to which there are at present no satisfactory answers. Frequently, partially amphibolised rocks contain relicts of clinopyroxene only, while completely amphibolised rocks of the same composition contain no pyroxene. It is difficult to write a straightforward reaction describing the degeneration of clinopyroxene since this mineral contains equal amounts of Ca and (Fe,Mg), whereas hornblende contains (Fe,Mg) in excess of Ca. No calcium rich minerals occur in significant quantities in the amphibolised rocks; formation of sphene is limited by the small amount of titanium available. It is possible to write reactions involving clinopyroxene and iron oxides, but the quantity of the latter required is in excess of the modal percentages present in the rocks. Alternatively, reactions involving albite, with the production of anorthite and relatively sodic hornblende, can be balanced on paper. Further detailed analyses of partially amphibolised rocks may detect fluctuations in hornblende chemistry resulting from such reactions. It is perhaps an indication of the mineralogical and chemical complexity of the rocks involved that these reactions do readily occur.

The second problem concerns the distribution of aluminium in the hornblende coronas. Initially (fig.5.1) aluminium is concentrated around the margins of the coronas, but after recrystallisation (fig.5.2) it is distributed evenly across the hornblende. It is not known whether this second process is accompanied by a decrease in the aluminium content of the hornblende at the margin of the corona, or whether the hornblende aggregate as a whole becomes enriched in aluminium during recrystallisation. Again, only careful analytical work will answer this question.

5.4 Minor reactions associated with amphibolisation.

It was mentioned earlier that certain other new minerals appear during amphibolisation, and these will be briefly discussed here.

The appearance of biotite in small amounts is apparently linked with the disappearance of the antiperthite blebs from the plagioclase. Biotite equilibria are discussed at length in the next chapter. Larger modal quantities of biotite occur in some of the steep limbs of Scourian folds (eg. samples 12 and 17) and it is not known at present whether this feature is a consequence of the bulk composition of these rocks or whether the biotite results from Laxfordian introduction of potash (cf. chapter 6 and discussion in chapter 9).

Zircon is recorded from only one set of samples (26-28) and its appearance can only be attributed to original rock composition.

The occurrence of small amounts of calcite as round grains within the plagioclase of some amphibolised granulites is at first curious, but can be explained by reference to some work by Barker (1965). Pyroxenes separated from a Scourian granulite (from the Lochinver district, see Evans 1963) were heated and the gases evolved collected and analysed. Barker found that approximately equal amounts of H₂O and CO₂, and lesser amounts of H₂ and CO, were evolved. The total quantity of gas evolved after heating at 1034°C for two hours was 5.3cm³ per gram of rock at STP. Release of CO₂ from the pyroxenes during amphibolisation could be sufficient to form calcite.

Epidote is often seen occurring as small grains along the boundaries between hornblende and plagioclase in the more mafic assemblages. Associated with the epidote are small amounts of tremolite and gedrite. It is suggested (section 7.1) that these minerals result from minor oxidation of the hornblende.

TABLE 5.4.

HORNBLLENDE PODS: MINERAL ANALYSES.

	SiO ₂	Al ₂ O ₃	TiO ₂	FeO	MgO	CaO	K ₂ O	total	mg
187H	43.4	11.1	1.5	13.1	13.9	11.7	0.8	95.5	51.0
187H	42.2	12.8	1.8	11.3	14.1	11.6	1.2	95.0	55.5
188H _c	46.2	10.2	0.4	10.6	14.9	11.5	0.6	94.4	58.4
188H _e	52.0	6.2	0.2	8.8	16.8	11.4	0.1	95.5	65.6
188HT	53.8	4.0	0.1	7.7	17.4	11.5	0.1	94.6	69.3
188B	42.1	15.7	1.6	10.5	18.6	0.0	6.0	94.6	64.0
188B	40.7	16.0	1.5	10.7	19.6	0.0	6.3	94.9	64.7
178H	43.9	10.9	0.9	16.0	11.4	11.5	0.6	95.2	41.5
178H	43.0	11.4	0.9	16.3	11.0	11.6	0.6	94.8	40.3
178H	41.6	12.5	1.2	16.8	10.1	11.3	0.8	94.3	37.6
162H	44.5	11.4	0.6	10.4	15.6	11.7	0.4	94.6	59.9
162H	42.3	13.0	0.9	10.6	14.6	11.3	0.7	93.7	57.9
162H	42.5	12.7	0.9	10.7	14.8	11.4	0.6	93.8	57.9
162HT	53.2	3.5	0.1	9.4	20.7	9.7	0.1	96.7	68.6
162HT	52.5	3.4	0.3	7.7	20.3	10.3	0.1	94.6	72.4
179H _c	44.7	11.8	0.4	11.9	14.7	12.2	1.1	97.1	55.3
179H _e	52.2	4.0	0.1	9.0	18.6	12.2	0.3	96.5	67.4
179H	48.6	5.8	0.1	8.8	18.2	12.4	0.3	94.4	67.4
179HT	56.1	2.8	0.1	7.5	17.6	12.6	0.05	96.8	69.9
179B	38.2	16.2	1.2	10.2	20.6	0.4	5.9	92.8	66.0

5.3 Hornblende pods.

Hornblende pods are a minor, though distinctive, component of the amphibolised granulites. Inspection of thin sections shows that these rocks have a variety of mineral assemblages. Modal analyses of some interesting assemblages are given in table 5.3. Not all of these are relevant to the present section since some of the hornblende pods have been biotised in Laxfordian shear zones.

A). Petrography.

Though the dominant mineral in these rocks is an amphibole, a number of samples contain relict orthopyroxene, clinopyroxene and plagioclase, and one sample (187) contains a virtually unaltered granulite facies assemblage.

In many of the samples where no relict minerals are present it is possible to recognise a number of distinct amphiboles, as follows:

- a) dark green to green brown hornblende containing many opaque inclusions,
- b) green pleochroic hornblende in optical continuity with a) but containing no inclusions,
- c) pale green to colourless amphibole, occasionally in optical continuity with the darker hornblende, but more usually occurring as aggregates between the hornblendes.

The appearance of opaque inclusions is a characteristic feature of the retrogression of the pargasitic hornblendes found in some of the mafic granulite facies assemblages (Muecke 1969). These inclusions are probably rutile and are retained within the host crystal. This feature enables originally pargasitic hornblendes to be tentatively identified in retrogressed assemblages.

The pale green amphibole (c above) was identified by analysis as tremolitic hornblende with a low alumina content (see table 5.4).

B). Source rock for the hornblende pods.

Sample 187 (table 5.3) falls into the hydrous mafic

division of the granulites set up by Muecke (1969). It is possible to predict the effects of retrogression on this assemblage according to a set scheme, and calculate the modal percentages of minerals in the resulting assemblage.

The composition of the orthopyroxene in 187 was estimated optically as En75. From the data presented by Muecke (1969, fig.4.2) an estimate of the composition of the coexisting clinopyroxene was made. The composition of the pargasitic hornblende is given in table 5.4.

The volume concentration of each atom in each phase is calculated and multiplied by the modal percent of the phase. Thus a total number for each atom present in 187 is obtained in a form that is easily manipulated. The atoms are then allotted to new minerals, the compositions of which have been taken from the analyses in table 5.4 of the hornblendes in the retrogressed assemblages. The following scheme of allotment is used:

- i) all AL is allotted to hornblende,
- ii) the remaining CA is allotted to tremolite,
- iii) excess FE is calculated as oxide,
- iv) the pargasitic hornblende is changed to hornblende by combination with tremolite,
- v) excess SI is calculated as quartz.

Following this scheme, sample 187 has been recalculated to the following mode:

hornblende	H1	39.0
tremolite	H2	29.7
hornblende derived from pargasitic		
hornblende	H3	20.3
quartz		6.5
haematite		4.5

A number of conclusions can be drawn from this calculation:

- a) The amount of quartz in the hornblende pod indicates the amount of anorthite present in the original granulite because the more anorthite that was present, the more would reaction

5.3 have been inhibited. For example, sample 177 can be recalculated according to the reactions presented in section 5.3 and shown to have contained about 9% anorthite initially.

b) The retrogression calculated above involves a volume increase of 60%.

c) If there is no thorough recrystallisation, a number of hornblendes of different compositions, derived from the different phases present initially, should be seen. (cf the petrographical note made above).

d) A two pyroxene, plagioclase, pargasite assemblage converges to a rock consisting almost completely of hornblende during amphibolisation.

C). Analyses of minerals from the hornblende pods.

Table 5.4 presents all the mineral analyses from the hornblende pods. Some of these analyses are from biotised pods which are discussed in the next chapter. These analyses have been grouped in one table for the sake of completeness. It has not been possible to obtain realistic analyses of the centres of some hornblende grains because of the large number of fine opaque inclusions in these areas. Only the clear green hornblende in optical continuity with these parts can be analysed.

The other analyses of concern here are those of the colourless amphibole mentioned in A) above. It is evident that this mineral is a hornblende with a low aluminium content, whereas the darker hornblendes have a much higher aluminium content.

From the analytical data it is concluded that two types of hornblende are present in many of the hornblende pods. It is not clear whether this is a result of the equilibrium coexistence of two hornblendes (separated by a compositional immiscibility gap) or of incomplete recrystallisation, resulting in non equilibrium assemblages. In view of the comments made above, the latter interpretation is preferred.

A detailed discussion of the mineral analyses will be presented in chapter 8.

5.6 The isochemistry of amphibolisation.

Evans (1963), after studying the amphibolisation of the Scourian granulites at Lochinver, concluded from his mineral analyses of a granulite and of an amphibolised granulite that the transformation of amphibolisation took place isochemically except for the addition of water.

In a similar way, by combining the mineral and modal analyses presented here, the amphibolisation of the Scourian granulites at Scourie is considered to be isochemical. The chemical and tectonic events at both Scourie and Lochinver during the late Scourian were thus very similar.

In an attempt to characterise the water introduced into the granulites, a large number of Geoscan wavelength scans on individual minerals have been prepared. No minor element contents were found in the hornblendes that could be attributed to introduction with the water. In particular, no significant variation is seen in the alkali, alkali earth and transition metal element contents of the hornblendes. Trace element data are not available.

It is concluded that during amphibolisation, water was the only significant chemical species introduced into the rock system.

CHAPTER 6.

REACTIONS INVOLVING BIOTITE.

Just as the chapter on amphibolisation was concerned with the reactions associated with Scourian structures, so this chapter on biotisation will describe mineral reactions seen mainly in samples from Laxfordian structures.

In the rocks studied here, assemblages in which biotite is a major phase are restricted in occurrence to Laxfordian shear zones and to the Tarbet Steep Belt. These biotite bearing rocks are strongly recrystallised and have characteristic mineral fabrics (see chapter 2). Because of the completeness of recrystallisation there is a general absence of textures which help to interpret the way in which biotite formed. There is thus very little petrographic detail that will add to the information already presented by Peach et al (1907) and Sutton and Watson (1951), and no systematic petrography is presented here.

The most important reaction to be discussed is the formation of biotite from hornblende. This reaction is recognised because modal analyses (table 6.1) show an increase in biotite content at the expense of modal hornblende in the shear zones relative to the samples collected adjacent to the shear zones.

In a later section the mineral reactions seen in a group of granulites referred to as the GK gneiss are discussed. These assemblages are of considerable importance because high alumina minerals are formed in the shear zones. The presence of these minerals enables an evaluation of the PT conditions during the formation of the shear zones to be made.

PLATE 6.1

Hornblende - biotite fabric in Laxfordian
shear zone

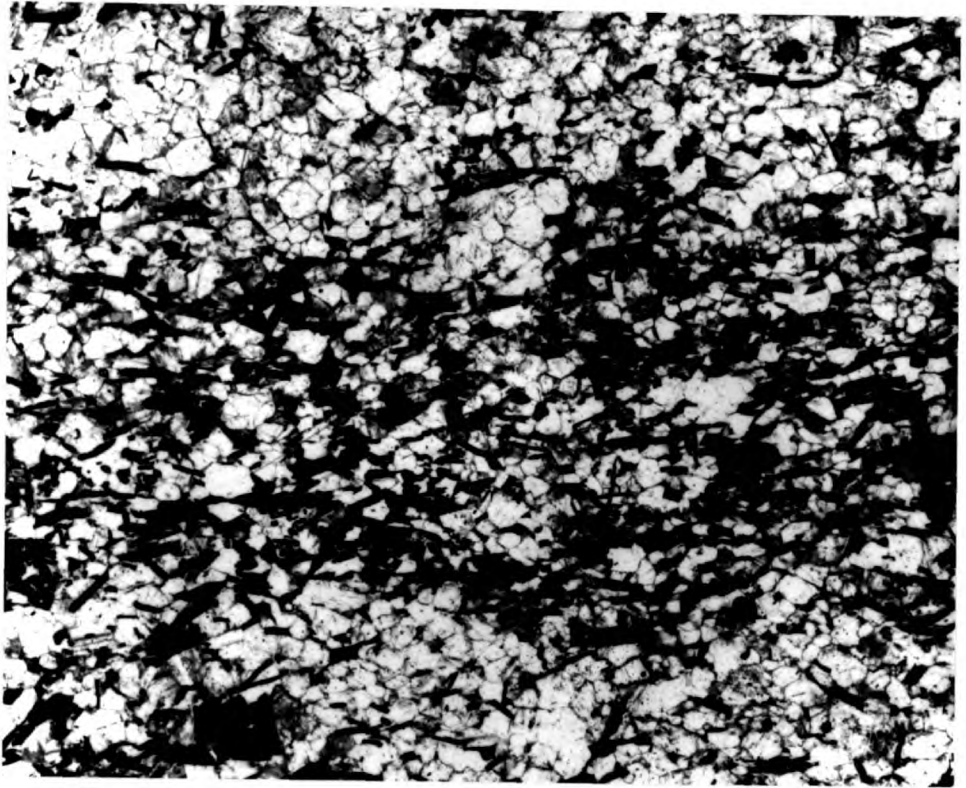


TABLE 6.1.

SHEAR ZONES AND ADJACENT ASSEMBLAGES: MODAL ANALYSES.

	<u>24_s</u>	<u>25_a</u>	<u>140_s</u>	<u>141_a</u>	<u>142_s</u>	<u>143_a</u>	<u>152_s</u>	<u>153_a</u>	<u>220_s</u>
Plag	52.8	51.1	57.1	50.7	62.3	64.2	49.3		62.6
Pap								54.0	
Kfsp									
Qtz	24.2	17.5	28.8	29.2	17.0	10.0	0.7	7.8	23.3
Cpx		7.5						11.7	
Hbl	9.5	21.3	11.5	19.6	8.5	20.0	43.9	21.9	5.6
Biot	13.3		2.5		10.1	5.2	3.1		7.9
Musc									
Epid									
Op	0.3	2.0	tr	0.5	1.5	0.7	2.5	4.3	0.6
Ap	tr	0.2	tr	tr	0.6		0.5	0.3	tr
Rut		0.4							
Cal									
counts	3600	3600	1500	3100	3200	3400	3300	3100	2800
An%			25		27	35	25	30	33

table 6.1 continued:

	<u>201_s</u>	<u>200_a</u>	164 _s	191 _s	195 _s	<u>19_s</u>	<u>18_a</u>	<u>2.7_s</u>	<u>2.2_a</u>
Plag	54.0	40.3	48.6	66.0	9.9	52.8	55.0	32.6	29.0
Pap									
Kfsp						4.0	6.9		
Qtz	19.4	25.2	15.6	18.8	8.8	28.6	32.0	8.9	5.5
Cpx									
Hbl	15.7	33.1	2.2		50.5			52.4	64.0
Biot	8.0		23.5	15.2	26.3	8.4	6.0	4.1	
Musc						6.1			
Epid	0.5		5.0		0.8				
Op	2.1	1.0	2.9		2.3	0.1	0.1	2.0	1.2
Ap	0.3	0.2	2.2	tr	tr	tr	tr		
Rut			tr		1.4				0.3
Cal						tr			
counts	2500	4300	4000	1000	1200	2600	2500	4100	3900
An%	25	29		24	27	12			

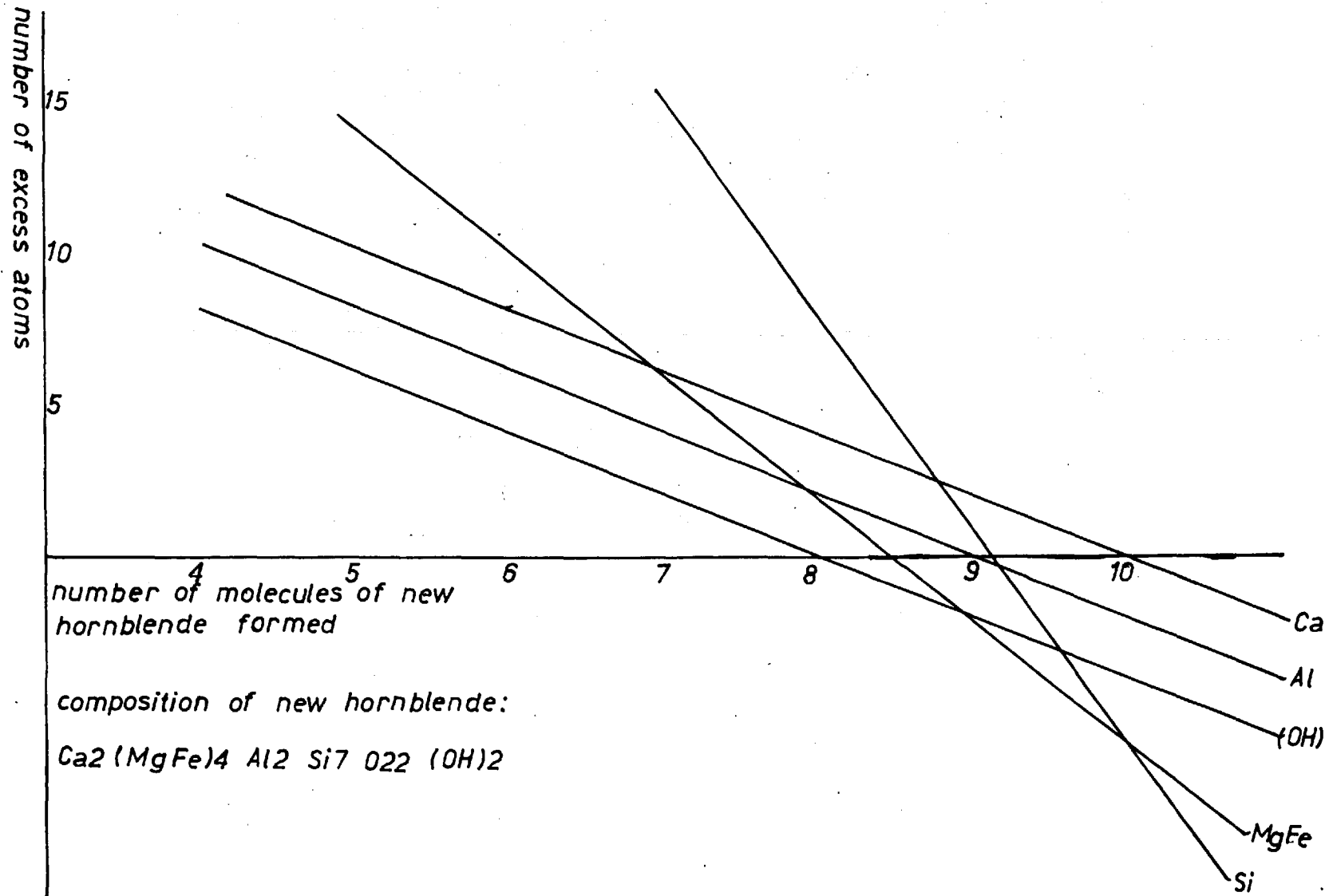


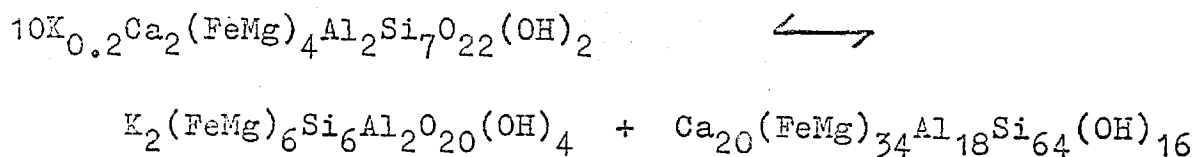
FIG 6.1

6.1 The transformation of hornblende to biotite.

Published whole rock analyses of the Scourian granulites (Muecke 1969) show that these rocks have very low potash contents. On average, the granulites could produce a maximum of 5 modal percent of biotite on the basis of this potash content. The amphibolised granulites have a similarly low potash content (Holland 1966, Holland and Lambert 1971). Since the Laxfordian shear zones cut across the banding of the amphibolised granulites, it is possible to compare assemblages from the shear zones with samples from equivalent rocks immediately adjacent to these zones.

Table 6.1 presents modal analyses of such a set of samples. It is immediately obvious that many of the shear zones contain larger quantities of biotite than the samples adjacent to the shear zones. For this reason the rocks cannot be linked by isochemical change with the adjacent rocks. This conclusion makes it difficult to write and evaluate conventional mineral reactions for the changes observed.

It is necessary first to consider further the formation of biotite isochemically within an amphibolised assemblage. This process is limited by the availability and balance of atoms other than potassium. The approach adopted here is to take a hornblende composition similar to the hornblendes in the amphibolised granulites (table 5.2) and to split it into a mineral of biotite composition and a new hornblende. The initial hornblende is considered to contain all the potassium present in the whole rock. The process of splitting used is governed by a) maximising the amount of biotite produced and b) minimising the excesses and deficiencies of the different atoms involved. The balance of atoms is particularly important with regard to AL, since no new AL bearing phases are seen in the assemblages. A close approximation to what is considered a probable reaction is as follows



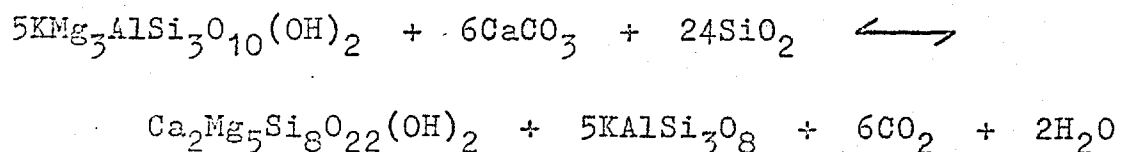
In fig.6.1 this reaction is presented graphically. The entire potash content of the hornblende is allotted to biotite and then a new hornblende is formed using up the excess atoms. In the example shown it is seen that only small excesses in Ca and Si are produced, and only small deficiencies in (FeMg) and (OH) incurred. It is assumed that both Fe and (OH) are available, or buffered, in a real system. A number of other examples have been worked in this way, and larger excesses and deficiencies, were encountered.

The significance of this example is that a rock containing 30 modal percent hornblende can produce 3 modal percent biotite and maintain an internal balance of atoms. The possibility of isochemical change of hornblende to biotite is thus limited. A further restriction on the formation of biotite exists because all the hornblendes analysed (tables 5.2, 6.3) contain potassium. This means that not all the potassium content of each whole rock is available for biotite formation.

6.2 The Carmichael evaluation.

It might be expected that if larger quantities of potash were present in a system, then the above reaction would not balance in terms of hornblende and biotite alone.

Carmichael (1970) defines a hornblende - biotite isograd in the pelitic schists of the Grenville Province, and writes the following ideal reaction:



This reaction is compatible with his observed assemblages (except that the schists contain hornblende and biotite instead of the idealised tremolite and phlogopite) and can be evaluated in terms of known experimental data on mineral stabilities.

Thus for conditions where $P_{\text{fluid}} = P_{\text{total}} = 1\text{Kb}$, and when the fluid is a mixture of H_2O and CO_2 in the correct stoichiometric proportions, this reaction takes place at about 550°C (see Carmichael 1970, figs. 23, 25).

An important consequence of this reaction is that large amounts of quartz and calcite are produced during the formation of biotite from hornblende. The Lewisian assemblages studied here are notable for the general absence of calcite and for the lack of evidence for increase in quartz content in the biotised assemblages.

To present the problem more clearly, one biotite bearing assemblage (sample 164) has been recalculated as a biotite free assemblage using a similar regression method to that employed in section 5.5B. In this case the scheme followed is:

- a) all FeMg is allotted to hornblende,
- b) all Ca is allotted to anorthite,
- c) all K is calculated as equivalent potash feldspar.

The recalculated mode of 164 is then:

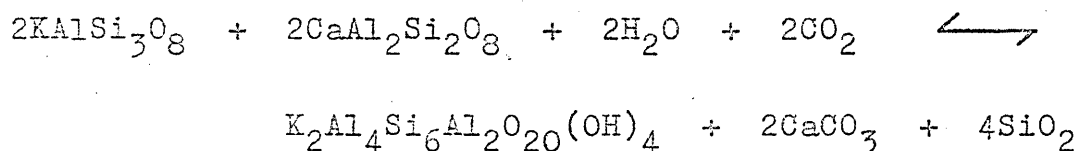
albite	30.2
anorthite	12.3
potash feldspar	17.0
quartz	3.5
hornblende	37.0

Because potash feldspar is essentially absent from the amphibolised granulite assemblages, this calculation is equivalent to saying that on the basis of constant AL content the biotite bearing assemblages have been enriched in potassium and depleted in calcium, while the silicon variation remains indeterminate.

It is not possible to write a balancing reaction for biotite formation without the appearance of a new calcium bearing phase, and since no such phase is observed (the plagioclase compositions are found to change very little, see table 6.1), it is tentatively proposed that the loss/gain metasomatism outlined above actually occurred.

6.3 The stability of muscovite.

The majority of assemblages contain no potash feldspar, and since it is unlikely that potassium was introduced as potash feldspar, the application of Carmichael's equilibrium reaction to biotised assemblages found here is limited. However, the Scourian granite sheets do contain microcline, though they do not contain hornblende. When these granites are cut by shear zones, mineral transformations are observed. Samples 18 and 19 (see table 6.1), collected from Sithean Mor, are taken from a sheet in such a situation and the following reaction can be written for the transformation seen in the shear zone:



Thermodynamic data are available for all the phases involved in this reaction. Thus

$$\begin{aligned} \Delta S &= \text{c.}280\text{cal/mol/deg} && \text{at } P_{\text{H}_2\text{O}}=5\text{Kb, } T=600^\circ\text{C} \\ \Delta G &= 92\text{Kcal} && \text{at } T=25^\circ\text{C} \end{aligned}$$

At a fluid pressure of 5Kb the equilibrium temperature of this reaction is calculated as 630°C . An error of $\pm 30^\circ\text{C}$ is inherent in the data used. The effect on equilibrium of other components in the phases (particularly albite in the anorthite) is unknown.

The stability of muscovite has been studied experimentally and the results of Althaus et al (1970) show that under the conditions $P_{\text{H}_2\text{O}} = P_{\text{total}} = 5\text{Kb}$, muscovite and quartz are unstable at temperatures greater than 710°C .

The presence of stable muscovite in a shear zone is probably the most reliable way of putting an upper limit on the temperature during deformation and metamorphism in the shear zone.

6.4 Biotised hornblende pods.

Reference to table 5.3 shows that a number of hornblende pods contain biotite. Of these, samples 179 and 184 were collected from shear zones with a strong Laxfordian LS fabric.

Sample 184 is from the margin of a pod, while sample 185 is from its centre. It is evident that only the margin has been biotised. Calcite appears as a new phase in the biotised rock, while there is an increase in the amount of quartz in the biotised rock relative to the hornblende rock. Thus this particular assemblage is in accordance with the reaction proposed by Carmichael (1970).

In contrast, sample 179 contains a large quantity of biotite, but no new calcium bearing phase is seen. This indicates that calcium and possibly silica have been lost from an initially hornblendic rock during biotisation. The general absence of calcite from most assemblages suggests that this is not merely a local phenomenon, but that the excess calcium has left the system altogether.

Mineral analyses (table 5.4) show that there is no substantial difference in the composition of many of the hornblendes and biotites from the different hornblende pods. In particular the calcium content of the hornblendes varies little.

6.5 The GK gneiss.

A distinctive group of gneisses has been mapped along the north side of Sithen Mor (see map, cover pocket) and is named the GK gneiss (from the presence of two distinctive minerals, garnet and kyanite). This rock can be seen in two varieties, first with a granulite facies assemblage of plagioclase-quartz-garnet(~~+~~hypersthene), and secondly, in shear zones, with assemblages containing large amounts of biotite. In the field the gneiss is easily distinguished by its rusty weathering and its massive character, with a general absence of mineral banding. This gneiss is superficially very similar

TABLE 6.2.

GK GNEISS: MODAL ANALYSES.

	56	55	53	52	53P	190
Plag	54.3	61.4	57.0	55.5	28.9	48.8
Qtz	17.4	6.1	3.15	10.25		
Gar	15.5	14.35	7.17	4.61	5.2	12.2
Hyp	7.8					
Biot	2.5	14.0	25.32	19.45	20.0	22.2
Musc			0.74	1.6	12.6	tr
Ky			4.26	5.66	10.0	2.3
Str			0.54	0.32	4.7	9.1
Sp		0.21	0.18	0.15	2.2	
Cor			0.10		0.6	0.5
Op	2.5	4.0	2.56	2.26	15.8	4.9
Ap				0.2		
counts	8900	22100	24300	10100	2500	4200

to some of the supposed metasedimentary relicts found in the Outer Hebrides (Coward et al 1969).

Since the assemblages found have not previously been recorded in this area, they will be treated in some detail. In addition, the minerals present are extremely important in helping to elucidate the chemical changes that have occurred in the shear zones.

A): Mineralogy.

Table 6.2 presents modal analyses of some samples of GK gneiss. Large counts were necessary to obtain accurate figures for the less common minerals, and to cover the obvious heterogeneity in some shear zone samples.

Sample 56 shows the least altered granulite assemblage. Small, poorly recrystallised biotites clustered around opaque grains, and a green "serpentine" mineral around hypersthene are the only indications of alteration. The plagioclase (composition An₄₀) is not zoned and Geoscan traverses detected no major element zonation in the garnets.

Sample 55, 53 and 52 were collected from a shear zone close to sample 56 in order of increasing deformation in this zone. Table 6.2 indicates that a rather surprising group of minerals are present in these rocks. Sample 190 is from a different shear zone.

The modal analyses, by covering a number of thin sections in each case, conceal the fact that certain minerals are always associated; the petrology of these rocks must be considered in terms of domains of only a few square millimetres each. The important petrographic observations made on the shear zone samples will now be presented.

The plagioclase is often seen as large crystals that appear to be unrecrystallised relicts of the granulite facies assemblage. They have the same composition as those in the granulite facies rocks. However, the plagioclase sometimes shows a peculiar mode of alteration which is visually analogous to an exsolution phenomenon. In a large feldspar crystal numerous small round

blebs appear, until eventually the original host crystal can no longer be made out, and only a mosaic of small crystals is seen. Optically it is evident that two mineral phases are present in this mosaic. Detailed X-ray diffraction studies and Geoscan analyses show that the two minerals are plagioclase and quartz. The plagioclase does not markedly change its composition during the development of this texture. Thus, if quartz is regarded as being exsolved, then some components of plagioclase must have been lost, and if it is regarded as replacing the plagioclase, then plagioclase as a phase must have been lost and presumably crystallised elsewhere. No evidence has been found to support the latter hypothesis, while the former process does not appear logical. The mode of alteration remains therefore to be explained.

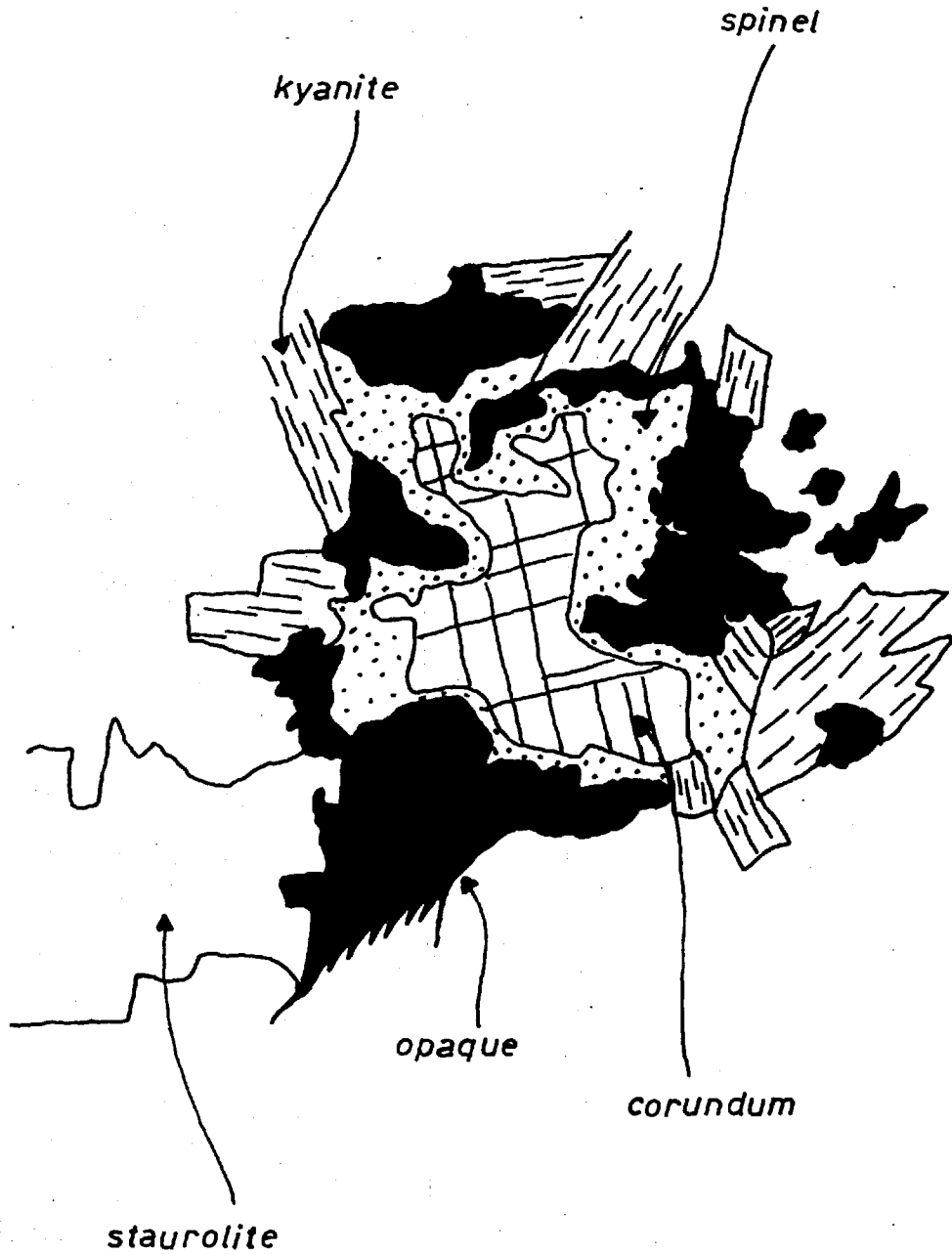
The garnets of the granulite facies assemblage are found, in the shear zones, to be broken and reduced in size and surrounded by sheaths of well crystallised red brown biotite. It is in these sheaths that the kyanite, staurolite, spinel and corundum, noted in the modes, are found.

Kyanite occurs frequently as single euhedral crystals dispersed throughout the biotite. It is never in contact with garnet. Staurolite, spinel and an opaque mineral (usually ilmenite) occur as small clusters of grains and are often intergrown. The spinel is green in colour and qualitative Geoscan work shows that it is a zinc-iron-aluminium variety. Corundum is the rarest of these aluminous minerals and occurs as occasional large crystals. In sample 190 the corundum is embedded in a very large staurolite crystal, while the relationship found in sample 53 is sketched in fig.6.2. The modal analysis of the biotite sheath containing the corundum in 53 is given in table 6.2 under 53P.

It is important to note that quartz and plagioclase are absent from the biotite sheaths, except at the margins of the sheaths where small quartz grains are sometimes seen within biotite crystals. The distribution of muscovite is similarly limited to the margins of the biotite sheaths and to the

RELATIONSHIP OF HIGH ALUMINA MINERALS

Sample 53

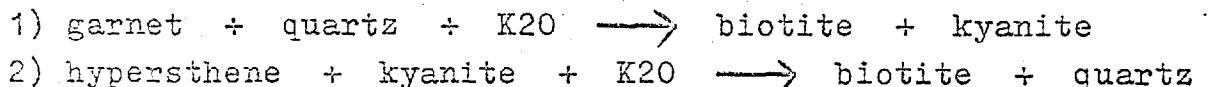


felsic parts of the rock. Muscovite is associated with opaque minerals, a relation to be discussed in the next chapter.

B). Mineral reactions.

The feasibility of a chemical reaction is perhaps too often judged by the ease with which it can be written explicitly in terms of known mineral species. Some work by Carmichael (1969) has shown that many complex reactions and catalysts may be involved in a mineral transformation, and this lessens the possibility of writing a unique chemical reaction that represents the total process. In the rocks being discussed here it is immediately apparent that the shear zone assemblages are considerably enriched in potassium relative to the granulite assemblages. The system is thus metasomatic and writing chemical reactions becomes more difficult. The following, then, must be considered as an idealisation that describes the observed petrography.

The two mafic minerals present in the granulite assemblage are considered to degenerate as follow:



Reaction 1) is written instead of the following

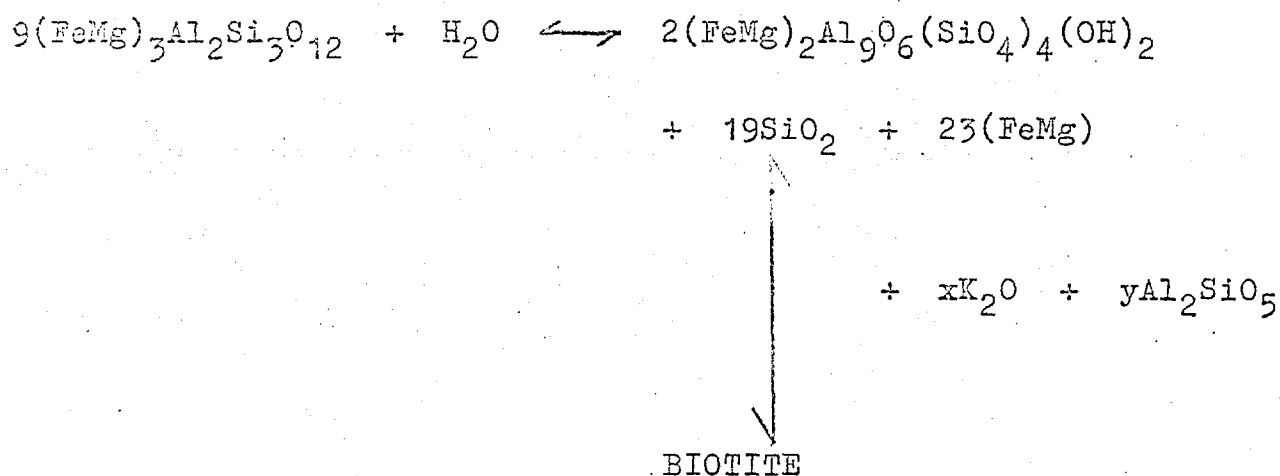


because analyses of the garnets (see table 6.4) show that they contain little of the grossular component. Garnet bearing assemblages are common in the Scourian granulites, and, except in the GK gneiss, these garnets contain about 20mol. percent grossular (Muecke 1969) and they would therefore degenerate according to reaction 3) above.

If garnet and hypersthene were present in equal amounts, then reaction 1) and 2) above would be coupled. However, garnet

occurs in excess of hypersthene, which is often absent from the granulite assemblage. The result of this is that there is a tendency for quartz to be consumed and for kyanite to appear as a reaction product. The modal analyses (table 6.2) clearly illustrate these trends. Under these conditions of silica consumption it is possible for a silica deficiency to arise within the biotite sheaths.

Staurolite producing reactions (see Richardson 1968) generally involve the production of quartz, and the following description allows the continued breakdown of garnet and formation of biotite. Staurolite production releases large amounts of quartz and thus lessens the degree of silica undersaturation.



There is no obvious textural relation between the staurolite and the garnet, and the reaction written above must remain an empirical description. In conjunction with the other reactions written above, it does indicate the probable evolution of these assemblages. Introduction of potash and water results in the appearance of biotite and kyanite, and a trend towards silica undersaturation is started. At a later stage, the formation of staurolite and quartz counteracts this silica undersaturation, at the same time permitting continued formation of biotite. The formation of staurolite will be accompanied by a

decrease in kyanite. The essential difference between samples 55 and 190 is the ratio of kyanite to staurolite (see table 6.2) and it is considered that sample 190 represents the later stage in the chemical evolution of these assemblages.

The corundum is always closely associated with staurolite, and its appearance is evidence for the development of silica undersaturation in certain domains of the rock, suggesting that the quartz consuming reactions proposed are correct in principle.

The scheme of reactions detailed above results from the introduction of potassium and water into the garnet bearing gneiss. The reactions therefore did not necessarily take place at equilibrium, and because of the lack of relevant thermodynamic data, no possible equilibrium states can be evaluated. However, the fact that kyanite and staurolite are individually stable phases places certain lower limits on the PT conditions during their formation.

The stability field of kyanite is still disputed, but a recent determination by Fyfe, based on a solubility method, is probably the most reasonable. This determination places the triple point in the system kyanite-andalusite-sillimanite at 2.5Kb and 450°C (Fyfe, personal communication).

Richardson (1968) has determined experimentally some invariant and univariant equilibria in the system Fe-Al-Si-O-OH. The stability field of staurolite determined by him is for quartz saturated mixtures only.

Richardson (1968) applies the Schreinemakers' construction (Schreinemakers 1965) to his experimental system. This construction is a method of determining invariant and univariant equilibria within a framework of two variables (these can either be pressure and temperature, or variation in the chemical potential of components such as water that are present in the system). Experimental determination of some of these equilibria enabled Richardson (1968) to fix the hypothetical equilibria in PT space. From this (see Richardson 1968, fig.4) it is apparent that reactions involving staurolite in quartz free

systems occur under much the same conditions as those in quartz saturated systems. Referring to the work of Richardson (1968), it can be seen that the absence of cordierite in the assemblages described here places a lower limit on the value of the fluid pressure of about 3Kb, and the presence of the minerals kyanite and staurolite indicates that the minimum values of fluid pressure and temperature were probably about 3.5Kb and 580°C respectively. The absence of chloritoid (produced from staurolite and almandine, see Richardson 1968, figs. 4,5) raises the minimum values of pressure and temperature to about 6Kb and 650°C.

6.6 Chemical analyses from biotised rocks.

Mineral analyses from biotised gneisses are presented in table 6.3, those from biotised hornblende pods in table 5.3 and those from the GK gneiss in table 6.4. A general discussion of these and other analyses is presented in chapter 8, and they are tabulated here for convenience. A number of coexisting pairs of hornblende and biotite have been analysed.

During a qualitative survey of biotite compositions, carried out with the aid of wavelength scans, significant quantities of barium were found in the biotites. The results of quantitative analyses for barium in a number of biotites are shown in table 6.5 (this table also shows the results of some manganese analyses that were carried out at the same time). Barium has not been found in any of the analysed phases that existed prior to biotisation and it is suggested that barium was introduced at the same time as potassium. A preliminary survey of whole rock samples of the GK gneiss by optical spectrography showed a real enrichment of barium in the shear zone samples relative to the unaltered granulite assemblages. The presence of barium is considered to characterise the assemblages resulting from Laxfordian metasomatism (in the area studied).

TABLE 6.3

BIOTISED GNEISS: MINERAL ANALYSES.

	SiO ₂	Al ₂ O ₃	TiO ₂	FeO	MgO	CaO	K ₂ O	total	mg
152H	45.0	10.7	0.8	15.1	12.2	11.6	0.5	95.9	44.8
152H	43.5	11.1	1.2	15.4	11.9	11.6	0.6	95.3	43.5
152H	44.3	11.4	1.0	15.7	11.8	11.7	0.6	96.5	42.8
180H	42.4	11.2	0.9	16.2	10.4	11.6	0.8	93.5	39.2
180H	43.0	10.9	0.9	16.5	10.6	11.5	0.8	94.2	39.0
180B	35.9	16.5	1.6	17.0	17.3	0.1	4.0	92.4	50.5
195H	39.6	12.2	1.3	18.8	9.1	11.6	1.2	93.8	32.7
195B	33.6	14.8	5.0	19.5	11.5	0.05	8.8	93.2	37.2
195B	36.0	15.0	5.1	19.8	11.5	0.05	8.8	96.2	36.7
195K	43.2	12.2	1.2	18.3	9.8	11.6	1.3	97.6	35.0
164B	35.7	17.5	2.7	17.4	12.4	0.05	9.0	94.7	41.6
164B	35.3	17.2	2.9	17.7	12.2	0.0	9.4	94.7	41.0
164H	40.2	13.1	1.4	19.4	8.7	11.5	1.1	95.4	31.0
164E	37.2	24.6	0.6	11.3	0.0	23.2	0.0	96.9	-
175B	34.6	16.1	4.9	19.9	9.4	0.0	9.5	94.4	32.1
142H	43.3	11.4	1.0	16.8	9.7	11.1	1.0	93.3	36.6
142H	42.8	11.1	1.0	17.4	10.0	11.2	1.0	94.5	36.6
142CB	32.4	16.8	1.5	19.7	15.6	0.3	3.4	89.7	44.2
142CB	33.2	17.0	1.2	19.8	15.8	0.3	3.4	90.7	44.4
201B	36.7	16.2	1.9	17.7	12.4	0.0	9.0	93.9	41.2
201B	37.0	16.4	1.9	16.0	13.3	0.05	9.2	93.8	45.2
201H	42.6	12.9	0.6	17.9	9.1	11.1	0.6	94.8	33.7
201H	43.2	12.7	0.9	17.6	9.5	10.9	0.8	95.6	35.1

TABLE 6.4.

GE GNEISS: MINERAL ANALYSES.

	SiO ₂	Al ₂ O ₃	TiO ₂	FeO	MnO	MgO	CaO	K ₂ O	total	mg
56G _c	38.6	21.4	0.2	26.7	0.8	10.1	1.8	0.0	99.6	27.4
56G _e	38.6	21.5	0.1	26.5	0.8	9.6	2.0	0.0	99.1	26.6
53B	34.6	17.9	4.1	18.3	nd	9.2	0.0	9.6	93.7	33.6
53B	34.8	18.2	3.8	20.1	nd	8.5	0.0	9.6	95.0	29.6
53B	34.6	17.7	4.1	18.4	nd	8.7	0.0	9.6	93.1	32.0
53M	45.7	32.1	1.4	2.4	nd	1.0	0.2	10.2	93.0	
53S	27.2	50.8	0.8	11.8	nd	1.6	0.0	0.0	92.2	

TABLE 6.5.

BARIUM AND MANGANESE ANALYSES OF SELECTED MINERALS.

<u>sample</u>	<u>BaO</u>	<u>MnO</u>	<u>mineral analysed</u>
12	1.44	nd	biotite single crystal
12	2.18	nd	biotite cluster
199	0.68	nd	biotite single crystal
199	1.02	nd	biotite single crystal
199	1.85	nd	biotite cluster
164	1.03	nd	biotite single crystal
164	2.10	nd	biotite single crystal
53	0.39	0.02	biotite in biotite-garnet band
53	0.40	0.01	biotite in plagioclase
53	0.35	0.01	biotite adjacent to muscovite
168	0.68	0.11	biotite against plag. and opaque
168	1.10	0.16	biotite within hornblende
168	0.03	0.24	hornblende adjacent to above biotite
175	1.29	0.20	biotite single crystal
175	1.24	0.22	biotite single crystal
175	1.67	0.19	biotite single crystal
17	0.48	0.10	biotite
17	0.60	0.09	biotite
17	0.60	0.11	biotite
17	0.00	0.23	hornblende
17	0.00	0.25	hornblende

TABLE 6.6.

GK. GNEISS WHOLE ROCK ANALYSES.

	52	53	55	56	171	173	174	194
SiO ₂	57.5	51.2	58.4	56.6	76.3	59.3	54.6	57.6
Al ₂ O ₃	19.4	20.1	18.8	17.1	11.3	16.0	16.3	16.0
Fe ₂ O ₃	4.6	4.9	2.8	1.9	0.6	3.8	3.1	2.7
FeO	5.1	7.3	5.9	8.9	4.9	7.7	11.5	9.2
MgO	2.6	3.8	3.1	4.3	1.6	4.0	4.6	4.3
CaO	2.2	3.0	4.1	4.5	2.0	3.7	4.1	3.1
Na ₂ O	4.2	3.4	3.9	3.7	2.2	2.6	2.4	3.9
K ₂ O	2.2	3.4	1.4	0.8	0.5	1.2	0.9	1.3
TiO ₂	1.1	1.7	1.0	1.0	0.1	0.9	1.2	1.0
MnO	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2
P ₂ O ₅	0.1	0.1	0.1	0.1	0.1	0.1	0.1	0.1
total	99.3	99.4	99.7	99.1	99.8	99.5	99.0	99.4
O.R.	64.4	57.5	56.3	29.8	19.0	49.5	35.0	37.5

CIPW. NORMS.

Qtz	14.2	3.1	13.9	9.7	51.0	21.7	14.0	11.1
Kfs	13.1	20.1	8.1	4.9	3.2	7.1	5.3	7.9
Plag	46.5	43.5	53.1	53.4	28.1	39.5	40.1	47.5
Hyp	10.4	16.4	14.8	24.2	12.7	19.8	28.5	23.9
Mag	6.6	7.2	4.1	2.7	0.8	5.5	4.5	4.0
Il	2.2	3.1	1.8	1.9	0.2	1.7	2.2	2.0
Ap	0.1	0.1	0.1	0.2	0.1	0.3	0.2	0.2
Cor	6.1	5.5	3.6	2.1	3.5	3.9	4.1	2.9

Professor Fyfe kindly obtained whole rock analyses of eight samples of the GK gneiss; the analyses were carried out by X-ray fluorescence at Manchester, and I would like to thank the analyst, Dr GC Brown, and Professor Fyfe for obtaining these results for me. The analyses (calculated on a dry basis) and the CIPW norms are shown in table 6.6. The significant points in this analysis set are as follows:

a) there is an increase in K₂O of up to 2.6%, and to a lesser degree in Na₂O, in the shear zone samples relative to the granulite samples;

b) the sequence 56-55-53-52 (ie. from granulite to the centre of a shear zone) shows a gradual decrease in the CaO content;

c) aluminium and titanium, the most immobile elements, show a slight enrichment in the shear zone samples;

d) the shear zone samples have the highest oxidation ratios, $2FE^{3+}/FE^{2+}+2FE^{3+}$.

Figs. 6.3A, B and 6.4 illustrate these correlations. Fig. 6.3A shows a plot of K₂O against Al₂O₃, and fig. 6.3B shows a plot of K₂O against TiO₂. These figures show that the slight enrichment of titanium and aluminium is indeed closely connected with the degree of metasomatism, this being conveniently indexed by the potash content of the rock. Similarly, fig. 6.4 shows a positive correlation between the oxidation state of the rock and the degree of metasomatism. This point will be returned to later in chapter 9.

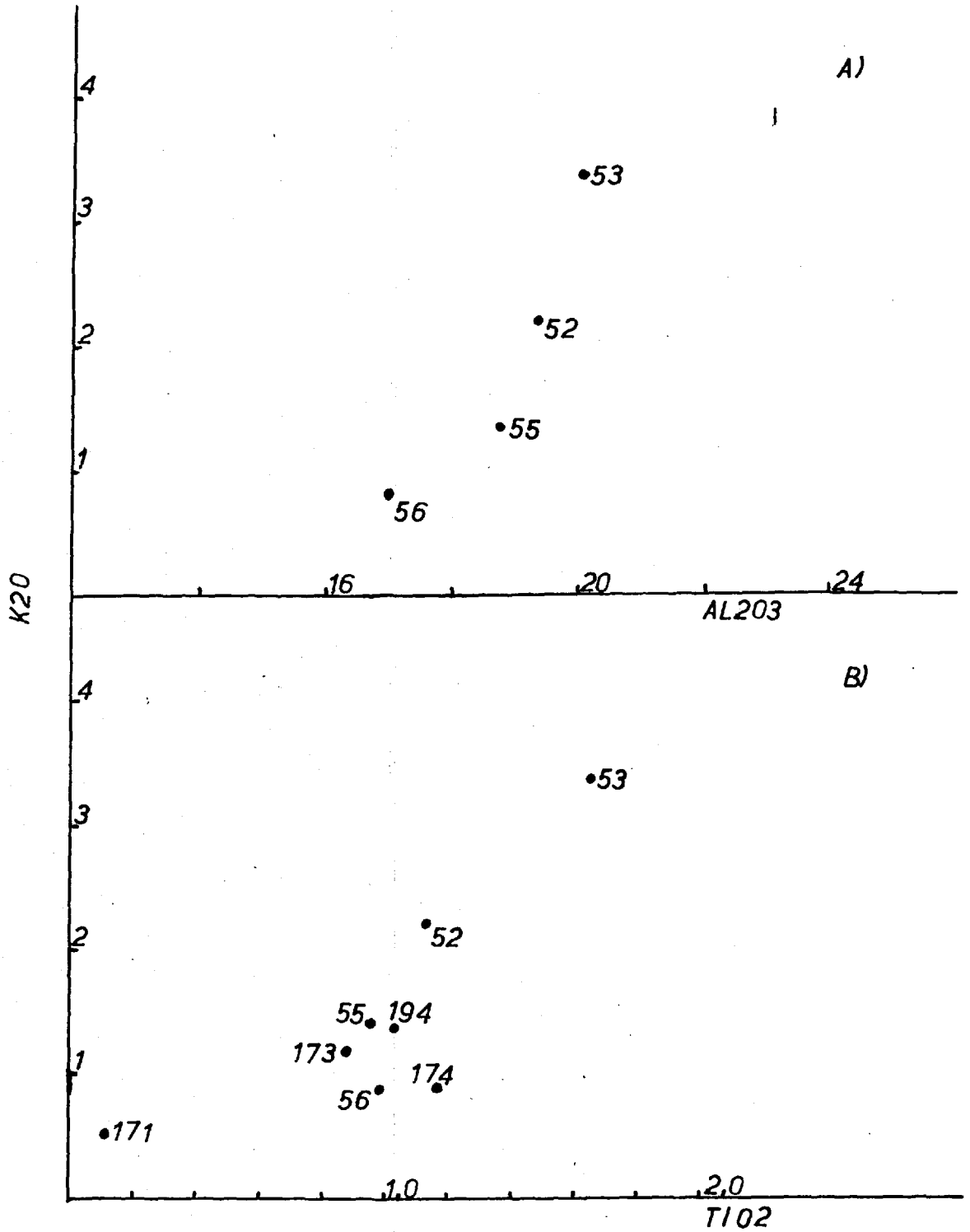
6.7 Shear zones cutting dolerite dykes.

Though the dolerite dykes are not well exposed in the area mapped at Scourie, shear zones are frequently seen crossing them. In particular, some shear zones can be traced from the gneiss into the dyke rock. When a shear zone containing large modal quantities of biotite is followed in this way it is found that there is an almost complete absence of biotite

GK GNEISS WHOLE ROCK ANALYSES

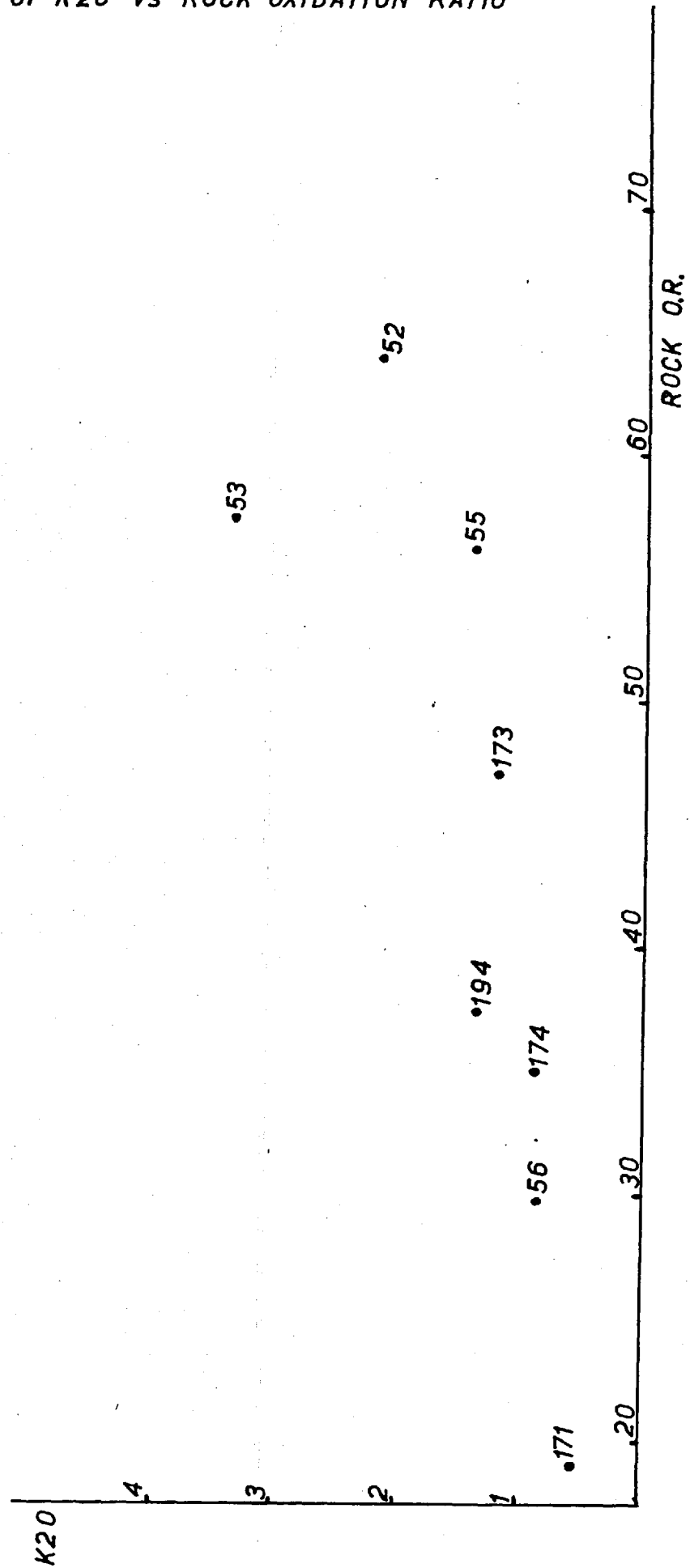
plots of A) K2O vs. AL2O3

B) K2O vs. TiO2



GK GNEISS WHOLE ROCK ANALYSES

plot of K2O vs ROCK OXIDATION RATIO



from the shear zone in the dyke. Burns (1966) presents whole rock analyses of many of the dykes in the Scourie-Laxford region, both deformed and undeformed dykes being represented. These analyses show that there is very little increase in potash content in the deformed dykes, and that the oxidation state of these rocks is no different from that of the granulites and amphibolised granulites.

The absence of biotite in the dyke shear zones is an enigma. Though the dolerites and amphibolised granulites have a very different bulk composition, they both contain hornblende. The chemical differences are thus expressed in different hornblende compositions in the two rock types (see O'Hara 1961 and Burns 1966 for mineral analyses from the dyke rocks). The dykes have hornblendes which are consistently more iron rich than the hornblendes in the amphibolised granulites. It is not clear why this compositional difference should inhibit the biotite forming reactions, but no other chemical differences between the minerals in the dykes and in the gneisses is apparent.

6.8 Discussion of the metasomatism.

From the data presented in this chapter it is concluded that the assemblages seen in the Laxfordian shear zones are the results of metasomatism. This metasomatism involved the introduction of water and potassium (and barium), and the loss of calcium. Silica metasomatism cannot be determined conclusively. It was the addition of water and potash, under the prevailing metamorphic conditions, that rendered pyroxene, garnet and hornblende in the rocks unstable. The change from hornblende to biotite has been considered in terms of Carmichael's (1970) ideal reaction, but in the biotisation of the GK gneiss it was found that addition of potash considered as K_2O was necessary to account for the minerals formed in these rocks (and in particular, for the silica undersaturation that developed). It has already been noted (section 3.3) that when potash

felspar^a is present in a Scourian assemblage, then this phase does not appear to be mobile. Thus the biotite forming reactions described here cannot be accurately evaluated in terms of equilibria involving potash felspar. This does not alter the essential nature of the reaction (ie. calcium producing), but only its thermodynamic interpretation.

The cessation of biotisation is assumed to have been due mainly to decline in the chemical potential of potassium in the metasomatic solution. This can be considered equivalent to saying that the potash in solution has been effectively used up and fixed within a mineral phase. In a system of ascending solutions, this will take place later at successively higher structural levels. The frequent coexistence of biotite and hornblende, and of garnet and biotite, could thus be not the result of establishment of equilibrium, but of cessation of reaction in a system of declining potash concentration. The exact point at which the reaction ceases in a particular rock will depend, among other things, on the composition of the minerals involved in the reaction. However, on a more general scale it should be possible to gain some knowledge of the original distribution of potassic solutions by plotting geographically a large number of whole rock determinations of potassium for assemblages containing biotite and hornblende. This has not been done in the present project, but a relevant observation is that the most potassic and the greatest number of shear zones occur around Sithean Mor. To the east, the number of shear zones rapidly declines, as does the importance of biotite in these zones. Northwards from Sithean Mor there is a similar but much more gradual decline. Thus already a possible three dimensional picture of shear zones is emerging.

If the metasomatic system proposed is to be accepted, then it is necessary to consider evidence for a source and a sink. That is, where did the water and potassium (and barium) come from, and where did the calcium go to? The complete evidence for discussions on source has not yet been presented. The fate of the excess calcium will be discussed here.

The presence of calcite in a few assemblages indicates that some carbon dioxide was present in the fluid phase during biotisation. The chemistry of the possible metasomatic fluid leaving the system could thus be considered in terms of CaO-H₂O-CO₂. It is well known that the solubility of calcite increases sympathetically with pressure, and that this is usually more than offset by the decreasing solubility with increasing temperature. Thus calcium metasomatism by solution of calcite is not a very viable proposition. However, the solubility of calcite must increase rapidly to the upper critical end point, where the mixture CaCO₃-H₂O becomes a melt. Complete experimental study of the melting relations of calcite at high water pressures is lacking, but the work of Wyllie and Tuttle (1959, 1960) should be noted. They showed that in the presence of water, calcite melts at 740°C when P_{H₂O}=1Kb, and at this pressure the minimum liquidus temperature in the system CaO-CaCO₃-H₂O is 675°C. At higher water pressures the minimum liquidus temperatures are depressed to 600°C or lower. In the system CaCO₃-H₂O-CO₂, where compositions include carbon dioxide, the effect of the additional CO₂ is to raise the liquidus temperatures by large amounts. Thus if melting of calcite is to occur, a fluid phase virtually free from CO₂ must be present.

The above data provide a means by which calcium may leave a system. Thus it is proposed that during the biotisation reactions, the calcium released from the hornblende never entered a mineral phase, but under the PT conditions prevailing at that time, formed a melt directly with the fluid phase. Even with some CO₂ present initially in the fluid phase, compositions would be unlikely to fall in the field H₂O-CO₂-CaCO₃, i.e. with excess CO₂, because the CO₂ present will be counterbalanced by the Ca released from the hornblende.

Calcite will crystallise from the melt when the temperature has dropped sufficiently, i.e. at higher structural levels. No evidence for this crystallisation has been found within the area studied, except for some calcite veins in the Tarbet Steep Belt. The age of these veins is not known.

6.9 Conclusion.

Biotisation was the result of introduction of potash and water. The chemical reactions occurred above the equilibrium of Carmichael's reaction (c.580°C at PH₂O=5Kb). The stability of aluminosilicates gives PH₂O=6Kb and T=650°C as minimum conditions during biotisation (and hence formation of the shear zones), while an upper limit of T=710°C at PH₂O=5Kb has been deduced from the stability of muscovite in some shear zones.

Loss of calcium in a melt phase is proposed, and this is compatible with the deduced PT conditions if the fluid phase is relatively poor in CO₂.

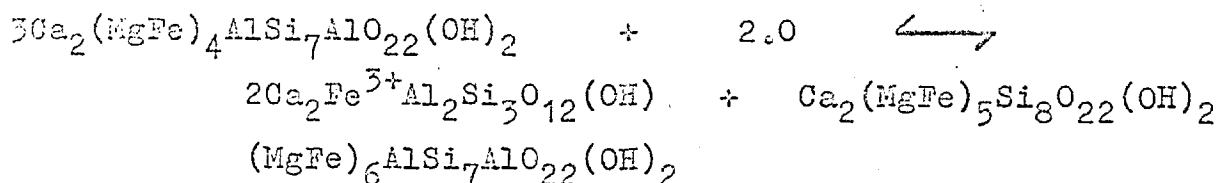
CHAPTER 7.

REACTIONS INVOLVING OXIDATION.

Consideration of oxidation in rocks is usually restricted to the oxidation of iron. The general principle of oxidation may therefore be stated as: iron bearing phases may be oxidised to an iron free phase plus iron oxide. The operation of this principle is well known in relation to biotite (see Eugster 1959), where the stability of biotite at different partial pressures of oxygen has been determined experimentally. This principle can be extended and the oxidation of hornblende considered; this also is an iron bearing hydrous phase. During the metasomatic events that have been outlined in the previous two chapters, there is evidence that both hornblende and biotite have been oxidised. Thus first of all the oxidation of hornblende during amphibolisation is briefly mentioned, and then the oxidation of hornblende and biotite in shear zones is discussed. Data from a suite of rocks collected in the Tarbet Steep Belt are then presented.

7.1 Oxidation during amphibolisation.

In section 5.5 it was mentioned that small quantities of epidote, tremolite and gedrite are sometimes associated with hornblende in the amphibolised gneisses (see for example sample 28, table 5.5). These minerals are considered to arise from minor oxidation of hornblende, and the following idealised scheme represents this reaction:



Analyses of the coexisting hornblende and tremolite in sample 28 are given in table 7.1, while the approximate formula of the gedrite in this assemblage is $Mg_4Fe_2Al_3Si_6O_{22}(OH)_2$ (Geoscan determination).

Since the above reaction is uncommon in the amphibolised rocks, it is concluded that breakdown of hornblende by oxidation during amphibolisation was not an important process. However, oxidation appears to have played an important role in the formation of hornblende from pyroxene. This topic has already been discussed in section 5.3.

7.2 Oxidation in shear zones.

Oxidation in shear zones affects both hornblende and biotite. In terms of the total mineral transformations seen in shear zones oxidation is a minor process, but it is important in view of its fairly consistent occurrence. The whole rock analyses presented in table 6.6 also show that the shear zone assemblages are oxidised relative to the unsheared assemblages. Three mineral associations are described here: biotite-potash feldspar, biotite-muscovite, and hornblende-tremolite-epidote.

A). Oxidation of hornblende.

Oxidation of hornblende is illustrated well by samples 123 and 126. Modal analyses of these samples are given in table 7.2, which shows that moderate quantities of tremolite and epidote are present in these assemblages. A similar reaction to that presented above might be thought to have ^{occurred} but two observed petrographic features point to a different interpretation. Firstly, gedrite is not found in the shear zones, and secondly, the volume ratio of epidote to new amphibole in sample 123 is approximately 1:1, whereas in the reaction written above it is approximately 1:2. The following reaction accounts for assemblage 123:

TABLE 7.1

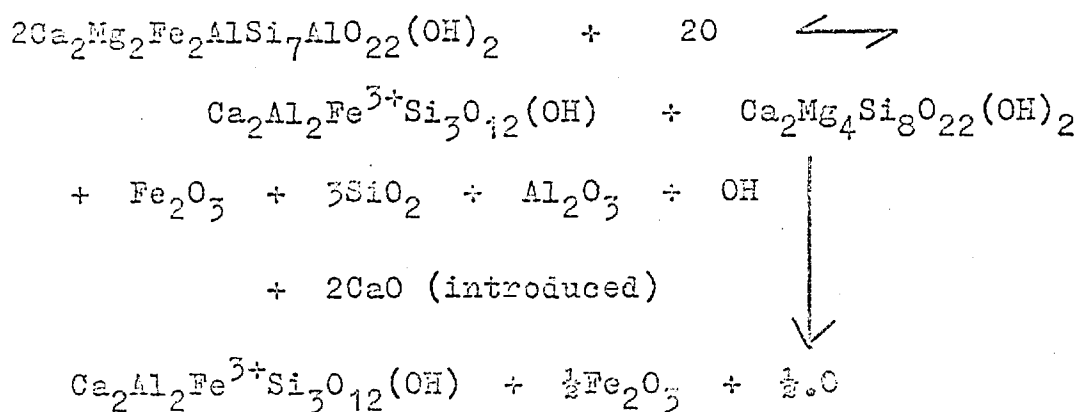
OXIDISED ASSEMBLAGES: MINERAL ANALYSES.

	SiO ₂	Al ₂ O ₃	TiO ₂	FeO	MgO	CaO	K ₂ O	total	mg
28H	41.1	12.2	1.0	17.1	9.2	11.6	0.8	93.0	35.0
28HT	53.4	2.5	0.1	9.0	18.0	12.4	0.05	95.4	66.5
123H	41.7	11.0	1.1	17.7	10.0	11.7	1.1	94.3	36.1
125HT	51.9	2.9	0.2	12.4	15.7	12.3	0.1	95.5	55.7
125H	43.1	11.2	0.9	17.9	10.0	11.5	1.2	95.8	35.7
123HT	54.1	2.3	0.2	12.4	16.7	12.4	0.1	98.2	57.3
123E	35.5	21.5	0.4	13.6	0.05	23.0	0.0	94.0	-
124H	43.8	10.4	1.1	15.0	12.4	11.5	0.7	94.9	45.3
124E	43.5	10.5	1.1	14.6	12.1	11.7	0.7	94.2	45.2
124H	44.4	10.5	0.9	14.8	11.9	11.5	0.6	94.6	44.4
126H	43.6	10.1	0.9	16.1	12.2	11.5	1.0	95.4	43.0
126H	44.6	9.6	0.8	16.3	12.1	11.4	0.9	95.7	42.7
126H	44.2	10.1	0.9	16.4	11.5	11.4	0.9	95.4	41.1
126E	38.4	23.2	0.4	11.1	0.1	22.5	0.0	96.7	-
127H	40.7	13.5	0.6	19.8	8.5	10.5	0.7	94.3	30.0
127H	41.0	13.1	0.6	19.5	8.7	10.4	0.7	94.0	30.9
127C	24.6	20.8	0.3	21.4	19.2	0.05	0.0	94.3	47.2

TABLE 7.2.

MODAL ANALYSES OF OXIDISED SHEAR ZONE.

	123	126
Plag	38.8	43.9
Qtz	7.1	11.5
Hbl	39.1	36.8
Trem	6.1	-
Epid	6.3	6.7
Biot	0.6	-
Op	1.4	tr
Rut	0.5	1.1
Ap	tr	tr
	7000	3500
counts		
An%	28	nd



In sample 123 haematite coexists with a titanium bearing oxide phase. Mineral analyses from samples 123 and 126 are given in table 7.1. The hornblende and tremolite coexist across optically sharp interfaces.

Two points need explaining concerning the reaction presented above. Firstly, both modes given in table 7.2 contain epidote, but only 123 contains tremolite. It is suggested that if the pair tremolite-hornblende recrystallised there would be a tendency for the much greater quantity of hornblende to resorb the lesser quantity of tremolite. If recrystallisation is the cause of the non-appearance of tremolite in sample 126, and this recrystallisation is coeval with the deformation, then the evidence suggests that oxidation was not entirely post-tectonic. There is also evidence that oxidation did not occur throughout the shear zone. For example, samples 127 and 128, collected from the same shear zone, contain no epidote. There is a small amount of well crystallised chlorite in sample 127. Sample 124 is from the same locality but shows only Scourian mineral fabrics. Mineral analyses from samples 124 and 127 are shown in table 7.1. There is no obvious difference between the compositions of the hornblendes in sample 124 and sample 127 that might be used to distinguish between Scourian hornblendes and the same hornblendes recrystallised in the Laxfordian shear zones (cf. section 8.4).

The second point in this discussion is that calcium was introduced into the reaction written above. It will also be noticed that though these assemblages are from a shear zone, they contain no biotite or other potassium bearing phase.

The calcium was introduced into the reaction so that the reaction described the observed modal ratio of epidote to tremolite. It will be recalled that in chapter 6 the loss of calcium from biotised shear zones was proposed, and that at a certain level the metasomatic solutions will be depleted in potassium, and biotisation will cease. It is consistent with this idea that samples 123 etc. were collected at a level in the shear zone where biotisation had ceased and where a little of the excess calcium released was fixed in epidote. If calcium had not been available then a reaction similar to that proposed in section 7.1 would be expected to have occurred during the oxidation of hornblende, with gedrite occurring as a reaction product. The cause of the reaction is thus not the availability of calcium but the effect of the partial pressure of oxygen.

B). Oxidation of biotite.

The other phase that is seen to be oxidised in the shear zones is biotite. Bearing in mind the general principle of oxidation, that iron free silicates and iron oxides are produced, two different phases can result from the oxidation of biotite. These are potassium feldspar and muscovite.

In many of the biotite bearing assemblages from the shear zones, small amounts of potash feldspar are seen to replace biotite. The replacement occurs along the edge of the biotite flakes and the original biotite grain boundary can still be seen. Though this feature is fairly common, the total amount of potash feldspar produced must be virtually negligible. Haematite is rare in these assemblages, the opaque mineral being an ilmenite (of unspecified composition). The equilibrium of the reaction

$$\text{annite} \rightleftharpoons \text{sanidine} + \text{magnetite}$$

has been studied experimentally (Eugster 1959). Extrapolation of his data to geologic systems must be treated with caution, but Eugster suggests that the experimentally determined results would not be very different from those that would be obtained from naturally occurring assemblages. At 600°C and $P_{H_2O}=4\text{Kb}$, partial pressures of oxygen of the order of 10^{-17} bars are necessary to oxidise annite to sanidine plus magnetite.

Oxidation of biotite^z to muscovite is not as widespread as oxidation to potash feldspar, and has only been recorded from the GK gneiss (see mode 53P, table 6.2). This reaction must be complex since the phases biotite-muscovite-ilmenite/haematite cannot possibly balance, and this is undoubtedly why muscovite has a limited occurrence in the shear zones studied.

Chinner (1960) was able to show that in a series of pelitic Dalradian rocks modal muscovite and haematite increased as modal biotite and garnet decreased, and thus he wrote a balancing reaction in terms of these phases.

In the GK gneiss it is not possible to tell if the phases garnet or kyanite are involved in the reaction since there is no textural relation between these phases and the muscovite. This suggests that oxidation occurred during continued recrystallisation of the assemblage. If garnet is involved in the reaction the general absence of this phase from most of the Scourian gneisses would explain the absence of muscovite as a product of the oxidation of biotite.

7.3 Mineral assemblages in the Tarbet Steep Belt.

The Tarbet Steep Belt was defined as the steep limb of a major Scourian fold in chapter 2. Judging from the mineral fabrics in this belt it is a zone of complete Laxfordian recrystallisation (excepting certain ultra-basic and basic bodies, see Sutton and Watson 1961), and there is no evidence for the present existence of minerals formed during the Scourian. A series of samples were collected to investigate the mineralogy of the southern part of this belt. The samples are from a coastal traverse starting south of the Tarbet Antiform in flat lying amphibolised granulites, and ending at Port Mor. Modal analyses of this suite are presented in table 7.3. Just north of Port Mor, the general character of the gneisses changes and from here northwards igneous bodies and brown gneisses are abundant (cf chapter 1). There is no evidence to suggest that the suite of rocks described here are other than the regenerated equivalents of the Scourian gneiss type seen in its unaltered form at Scourie.

TABLE 7.3.

TARBET SUITE: MODAL ANALYSES.

	42	43	44	45	46	47	199	100	102	103	202
Plag	37.4	45.0	55.6	48.3	57.4	42.3	48.6	49.5	45.5	47.0	9.7
Pap		10.8									
Kfsp						7.0					
Qtz	6.4	22.8	19.1	10.0	25.8	34.5	27.6	28.4	38.1	15.3	
Hbl	54.2	12.8	16.0	30.4	4.5	tr		12.8		26.5	48.7
Biot		8.0	7.5	8.6	11.7	4.4	18.2	7.5	13.8		31.9
Musc						11.6					
Epid							4.7	0.2	2.1	4.4	2.4
Op	1.8	0.2	1.4	2.4	0.5	0.2	0.9	0.9	0.5	0.6	0.6
Haem	0.3		tr	tr							
Rut	tr		tr	tr	tr	tr					
Ap	tr	0.2	0.4	0.3	0.3		tr	tr	tr	0.4	0.2
Cal		0.2		tr						2.2	4.3
Chl						tr	tr	0.7		2.9	
Sph										1.7	2.2

counts 6600 7400 5000 5200 4700 4900 5300 4700 3600 4700 7800

An% 38 28 31 34 25 4 24 25 28 32 30

TABLE 7.4.

TARBET STEEP BELT: MINERAL ANALYSES.

	SiO ₂	Al ₂ O ₃	TiO ₂	FeO	MgO	MnO	CaO	K ₂ O	total	mg
199B	35.3	17.1	2.8	18.1	12.7	0.3	0.05	9.4	95.7	41.3
199B	34.5	16.4	3.5	18.3	13.1	0.3	0.1	8.4	94.6	41.7
199E	36.8	24.2	0.5	11.0	0.05	0.2	22.8	0.0	95.5	-
100H	42.9	11.8	1.3	17.3	10.7	0.3	11.9	0.9	97.1	38.3
100H	43.6	11.8	1.5	17.3	10.7	0.3	12.1	0.9	98.2	38.2
100B	35.2	17.1	2.5	16.3	14.0	0.1	0.05	8.3	93.5	46.2
100E	36.4	17.1	2.5	17.1	14.1	0.2	0.05	8.3	95.7	45.1
100C	24.6	21.6	0.3	19.6	21.1	0.2	0.0	0.0	87.4	45.1
103H	40.5	11.6	1.3	19.0	9.5	nd	11.2	1.1	94.2	33.3
103H	40.6	11.5	1.2	18.6	9.8	nd	11.6	1.1	94.4	34.6
103H	41.7	11.4	1.2	18.7	9.8	nd	11.6	1.1	95.5	34.3
103E	36.6	24.2	0.5	11.9	0.0	nd	22.3	0.0	95.5	-
202H	42.6	12.7	0.6	13.1	13.1	nd	13.0	1.3	96.4	50.0
202H	44.8	11.3	0.5	12.6	13.6	nd	13.0	1.1	96.9	51.9
202B	39.5	16.4	1.5	12.6	16.3	nd	0.0	9.5	95.9	56.4
202B	39.2	16.7	1.6	12.5	17.5	nd	0.0	9.2	96.7	58.3
202SPH	30.2	3.0	36.7	0.6	0.0	nd	27.8	0.0	98.3	-

Samples 42-46 show only incomplete Laxfordian recrystallisation, as suggested by the presence of granular hornblendes containing small quartz grains, and by the presence of blebs of potash feldspar in some of the plagioclase grains (a feature of Scourian plagioclases) (see chapter 5). Samples 199 and 100-103 show complete Laxfordian recrystallisation and were collected north of the Tarbet Antiform. Sample 47 is an example of the concordant Scourian granite sheets and is included to show the continued identity of these assemblages after Laxfordian recrystallisation (see Muecke 1969 for the original work on these Scourian granite sheets).

It is apparent that the amount of biotite in these assemblages varies widely and assemblages without biotite (eg sample 103) are not uncommon. The biotite is generally pleochroic from dark olive green to very pale brown, and frequently contains opaque inclusions (these inclusions are not entered under 'Op' in table 7.3). The pleochroism of these biotites differentiates them from the biotites in the shear zones to the south, where they are pleochroic in shades of brown and red brown. Dear et al (1966) suggest that the green colour of biotite may be due to the presence of ferric iron.

In addition the Laxfordian assemblages are marked by the presence of epidote and occasionally of large grains of well crystallised chlorite. Coexisting hornblende and tremolite are sometimes seen, though the latter phase has probably been more generally homogenised into the hornblende by recrystallisation in many samples. It is suggested that the assemblages from the Tarbet Steep Belt show the effects of Laxfordian oxidation similar to those described in section 7.2 above.

Analyses of minerals from the Tarbet Steep Belt are given in table 7.4.

7.4 A note on opaque minerals.

Sulphide minerals (usually pyrite) are sometimes present in the Scourian granulites, but these were altered to or rimmed by ilmenite during amphibolisation. It is then possible to

consider the opaque phases in terms of the oxide compositional field $\text{FeO}-\text{Fe}_2\text{O}_3-\text{TiO}_2$. Though detailed oxide identifications have not been made, and no quantitative analyses are available, two broad oxide associations have been recognised. These are:

- a) ilmenite or ilmenite and rutile in the amphibolised Scourian assemblages,
- b) ilmenite (magnetite) and haematite in the Laxfordian assemblages.

These two associations would represent different oxidation trends and would not be buffered, except for the possible assemblages containing the oxides magnetite-haematite, to changes in partial pressure of oxygen.

The detailed study of rock and mineral oxidation should be a topic for future geochemical research.

CHAPTER 8.

A DISCUSSION OF THE MINERAL ANALYSES.

Mineral compositions in metamorphic rocks are usually considered in terms of five variables, rock composition, total pressure, temperature, water fugacity and Oxygen fugacity. In the case of the chemically complex minerals hornblende and biotite it is often difficult to evaluate separately the effect of each variable on mineral composition. This problem can be overcome to a certain extent by studying a specific rock type along one defined isograd of progressive metamorphism (eg. Hounslow and Moore 1967, Carmichael 1970). Such studies, to be complete, would entail analysis of whole rocks, individual minerals and opaque phases. In the absence of works as comprehensive as this interpretation of chemical data is done mainly by analogy with other analytical results.

In the present study, the time available has permitted only mineral analyses to be carried out. By discussing all the analyses in this chapter it is possible to compare minerals sampled from different tectonic structures and to attempt in particular to differentiate between metamorphisms associated with the Scourian and Laxfordian structures.

8.1 A note towards the definition of a metamorphic system.

Perhaps the greatest significance of the mineral analyses presented here arises from the fact that the minerals are from rocks that were initially high grade metamorphic rocks (granulite facies) and that have been subsequently remetamorphosed under different conditions. These metamorphic systems would thus not be included in the general category of progressive metamorphic systems on which the bulk of research in metamorphic petrology

has been concentrated.

A number of concepts that have arisen from the study of progressive metamorphic systems have often been thought to characterise all metamorphic systems. Significant ideas are contained in the terms 'equilibrium', 'isograd' and 'closed system'. These concepts are apparently applicable during progressive metamorphism, and for this reason alone progressive metamorphism must be considered as a special case of the metamorphic system.

The attainment of equilibrium between mineral phases in a rock has been judged in two ways a) that of physical equilibrium (Kretz 1966, 1969, Vernon 1968, 1970, Flinn 1969), and b) that of chemical equilibrium (Kretz 1959, Saxena 1969). The two approaches differ in the theoretical basis for the concept of equilibrium in each; in a natural rock it is realised that they are two ways of describing the same phenomenon.

A closed system is defined as one which neither loses nor gains matter from other external systems. A geologic system will in general be more likely to reach equilibrium if it is closed. Petrologic studies have shown that many progressive metamorphic systems can be considered to have been closed with respect to all components except a fluid phase consisting mainly of water.

In a system that is closed and at equilibrium it is possible to define isograds and to evaluate them thermodynamically. An isograd defines a particular reaction between mineral phases and is fixed in PT space. The assumption of equilibrium precludes the possibility that the position of the isograd mapped out in a metamorphic terrain could be partly determined by the rate of the mineral reaction.

Once a system is recognised as having been open the concepts of equilibrium and of the isograd are no longer immediately applicable. During a metasomatic event the possibility arises of continued mineral reaction due to the introduction of material. The areal extent of different assemblages will then reflect the extent of metasomatism, and will not in general

define isograds in the accepted sense. For example, in an amphibolised granulite facies terrain the coexistence of pyroxene and hornblende defines only the quantity of water that has been fixed in the system. The two minerals may or may not have been in equilibrium in a particular sample, but this equilibrium does not help to define P_{H_2O} or T during the recrystallisation of this sample.

It has been shown that the systems studied here were open systems during metamorphism. Interpretations of the mineral analyses must therefore be made with caution. Topics discussed in the following sections include the possible control of rock oxidation state on mineral compositions and the relation between hornblende and the biotite derived from it.

8.2 The relation between rock oxidation and mineral chemistry.

Chinner (1960) describes the effect of differing rock oxidation states on the composition of minerals in a series of pelitic schists from the Dalradian. Similarly Hounslow and Moore (1967) have studied the rock and mineral compositions of a number of Grenville schists from the staurolite zone. Three general statements can be made from these works:

- a) the compositions of minerals (garnet and biotite) do not correlate with the bulk rock composition, but they do correlate with the rock oxidation ratio,
- b) with increasing rock oxidation ratio, the $Mg/Mg+Fe$ ratio of biotite increases,
- c) with increasing rock oxidation ratio the absolute amount of titanium in biotite decreases.

If the Mg ratio of biotite is related to the rock oxidation ratio then the latter becomes an important compositional variable. The changing titanium contents could reflect a changing ease of substitution in the lattice as the amount of iron decreases. It could also be due (see Chinner 1960, p201) to preservation of a constant titanium distribution between the biotite and the oxide phase, the latter increasing its Fe/Ti ratio with increasing oxidation. If this second alternative is

correct then the titanium content of biotite could provide a means of estimating relative oxidation states in a suite of rocks.

Figs. 8.1 and 8.2 show plots of TiO_2 against mg ratio for biotites and hornblendes respectively, while fig. 8.3 is a similar plot for coexisting hornblendes and biotites only. (In these and other figures analyses from rocks of extreme composition such as the hornblende pods have been excluded). It is evident that there is a strong correlation between the parameters plotted. Since the assemblages analysed all contain an oxide phase (ilmenite) the hornblende and biotite may, at first, be considered to be saturated in titanium, and it remains to deduce the control on saturation level.

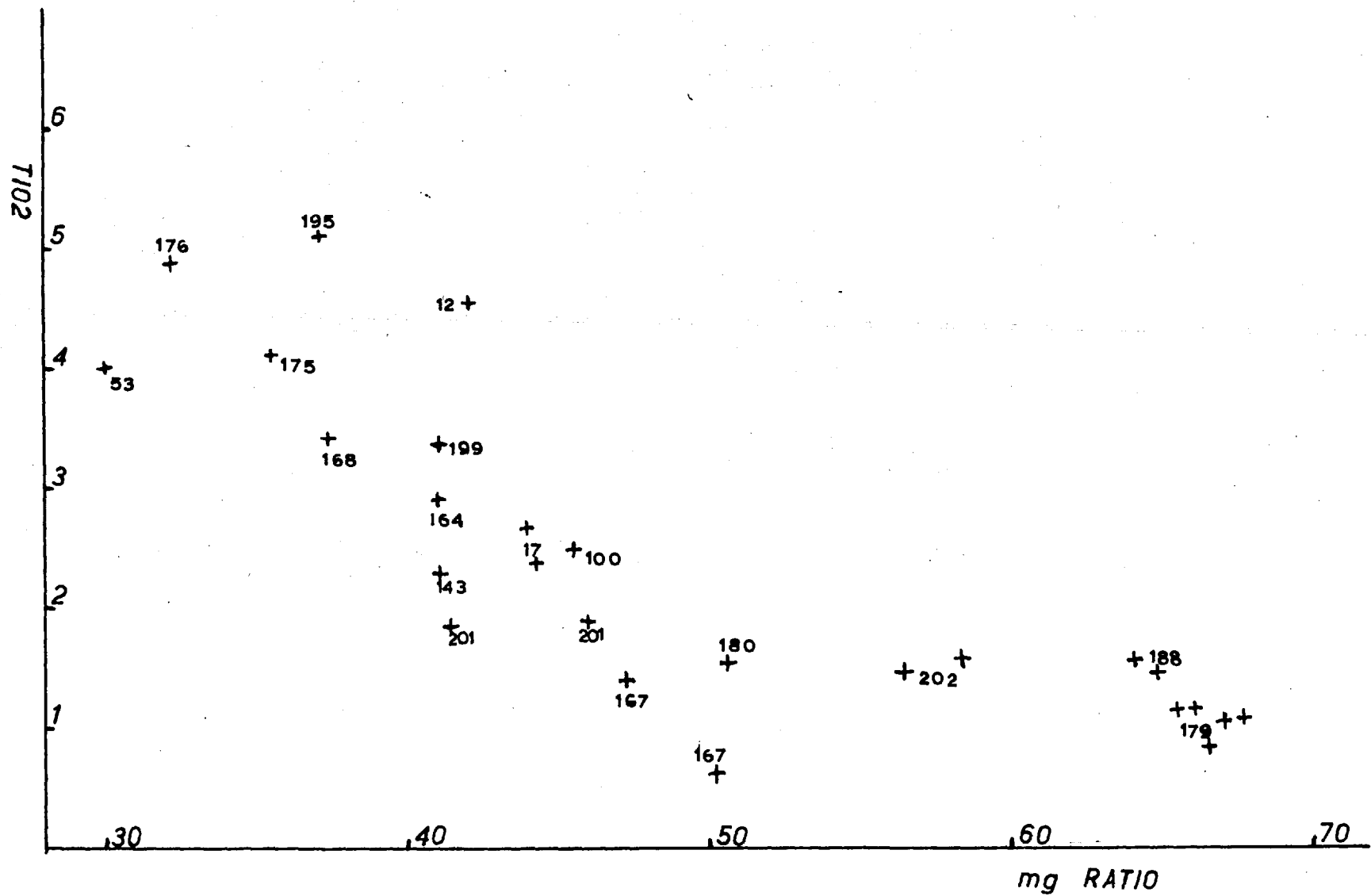
8.3 The significance of potassium and titanium in biotite and hornblende.

Without whole rock analyses to complement the mineral analyses made in this study it is not possible to say anything definite about the oxidation state in different assemblages.

If the major element content in the minerals is controlled to some extent by the host rock chemistry then these parameters are of little use in determining trends that reflect the physical variables PT during metamorphism. The possibility that elements present in small amounts in minerals may be present to saturation level, and the concentrations therefore independent of rock chemistry, has to be investigated. Elements that can be usefully considered in this category include Ca, Mn in garnet, Ti, Na in biotite, Ti, K, Na in hornblende, Ti, Fe, Na in muscovite. In the present study only data for Ti and K in hornblende and biotite are available.

Leake (1965) considered that the amount of titanium contained in a calciferous amphibole increased as the temperature of crystallisation increased. This statement is valid only as a generalisation when igneous and metamorphic amphiboles are being compared (see Leake 1965 figs. 1, 2). Leake (1965) points out that the hornblendes considered should come from biotite

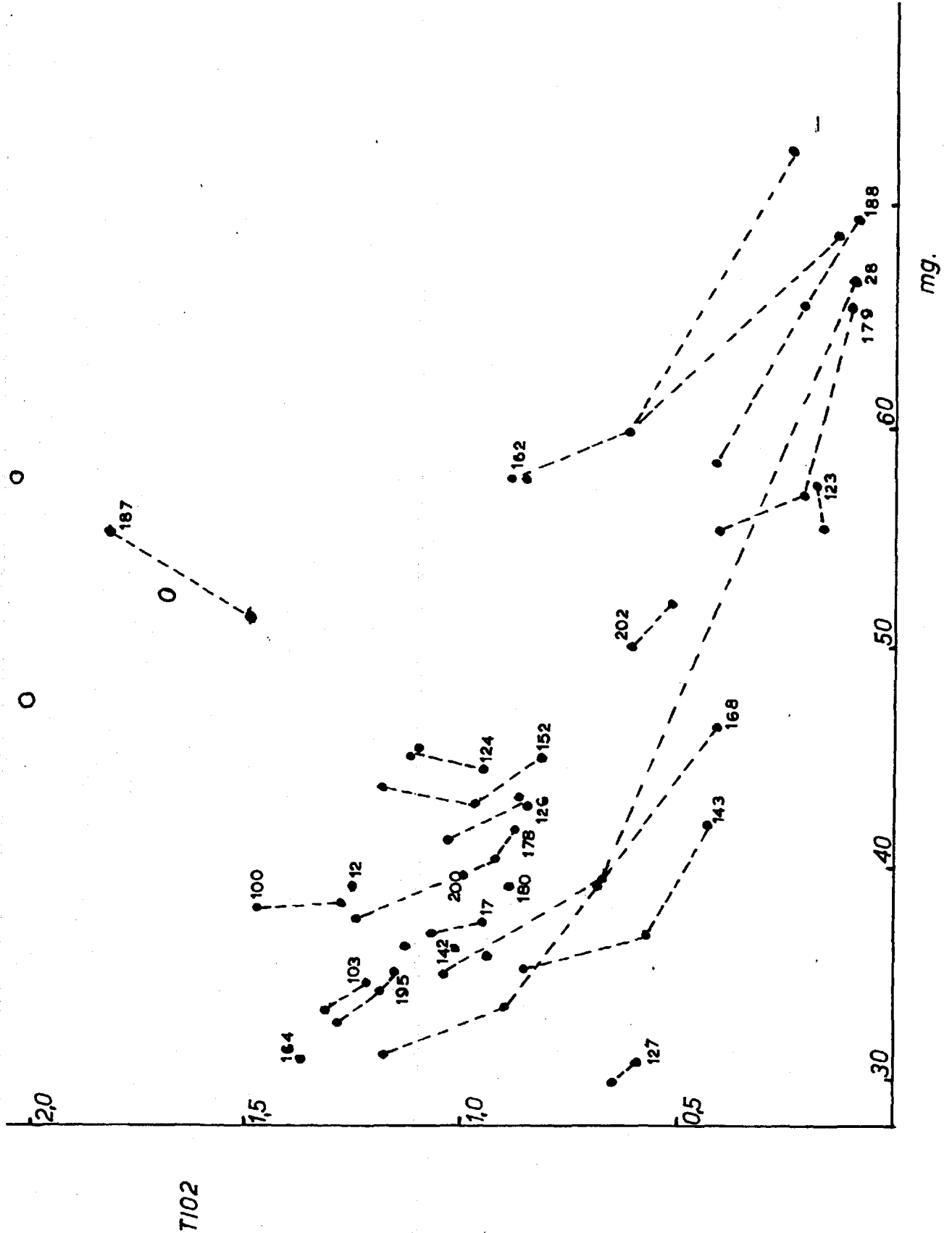
BIOTITE ANALYSES
Plot of T102 vs. mg



HORNBLende ANALYSES

plot of T102 vs. mg

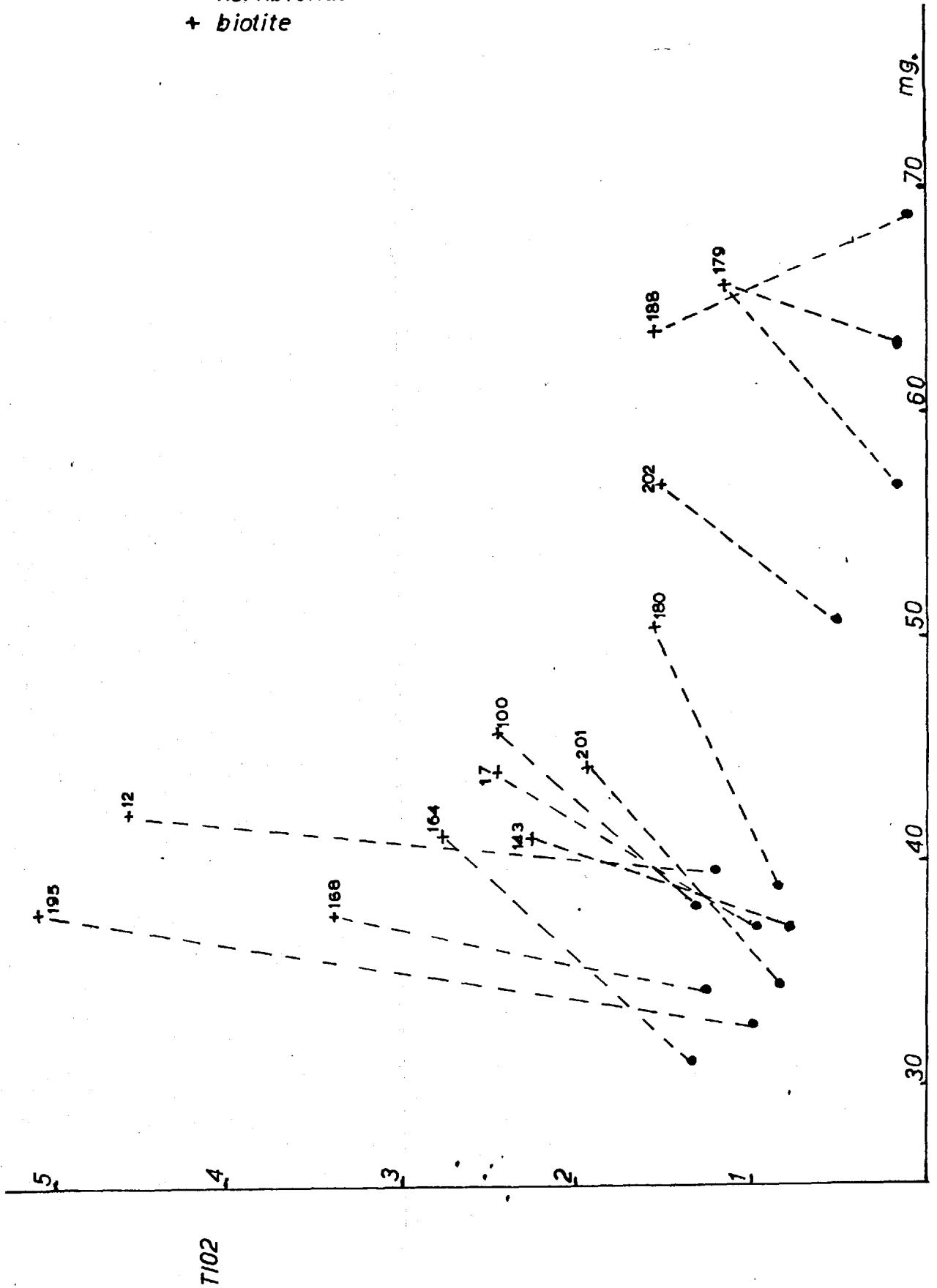
o Muecke 1969



HORNBLENDE AND BIOTITE ANALYSES

plot of T102 vs mg for coexisting pairs

- hornblende
- + biotite



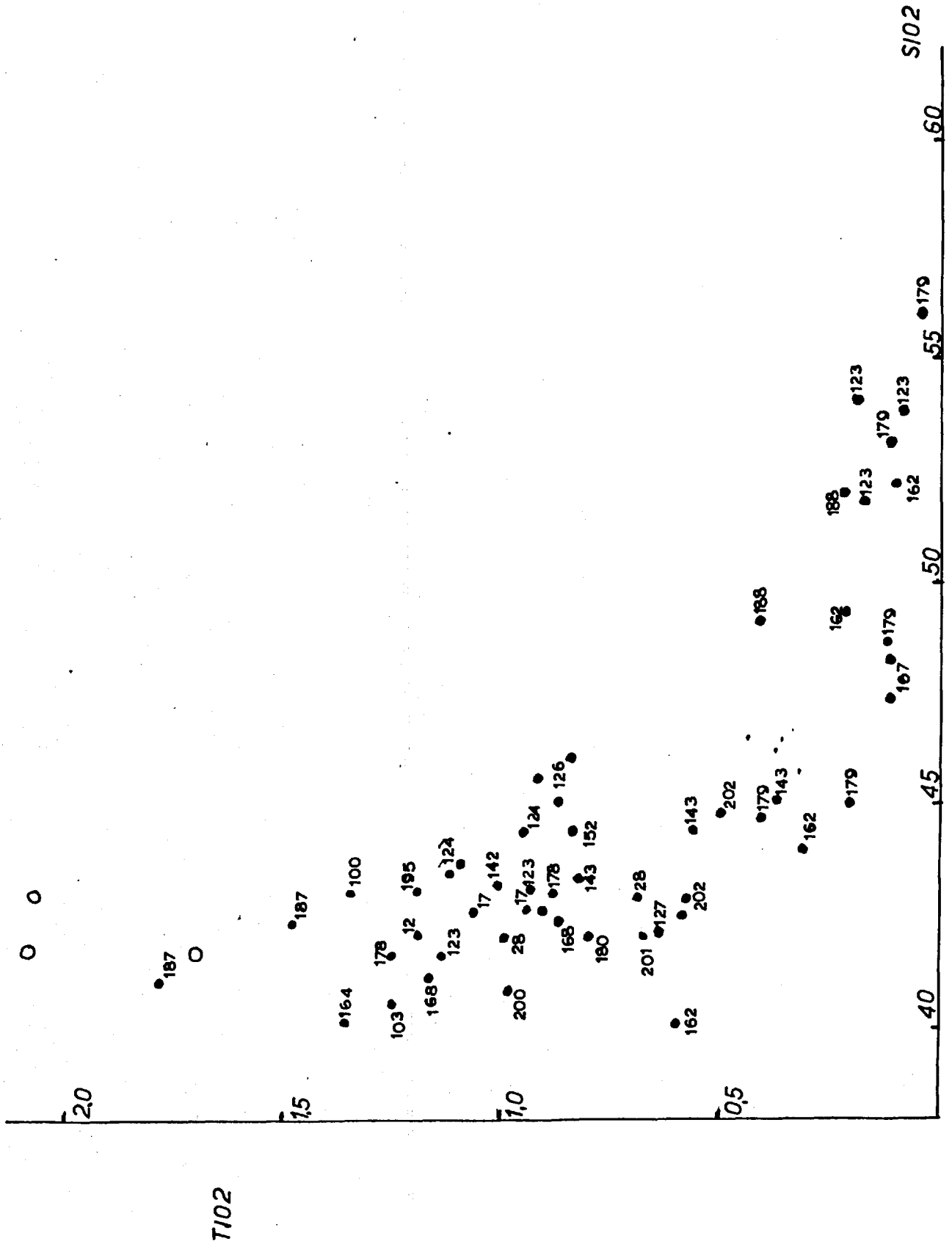
free assemblages since titanium is partitioned favourably into biotite. The assemblages should also contain ilmenite, the assumption being that the hornblende will then be saturated in titanium. Binns bases an objection to this premise on some histograms (Binns 1969b fig.1), which show high central tendencies for Ti contents of hornblendes but also show pronounced tails towards the higher Ti side of each histogram. It is necessary to consider the possibility that titanium will be partitioned between the hornblende and oxide phases and that with a limited amount of titanium in a rock the modal quantity of oxide present is a critical factor in determining the titanium content of the coexisting hornblende. This objection is perhaps overruled by Leake's (1965) observation that there is no correlation between the titanium content of the rock and of the hornblende in a suite of samples from the Adirondacks. Adequate solution of this problem requires that whole rock, mineral and modal analyses are available for the same suite of rocks. Oxidation will affect the composition of the oxide phase and the titanium content of the hornblende may be expected to change in much the same way as the titanium content of biotite changes with oxidation state (see Chinner 1960). It will be realised that when titanium is distributed between three phases - hornblende, biotite, oxide - no direct information on PT conditions of crystallisation can be gained from analytical data.

Despite the problems discussed above, fig.8.4 shows a plot of TiO₂ against SiO₂ (cf. Leake 1965) for all the hornblendes analysed. Also shown are some analyses of granulite facies hornblendes taken from Evans (1963) and Muecke (1969). It is evident that there is a correlation, but there is no clear separation between the hornblendes from samples with Scourian fabrics and those from samples with Laxfordian fabrics. The hornblendes from the granulite facies contain the greatest amount of titanium and this may be consistent with their having crystallised at higher temperatures.

The significance of the potassium content of hornblende is not clear. The fact that potassium must be distributed between

HORNBLLENDE ANALYSES

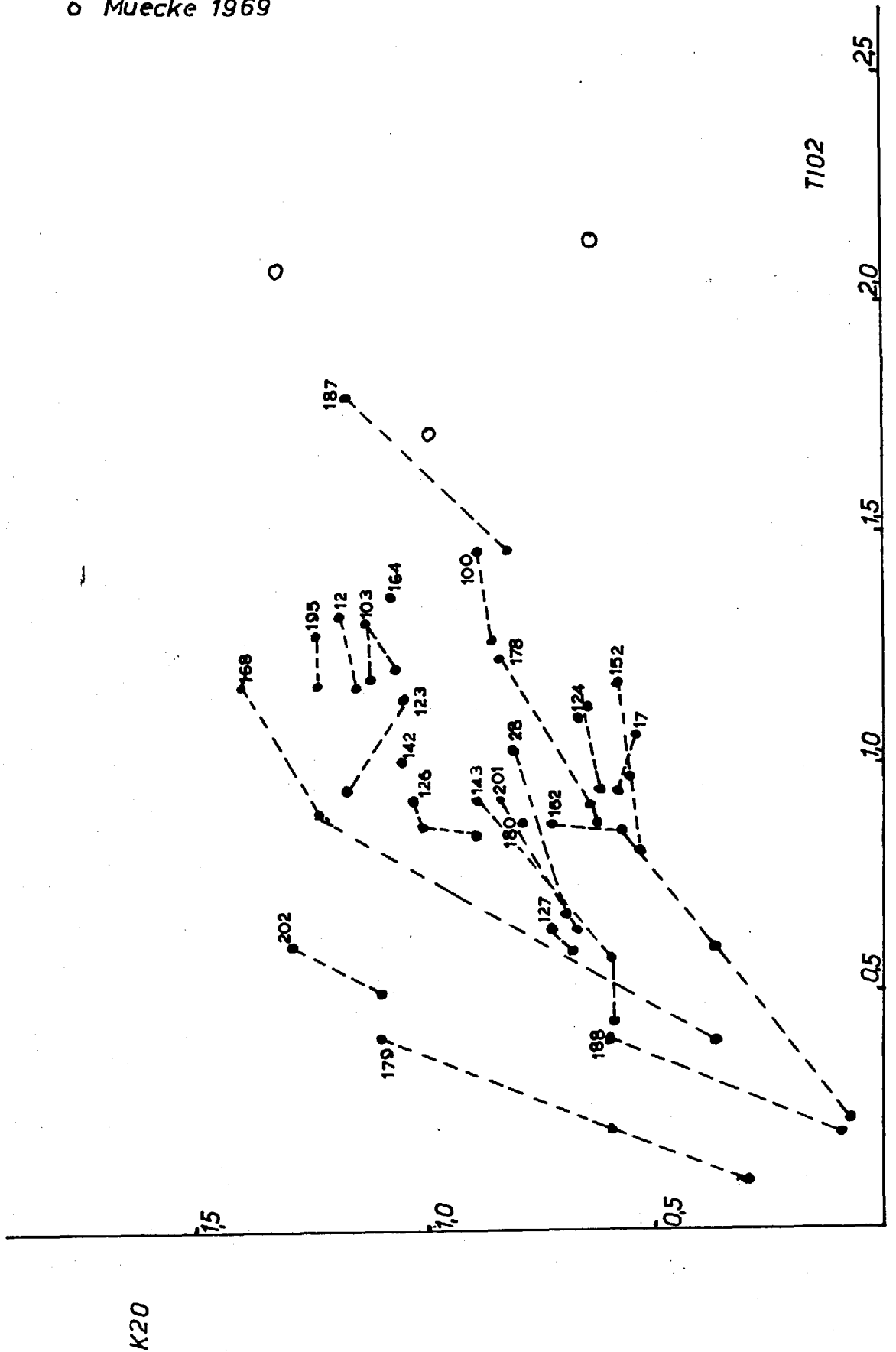
plot of T102 vs. S102
o Muecke 1969



HORNBLLENDE ANALYSES

plot of K20 vs. T102

o Muecke 1969



potash feldspar, biotite, muscovite and hornblende in many metamorphic rocks renders the situation as complex as that encountered above when the distribution of titanium was considered. In the present study the only potassium bearing phases in the assemblages are biotite and hornblende; the potassium contents of the plagioclase was found to be very low (in general below the limit of accurate detection of the Geoscan). In the biotite potassium is a major component, while in hornblende it is a minor component. Thus the possibility arises that the distribution ratio between the two phases is partly determined by the bulk composition of the rock, rather than TP conditions alone. A plot of K₂O in hornblende against K₂O in biotite showed no consistent pattern.

Fig.8.5 is a plot of K₂O against TiO₂ for all hornblendes and has been included because the tie lines, joining analyses from the same assemblage, show a fairly consistent trend. The significance of this is not known, but the varying K₂O and TiO₂ contents of the hornblendes from one assemblage is indicative of a lack of attainment of equilibrium in these assemblages.

8.4 A statistical study of hornblende compositions.

It is considered that sufficient analyses of aluminous hornblendes have been made to make a statistical treatment of populations meaningful. In all, forty analyses of aluminous hornblendes have been presented, twenty four of these coming from rocks with Laxfordian fabrics, and the remaining sixteen from rocks with Scourian fabrics. These groups are named S+L, L, and S respectively. Analyses from the ultramafic hornblende pods and analyses of the low alumina, or tremolitic, hornblendes have been excluded, since these form obvious distinct compositional groups.

The analytical data, input in three groups - S+L, L, S - were run with the computer programme 'Gunngeocor'. This programme outputs graphs of element pairs plotted in terms of standardised coefficients (ie. after this transformation, the

mean of each variable is zero and its standard deviation is unity). This procedure permits easy visual appreciation of correlations that may exist between pairs of elements.

Information gained from this work falls into two groups:

- a) parameters of central tendency, which show the characters of one population and enable different populations to be compared
- b) parameters which measure the correlations between elements in one population.

Table 8.1 presents some of these results. In the first part of the table are listed the arithmetic means (M), the standard deviation (SD), the coefficients of variation (CV), for each element oxide and the mg ratio in each of the data groups defined above; also given is the standard error of difference between groups S and L (SED). The second part of the table is a correlation matrix for oxide pairs in the group S+L. In this matrix a figure of ± 1 indicates that a perfect functional relationship exists between the oxide pair, while a figure of 0 indicates a complete lack of correlation. With a population sample of forty figures greater than 0.8 can be considered significant, while figures in the range 0.6-0.8 can probably be assigned some meaning.

Considering the parameters of central tendency it can be seen that there is very little difference in composition between the Scourian and Laxfordian hornblendes. The only arithmetic means that lie outside the standard error of difference between groups S and L are those for FeO, K₂O and mg. However these later are only just outside the range of the standard error of difference and it would be safer to assume at present that groups S and L belong statistically to the same population. This conclusion is interesting because of the large mineralogical differences that exist between the Scourian and Laxfordian assemblages. It might be expected that during the formation of biotite from hornblende and during recrystallisation of both phases the composition of the hornblende would change, mainly through the establishment of new partition coefficients between the hornblende and the new phase biotite. Such an effect is not apparent. It can only be concluded that the interaction

TABLE 8.1

MEANS, STANDARD DEVIATIONS AND COEFFICIENTS OF VARIATION
FOR HORNBLÉNDE ANALYSIS GROUPS.

		SiO ₂	Al ₂ O ₃	TiO ₂	FeO	MgO	CaO	K ₂ O	mg
M.	L	42.6	11.6	1.0	17.5	10.3	11.4	0.9	37.0
	S	42.7	11.8	1.0	16.6	10.4	11.5	0.8	38.5
	L+S	42.6	11.6	1.0	17.1	10.3	11.4	0.9	37.6
SD.	L	1.5	1.0	0.2	1.3	1.1	0.4	0.2	4.3
	S	1.4	0.9	0.2	1.2	1.1	0.3	0.3	4.1
	L+S	1.4	0.9	0.2	1.4	1.1	0.3	0.2	4.2
CV.	L	3.5	8.6	23.2	7.7	11.2	3.5	23.9	11.6
	S	5.2	7.4	24.9	7.4	10.2	2.3	35.5	10.6
	L+S	3.4	8.1	23.8	7.9	10.7	3.1	28.8	11.2

STANDARD ERROR OF DIFFERENCE BETWEEN GROUPS S AND L.

SiO ₂	0.46	MgO	0.354
Al ₂ O ₃	0.297	CaO	0.106
TiO ₂	0.077	K ₂ O	0.083
FeO	0.413	mg	1.35

CORRELATION MATRIX FOR S+L GROUP (40 SAMPLES).

	K ₂ O	CaO	MgO	FeO	TiO ₂	Al ₂ O ₃
SiO ₂	-0.59	0.02	0.70	-0.64	-0.38	-0.63
Al ₂ O ₃	0.12	-0.26	-0.83	0.67	0.04	
TiO ₂	0.51	0.64	-0.03	0.11		
FeO	0.48	-0.33	-0.88			
MgO	-0.40	0.33				
CaO	0.20					

of the physical and compositional variables is so complex as to render the composition of the hornblendes of little value in attempting to unravel the possible changes in the physical variables (PTetc.) during metamorphism of Scourian assemblages in the Laxfordian.

In view of the apparent similarity of groups S and L, only one correlation matrix has been presented, for the combined group S+L. Correlations between elements are generally interpreted on the basis of atomic substitutions within the crystal lattice. Substitutions that occur commonly in hornblendes include:

- a) Mg --- Fe²⁺
- b) Si --- Al, Fe²⁺ --- Al
- c) Si --- Al, Mg --- Al
- d) Si --- Al, Fe²⁺ --- Fe³⁺
- e) Fe³⁺ --- Al

The three pairs of coupled substitutions (b,c,d) are interesting because different correlation matrices would result from the individual and separate occurrence of each substitution. Thus for b) there will be a moderate correlation between both Si-Al and Fe-Al, for c) there will be a moderate correlation between Si-Al and Mg-Al, while for d) there will be a correlation between Si-Al, Fe-Al and Fe-Si. Parts of the correlation matrices for the individual groups S and L are given below:

Group S				Group L			
	Mg	Fe	Al		Mg	Fe	Al
Si	-0.52	-0.30	-0.70	Si	0.80	-0.85	-0.61
Al	-0.86	0.81		Al	-0.83	0.73	

There is an interesting difference between the coefficients for Mg-Si. Coupled with the strong negative correlation for Fe-Si in group L this suggests that there is a higher percentage of the aluminium in the octahedral sites in group L, or that a substitution Fe²⁺-Fe³⁺ is present in this group. Detailed analysis of the oxidation state of iron and calculation of structural formulae would be necessary before any firm conclusions could be made.

CHAPTER 9.

THE RELATIONS BETWEEN METAMORPHIC REACTIONS AND
TECTONIC STRUCTURE.

The previous chapters have been concerned with the presentation of data with little emphasis on discussion. It was noted a number of times that metamorphic and tectonic processes were closely associated during the evolution of the rocks studied. It is now necessary to discuss in detail the relationships between these two groups of processes. It is inevitable that cyclic arguments will result in such a discussion unless the premises of arguments are clearly defined. For this reason, the following short summary is presented. These statements are then taken to be true in the discussion that follows. Any generalisation of the discussion to include geologic systems other than those studied here will thus rest on proof of the truth of these statements in each case.

It is considered that enough evidence has been presented to show that the Scourian granulites have been altered chemically during two metasomatic ^{events}, and that tectonic deformation coeval with each event affected the gneisses. In particular, the shear zones studied are zones of anomalously high deformation and of anomalously high chemical change.

Consider the following cause and effect:

- a) deformation and metamorphism are coeval,
- b) high deformation and high chemical change are associated.

If a) is a sufficient condition of b), then when a) occurs b) must follow. If a) is a necessary condition of b), then b) cannot occur without a), (see Ayer 1956, p170). Only the sufficiency condition is fulfilled and it follows that the possibility of high strain zones being superposed on zones of

high chemical change cannot be logically excluded. The relationship of the mineral fabrics to the strain states measured in shear zones (Ramsay and Graham 1970) excludes the possibility that chemical changes were superposed on high strain zones.

Despite the logical inadequacy of the argument, it is assumed that the occurrence of chemical change and strain in shear zones is a sufficient reason for believing that deformation and metamorphism were coeval, and this discussion is based on this assumption. In retrospect, this intuitive step may seem to have been justified.

9.1 The concept of fugacity: fluid diffusion in rocks.

The scheme of mineral reactions outlined in earlier chapters involves the progressive hydration of assemblages during the formation of hornblende and biotite from initially pyroxene bearing rocks. It is essential to be able to define as accurately as possible the mechanism by which the water was introduced into the rocks studied, and to evaluate the consequences of such a mechanism occurring in a defined system.

The concept of fugacity needs careful definition, since the common meaning of something fugacious is evanescent or fleeting. The fugacity of a fluid would thus imply some measure of its mobility; this would be misleading.

Thermodynamically, the fugacity of a fluid is defined concisely. For an ideal gas, there exists the relation

$$\mu = \mu^+_{(T)} + RT \ln P \quad 9.1)$$

where P is the partial pressure of the gas phase. For a non-ideal gas (fluid), it is convenient to define a fictitious pressure, called the fugacity, to which the chemical potential, μ , of the fluid bears the same linear relationship:

$$\mu = \mu^+_{(T)} + RT \ln f \quad 9.2)$$

where f is the fugacity of the fluid.

The mobility of a component in a system is also thermodynamically defined, and this definition involves some knowledge of the kinetics of processes occurring in the system. The conservation of mass in a continuous system is expressed by the equation of continuity of density, ρ

$$\frac{\partial \rho}{\partial x} = -\text{div} \rho w \quad 9.3)$$

where w is the macroscopic velocity. For a mixture, w is related to the macroscopic velocities of the different components, i , by

$$w = \left(\sum_i \rho_i w_i \right) / \rho \quad 9.4)$$

Thus w is the velocity of the centre of gravity of an element of volume.

For a two component system in which irreversible diffusion alone takes place, it can be shown that only one independent irreversible process can occur, and that the entropy source strength for this simple system will be

$$(SS) = \frac{1}{T} \left(F_1 - \frac{\partial u}{\partial x} \right) \rho_1 (w_1 - w_2) > 0 \quad 9.5)$$

(see Prigogine 1967, chapter 3; the symbol (SS) replaces σ of DeGroot and Mazur 1962 and Prigogine 1967).

F_1 is the force per unit mass acting on component one, u_1 is the chemical potential per unit mass of component one, and w_1, w_2 are the macroscopic velocities of components one and two. For convenience, the forces and concentration gradients have been assumed to act along the x cartesian coordinate only. More general equations may be found in the works referred to above.

The phenomenological relation (ie. the relation between the flux of a process and the generalised thermodynamic force causing the process) corresponding to equation 9.5 is given by

$$\dot{\rho}_1 (w_1 - w_2) = \frac{L}{T} (F_1 - \frac{\partial u}{\partial x}) \quad 9.6)$$

Substituting equ.9.2 in equ.9.6 gives

$$C_1 (w_1 - w_2) = \frac{L}{T} (F_1 M_1 - RT \frac{d \ln f}{dx}) \quad 9.7)$$

where C_1 is the concentration of component one in the system, and M_1 is its molecular mass.

For a uniform system, ie. $\frac{d \ln f}{dx} = 0$,

$$(w_1 - w_2) = \frac{L}{TC_1} F_1 M_1 \quad 9.8)$$

The coefficient of proportionality, L/TC_1 , is called the mobility of component one, and has dimensions of velocity per unit force on a mole.

Consider the system under discussion to be made up of a rigid framework (component 2) and a number of interconnected voids (component 1). Equation 9.7 can be rewritten

$$C_1 w_1 = \frac{L}{T} (a P_x - RT \frac{d \ln f}{dx}) \quad 9.9)$$

where P_x is the pressure acting on the fluid along the x coordinate, and a is the cross-sectional area of the voids normal to x.

The magnitude of the pressure acting on the fluid can be considered, on such a scale that the system embraces an effectively infinite number of voids, as a measure of the fluid source strength. Knowledge of the variation in space of the three dimensional tensor P would enable the directions of fluid diffusion to be predicted.

The interesting point about diffusion of fluid through metamorphic systems stems from the realisation that there is a discontinuity in equ.9.9, and indeed this equation is no

longer directly applicable when the fluid 'source strength' exceeds a certain limit. Beyond this limit the framework is no longer rigid and processes of hydraulic fracturing (see Secor 1969) occur. The main effect of hydraulic fracturing is to cause an increase in the size of the voids (grain boundary spaces) in the system, thus permitting faster rates of fluid diffusion (the fluid can be considered as creating its own permeability). The discontinuity in equ. 9.9 exists because there is a large increase in the size of the voids for a small increment in the value of the function P across the limiting region.

Little is known about the permeability of metamorphic systems, but McLean (1965) makes the pertinent observation that grain boundary spaces between silicate minerals will generally be too small to allow free transport of water molecules. Grain boundary spaces are likely to decrease in size with increasing metamorphic grade since the stable minerals have progressively closer packed lattices. The problem of free diffusion is made more acute if water exists as a polymer of H_2O units. These problems described by McLean (1965) do not seem to have been investigated further by petrologists. Even so, they constitute sound evidence against the hypothesis that the fluid phase in metamorphic rocks diffuses passively along intergranular spaces in response to gradients in chemical potential (Gresens 1966), particularly in the higher grades of metamorphism. Discussions with Professor Fyfe have led to similar conclusions. The fluid phase in a rock cannot be considered as perfectly mobile if only chemical processes of diffusion operate; it is impossible to say how mobile the fluid was in any system, and misleading to consider it as immobile.

The work of Secor (1969) made a real contribution to the problem of fluid diffusion by presenting a detailed analysis of the phenomenon of hydraulic fracturing. Briefly, the analysis was based on the Griffith criterion for brittle failure and crack propagation. As the value of the fluid pressure in a system increases, the values of the effective principal

compressive stresses acting on the system are decreased. When the value of the least effective principal stress equals the tensile strength of the grain boundaries in the system, then brittle failure in tension of the grain boundaries occurs. Assuming the usual relation between the stresses at failure (Price 1966), it is realised that both tensile failure or shear failure may occur, the mode of failure in a particular instance being governed by the principal stress difference. It is suggested that if geological evidence exists that moderate to large quantities of fluid have been added to or passed through a metamorphic system, then processes of chemical diffusion alone cannot adequately account for this transport, and processes analogous to hydraulic fracturing must be considered. The implications of a mechanical mode of fluid transport in terms of the relations between deformation and metamorphism are discussed in a later section of this chapter.

9.2 Amphibolisation as a metasomatic event.

Amphibolisation of a granulite facies terrain involves addition of water to pyroxene bearing assemblages, and this system is defined as being thermodynamically open. During the hydration of pyroxene bearing rock which is now observed to contain 30 modal percent hornblende, approximately 0.5g of water per 100g of rock has been fixed in the hornblende. The presence of relicts of pyroxene in some samples is interpreted as indicating that only a limited amount of water was added to these parts of the granulites.

At any temperature and total pressure there will be a certain pressure of water with which pyroxene can stably coexist. If the water pressure exceeds this value, then the pyroxene will be amphibolised. The pyroxene stability curves evaluated in section 5.3 apply mostly to the coexistence of pyroxene and water when $P_{H_2O} = P_{total}$. It is possible for large quantities of water to pass through a granulite system and to leave no trace, provided a mechanism of water transport below

the critical pressure is available. The fact that there is, on a regional scale, a rapid gradation between dominantly pyroxene and dominantly hornblende bearing assemblages suggests a) that a regional front of amphibolisation can be drawn, and b) that this line indicates the position where P_{H_2O} dropped below the critical pressure necessary to cause amphibolisation under the PT conditions in existence at that time. If the system considered was one of progressive metamorphism, then included in b) would be the more normal hornblende-pyroxene isograd.

It is impossible to interpret the observed hydration or deformation of the granulites as either a cause or an effect, and the two processes are considered to have occurred contemporaneously. Some of the factors controlling the extent of hydration and deformation in such systems are:

a) Time duration of the water source strength, which can be thought of in terms of the rate of production of water available for transport. Continued source strength is necessary for both chemical and mechanical diffusion processes to occur to a noticeable extent.

b) Rate of diffusion of water. A gradient in the chemical potential of water will tend to be maintained between the water source and the granulite system because water is used up in the formation of hornblende from the granulite facies pyroxenes. The extent of hydration will depend on the rate of diffusion, and, as discussed in the previous section, diffusion is considered to take place most rapidly by a mechanical process. Seen in the light of this concept, the amphibolisation front mentioned above represents the limit beyond which water did not diffuse because P_{H_2O} was no longer sufficiently high to cause hydraulic fracturing.

c) Strain rate during the formation of the associated tectonic structures. Textural evidence from the rocks sampled suggests that amphibolisation of pyroxenes was limited to the time period during which the associated tectonic structures (see chapter 2) formed. If, as suggested, the processes of hydration and deformation are completely coeval, then evaluation of the strain

rate during deformation places limits on the time during which known quantities of water were added to the system.

Before some of the general points raised in this section are discussed, the Laxfordian event of biotisation associated with the shear zones will be considered in a manner similar to that laid out above.

9.3 Biotisation as a metasomatic event.

Considering only the area in which shear zones are found, it is realised that large modal quantities of biotite are confined to these shear zones. Formation of biotite from hornblende (and pyroxene) involves further hydration and also fixing of potash. The biotite fabrics developed within the shear zones suggest that the biotite crystallised entirely during deformation.

At the margins of the shear zones a biotite bearing rock passes rapidly outwards into a hornblende or pyroxene bearing rock lacking Laxfordian fabrics. It is inferred that sharp gradients in fluid pressure must have existed at these margins. While fluid must have migrated along the shear zones, there is little evidence for substantial lateral migration out of the shear zones. The metastable existence of gradients of fluid pressure is in agreement with the hypothesis that chemical diffusion plays a subordinate role in fluid transport.

Biotite is not restricted entirely in its occurrence to shear zones. In particular, biotite has been recorded in a number of samples from the steep limbs of asymmetric Scourian folds. The biotite in these samples appears as unoriented clusters, and it is difficult to tell immediately if it is of the same age as the biotite found in the shear zones. There are no apparent differences in chemistry between the two sets of biotite. They both contain barium as a minor element, and this is considered to characterise the assemblages resulting from Laxfordian metasomatism (see section 6.6).

The difference between the two modes of occurrence of Laxfordian biotite is interpreted in terms of the kinetic model of fluid diffusion presented in section 9.1. Arguments presented in section 9.5 leave little doubt that fluid diffusion in the shear zones occurred by a mechanical process at high fluid fugacity. Slower chemical diffusion at lower but variable fugacity or interrupted hydraulic fracturing in some parts of the system adjacent to the shear zones led to a much slower rate of formation of biotite, resulting in the ragged, unoriented clusters found in these areas. This notion is entirely consistent with the thermodynamic theory of the kinetic relations between diffusion and reaction rate that will be presented in section 9.7.

Finally, attention is drawn to a set of shear zones developed in the greenstone belt at Yellowknife, NWT, Canada. These shear zones cut across unmetamorphosed lava flows, in which igneous textures are easily recognised, and consist of chlorite-sericite schists. The association of deformation and metamorphism in zones is strikingly similar to the relations seen at Scourie. From a classical point of view, the shear zones at Yellowknife are at higher metamorphic grades than the adjacent lavas; the physical variable determining this difference must be the fluid fugacity. The analysis of such situations must be made through the use of thermodynamics, employing the notions of the irreversibility and coupling of processes, and the open nature of the systems being considered. The sections following investigate in detail these ideas.

9.4 Work done and heat produced in a shear zone.

The work done during a deformation may be estimated by measuring the area under the stress-strain curve. In an elastic solid the work done is $\frac{1}{2}\sigma\epsilon$ (Ramsay 1967, p287). In viscous and plastic materials the total strain in the material is time dependent, and the total work done under conditions of constant

stress difference is closely approximated by the product of the stress and finite strain tensors.

Any attempt to estimate the work done in the production of various tectonic structures now observed requires knowledge of the complete deformation history of the rocks, including such parameters as rock viscosity, the differential stress and the constitutive law obeyed during each increment of deformation.

Price (1971) discusses the relations between experimental and field data; in particular, he suggests various reasonable values for the strain rate and stress difference that could have existed in some tectonic environments.

Estimation of the work done in a shear zone has been done by assuming values for the strain rate and stress difference (10^{-10} /sec. and 100bars respectively), considered reasonable in the light of Price's (1971) discussion, and also by assuming a value for the finite natural strain in the shear zone of 2.3 (this is equivalent to a strain ratio of 100/1), (cf. Ramsay and Graham 1970). This model does not and cannot take account of such features as the development of fabric after a certain number of increments of strain, etc.

If the material in the shear zone deformed as a linear Newtonian liquid, then it would have had, under the conditions given, a viscosity of 10^{15} poise. A shear zone with a finite natural strain of 2.3 would then have formed in $0.6 \cdot 10^3$ years, and the work done would be 230bars or $2.5 \cdot 10^{-3}$ cal/cm³/year. The latter figure will only be slightly greater than the heat produced during deformation.

This discussion has only considered mechanical work and its equivalent mechanical heat. There exists the real possibility that chemical heat and work may contribute to the total heat produced and work done in a shear zone.

The metamorphic reactions that occurred during formation of the shear zones have been described, the essence of these reactions being that water and potash are fixed in the new assemblages. These reactions are also exothermic since the

formation of biotite/^{takes}place spontaneously below the probable equilibrium of the reaction. An estimate of the energy released during reaction can be made by using the Gibbs function at constant pressure, this being written as:

$$\Delta G = -\Delta SdT$$

Taking ΔS as 50cal/mol/deg for the biotite forming reactions, we have $\Delta G = 12.5\text{Kcal/mol}$ if the reaction takes place at 250°C below equilibrium under the conditions $P_{\text{H}_2\text{O}} = P_{\text{total}}$. In a rock containing about 30 modal percent biotite, this is equivalent to an energy release of approximately 14cal/cm^3 . If the reaction rate was constant throughout the $0.6 \cdot 10^3$ years during which the shear zone formed, there is an energy release of $2.3 \cdot 10^{-2}$ cal/cm³/year. It will be noticed that this figure is one order of magnitude greater than the estimate of mechanical work made above.

The muscovite producing reaction described in chapter 6 occurred at an estimated 70°C below equilibrium, but the entropy change during this reaction is sufficiently large to make the energy released about the same as that already calculated above.

One interesting difference between the reactions involved in biotisation and those involved in amphibolisation is that during the latter the inferred changes in entropy are very small (see section 5.3). Thus during amphibolisation of pyroxene the rate of release of chemical energy will be very low.

The total heat produced in a shear zone during its formation will be made up of two parts, a mechanical and a chemical contribution. Calculation of the resultant temperature rise would necessitate certain assumptions about the rates of heat production and would involve solution of a non steady state equation. This has not been attempted.

The significance of the released chemical energy in relation to the mechanical work done is difficult to estimate numerically. Can the released chemical energy contribute to the work term in a system? The answer to this question must be

in the affirmative. The difference between pressure-volume and entropy-temperature terms in the thermodynamic equations of state is that the former expresses work done on a system, while the latter expresses work done by a system. The importance of the work done by a system is that it lowers the activation levels for processes that result in work being done on the system. Put very crudely, this means that the breakdown of minerals during reaction permits an increased rate of mechanical working relative to the system where no such chemical breakdown is occurring.

Considering the two energy terms as simply additive (2cal/cm^3 plus 14cal/cm^3 equals approximately 700bars) and allowing for the continued existence and constancy of the rock viscosity, the strain rate evaluated is $7 \cdot 10^{-10}/\text{sec}$. A shear zone with a finite natural strain of 2.3 would then form in $0.11 \cdot 10^3$ years instead of the $0.6 \cdot 10^3$ years estimated above. This must be considered an approximation, but it does indicate the important contribution that mineral reactions could make to the total work done in systems such as shear zones similar to those considered here. Derivation of a functional relationship between work terms seems difficult at this stage.

9.5 The importance of oxidation in shear zones.

The relations between the different iron oxides at varying temperatures and oxygen fugacities have been determined experimentally (Eugster 1959). Thus at 600°C the coexistence of the pair magnetite-haematite means that the partial pressure of oxygen is buffered at about 10^{-14} bars. In any system in which a fluid water phase is present there will be a certain partial pressure of oxygen resulting from the equilibrium dissociation of this water. The dissociation constants for water at different temperatures are known thermodynamic constants. In a closed system it is evident that the actual amount of water present determines the amount of oxygen that is available for oxidation in that system, since as oxygen is used up, the partial pressure

of hydrogen increases and prevents further dissociation of water. By increasing the amount of water relative to solid phases in the system, further oxidation can be accomplished. An alternative way of considering this is to assume that the system was open to water - the observed oxidation then defines the total water to rock ratio, ie. the quantity of water that has passed through a given quantity of rock.

The whole rock analyses presented in table 6.6 indicate that the shear zone assemblages are more oxidised than the unaltered assemblages. The amount of oxidation correlates well with the amount of potash in the assemblage, which can be considered as a convenient index of the extent of metasomatism.

Haematite has only been recorded in a few assemblages studied here, and it generally forms from a titanium bearing phase. For convenience, the oxygen partial pressure has been set at 10^{-14} bars (ie. the magnetite-haematite equilibrium) in the calculations below. This value must be considered as a maximum for the shear zones studied here. The quantity of iron that has been oxidised is taken from the analyses in table 6.6.

about 4% FeO is oxidised to Fe₂O₃
ie. 0.11g FeO is oxidised per cm³ of rock

P_{O₂} is buffered at 10^{-14} bars
assume water phase is pure H₂O
no. of mols. H₂ = twice no. mols. of O₂

the dissociation constant for water is given by:

$$1/K = \frac{(H_2)^2(O_2)}{(H_2O)^2}$$

at 600°C logK = 11.5
 K = 3.2 . 10¹¹
 P_{O₂} = 10⁻¹⁴ bars
assume P_{H₂O} = 6Kb.
ie. fugacity H₂O = 4100bars

Using this data, the partial pressure of hydrogen may be calculated.

$$\text{thus } P_{\text{H}_2} = 0.128\text{bars}$$



72g FeO uses $\frac{1}{4}$ mol. O₂

$$\text{ie. } 0.11\text{g FeO uses } \frac{0.25 \cdot 0.11}{72} \text{ mols. O}_2$$

This will produce $x = 0.5 \cdot 0.11/72$ mols. H₂ at 600°C, 0.128bars
Under these conditions, x mols. H₂ occupies 220cm³.

The volume of hydrogen is approximately the same as the volume of water-

$$\text{thus quantity of H}_2\text{O} = 220\text{cm}^3/\text{cm}^3 \text{ rock}$$

$$\text{at } 600^\circ\text{C and } P_{\text{H}_2\text{O}} = 6\text{Kb}$$

$$\text{ie. } 185\text{g H}_2\text{O}/\text{cm}^3 \text{ rock}$$

$$66\text{g H}_2\text{O}/1\text{g rock}$$

If oxygen partial pressures of 10^{-16} and 10^{-18} bars had been used, then the quantities of water added would have been 22 and 2.2 cm³H₂O/cm³ rock respectively.

Within the limiting assumptions of the calculation, it can be seen that moderately large quantities of water are necessary to cause oxidation with the production of haematite in a shear zone. Even if the system were buffered at some lower partial pressure of oxygen, significant quantities of water are needed.

It is interesting to compare these figures with the results of the similar, but much simplified, calculation presented in section 5.3. In general, a smaller quantity of water is necessary to cause oxidation during the formation of hornblende. Indeed, this quantity may approach the same order of magnitude as the quantity of water actually locked up in the hornblende (for a rock containing 30 modal percent hornblende this is approximately 0.005g H₂O/1g rock).

There is an independent way of checking the figures for the quantity of water added to the shear zones. During metasomatism, a certain amount of potash has been introduced into the shear zones (see table 6.6). The probable identity of this potash in solution is potassium silicate, and although accurate solubility data are not available for this, a figure of 1g/1000g H₂O is probably reasonable for the conditions being considered (Professor Fyfe, personal communication). Thus for an increase in a rock of 2.6wt% K₂O, 26g H₂O/1g rock are required. This order of magnitude is consistent with the calculation based on the oxygen partial pressure.

In concluding this section, attention is drawn to two differences between the events of amphibolisation and biotisation - for the former, less chemical energy was released during reaction, and less water was added to the system. It is possible that these differences have determined the nature of the tectonic events, at least in part, that are associated with each metasomatism.

9.6 An instability leading to the generation of shear zones.

It is considered that the data presented in the previous two sections suggests that moderate to large quantities of water passed through the shear zones, and that this occurred during the relatively short time of formation of these shear zones. It is also considered that the only mode of fluid transport compatible with these data is a mechanical process analogous to hydraulic fracturing (this term is used in the broad sense, and describes failure in both tension and shear).

The consequences of the operation of processes of hydraulic fracturing can be listed as follows:

a) Brittle failure occurs on the scale of individual mineral grains, and along the boundaries of or across these grains. There is no direct relation between failure on this scale and failure of the rock mass as a whole.

b) The stress distribution across an aggregate of rigid grains may be very irregular (see Price 1966 plate 1). However, at high temperatures and confining pressures, it seems likely that grains will no longer make contact at points, but most likely over the whole grain boundary area. It is then possible that the stress distribution across a crystal aggregate may be more regular. The predominance of certain orientations of microfractures in quartz and feldspar crystals in many of the rocks studied here supports the view that stresses were fairly regularly oriented across mineral aggregates. The formation of regular fracture sets during hydraulic fracturing will impart an anisotropy to the rock, and this anisotropy may have a profound influence on the subsequent macro-behaviour of the rock system.

c) Fracturing along grain boundaries by high fluid pressures overcomes much of the cohesion of the rock, or alternatively, this cohesion is the main activation barrier to deformation.

The rheological properties of the rock will thus be profoundly changed whilst hydraulic fracturing is actively occurring within that rock. The failure of grain boundaries can be adequately described in terms of brittle failure by utilising the Mohr circle and Mohr envelope constructions. Macroscopic failure resulting in large scale (relative to the grain size) shear zones cannot be described in such a simple manner.

Undoubtedly, shear zones form in a number of different ways, but it is possible that many are the macroscopic result of a prolonged history of small scale fracturing. This history of fracturing is an expression of mechanical cataclasis of grains, and hydraulic fracturing is a particular mechanism by which such cataclasis may take place. A recent paper by Lajtai (1970) records in detail the fracture history leading to the formation of a macroscopic shear zone during the experimental deformation of plaster. The shear zone results from the coalescence of earlier tensile and shear fractures. The sequence of fracturing during loading of the plaster blocks is interesting and can be summarised as follows:

a) initial formation of tensile fractures; these fractures do not reduce the load carrying capacity of the sample and do not register on the load versus axial strain diagram;

b) with increasing load shear fractures form, and axial strain becomes apparent;

c) coalescence of fractures occurs by further shear and tensile failure;

d) at maximum load the macroscopic shear zone is not present;

e) the shear zone develops and incorporates pre-existing smaller scale fractures.

Lajtai (1970) suggests that the brittle strength of the material is related to the resistance to shear stress within the material. His results also suggest that less work is done during the formation of tensile fractures than during formation of shear fractures, because tensile fractures invariably develop first at lower load values than those at which shear fractures develop. There will in general be a smaller loss of potential energy during tensile fracturing.

Bombalakis (1968) has studied photoelastically the coalescence of some simple crack arrays. He noted that within a particular array, the fracture spacing was an important parameter in determining how easily coalescence occurred. In some natural rocks, the fracture spacing may be largely controlled by the grain size and a detailed analysis may show that in some cases this parameter controls the points at which macroscopic shear zones are initiated.

It is also possible for the initial phase of fracturing to granulate the rock to an extent that a macroscopic shear zone develops directly. This, presumably, is determined by the rate of loading and the rate of fracturing within a given volume of rock.

The shear zones studied at Scourie show no development of large scale discontinuities within them. The early Scourian banding may be traced continuously across some of the structures, and these shear zones must be considered as macroscopically continuous, the strains developed within them

being described by the compatibility requirements for finite heterogeneous strains. For this reason the shear zones cannot be treated in terms of the theory of brittle failure alone.

The stress-strain curve for a rock subject to hydraulic fracturing and in which a shear zone develops will in general have a yield point. A generalised yield criterion, being a combination of the brittle and plastic criteria, will describe the rheological properties of the rock. Whatever the analytical form of this yield criterion, it can be appreciated that shear zones will propagate from zones of intense small scale fracturing where certain relations between the principal deviatoric stresses are satisfied. The theory of brittle failure alone predicts that a certain magnitude of the principal stress difference must be reached before shear failure occurs. This very simplified notion has been adopted as a general working criterion for yield in the rocks being discussed, and is used in a later section where the relation of the shear zones to the regional tectonic structure is analysed.

9.7 Entropy production and irreversible processes.

The most general thermodynamic statements take account of the fact that systems considered are often open systems, and that the processes occurring within these systems are usually irreversible. For any system, the state function S , the entropy, can be introduced. The variation in entropy dS may be written as the sum of two terms

$$dS = d_e S + d_i S \quad 9.10)$$

where $d_e S$ is the entropy supplied to the system by its surroundings, and $d_i S$ is the entropy produced inside the system. The second law of thermodynamics states that $d_i S$ must be zero for reversible transformations and positive for spontaneous irreversible transformations of the system:

$$d_i S \geq 0 \quad 9.11)$$

The entropy supplied to the system, $d_e S$, may be positive, zero or negative, depending on the interaction of the system with its surroundings.

Irreversible processes are described by the expressions defining the entropy production per unit time for these processes. The entropy production takes the general form of a bilinear equation, and is written as the sum of the products of the generalised forces (X) and the corresponding fluxes (J) of the irreversible processes:

$$\frac{d_i S}{dt} = \dot{S} = \sum_k J_k X_k > 0 \quad 9.12)$$

Thus it is both necessary and sufficient to have a knowledge of the rate at which a process takes place and of the force causing this process in order to be able to define completely this irreversible process. A system can often be described by considering processes of heat transfer, transfer of mass (diffusion), viscous flow, and chemical reaction. DeGroot and Mazur (1962) present general equations for the entropy production and entropy flow in such systems.

Any spontaneous irreversible process results in an increase of the entropy for the system in which the process occurs. Consider two irreversible processes, the one spontaneous, the other not. The entropy production for the former will be positive, for the latter negative. If the net entropy production is positive, then the unnatural process may occur in spite of its negative entropy production and because of the positive entropy production of the natural process. This is the basis of the concept of coupled irreversible processes.

It is immediately apparent that these ideas can be applied to the natural deformation of rocks; this will be illustrated with reference to the shear zones being studied here. The irreversible processes that are relevant to the discussion are:

- a) introduction of chemical species,
- b) chemical reactions,
- c) finite deformation.

The first two processes are considered to occur spontaneously, for reasons already discussed. The close association within shear zones of high strain and the products of mineral reactions resulting from metasomatism, and the absence of visible deformation in areas where such reactions have not occurred suggest that the processes of deformation, considered alone, were not spontaneous processes until chemical reactions had been initiated in the system. It remains to be shown that such a situation can exist.

In the following equations (based on Prigogine 1967, and DeGroot and Mazur 1962), all "dotted" symbols refer to the time derivatives of these quantities.

The rate of entropy production resulting from a spontaneous chemical reaction in a closed system is given by

$$\dot{S} = \frac{1}{T} (Av) > 0 \quad 9.13)$$

where A is the affinity of the reaction and v the reaction rate. Associated with this will be a rate of entropy flow resulting from interactions of the system with the exterior, and given by

$$\dot{S} = \frac{\dot{Q}}{T} \quad 9.14)$$

Q being a measure of the heat received. The total rate of entropy change is given by the sum of equ. 9.13 and 9.14:

$$T\dot{S} = \dot{Q} + Av > 0 \quad 9.15)$$

Equ. 9.15 describes completely the conditions for the spontaneous occurrence of an irreversible chemical reaction in a closed system. At equilibrium, the process becomes reversible and

equ. 9.15 reduces to

$$T\dot{S} = \dot{Q} \quad 9.16)$$

the familiar form of the second law.

The affinity A which appears in equ. 9.15 needs further definition. The affinity of a reaction is related to the chemical potentials of the components in the reaction by

$$A = -\sum c_i \mu_i \quad 9.17)$$

where μ is the chemical potential of component i which has a stoichiometric coefficient c in the reaction. The chemical potential of a species in an ideal system is defined, using standard thermodynamic symbols (Denbigh 1966, p79,111), as

$$\mu_i = \left(\frac{\partial G}{\partial n_i}\right)_{P,T,n_j} = \mu_{i(P,T)}^+ + RT \ln x_i \quad 9.18$$

Substitution of 9.18 in 9.17 gives two explicit definitions of the affinity:

$$A = -\sum c_i \left(\frac{\partial G}{\partial n_i}\right)_{P,T,n_j} \quad 9.19)$$

$$A = -\sum c_i \mu_{i(P,T)}^+ - RT \sum c_i \ln x_i \quad 9.20)$$

By introducing r , the degree of advancement of the reaction, defined by

$$dn_i = c_i dr, \quad 9.21)$$

equation 9.19 becomes

$$A = -\left(\frac{\partial G}{\partial r}\right)_{P,T} \quad 9.22)$$

The affinity is now expressed as a partial derivative of the Gibbs free energy, G .

Equation 9.20 may be modified by introducing the equilibrium constant, $K(P,T)$, defined by (Denbigh 1966, p140)

$$RT \ln K(P,T) = - \sum c_i \mu_i^+(P,T) \quad 9.23)$$

Equ. 9.20 may then be rewritten as

$$A = RT \ln \frac{K(P,T)}{N_1^c \dots N_i^c} \quad 9.24)$$

Finally, the rate of a chemical reaction can be defined by means of the laws of conservation of mass:

$$\rho \frac{\partial x_i}{\partial t} = c_i v \quad 9.25)$$

the only new symbol being the density, ρ .

In an open system, the entropy production resulting from transfer of energy and matter into the system must be considered. This is most easily defined by the following equation:

$$\dot{T}S = \dot{F} + \sum_i \mu_i \dot{n}_i \quad 9.26)$$

F denotes the resultant flow of energy, and \dot{n} defines the rate at which component i is supplied to the system. If diffusion occurs spontaneously and irreversibly, the quantity \dot{S} will, of course, be greater than zero.

The above equations describe some of the general notions of the thermodynamics of irreversible processes. Some simple calculations are now presented where rates of entropy production have been estimated, and the coupling of mechanical deformation with hydraulic fracturing is investigated. Many physical parameters have been assigned actual values in the calculations, and it is important to appreciate the results of varying the values of these parameters. The calculations are intended to

illustrate principles rather than provide a concise analysis. At this stage it would be impossible to evaluate accurately rates of entropy production.

First, the diffusion of water will be considered.

Let the rock porosity = 0.1%

ie. 1cm^3 of rock contains 0.001cm^3 fluid

During shearing, 1cm^3 rock has 1g fluid passing through it.

The specific volume of water at 600°C , 6Kb is $1.2\text{cm}^3/\text{g}$.

Assume that the life of a shear zone is 10^3 years

In 10^3 years, $1.2\text{cm}^3\text{H}_2\text{O}$ passes through 0.001cm^3 rock

ie. the rate of flow at a point is $1.4 \cdot 10^{-9}$ mol/sec

The chemical potential is defined as

$$\mu = \mu^+ + RT \ln f$$

ie. $\mu/T \sim R \ln f$

Under the physical conditions specified

$$R \ln f = 660 \text{ cm}^3 \text{ bar}^2 / \text{deg/mol}$$

Thus the entropy production resulting from the diffusion of water is

$$\dot{S} = 2.4 \cdot 10^{-5} \text{ cal/deg/mol/sec.}$$

This calculation ignores the change in entropy resulting from flow of heat, and considers only diffusion of water (fluid) through a defined porosity and under the assumption that the rock permeability need not be accounted for. This recalls the system that was analysed in section 9.1 - a rigid framework with interconnecting pore spaces. The real system (ie, the shear zones) that is being considered is complicated by the fact that both water and potash were introduced into the shear zones. Also much of the water must have entered and subsequently left the system since it is not all accounted for as mineralogical water still contained in the system; for this reason, entropy increases will have been reduced by corresponding entropy losses.

It is now necessary to consider the entropy production resulting from processes of hydraulic fracturing. This is a spontaneous process by virtue of the high fluid pressure. Tensile

failure occurs when the effective least principal stress equals the tensile strength of the rock (grain/grain boundary), and it is necessary to assume that this condition is maintained.

The grains making up the rock are taken to be 0.1cm in diameter (ie. 1cm^3 of rock contains about 10^3 grains), and the length of each fracture formed is assumed to be approximately the same as the grain diameter, ie. 0.1cm.

The fractures are assumed to be ellipsoidal in shape, with dimensions

$$a/b/c = 10/10/1$$

ie. the volume of one fracture = 10^{-4}cm^3 (approx.).

The rate of flow of the fluid is taken, as above, to be $1.4 \cdot 10^{-9}\text{mol/sec}$. (this effectively defines a mean porosity, the one that was assumed above).

Therefore the rate of fracturing = $3 \cdot 10^{-4}/\text{sec}$,

ie. one fracture forms every $3 \cdot 10^3\text{sec}$.

If the tensile strength of the grain/grain boundary, under the conditions being considered, is taken as one bar, then there will be an entropy production of

$$\dot{s} = 3 \cdot 10^{-6} \text{ bars/deg/sec}$$

It is clear that fracturing resulting from high fluid pressures will lower the activation level for bulk deformation in a system under deviatoric stress. The entropy production during a viscous deformation is given by the product of the principal deviatoric stress tensor and the principal strain rate tensor. To judge the extent to which the processes of fracturing and viscous deformation may be coupled, it is only necessary to assume values for the deviatoric stresses, and equate the expressions for the entropy production of each process. The unknown parameter, the strain rate, is then calculated. It is assumed that each fracture has a life of one second before it is closed again, and that only during this time can deformation occur. The following values of the deviatoric stresses are used:

$$\sigma_1' = 500\text{bars}, \quad \sigma_2' = 400\text{bars}, \quad \sigma_3' = 100\text{bars}$$

Thus the strain rate, for the one second duration of each fracture subsequently averaged over a period of time, is

$$\dot{\epsilon} = 3 \cdot 10^{-7} / \text{sec}$$

This value of the strain rate must be considered a maximum, within the assumed parameters of the calculation, because it was evaluated on the basis of a net entropy production of zero. It does, however, indicate that the process of hydraulic fracturing could enable moderate deformation to occur in systems where, without fracturing, deformation would not take place. Attention is also drawn to the relatively large values of the deviatoric stresses chosen above.

The lack of precise knowledge regarding the values of parameters such as the tensile strength of grain boundaries at high temperatures, the rates of fluid flow, and the time duration of open fractures (values of these parameters were assumed completely intuitively) leads to a large range in the possible values of the entropy production. However, both diffusion and chemical reactions (including recrystallisation) have been completely ignored, and both of these groups of processes will result in a positive entropy production.

It is therefore concluded that introduction of water (and dissolved potash) into the system by means of mechanical fluid fracturing resulted in a) breaking of grain boundaries, and b) chemical reactions and mineralogical reorganisation. Only because of the spontaneous nature of these processes was it possible for coupling between them and viscous deformation to occur. The result is the close association in both time and space of areas of high strain and extensive chemical/mineralogical change.

CHAPTER 10.

A REGIONAL MODEL.10.1 Introduction.

The recent work of Beach et al (in preparation) has shown the real existence of a three fold division of the Lewisian basement in the Scourie region beyond that shown by Peach et al (1907) and Sutton and Watson (1951) and implied by Holland (1966) and Holland and Lambert (1971). The divisions are

- a) granite gneiss and migmatite north of Laxford, with a concentration of granite sheets along Laxford and Stack,
- b) gently dipping pyroxene bearing banded gneiss south of a line joining Scourie and Ben Auskaird
- c) an area of strongly deformed hornblende gneisses, containing a large number of igneous bodies and strips of gneiss of supposed metasedimentary origin, lying between a) and b).

All three gneiss groups contain members of the Scourie dyke suite.

There is a real concentration of igneous bodies in this region, and these have a large range in composition (see section 1.4). They appear to have been in existence prior to the granulite metamorphism that closed at about 2600my. It is possible that two different ages of basement will be recognised in this region - an older basement that was intruded by an igneous complex - both of which are pre-2600my.

The question arises as to whether the late Scourian (amphibolite facies) deformation is coincidentally present in the region of this earlier igneous complex or whether there is some fundamental crustal inhomogeneity genetically linking the two features. Similarly, is the presence of a granite-migmatite front in this region a coincidence?

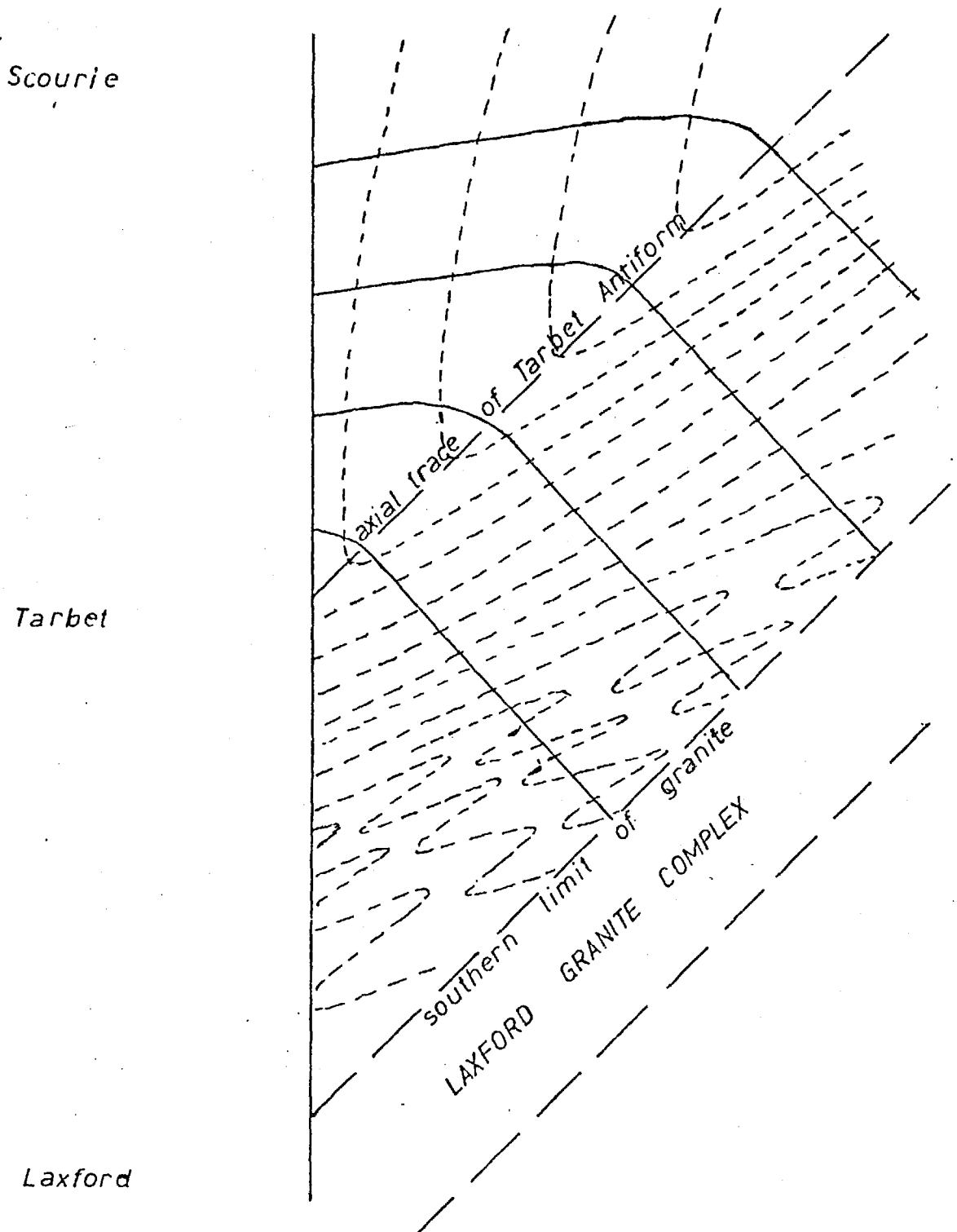
The southern limit of the granite sheets is a remarkably sharp line, termed the Ben Stack line; Peach et al (1907, p126) noted that this line was regionally discordant to the gneissic foliation. According to the hypothesis of Holland (1966), the Ben Stack line marks the contact between amphibolised Scourian granulite type and the Laxford gneiss type; these two rock groups have been shown to have different bulk chemical compositions, (Holland and Lambert 1971), though it is difficult to interpret this difference. Holland (1966) proposed that the Laxford group represented a metasedimentary suite that had been infolded into the basement and had undergone partial melting. The structure of this area is contrary to this notion; indeed, the granite gneisses now occupy a structurally lower level than the granulites and amphibolised granulites (Beach et al). If the Laxford group was once a supercrustal series, it is necessary to invoke overthrusting of the Scourian rocks in order that the present day spatial arrangement be reached; in this model the Ben Stack line becomes a thrust plane. Alternatively, the Ben Stack line could represent a true migmatite front in the sense of an upward moving front of granite formation. It is then necessary to explain the concentration of intrusive granite sheets along the southern boundary of this migmatite area, and the very sharp transition to non granitic material at this southern boundary.

Whatever the answer to these problems, it is proposed only to consider a model where the granites are assumed to occupy approximately their present spatial position, and to present some ideas concerning the development of shear zones at Scourie in relation to the granite complex at Laxford. That the generation of shear zones is undeniably linked with the granite forming metamorphism is shown by the metasomatism that is associated with the shear zones. Thus the basic premise of the model is that the water and potash (and also minor elements) that have been added to the rocks in the shear zones were derived from the Laxfordian granite complex. Knowledge

Section between Scourie and Laxford
showing principal structural units and
generalised orientation of stress
trajectories (σ_1)

FIG 10.1

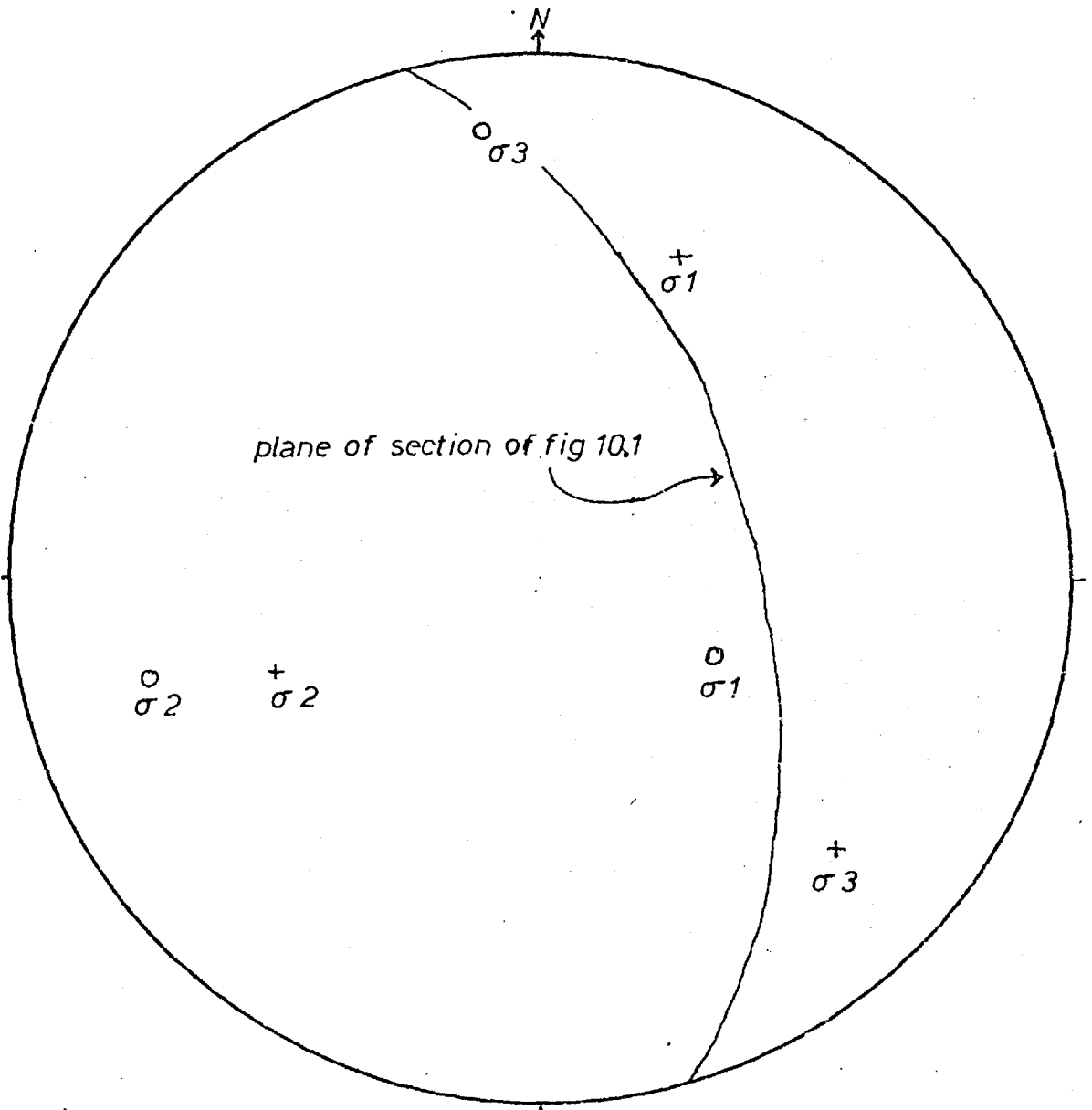
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Deduced orientation of principal stresses during the main Laxfordian deformation and metamorphism

FIG 10.2

- + Tarbet Sleep Belt
- o Scourie shear zones



of the detailed age relations between shear zones and granite complex requires that further age determinations be carried out in this region. At present the only data available are those of Lambert and Holland (1970); they found that the peak of metamorphism at Laxford occurred at 1850my.

Further information essential to the basic concept of the model has been deduced from field evidence (presented in chapter 2). The presence in places of a conjugate set of shear zones enables the direction of the principal stresses giving rise to these shear zones to be defined. Also, it has been shown that the dominant Laxfordian strains in the Tarbet area were of a flattening type. It has therefore been assumed that the maximum compressive stress in this region was approximately at right angles to the final position of the foliation. Fig.10.1 shows an idealised two dimensional section of the region, in which the principal structural units and the maximum compressive stress trajectories are drawn. The stereogram (fig.10.2) shows the approximate orientations of the deduced principal stress axes, and also the orientation of the plane of section in fig.10.1.

The development of the model is based where possible on analytical procedures, but it must be emphasised that the model is wholly contrived. The analysis that follows contains no fundamental truth by virtue of it being analytical, but is merely the most concise way of expressing some ideas. Assumptions made have been clearly stated where they arise.

10.2 Stress function analysis.

The following two dimensional stress function, F_1 , has been chosen since it gives rise to a pattern of trajectories similar to that drawn in fig.10.1:

$$F_1 = \frac{x^3}{3} - xy \quad (10.1)$$

This function has been chosen without any regard for boundary conditions, and discussion is therefore confined to a region defined by $-5 \leq x \leq +5$, $0 \leq y \leq 70$. Otherwise it is a perfectly valid function since it automatically satisfies the stress equilibrium equations and the biharmonic compatibility equation. In a cartesian (x,y) space, the following stress components can be derived:

$$\begin{aligned}\sigma_x &= \frac{\partial^2 F_1}{\partial y^2} = 0 \\ \sigma_y &= \frac{\partial^2 F_1}{\partial x^2} = +2x \\ \tau_{xy} &= \frac{\partial^2 F_1}{\partial x \partial y} = +1\end{aligned}\quad 10.2)$$

that is

$$\begin{aligned}\sigma_1 &= x + (x^2 + 1)^{\frac{1}{2}} \\ \sigma_3 &= x - (x^2 + 1)^{\frac{1}{2}}\end{aligned}\quad 10.3)$$

$$\begin{aligned}\sigma_1 - \sigma_3 &= 2(x^2 + 1)^{\frac{1}{2}} \\ \frac{1}{2}(\sigma_1 + \sigma_3) &= x\end{aligned}\quad 10.4)$$

The equation of the principal stress trajectories can be evaluated by solving the following equation:

$$\tan 2\theta = \frac{2\tau_{xy}}{\sigma_x - \sigma_y} = \frac{2\tan\theta}{1 - \tan^2\theta}\quad 10.5)$$

$$\text{therefore } 2\tan^2\theta - 4x\tan\theta - 2 = 0\quad 10.6)$$

The solution of this equation is

$$\frac{dy}{dx} = \tan\theta = x + (x^2 + 1)^{\frac{1}{2}} \quad 10.7)$$

This is a standard integral form, and on integration, gives

$$2y = x^2 + x(x^2 + 1)^{\frac{1}{2}} + \ln(x + (x^2 + 1)^{\frac{1}{2}}) + K1 \quad 10.8)$$

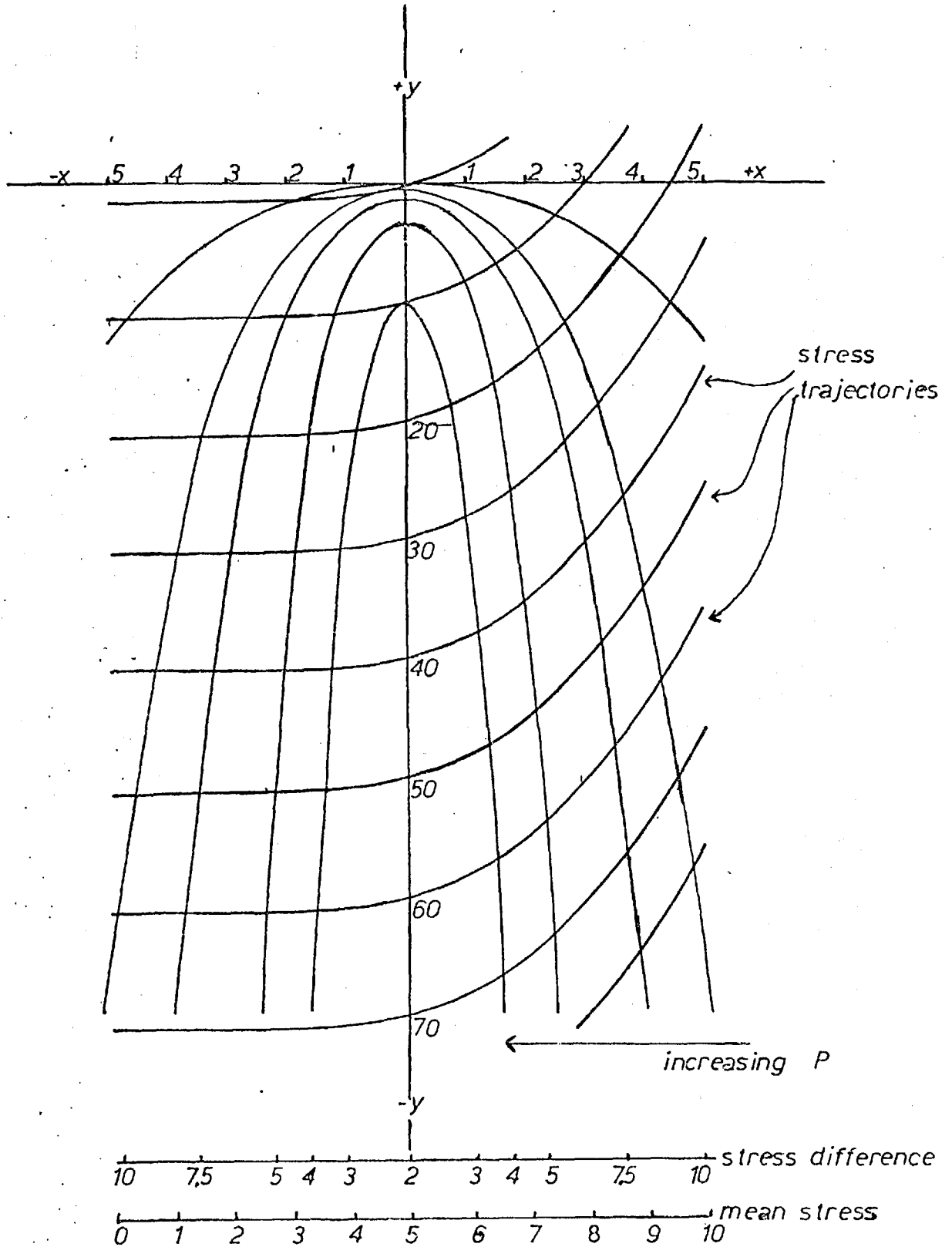
Part of the family of curves defined by 10.8 are shown on fig.10.3. Also shown are arbitrary scales of the mean stress and of the stress difference, as defined by equ.10.4; note that these parameters are dependent only on the x coordinate.

If the stress pattern of fig.10.3 was superposed on a linear homogeneous material, then after a given time the minimum amount of deformation would be at $x=0$ and it would increase away from this line. However, the system being considered had neither linear nor homogeneous rheological properties. The region of the Tarbet Steep Belt records large Laxfordian strains, while south of the Tarbet antiform Laxfordian strains become low to negligible. In fact, it is this marked heterogeneity that must be the cause of the stress distribution deduced in fig.10.1. There is effectively a stress concentration in the resistant and undeformed Scourie block. In fig.10.3, the line $x=0$ corresponds approximately to the axial trace of the Tarbet antiform, while the region in $-x$ corresponds to the Tarbet Steep Belt - Laxford area.

It has been assumed that the stress trajectories are very nearly straight lines between Tarbet and Laxford. Though this assumption is not essential to the model, without it another stress distribution would have to be fitted to the Laxford region, and at the present time there is no way of telling what sort of perturbations, if any, there were in the virtually linear stress field assumed here for this region. It is also

Theoretically derived stress trajectories (σ_1) and lines of equal pressure (parabolic curves)

FIG 10.3



assumed that the southern limit of the granite complex is a planar surface to depth. Again, this is not an essential assumption, but without knowledge of the geometry of the surface, from seismic or other work, it is the simplest.

10.3 The source and motion of the fluid phase.

Tuttle and Bowen (1958) suggested that granite melts were saturated in water and formed at the minimum melting eutectic in the system Q-Ab-Or. More recently, Fyfe (1970) and Brown and Fyfe (1970) have suggested that granite melts are formed at temperatures above the eutectic point and contain only 1-2% water. This water is considered to be derived from the breakdown of hydrous minerals (muscovite-biotite-hornblende) during progressive metamorphism. Much of this water is released into the surrounding rocks when the granite melt solidifies. However, if separation of the granite melt from its source region has occurred, then the water contained in the melt will also be removed from its source region, and subsequently concentrated in the region of solidification of the melt. Such a process seems to have occurred at Laxford, where granite is concentrated in a linear zone about 1km wide, and consisting of 70-80% by volume of granite. If the granite magma contained 1% water, then 1km^3 of rock in this region will have contained about $7 \cdot 10^6 \text{m}^3$ of released water. During solidification of the granite magma this region will have been a zone of high water pressure bordered above and to the south by a region of low water pressure (ie. the hornblende gneiss of Scourian age). This unstable configuration would be made more stable if the water diffused out of the zone of high water pressure.

It was considered in the previous chapter that the most likely mechanism by which large quantities of water diffused over large distances was a process analogous to hydraulic fracturing. Failure in shear or in tension results from high fluid pressures reducing the effective principal stresses. In a homogeneous rock, the fluid will move outward from its source

along the direction of maximum compressive stress. It is at first difficult to relate the motion of fluid in microfractures (of the same size as individual grains) to a regional stress pattern. However, Beach (1969) showed that a set of healed microfractures had a constant orientation in a number of large quartz grains of differing shapes and lattice orientations. Similarly, a preliminary study of samples from the Tarbet Steep Belt reveals a set of healed microfractures with a constant orientation at right angles to the foliation and approximately normal to the finite stretching direction in these rocks. It is therefore suggested that the orientation of microfractures does reflect the overall orientation of the principal stresses in a system.

This discussion leads to one of the basic assumptions of the model -- that the regional stress trajectories shown in fig.10.1 define the bulk paths of fluid motion in the system. This assumption does not take into account the effect of anisotropies, such as foliation, in the rocks, which may affect diffusion paths. However, the large number of microfractures seen at right angles to the foliation suggests that there was a component of diffusion in this direction that cannot be ignored.

Considering only the fluid diffusing along the defined stress trajectories, the equations of these stress trajectories are then made to define stream lines of fluid flow. The following stream function is then defined:

$$F_2 = x^2 + x(x^2 + 1)^{\frac{1}{2}} + \ln(x + (x^2 + 1)^{\frac{1}{2}}) + K_2 - 2y \quad 10.9)$$

Derivation of a pressure function from a stream function is a standard procedure, and the following is obtained

$$P = \frac{y}{2(x^2 + 1)^{1.5}} + K_3 \quad 10.10)$$

Part of the family of curves $P = \text{constant}$ is shown on fig.10.3.

The stream function above is derived purely from a knowledge of the stream lines, that is, the assumption made in taking this step is that the fluid motion can be adequately described by a stream function. It is difficult to judge how good an approximation this is, but equ.10.10 serves a useful descriptive purpose.

In any system, there will be an upper limit to the value of the water pressure, since at a certain value fracture in shear or tension will occur. Thus the pressure variations in the moving fluid predicted by equ.10.10 are fictive. Since the mechanism of transport everywhere involves fracturing resulting from high fluid pressures (this model would not have been devised if this was not the case), equ.10.10 can be interpreted as a measure of the rate of working at a particular point.

For any specified brittle failure criterion, tensile failure will occur at lower values of the stress difference than shear failure. In a system where microfractures are continually opening and closing, the displacements resulting from shear failure will generally be larger than those resulting from tensile failure. Thus, as a generalisation, it will be assumed that more work is done during shear failure than during tensile failure.

Inspection of fig.10.3 shows that the stress difference reaches a minimum at $x=0$, and that it is at $x=0$ ($y=\text{constant}$) that P is a maximum. It might be expected that in this region the work done in forming one fracture would be a minimum, and since the rate of working is a maximum, the largest number of fractures per unit time would be formed here.

A complication is introduced because of the likely heterogeneity of the failure criterion in the whole region being discussed. The simplest approximation would assume that the region could be described in terms of two failure criteria (with regard to the formation of microfractures) - one for the

region $x = -5 \rightarrow 0$, and one for the region $x = 0 \rightarrow +5$. Evidently it was difficult for water to penetrate very far beyond $x = 0$, judging from the lack of uniform Laxfordian hydration of this region compared with the Tarbet Steep Belt.

It is suggested that the possible build up of fluid at $x = 0$ (ignoring diffusion upwards parallel to the foliation) and the history of intense fracturing in this region resulted in the development of instabilities leading to the formation of macro-failure planes (cf. Lajtai 1971, and section 9.6). Thus the first formed shear zones would have propagated outwards from the region $x = 0$ and permitted further migration of the fluid away from its source region into a region of low water content. This model predicts (see fig.10.1) that the fluids passing through the shear zones at Scourie were derived from the granite source at a vertical depth of 4-5km. below the present day land surface. The decrease in the number of shear zones and the biotite content of individual shear zones northward from Scourie may be related to real vertical variations in temperature in the source region.

10.4 End.

Any attempt at formulating a regional model involves making assumptions and expressing ideas that are perhaps partly intuitive in origin. This chapter is no exception. Like many ideas, it will have served its purpose if it prompts constructive criticism.

PERSPECTIVE.

"This, then, is why I have brought you so far..."

Many plagioclase crystals in Scourian granulite facies rocks contain bubble inclusions. The tiny bubbles in these small cavities are in perpetual oscillation - some typical figures are

diameter of hole = 6microns

diameter of bubble = 1micron

no. of oscillations per second = 5

ie. distance travelled in one second = 25microns

It is assumed that these figures have remained constant since the crystal formed (at least 2600my) - these assumptions are obviously not correct, but at higher than room temperatures the increased bubble diameter is compensated by a faster oscillation rate.

Using these data, it is found that the bubble has travelled $1.95 \cdot 10^9$ km. so far in its life. The mean distance between the earth and sun is $1.495 \cdot 10^8$ km - hence the bubble has already travelled 13 times this distance.

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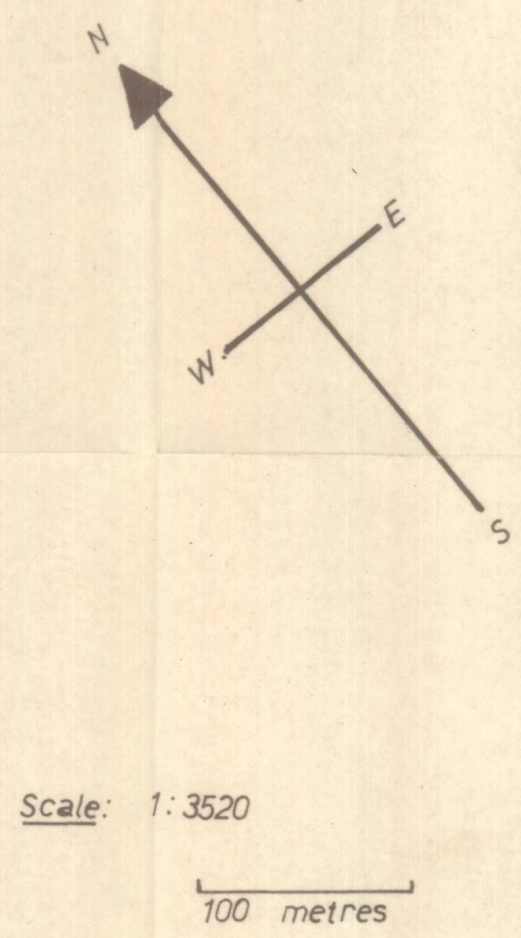
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A Map of the area between Scourie and Tarbet in the District of Eddrachillis, Sutherland, Scotland.



- | | | |
|---|--|--|
| <ul style="list-style-type: none"> — mineral banding → lineation L1 — trend lines — foliation → lineation L2 — trend lines — dyke contact F fault - - - inferred fault | <p>Scourian and Pre dyke fabrics</p> <p>Laxfordian fabrics</p> | <ul style="list-style-type: none"> ultra basic bodies in Scourian Scourie Dyke Suite GK gneiss shear zones steep limbs of pre dyke asymmetric folds |
|---|--|--|

NATURAL ARCH DYKE.

N. ←

1 metre.

- gneiss foliation
- gneiss lineation
- dyke foliation
- dyke lineation
- ∩ dyke contact
- ⇒ fold axis of dyke contact

