

GEOCHEMISTRY AND PETROLOGY OF LAYERED BASIC INTRUSIONS
IN THE LEWISIAN COMPLEX AROUND SCOURIE

by

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To my Mother and Father

ABSTRACT

Layered mafic-ultramafic intrusions form concordant sill-like bodies within the acid gneisses and granulites of the Lewisian Complex around Scourie, N.W. Scotland. These were initially metamorphosed during the Scourian, approximately 2,900 million years ago.

The main components of the intrusions are ultramafites, consisting of hornblende-pyroxenites; mafic gabbros, comprising garnet - two pyroxene - plagioclase gneisses; and sulphide-rich mafic gabbros. These rock types are repeated in cyclic units and provide a means of determining the original way-up of the bodies. Using these criteria, an early period of isoclinal folding has been recognised. The intrusions on Scouriemore are folded in such a way that the initial base is now central to the outcrop.

Petrographic and electron microprobe examination of mineral phases has revealed three periods of granulite facies mineral growth and one of amphibolite facies. The peak of metamorphism is estimated as being of high-pressure granulite facies type, at approximately 1000°C and 15Kb total pressure with a low P_{H_2O} and oxygen fugacity. This implies a depth of metamorphism of 40-50 kilometres and a geothermal gradient of 20°-30°C/km.

XRF analyses of samples from one selected intrusion indicate that the rocks are tholeiitic, their differentiation being controlled by the fractional crystallisation of aluminous pyroxenes and minor olivine at 9-10 kb total pressure with a low P_{H_2O} and oxygen fugacity. The precipitation of sulphides was caused by periodic influxes of new magma. Rare-earth element distributions are flat and relatively unfractionated compared with chondritic abundances.

XRF and rare-earth element analyses of minor granodiorites and

adamellites associated with the layered intrusions suggest that they were derived as small-volume partial melts of garnet granulites.

The layered intrusions are interpreted as being the initial stages of the development of an Archaean Island Arc system.

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CHAPTER 1

INTRODUCTION

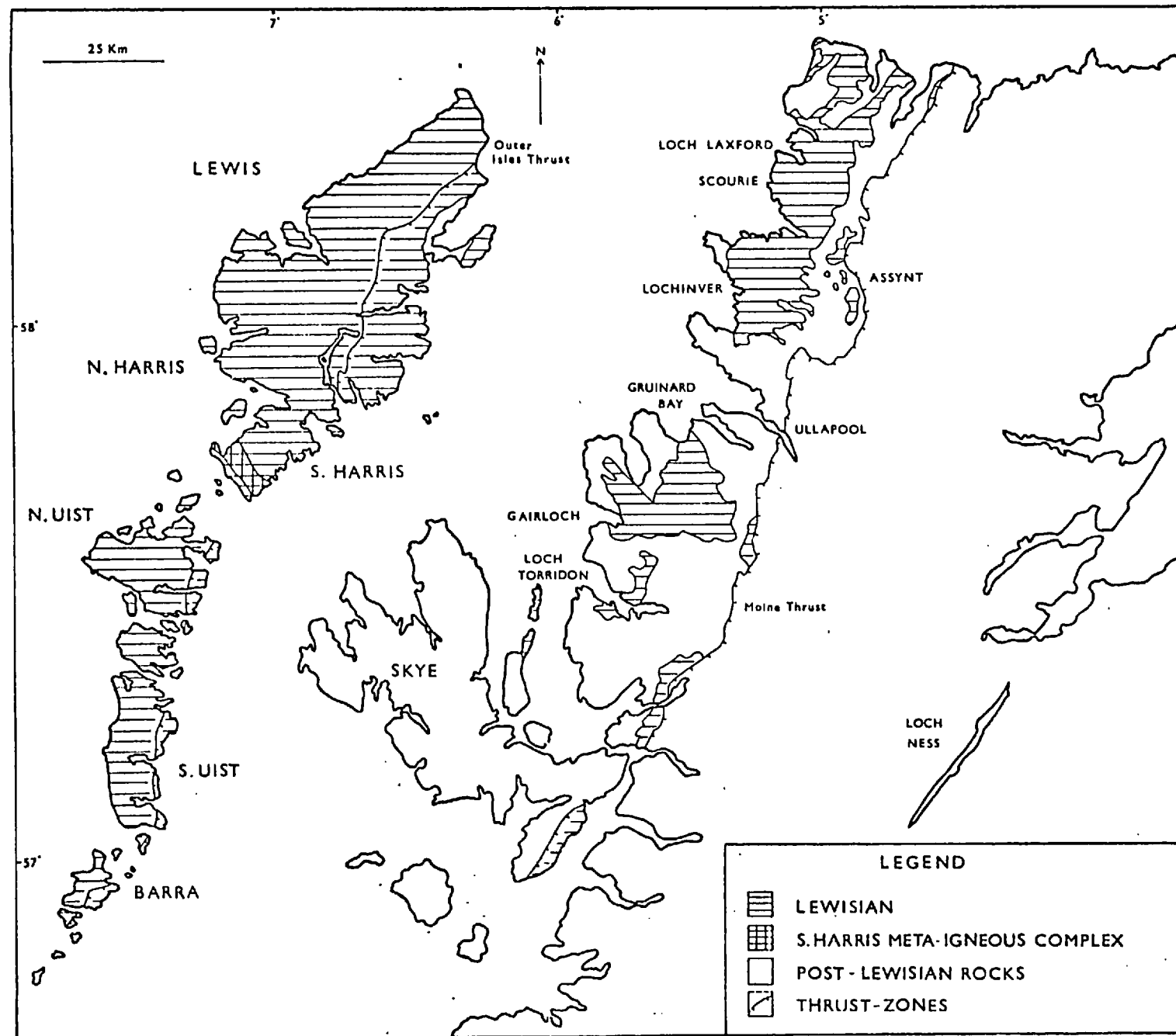
This study is concerned with an assemblage of metamorphosed, layered mafic and ultramafic rocks occurring in the Lewisian Complex around Scourie, N.W. Scotland. Emphasis is placed on the geochemical history of these rocks, and their role in defining the early evolution of the Lewisian Complex as a whole.

1.1 GENERAL GEOLOGY OF THE LEWISIAN COMPLEX

The Lewisian Complex (Fig. 1.1) forms the most easterly exposed portion of the North Atlantic Archaean Craton (Bridgwater et al., 1973). It constitutes the basement upon which later sedimentary successions, namely the Torridonian Sandstone and Cambrian quartzites, were deposited. It extends from Cape Wrath in the north to the isles of Coll and Tiree in the south. It is a polycyclic gneiss terrain, comprising quartzofeldspathic gneisses, minor metasediments, and meta-igneous rocks.

The first major study of Lewisian geology was completed by members of the Geological Survey of Great Britain in 1907 (Peach et al., 1907). They sub-divided the region into a 'Fundamental Complex' and 'Later Intrusive Igneous Rocks'. This was taken further by Sutton and Watson (Sutton and Watson, 1951), who defined the terms 'Scourian' and 'Laxfordian' for episodes of deformation and metamorphism separated by a period of basic dyke intrusion (the Scourie Dykes). This classification has formed the basis for all subsequent research into the Lewisian. Isotopic age-dating (Giletti et al., 1961) confirmed Sutton and Watson's conclusions and gave the first quantitative evidence of the great age of the rocks concerned. Further structural and geochemical

Fig. 1.1 The Lewisian Complex.



studies in the last fifteen years have served to refine Lewisian chronology even further (see Table 1.1). Notable contributions include the proposals for recognition of 'Badcallian' (Park, 1970) and 'Inverian' (Evans, 1965) phases to denote the early granulite facies metamorphism and its later retrogression, respectively. Major advances in isotopic age-dating, particularly by Moorbath and co-workers have integrated this research into a coherent chronology (Moorbath and Park, 1971; Moorbath et al., 1969; Chapman and Moorbath, 1977). Correlations between the Outer Isles and mainland Lewisian geology have been carried out by Dearnley (1962) and Coward et al. (1970).

1.2 BASIC ROCKS IN THE LEWISIAN

Lithologically, the Lewisian Complex is extremely heterogeneous, rock types varying from acid to ultrabasic in character. Basicgneisses however, form a significant proportion of the area, occurring in a variety of forms (Table 1.2).

Mafic bands concordant with the gneissic foliation may vary in thickness from tens of metres down to millimetres and consist of varying proportions of plagioclase, hornblende, pyroxene, and quartz. Generally, garnet is absent.

Small, orbicular pods of mafic or ultramafic gneiss, mostly less than two metres in diameter occur within acid gneisses, the foliation in the latter flowing evenly over the 'eyes' of mafic/ultramafic material. Hornblende is often the sole mineral within these pods, although pyroxene, plagioclase, and biotite may be present.

Bodies of cross-cutting meta-dolerite (in varying degrees of metamorphism) are common in the Lewisian (the Scourie Dykes). Their structural state varies directly with the amount of Laxford-

Table 1.1 Summary of events in the Lewisian (after Watson, 1975).

	<u>VOLCANIC & SEDIMENTARY ROCKS</u>	<u>INTRUSIVE ROCKS OTHER THAN GRANITE</u>	<u>PHASES OF DEFORMATION METAMORPHISM & MIGMATISATION</u>	<u>GRANITE & PEGMATITE</u>
LALFORDIAN 2200-1600 m.y.		lamprophyric dykes	metamorphism & migmatisation 1850-1700 m.y.	pegmatites granites & pegmatites
			repeated deformation and metamorphism of amphibolite or granulite facies	
		SCOURIE DYKE SWARM (tholeiites)		
	Loch Maree Group ?		INVERIAN EPISODE local deformation, metamorphism	pegmatites 2600-2400 m.y.
SCOURIAN 2800-2200 m.y.		diorites etc. (Barra)	BADCALLIAN EPISODE 2800 m.y. gneiss-forming metamorphism granulite or amphibolite facies interleaving of supracrustal, intrusive and gneissose layers	granites early tonalites ?
	supracrustal units	basic/ultrabasic & anorthositic		
PRE-SCOURIAN			formation of early gneiss complex ?	

Table 1.2 Basic rocks in the Lewisian.

	<u>MINERAL ASSEMBLAGE</u>	<u>FORM</u>	<u>AGE (M.Y.)</u>
Mafic gneiss bands	CPX + OPX + PLAG ± QZ ± GT ± HB	concordant foliae (1-2 mm in thick- ness) and bands (tens of metres in thickness)	2800
Mafic pods	CPX + OPX ± PLAG + HB ± BI	sub-spherical balls (10 cms - 2 m in diameter)	2800
Mafic components of layered mafic-ultramafic sheets	CPX + OPX + PLAG ± GT ± HB ± SP	pods and sheets (up to hundreds of metres in length) pronounced layering	2800
Scourie Dykes	CPX + PLAG + OPX ± GT ± HB	discordant dykes (up to tens of metres wide)	2500-2200

ian re-working. The dykes are abundant in the central Scourian block (Loch Laxford to Loch Maree), but less easy to recognise in the Laxfordian Complex (Davies and Watson, 1977).

Layered sheets of ultramafic and mafic rock types form a major component of, and are distributed throughout, the complex on the Scottish mainland (Bowes et al., 1964; Davies, 1976; Davies and Watson, 1977). Pyroxenites and garnetiferous basic gneisses are the main lithological types. They occur as large pods or sheets, characterised by a pronounced layering of the mafic and ultramafic components. The bodies are cut discordantly by Scourie Dykes and are therefore Scourian in age. Commonly associated with these rocks are granitic sheets which form thin, generally concordant bands within the mafic rock types. These granitic rocks are more conspicuous within the layered mafic-ultramafic intrusions, but do occur outside these bodies (Davies, 1975). The mafic-ultramafic assemblage also frequently occur within belts of presumed meta-sedimentary material (Sutton and Watson, 1951; Davies, 1974b; Okeke, 1978).

1.3 HISTORY OF RESEARCH OF THE LAYERED MAFIC-ULTRAMAFIC SHEETS

Of all the lithological units within the Lewisian Complex, the mafic-ultramafic sheets are probably the most thoroughly documented. Despite this plethora of information, the original nature of these rocks remains problematical and controversial.

The early Geological Survey workers regarded the mafic and ultramafic members of the Lewisian as older than the enclosing acid gneisses, and concluded that the separation of the two major portions had occurred by differentiation of a 'mother liquor', subsequently modified by plastic deformation (Peach et al., 1907, p.35).

Sutton and Watson (1951) modified this unwieldy concept to suggest that plutonic basic and ultrabasic intrusions were invaded by mobilised acid material (the host acid gneisses) during metamorphism to produce the scattered distribution of basic pods.

From a detailed petrological study, solely in the Scourie area, O'Hara (1961a; 1965, 1966) concluded that the assemblage had developed as zoned intrusions, the basic rocks being an 'aureole' formed between igneous peridotite and acid gneisses under granulite facies conditions.

A markedly different origin for the basic and ultrabasic masses was proposed in a series of papers by Bowes, Wright, and Park (1961, 1964, 1966), and Bowes, Skinner, and Wright (1970). A survey of field relationships (with minor geochemical data) of a large number of the bodies led them to suggest (initially, at least) that they represented disrupted stratiform igneous intrusions, preserving relict igneous mineralogy and cumulate textures (Bowes et al., 1961). In subsequent papers (Bowes, Wright, and Park, 1964, 1966; Bowes, Skinner, and Wright, 1970), their original concepts were revised to concede that the rocks were wholly metamorphic, yet preserving cumulate structures of a pre-metamorphic, igneous origin. Whole-rock chemistry was compared with that of un-metamorphosed layered basic intrusions. Bowes et al. also recognised anorthositic rocks in a number of the intrusions.

The wildly contrasting views of O'Hara and Bowes et al. were the subject of much subsequent controversial discussion (O'Hara, 1965, 1966; Bowes et al., 1966).

Bowdidge set out to resolve these alternate versions with a geochemical and petrological study of several basic and ultrabasic masses between Scourie and Kylesku (Bowdidge, 1969). He regarded the ultrabasic members as either olivine-orthopyroxene-amphibole cumulates or metasomatised, originally homogeneous, peridotites.

The garnetiferous basic gneisses could have been derived from basic igneous rocks if hornblende had been a fractionating phase from the original magma. Chemical variation did not appear to be consistent with the possibility that the ultrabasic and basic portions of the suite were co-magmatic. He considered the basic gneisses to have formed by 'metasomatic interchange' between the garnetiferous basic and acid gneiss. Concluding, Bowdidge decided that the assemblage did not seem to fall into either simple category of 'layered' or 'zoned' intrusion.

Davies (1974a, 1974b, 1976) has described the field relations, petrography, and structure of several mafic and ultramafic masses between Scourie and Loch Laxford. He has suggested that these bodies represent fragmented layered intrusions, preserving original 'way-up' sequences ranging from ultramafic to anorthositic, and relict igneous cumulate textures. He emphasised the common association between the bodies and belts of metasedimentary material. Using his 'way-up' criteria, the intrusions define large-scale tectonic structures when mapped over a wide area.

Davies and Watson (1977) have correlated basic-ultrabasic bodies in the type Laxfordian Complex (north of Loch Laxford) with the better described masses of the Scourian Complex.

O'Hara re-investigated the mineral chemistry of the bodies around Scourie to estimate conditions of the Scourian metamorphism (O'Hara, 1977a; 1977b; O'Hara and Yarwood, 1978). His best estimate of the peak of metamorphism was 1250°C and 15 Kb.

1.4 PRESENT STUDY

From the historical perspective of the mafic-ultramafic sheets, it is evident that several diverse hypotheses of origin are available in the geological literature. The present project was designed

to reconcile or reject the numerous petrogenetic theories outlined previously. Detailed field-mapping (1:360) of selected sheets and sampling of one such sheet was undertaken to enable the study to be carried out. A number of rock and mineral analyses were made to elucidate not only any pre-metamorphic chemical variations (igneous fractionation trends, relict igneous mineralogy), but also to determine the nature of the early Scourian metamorphism, particularly in relation to structural sequence.

1.5 LITHOLOGICAL NOMENCLATURE

All the rocks dealt with in this thesis are metamorphic in nature. As such, each rock-type should be prefixed by the term meta-. This however, leads to an awkward terminology and has been discarded in favour of a pseudo-igneous classification based on modal mineralogy (Table 1.3). To aid comparison between the meta-igneous rock types of the Scottish mainland, the Outer Hebrides, and Greenland, the classification of Witty (1975) (in part after Windley et al., 1973) has been adopted.

Table 1.3 Rock terminology (after Wilty, 1975).

<u>ROCK TYPE</u>	<u>% MAFIC</u>	<u>MINERALS</u>
Anorthosite	0-10	
Gabbroic Anorthosite	10-20	
Anorthositic Gabbro	20-35	
Gabbro	35-65	
Mafic Gabbro	65-90	
Ultramafites	90-100	{ - Dunite Olivine 90-100 - Peridotite Olivine + Pyroxene 90-100 Olivine < 90, Pyroxene < 90 - Pyroxenite Pyroxene 90-100 - Amphibole Dunite Amphibole 10-50; Olivine + Amphibole 90-100 - Amphibole Peridotite Amphibole 10-50, Olivine + Pyroxene 50-90; Olivine, Pyroxene < 90 - Amphibole Pyroxenite Amphibole 10-50; Pyroxene + Amphibole 90-100 - Hornblendite Amphibole 50-100 - Garnetite Garnet 50-100

CHAPTER 2

THE LAYERED MAFIC-ULTRAMAFIC SHEETS AND ASSOCIATED ROCKS

Work has been concentrated in the immediate Scourie district and in particular, the Scouriemore headland (fig.2.1). This small area probably contains the least retrogressed portion of the Scourian granulite facies gneisses (Muecke 1969) and as such is the best region for studying early structural or chemical trends within the layered mafic-ultramafic sheets.

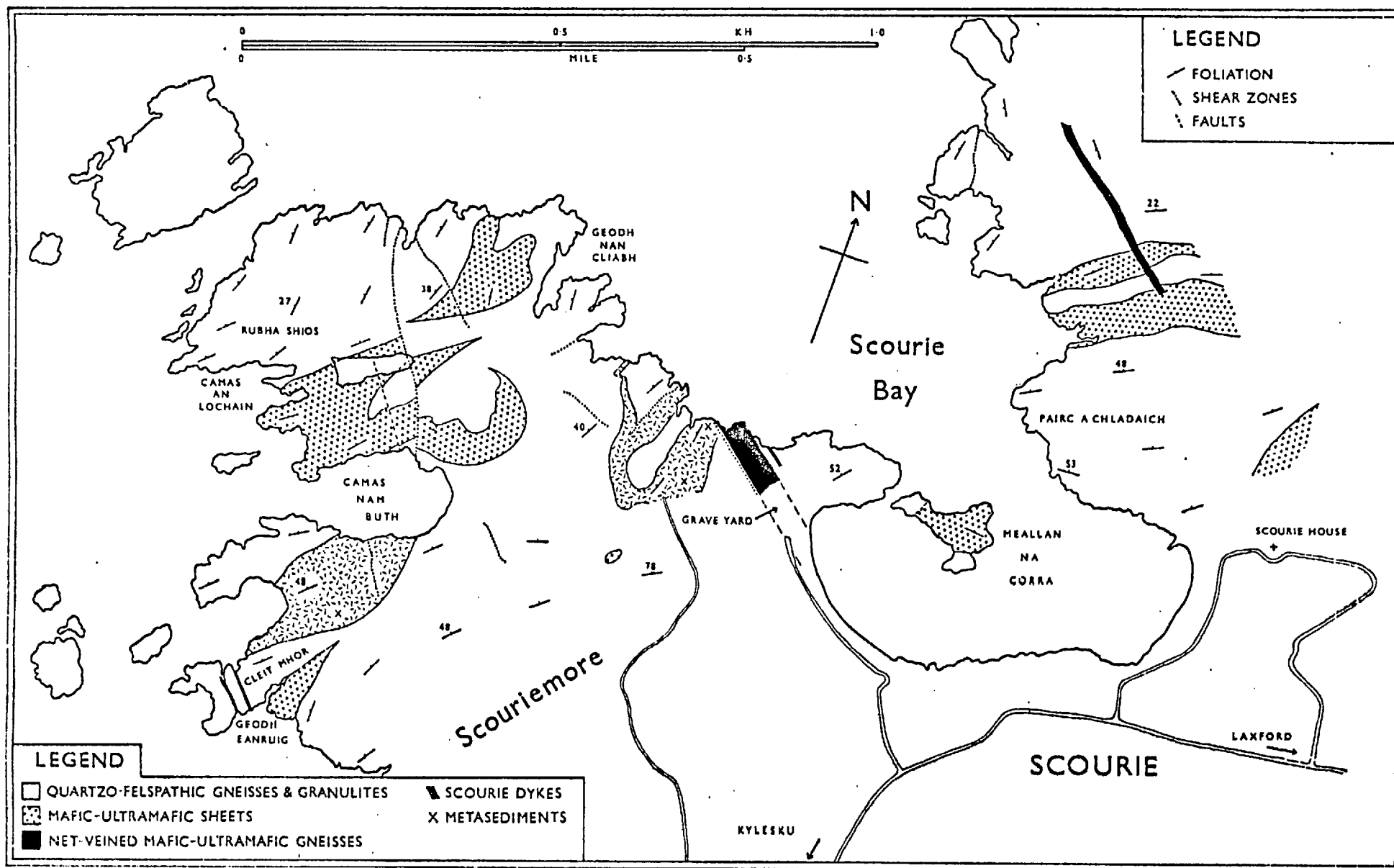
The main features of interest are:-

A). THE LAYERED MAFIC-ULTRAMAFIC SHEETS. These will be described by means of an account of the field relations of the masses at Camas nam Buth, Geodh'Eanruig and Pairc a'Chladaich. Petrographic data concern mainly the Camas nam Buth sheet.

B). THE GRANITE SHEETS. These are pre-Scourie Dyke in age and form intrusive sub-concordant layers within the layered mafic-ultramafic sheets. Granitic material of similar age is an integral part of the net-veined mafic-ultramafic sheets.

C). THE NET-VEINED MAFIC-ULTRAMAFIC SHEETS. Lithological types present within these units are similar to the layered sheets, although the former lack the distinctive banding of the layered sheets and are also characterised by networks of intrusive granite veins. The form of the units and age relationships are also comparable with the layered bodies. Chemical evidence (to be presented in chapter four) suggests different origins for the net-veined and layered mafic-ultramafic sheets. Pods of presumed metasedimentary material occur solely within these units in the Scourie region.

Fig. 2.1 The geology of the Scourie district.



2.1 THE LAYERED MAFIC-ULTRAMAFIC SHEETS

2.1.1 GENERAL FEATURES

These form concordant, sill-like bodies within the acid gneisses and granulites. An idea of their distribution within the Scourie district may be gained from fig. 2.1. Thicknesses of the units are of the order of 10-20 metres, whilst outcrop length along strike is generally 50-200 metres. Approximately eighty percent of the volume consists of mafic gabbros, the rest being ultramafite. No anorthosite has been recognised. The mafic gabbros and ultramafites are inter-banded, forming discrete, concordant layers. The thickest development of ultramafite however, is commonly within the central portion of the sheets. Mineral grading of layers is absent, but despite this, a "way-up" within the intrusions may be recognised from the following:-

- 1). Thin horizons of sulphide-rich mafic gabbros are a common feature of the sheets. Petrographic evidence suggests that the sulphides were co-genetic with the original emplacement of the sheets as igneous bodies. They are concordant with the gross layering and usually occur adjacent to ultramafite bands. By analogy with un-metamorphosed layered basic intrusions, it is assumed that the ultramafites formed the base and the sulphides the top of what was originally a gravity-stratified unit. These accumulations of sulphides at certain horizons are also indicative of multiple injection of magma (Campbell, 1973, 1977; Buchanan, 1976a, 1976b). Several units may be present in any one sheet. Discussion of the precipitation of sulphides will be discussed in more detail in chapter four.
- 2). Geochemical evidence from the Camas nam Buth sheet shows that each stratified unit displays variation in the following parameters from the base to the top (figs.2.2 and 2.3):-

- (i) An increase in the $\text{FeO}/\text{FeO}+\text{MgO}$ ratio, indicative of

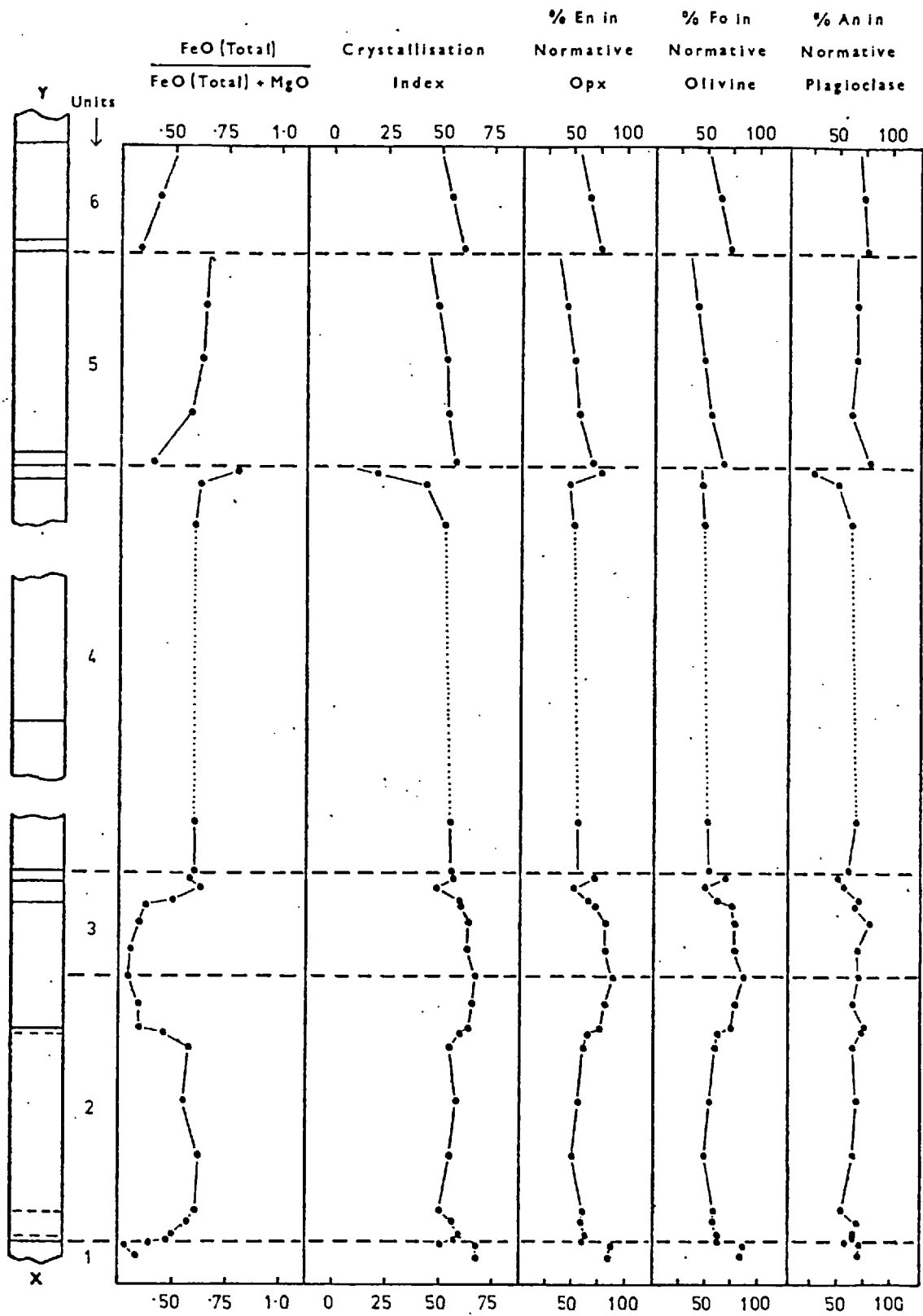


Fig. 2.2 Major element variation across the Camas namuth sheet.

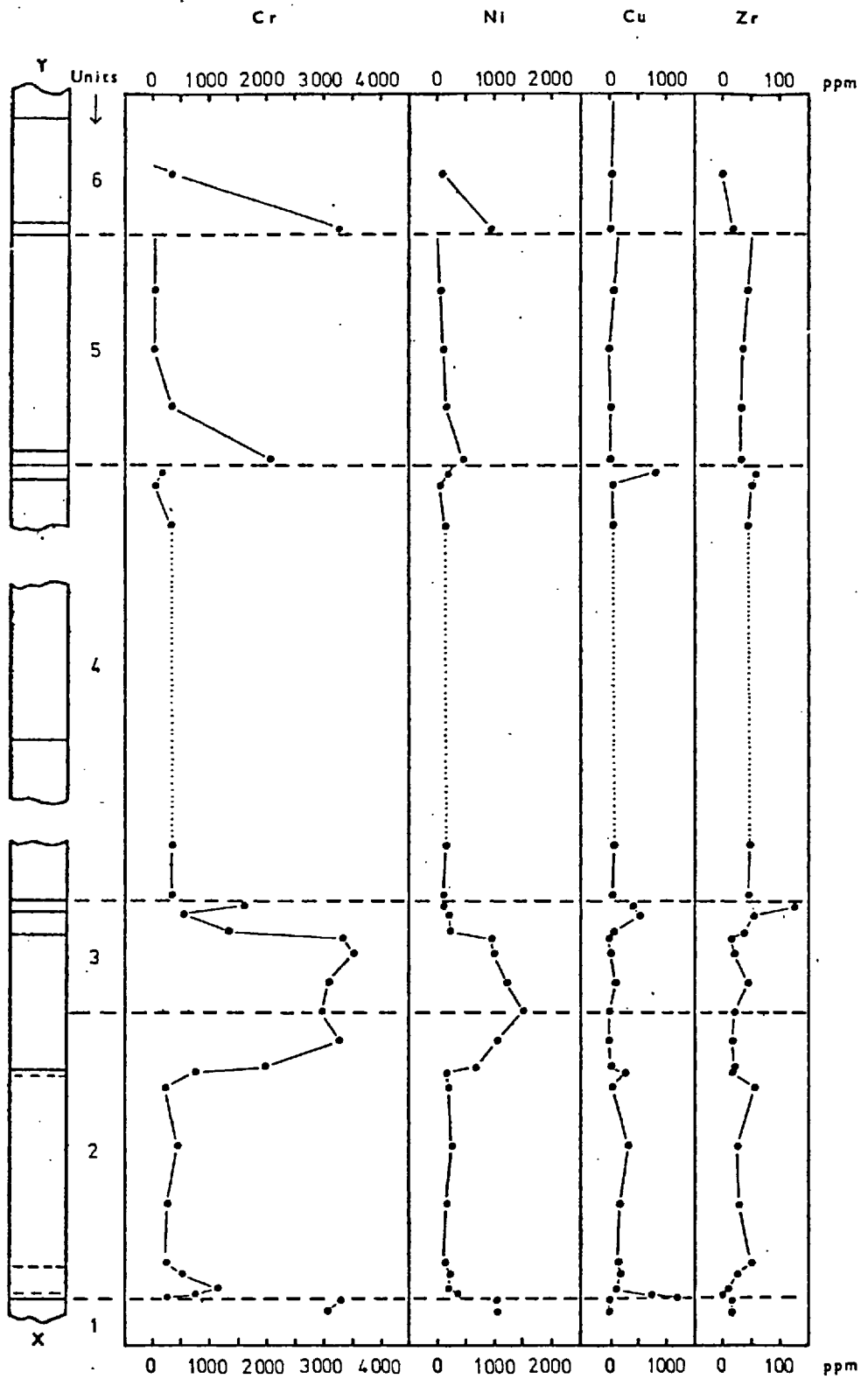


Fig. 2.3 Trace element variation across the Camas nem Butte sheet.

differentiation in tholeiitic suites (Miyashiro, 1974).

(ii) A decrease in the concentrations of nickel and chromium, which from crystal field theory partition strongly into early olivines and pyroxenes, respectively.

(iii) An increase in the amount of zirconium.

(iv) An increase in the iron molecule of the normative minerals hypersthene and olivine.

(v) A decrease in the crystallisation index (Poldervaart and Parker, 1964) and the anorthite content of normative plagioclase.

(vi) Concentration of copper in the sulphide horizons at the top of each unit.

The primary way-upness criteria defined above have been modified by an early period of isoclinal folding. Thus the base now occurs central to the outcrop of the sheet in an anticlinal arrangement of layers, so units 3,4,5 and 6 are the correct way up, and units 1 and 2 are upside-down. It is assumed that the central position of the thickest ultramafite layer in all the intrusions studied is due to the early period of anticlinal folding.

A similar structural-geochemical relationship has been described from West Greenland by Windley et al. (1973). Here, a large Archaean anorthositic intrusion (the Fiskenaasset Complex) has been isoclinally folded to produce geochemical variations similar to those observed from Camas nam Buth. Trends of differentiation are such that the intrusion appears to have been synclinally folded. Thus the ultramafic portions are external to the complex.

Macro-rhythmic units in layered intrusions are usually of the order of 30-300 metres thick (Campbell, 1977), whereas the Camas nam Buth units are only six metres thick at most. O'Hara (1965) has

already pointed out the difficulties in achieving the geochemical variations observed in the sheets by cumulus igneous processes in such small thicknesses. However, it seems inconceivable that all the features described above were due to any process other than igneous gravity stratification. Thus it is necessary to invoke considerable tectonic thinning to explain the small thicknesses of rocks exposed. Thinning would also explain the lack of direct comparison in lithological types and thicknesses between units 2 and 3, and 1 and 5.

An account of the macroscopic and hand-specimen characteristics of the principal rock-types present within the bodies is given below. These descriptions refer to unretrogressed granulites, generally lacking obvious effects of pronounced deformation.

Mafic gabbros are the most abundant lithological type in the layered sheets. These mafic rocks are medium to coarse grained, have a granoblastic or porphyroblastic texture and weather to a brown surface. They are massive, the poor fabric being defined only by occasional feldspathic or mafic schlieren. They are uniformly rich in garnet porphyroblasts, except at the contacts with the ultramafites, where a fine grained, mafic variety poor in plagioclase and garnet exists (approximately 10cms. thick). With this specific exception they are remarkably homogeneous in terms of texture, mineral content and colour index. Garnet porphyroblasts are a startling crimson in pigmentation, averaging 1-2mm. in diameter (exceptionally up to 10cms. in diameter) and their effect is enhanced by a 'collar' of plagioclase grains. Thin 1-2mm. feldspathic veins occasionally traverse obliquely across the banding in these rocks.

The sulphide-rich mafic gabbros are rarely fresh and occur as thin (10-20cms.) yellow-brown weathering, concordant layers

within the sheets. They are mineralogically similar to the gabbroic rocks described above apart from the presence of abundant pyrite, pyrrhotite, chalcopyrite, magnetite and ilmenite. Some contain quartz as a minor phase.

The ultramafites are generally hornblende-pyroxenites. Rarely is the olivine content high enough to describe any of the rocks as peridotite, although peridotites and even dunites have been reported elsewhere in the mainland Lewisian (Bowes et al., 1964; Davies, 1974b). The rocks are medium to coarse grained and weather dark brown to black. On hand specimen scale they are granoblastic in texture, but on the scale of an outcrop are lithologically banded, layers of the order of 20cms. in thickness being defined by variations in proportions of olivine and amphibole. Occasionally, finely banded horizons alternate with massive layers. Bands of coarse to pegmatitic hornblendite may occur within pyroxenites. Wedging out of layering along strike is common (Plates 2.1, 2.2 and 2.3). These structures were thought by Bowes et al. (1964) to be due to the effects of cumulate igneous layering. However, O'Hara (1965, 1966) showed that such structures were compatible with a tectonic origin and due to the shearing out of the fold-limbs of small-scale isoclines. Thus although the banding within the ultramafites may be ascribed to the effects of tectonic processes, the gross layering within the sheets as a whole relates to a primary igneous layering as indicated by the position of ultramafic and sulphide bands, and the geochemical trends described above.

2.1.2 FIELD RELATIONS

Aspects of deformation, metamorphism and igneous intrusion for the Scourie area are summarised in Table 2.1. Although structure



Plate 2.1 'Cross-bedding' structures of Bowes et al.(1964).
Craggs north-east of Camas nam Buth.



Plate 2.2 Finely banded hornblende-pyroxenite, Meallan
na Corra.



Plate 2.3 Alternation of massive and finely banded layers in hornblende-pyroxenite. Craggs north-east of Camas nam Buth.



Plate 2.4 F_3^2 synform. Coast, north-east of Sgeir Camas nam Buth headland.

Table 2.1 Summary of metamorphism, igneous activity, and deformation in the Scourie area.

		AGE (M.Y.)	METAMORPHIC EVENTS	INTRUSIVES & MINERALISATION	DEFORMATION
SCOURIAN	BADCALLIAN	3000 ?		Layered basic intrusions and associated sulphides	
		2900-2700	M ₁ High Pressure Granulite Facies M ₂ Medium Pressure Granulite Facies	Granite sheets	? F ₁ Isoclinal folding of layered intrusions F ₂ NE-SW large scale folding (isoclinal)
	LATE SCOURIAN or INVERIAN	2600-2400	M ₃ Upper Amphibolite Facies M ₄ Medium Pressure Granulite Facies	T ₁ Pegmatites Sulphide veins	F ₃ ¹ NW-SE Open warps F ₃ ² NE-SW Monoclines and tight folds
		2500-2100		Scourie Dykes	
LAXFORDIAN		1850-1700 ?			F ₄ NW-SE Shears

and metamorphism will be discussed in detail in chapter three, a brief explanation of Table 2.1 will help the reader to follow the descriptions of field relations.

The earliest fabric recognised is the compositional layering observed in the layered mafic-ultramafic sheets and has been designated S_1 . The accompanying isoclinal folding, producing the anticlinal arrangement of layers in the intrusions is therefore F_1 . Occasional evidence for a diffuse S_1 layering in the acid gneisses may be observed in small-scale intrafolial folds.

The main gently-dipping fabric of the acid gneisses has been termed S_2 . This develops parallel to the axial planes of large-scale recumbent isoclines (F_2) which fold the S_1 compositional layering. The main effects of the deformation is to produce small-scale intrafolial folds, e.g. the folds described from the layered sheets by O'Hara (1965) mentioned above. There is an associated weak lineation plunging to the north or north-west.

The effects of the D_3 and D_4 deformations are very localised, generally present as tight folds, monoclines or shear-zones. Steeply dipping fabrics may be developed parallel to these structures.

(i) CAMAS NAM BUTH

This body is the thickest and best exposed in the area and forms a concordant sheet within the host acid gneisses (Map 1). It is well exposed along the shoreline of Camas nam Buth and Camas an Lochain, and crags of mafic and ultramafic material may be followed inland for some distance. It is approximately 300 x 150 metres in extent, trending NE-SW in orientation.

The outcrop of the sheet traces out a large early Scourian (F_2) recumbent isoclinal synform which plunges at a low angle to the north-west. Dips of banding and foliation are quite gentle ($20^\circ - 30^\circ$) in the eastern limb, but later F_3^2 folding steepens

foliations towards the isocline nose (70° – 80°).

A detailed section has been made across the eastern fold limb (Section X-Y on map 1) and the petrological information is presented in fig. 2.4. This section may be followed for 100 metres along strike, but is complicated by the later F_3^2 folding north-westwards of Sgeir Camas nam Buth headland (plate 2.4).

In this eastern fold limb, the ultramafites have a gently dipping S_1 foliation, preserving a granulite facies mineral assemblage (Plate 2.5) and the gabbroic rocks are massive (Plates 2.6 and 2.7). As the outcrop is followed westwards towards the fold hinge, the banding within the ultramafites is lost, to be replaced by a homogeneous, strongly foliated and retrogressed type of ultramafite (Plate 2.8). The mineral assemblage is characterised by hydrous phases such as serpentine, tremolite and actinolite. On the other hand, the mafic rocks become highly foliated and segregated, producing a pseudo-igneous stratification of mafic and felspathic components (Plate 2.9). These changes are due to the presence of F_3^2 folding within the F_2 isocline fold nose. A steep (60° – 80°) foliation parallel to the axial planes is developed which wraps around the earlier M_1/M_2 garnet porphyroblasts in the mafic rocks (Plate 2.10). An accompanying coarse lineation plunges gently to the south-west. The foliation and segregation associated with this later phase of folding is misleading and may be confused with an original igneous layering. The distinction of the two structures is of prime importance when comparing the features of the layered sheets around Scourie with those further north towards Loch Laxford where later deformation is considerably more pronounced. It also clearly emphasises the need to choose areas lacking Late Scourian (Inverian) or Laxfordian deformation and metamorphism to study the earliest textural and

Fig. 2.4 The Camas nam Buth layered mafic-ultramafic sheet.

Measured section X-Y

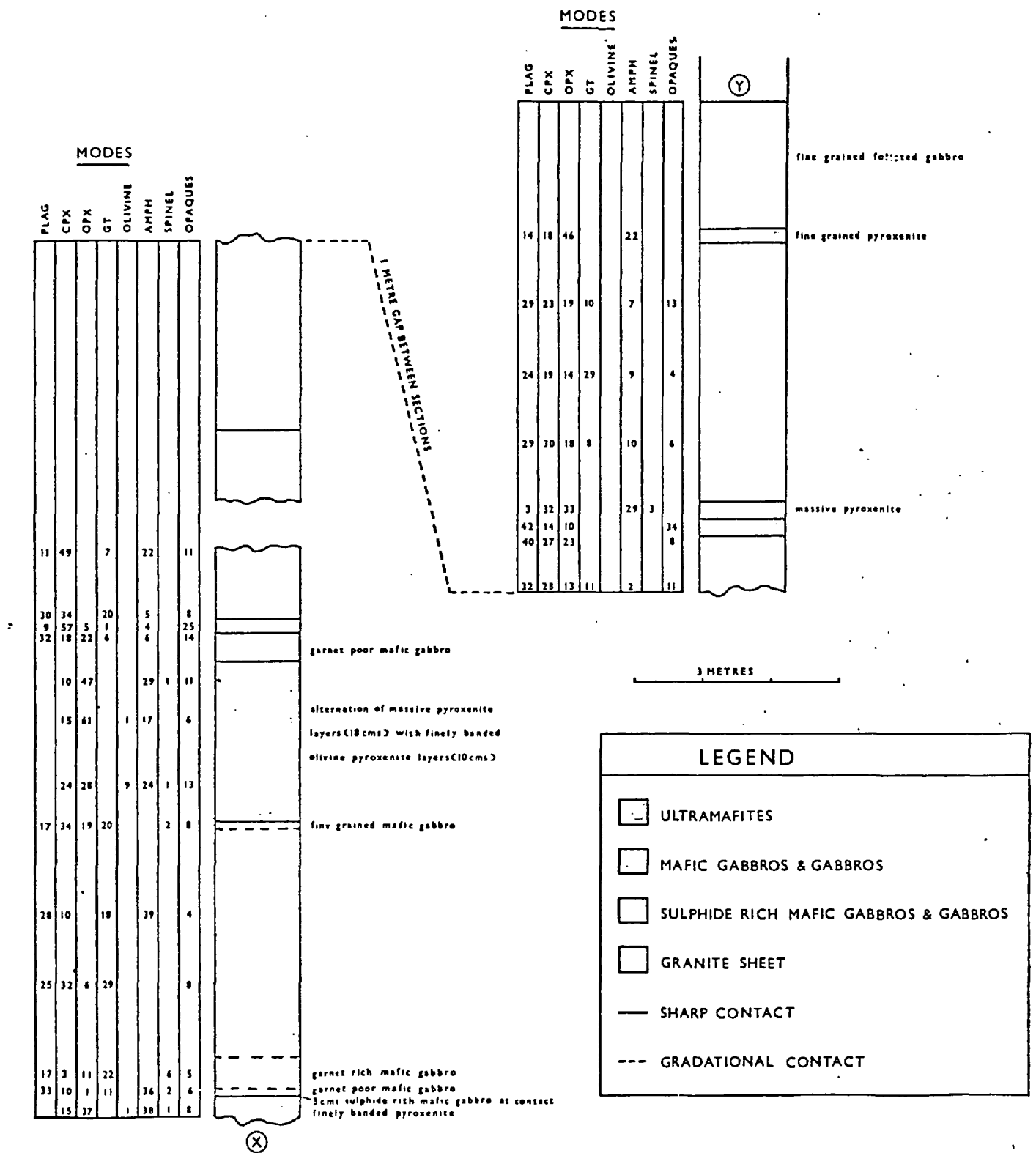




Plate 2.5 Alternation of massive and finely banded pyroxenites. Crags north-east of Camas nam Buth.



Plate 2.6 Massive mafic gabbro. Shoreline, north of Camas nam Buth.



Plate 2.7

Hammer handle rests on banded hornblende- pyroxenite. Massive mafic gabbro is above this. Note the thin light-weathering sulphide-rich mafic gabbro at the contact. Inland, north-east of Camas nam Buth.



Plate 2.8 Green-weathering, massive serpentinised ultramafite. Note the rusty-weathering sulphide veins traversing the foliation. North of Sgeir Camas nam Buth headland.



Plate 2.9 Development of steep S_3^2 foliation and mineral segregation within gabbro. Note the large M_1/M_2 garnet porphyroblasts. North-east of Sgeir Camas nam Buth headland.



Plate 2.10 Mafic schlieren defining S_3^2 foliation, wrapping around garnet porphyroblasts in mafic gabbro. North-east of Sgeir Camas nam Buth headland.

geochemical features of the layered mafic-ultramafic sheets.

Only the 'upper', north-western contact of the sheet within the acid granulites is exposed. This is apparently concordant, being gradational over 1-2 metres. The basic rocks develop a marked planar fabric and are reduced in grain-size towards the contact, suggesting differential movement during deformation between the gabbros and acid granulites. Sericitisation of feldspar is also evident along the margin, giving a whitish appearance in hand-specimen.

The north-western limb of the isocline is poorly exposed and is thinner than the eastern limb.

Minor rusty-weathering zones and pods of quartz-pyrite material line joints within the isoclinal fold core. These cross-cut foliations and lithological contacts and appear to be a late-stage phenomenon.

(ii) GEODH' EANRUIG

A small, wedge-shaped body (100 x 40 metres at most) elongate in a NNE - SSW direction is exposed in coastal cliffs with sporadic outcrop inland (fig 2.5). Within the margins of the sheet, the rocks are folded by an F_3^2 'S' shaped fold, plunging to the south-west. A large branching pegmatite (π_1 of Sutton and Watson, 1951) has been emplaced along the fold axial trace and appears to be synchronous with the folding. This pegmatite yielded Rb-Sr mineral age of 2460 ± 130 million years (recalculated isochron of Giletti et al., 1961 from Watson, 1975).

The lithological layering (S_1) is discordant with the surrounding acid granulites on east and west contacts, although this relationship may be obscured by small scale shearing parallel to the contacts. Wedging out of layers occurs along the eastern contact. Similar relationships between meta-igneous banding and

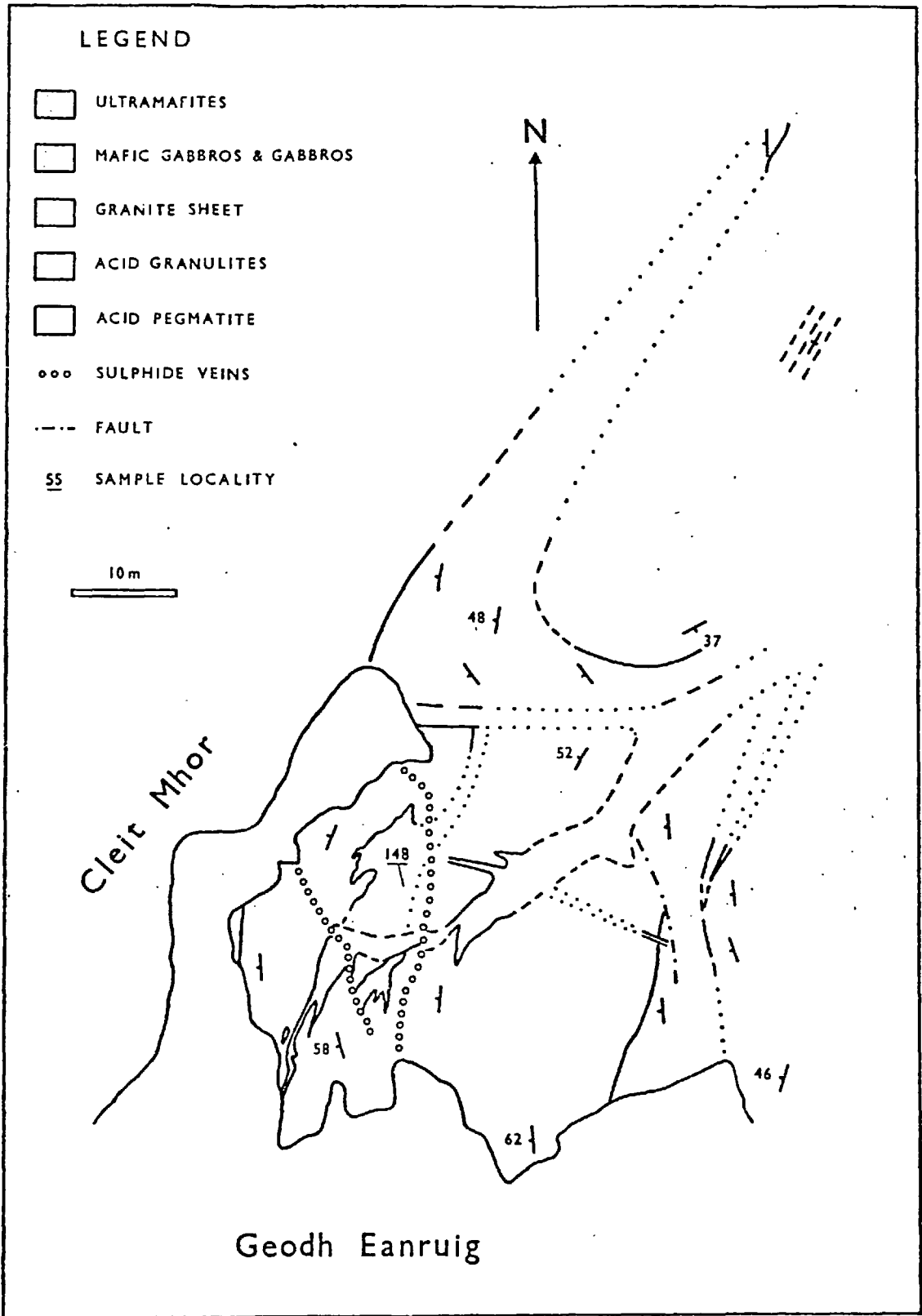


fig 2.5 The Geodh' Eanruig layered mafic-ultramafic sheet.

regional foliation have been described by Graham and Coward (1973) and Witty (1975).

The main ultramafic layer is central to the outcrop and preserves a pronounced layering. Coarse pyroxenites and hornblendites are the main constituents. Some retrogression to talc, tremolite and serpentine has taken place at contacts with the enclosing gneisses. The ultramafites are quite concordant within the mafic gabbros. A steep foliation (S_3^2) in the gabbroic rocks becomes more apparent in the inland outcrops. Gabbros in contact with the large acid pegmatite are generally amphibolitised for 5-10 cms. in thickness (Plate 2.12). Sulphide veining (quartz-pyrite) occurs along the minor dislocations which cut the S_1 banding, the granite sheet and the acid pegmatite (fig. 2.5).

(iii) PAIRC A'CHLADAICH

The sheet is exposed in two coastal inlets, on the north side of Scourie Bay (fig. 2.6). The margins of the sheet may be traced inland for approximately five hundred metres, although lithological layering is only discernible along the coastline where exposure is good. The sheet is divided into two portions by a wedge of acid gneisses, the eastern portion being the thicker of the two. Minor 's' and 'z' shaped folds in the surrounding acid gneisses plunge at a low angle to the north-west. A weak amphibole lineation in the ultramafites plunges in the same direction. These two outcrops constitute the limbs of a recumbent isoclinal synform (F_2), plunging to the north-west. The sheet is concordant within the acid gneisses and contacts exposed along the eastern side are sharp. This contact has been folded by minor F_3^2 folds, plunging twenty to thirty degrees to the south-west.

Both outcrops contain a thick layer of a poorly banded ultramafite in a central position. Gabbroic rocks are more



Plate 2.11 Banded ultramafite, Geodh' Eanruig.



Plate 2.12 Acid pegmatite (Π_1), intruding mafic gabbros, Geodh' Eanruig.



Plate 2.13 Hammer handle resting on a discordant sulphide vein, Geodh' Eanruig.

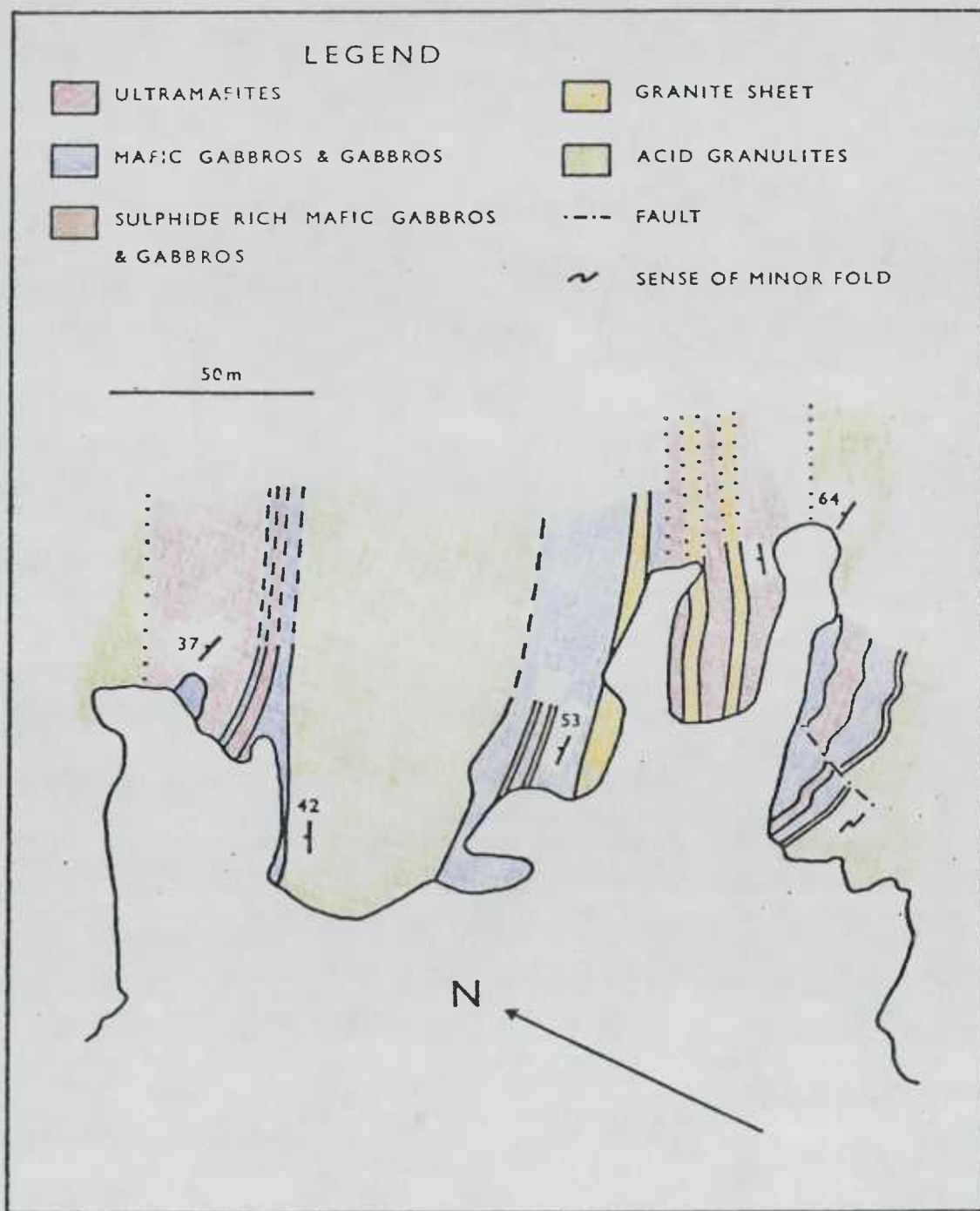


Fig. 2.6 The Pairc a' Chladaich layered mafic-ultramafic sheet.

voluminous, but are unevenly distributed between the two fold-limbs. The mafic rocks are well foliated throughout, again accompanied by a sericitisation of plagioclase feldspar.

2.1.3 PETROGRAPHY

The modal mineralogy of section X-Y (Map 1) from the Camas nam Butth layered sheet is summarised in fig. 2.7. The possible mineral assemblages of the rock-types studied are presented in Table 2.2.

(i) ULTRAMAFITES

The mineral assemblage falls into one of two categories:-

- 1). CPX + OPX + AMPH + SP + MAG $\pm$ OL $\pm$ PYR $\pm$ CHALC $\pm$ PYRRH $\pm$ PENT
(Samples 74,75,86,87,88,89,90)

This assemblage is present in samples with low FeO/FeO + MgO ratios.

- 2). CPX + OPX + AMPH + SP + PLAG $\pm$ GT $\pm$ MAG $\pm$ PYR
(Samples 85,104,108)

This assemblage is present in samples with higher FeO/FeO + MgO ratios.

The above mineral assemblages may be slightly modified by retrograde metamorphism in such a way that:-

CPX $\rightarrow$ ACT/TREM; OL $\rightarrow$ SERP + MAG; OPX $\rightarrow$ TREM; OPX/CPX $\rightarrow$ BI;
PLAG $\rightarrow$ ZO.

Alteration is generally restricted to fringes around primary grains.

The plagioclase-free ultramafites have a medium grain size (0.5 - 0.75mm.) with an equigranular, granoblastic texture. Annealed triple point junctions between phases are typical. A vague lineation defined by olivine or serpentine layers may be present. Each phase appears to be randomly distributed throughout

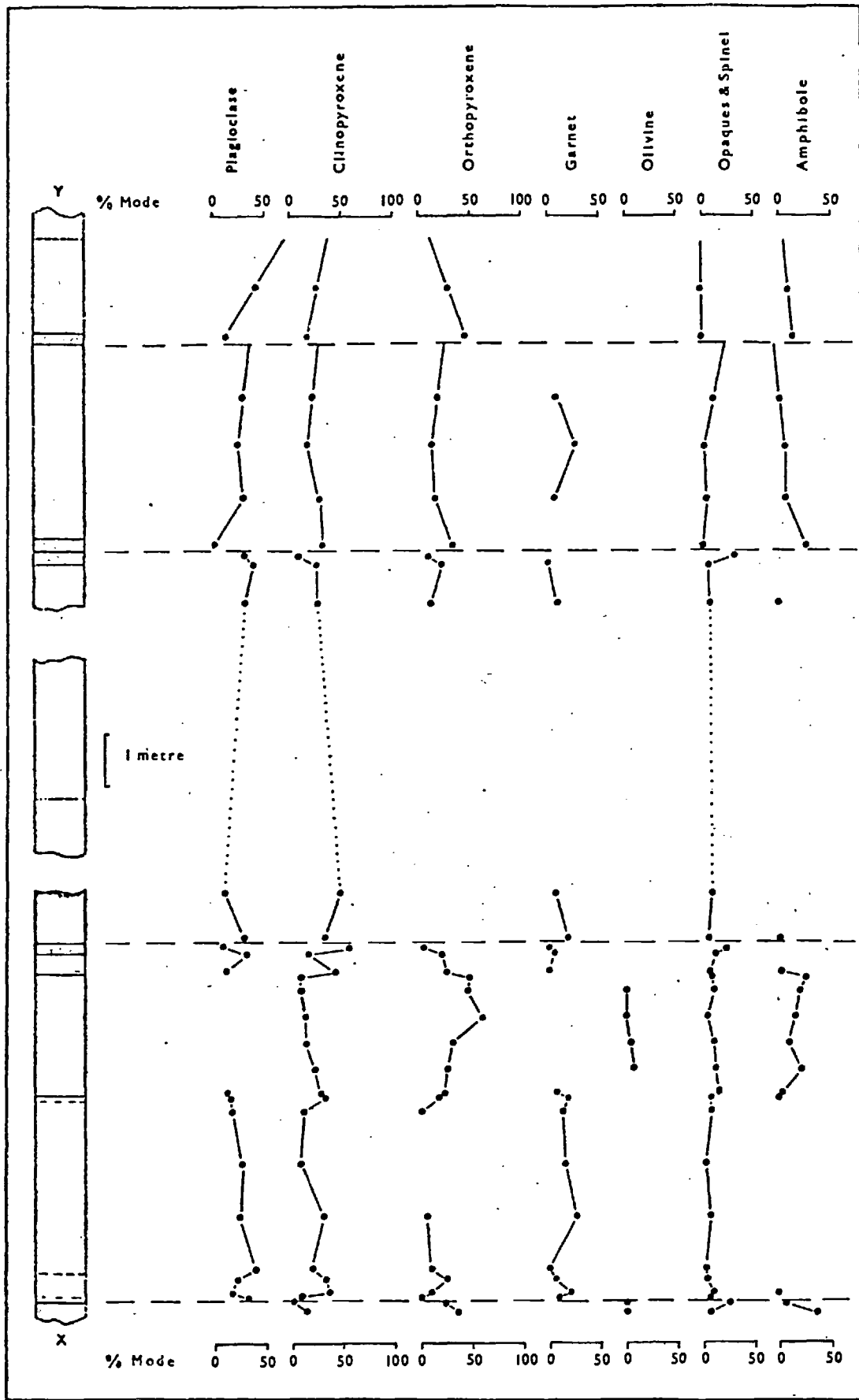


Fig. 2.7 Modal mineralogy of section X-Y.

Table 2.2 Mineral assemblages.
x - always present, o - possibly present.

	Ultramafites without plagioclase	Ultramafites with plagioclase	Mafic gabbros (low F/PM)	Mafic gabbros (high F/PM)	Sulphide-rich Mafic gabbros	Granite sheets	Ultramafites from net- veined sheets	Mafic gabbros from net- veined sheet
Olivine	o							
Orthopyroxene	x	x	x	x	x	o	x	x
Clinopyroxene	x	x	x	x	x	o	x	x
Plagioclase		x	x	x	x		x	x
Garnet		o	x	x	o			x
Spinel	x	x	x					
Pargasite	x	x	x	x	x		x	x
Quartz					o	x		
Mesoperthite						o		
Antiperthite						o		
Apatite						o		
Magnetite	x	x	x	x	x	x	o	x
Ilmenite	o		o	x	x		o	x
Pyrite	o	o	o	o	x			
Chalcopyrite	o		o	o	x			
Pyrrhotite	o		o	o	x			
Pentlandite	o							
Actinolite	o	o	o	o	o	o	o	o
Tremolite	o				o			
Scapolite			o	o	o			
Biotite	o	o	o	o	o	o	o	o
Serpentine	o							
Epidote					o			
Sericite		o	o	o	o	o	o	o

with an absence of reaction rims or corona structures (Plate 2.14). It is thought that this represents an equilibrium granulite facies assemblage.

Orthopyroxenes are bronzites ($\text{En}_{80-83} \text{Wo} < 1\%$) that are strongly pleochroic from pale green to pink (all mineral analyses are determined by electron microprobe analysis. See Appendix II). Grains are sub-idioblastic. A schiller structure of fine rods of iron oxide is common. Small granules of iron ore may also be present along grain boundaries. No exsolution of a calcium-rich pyroxene has been observed.

Clinopyroxenes are diopsidic ($\text{En}_{42-44} \text{Wo}_{48-49}$) and are colourless or very pale green in thin section. The colour is emphasised by the presence of profuse platelets of opaque iron oxide. These platelets are absent from grain rims, the density and coarseness of the inclusions is mirrored by chemical zoning of point analyses between rims and cores (FeO approximately 0.5 - 1.0% higher in the cores than in the rims). This is presumably a consequence of re-equilibration of the grain rims through cooling and recovery by intergranular diffusion, rather than by the exsolution mechanism exhibited by the grain cores. There is no exsolution of a calcium-poor pyroxene.

The amphibole is pargasitic in composition and appears to be part of the initial assemblage rather than diaphoritic in origin. Grains are often surrounded by fine blebs and plates of iron oxide. These plates may be intracrystalline and crystallographically oriented.

Spinel is olive-green to opaque, forming rounded grains in association with magnetite. The spinel is a pleonaste ($\text{Mg}/\text{Fe} \approx 1.3$) in composition.

Olivines are forsteritic, in the range Fo_{82-85} . Grains are

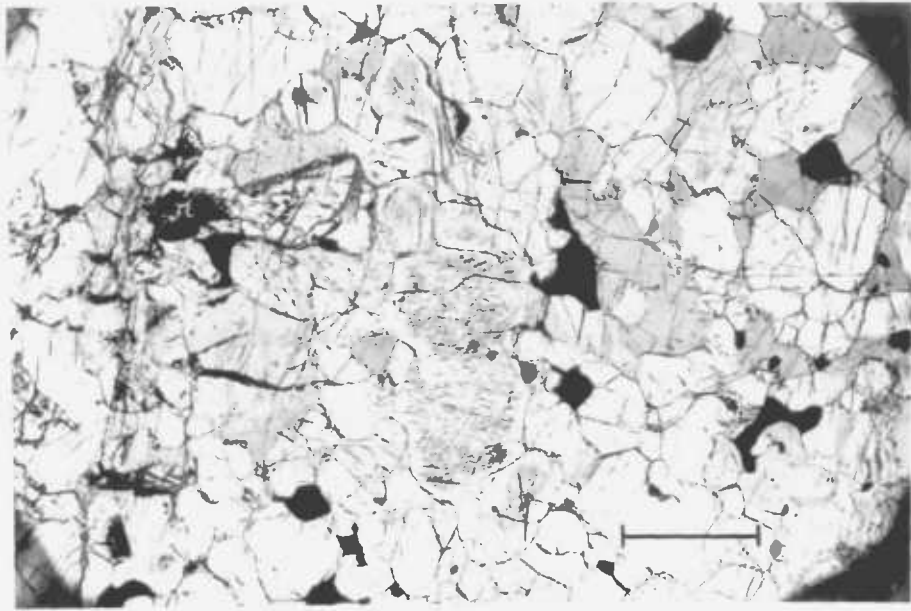


Plate 2.14 Granoblastic aggregate of clinopyroxene, orthopyroxene, amphibole, olivine, spinel, and magnetite in a hornblende-pyroxenite. Scale bar is 1mm. Sample 74, plane-polarised light.

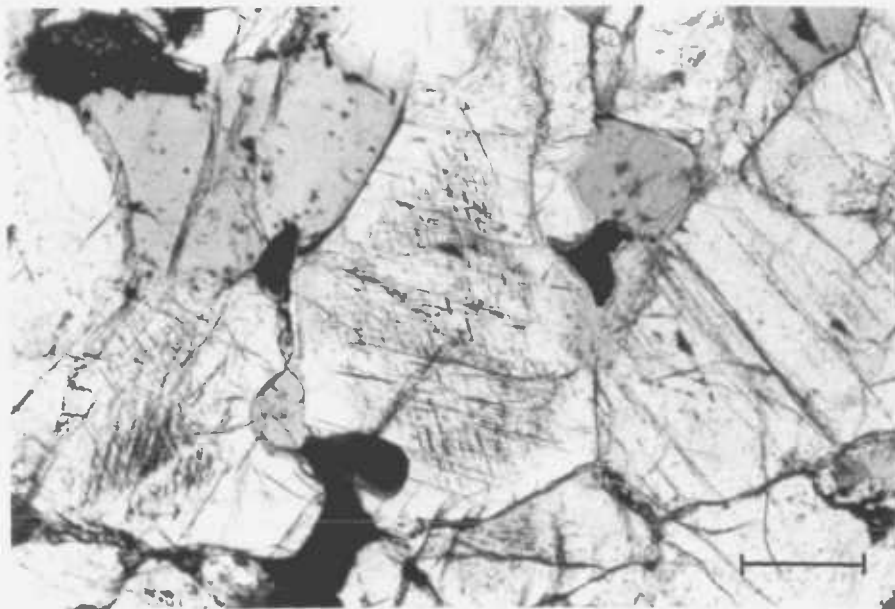


Plate 2.15 A feldspar (centre) from a hornblende-pyroxenite showing exsolution of platelets of iron oxide in two crystallographic directions. Note the concentration of these platelets in the core of the grain. Scale bar is 0.25mm. Sample 74, plane-polarised light.

rounded and sub-idioblastic. Alteration cracks are lined with dusty aggregates of opaque iron oxide.

Magnetite is the common opaque mineral, forming irregularly shaped blebs (in association with green spinel). Sulphides are of minor importance. Nickel-bearing sulphides are rare, although pyrrhotite has exsolved elongate blebs of Pentlandite. The sulphides show the same textural features as the silicates (equant grains, annealed triple-point boundaries) and are assumed to be co-genetic with the original emplacement of the ultramafic rocks.

The plagioclase bearing ultramafites (Samples 85, 104, 108) have petrographic features similar to those described above, except for an absence of olivine and the presence of pseudomorphs after garnet in the form of:-

- i). Amoeboid intergrowths of bright green hercynite ($Mg/Fe \approx 0.9$), orthopyroxene (En_{73}) and plagioclase (An_{62}). These are present in samples 104 and 108 (Plates 2.16 and 2.17).
- ii). Symplectites of green hercynite ($Mg/Fe \approx 0.7$), plagioclase (An_{59}), hypersthene (En_{73}) and garnet ($Am_{47}Py_{34}Gr_{17}$). These are both present in sample 85.

Sample 85 is important in the interpretation of spinel plagioclase $\pm$ orthopyroxene $\pm$ garnet intergrowths, both in the ultramafites and mafic gabbros. It was taken from the contact of ultramafite and mafic gabbro (see Map 1) and shows features of both rock-types. MgO content is high (22%), but the sample exhibits garnet breakdown textures usually associated with the mafic gabbros (see 2.1.2 (ii) below). A thin section cut normal to the ultramafite mafic gabbro contact illustrates features associated with garnet instability.

The fabric is inequigranular and granoblastic. A matrix of medium grained (1.5 - 2.0mm.) idioblastic silicates surrounds patches



Plate 2.16 An orthopyroxene (grey)-spinel (dark grey-black)-plagioclase (white) symplectite (after garnet) from sample 85. The spinel is present as amoeboid lobes within the orthopyroxene. Coarse, relict orthopyroxenes and clinopyroxenes surround the fine grained symplectite area. Scale bar is 0.25mm. Plane-polarised light.

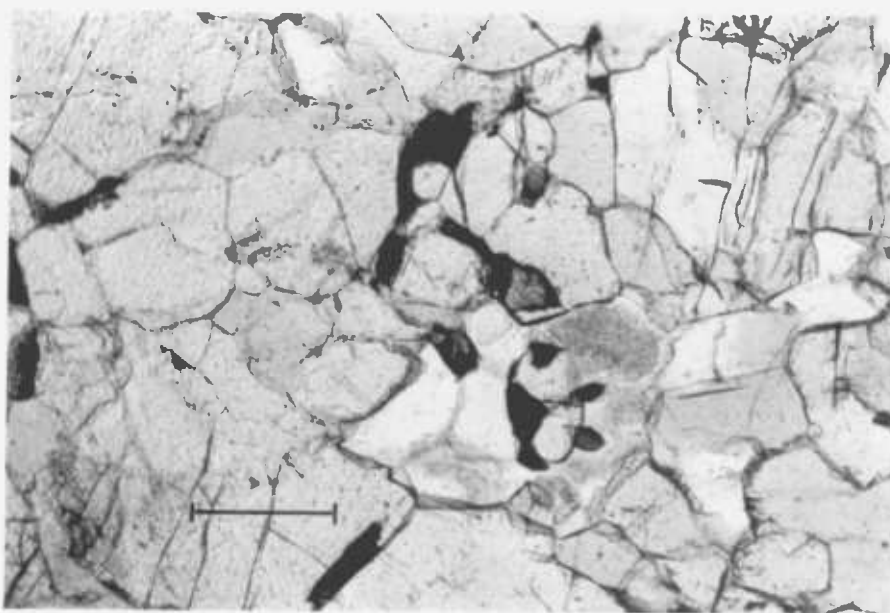


Plate 2.17 Amoeboid intergrowth of orthopyroxene (medium grey) with spinel (dark grey-black) and plagioclase (white-grey) in an ultramafite. Note the clouding of the plagioclase at grain boundaries. Scale bar is 0.25mm. Sample 104, plane-polarised light.

of fine grained (0.1mm.) symplectites. The mineralogical constituents of these symplectites vary across the thin-section. They are garnet-free towards the ultramafite portion and become progressively more garnetiferous away from it. Partitioning of Mg and Fe between phases varies accordingly. As intensive variables (P,T) may be assumed constant over the slide, garnet stability appears to be strongly dependent on variations in the whole-rock FeO/FeO+MgO ratio (to be discussed in more detail in chapter three).

(ii) MAFIC GABBROS

The description given below is also applicable to the sulphide-rich mafic gabbros.

Mineral assemblages present are:-

- 1). CPX + OPX + GT + PLAG + SP $\pm$ AMPH + MAG $\pm$ ILM $\pm$ CHALC $\pm$ PYR $\pm$ PYRRH
(Samples 77,78,84,91)

This assemblage is present within magnesian mafic gabbros (developed adjacent to the contacts with ultramafic rock types), with low FeO/FeO + MgO ratios.

- 2). CPX + OPX + GT + PLAG $\pm$ AMPH + MAG + ILM $\pm$ CHALC $\pm$ PYR $\pm$ PYRRH $\pm$ COV
(Samples 79-83, 92-95, 101-102, 105-107, 109)

This assemblage is present within mafic gabbros with generally high FeO/FeO + MgO ratios.

Modifications of the above by retrogression may result in the following:-

CPX/OPX $\rightarrow$ TREM/ACT/EPID; PLAG $\rightarrow$ SCAP/ZO/SERIC; GT/OP $\rightarrow$ BI;

GT/PLAG $\rightarrow$ MUSC

The magnesian mafic gabbros are medium to fine grained with an equigranular, granoblastic fabric. They are characterised by the presence of two textural areas:-

- 1). A coarse grained (1.5 - 2.0mm.) equigranular mosaic of idioblastic

clinopyroxene forms essentially a monomineralic matrix (Plate 2.18) This is believed to represent a relict granulite facies equilibrium texture (the M_1 metamorphism, see chapter three).

2). Within this matrix are rounded zones of fine grained (0.1mm.) orthopyroxene, plagioclase, green spinel and opaques which are usually intergrown in a complex fashion. Relict garnet granules may form cores to these areas. They are thought to mark the former presence of garnet porphyroblasts (Bowdidge, 1969; O'Hara, 1977a; O'Hara and Yarwood, 1978).

Clinopyroxenes are green salites ($En_{34}Wo_{46}$ - $En_{40}Wo_{46}$). Grains show lamellar exsolution (20mm thick) of orthopyroxene. These lamellae may be bent and strained. Fine granules of magnetite may trace grain boundaries.

Garnet is generally granular and fine grained, but corroded porphyroblasts may occasionally be present. Compositions range from $Am_{51}Py_{27}Gr_{19}$ to $Am_{48}Py_{34}Gr_{15}$. They are bright pink in colour and generally free of inclusions. Lobes may poikilitically include spinel, hypersthene or plagioclase. Colour zoning is sometimes present, the rims being darker than the cores, but this zoning has not been identified chemically. (Plate 2.19)

Orthopyroxenes are strongly pleochroic from colourless to red-pink and are hypersthene (En_{63-70}). They are fine grained and xenoblastic, occurring only within areas of garnet corrosion. There is a tendency for them to 'armour' relict salites at the external margin of garnet breakdown areas. No exsolution of a Ca-rich pyroxene is visible.

Spinel is emerald-green hercynites ($Mg/Fe=0.7-0.9$) and form rounded xenoblastic grains either adjacent to, or surrounded by, garnet. Composite grains with magnetite are common.

Plagioclase is fine grained, of composition An_{51-74}

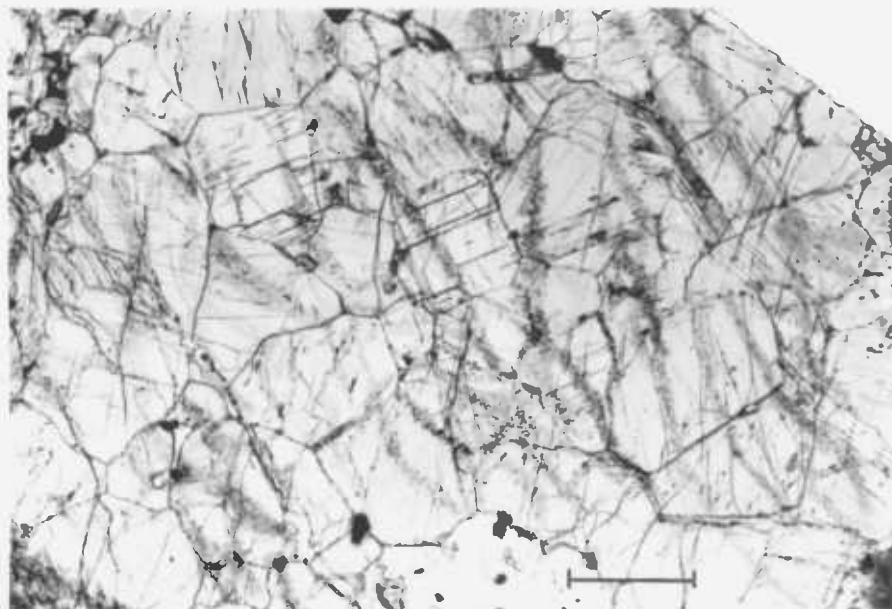


Plate 2.18 Granoblastic aggregate of relict clinopyroxenes in a mafic gabbro. Note the annealed triple-point grain boundaries. Scale bar is 1mm. Sample 78, plane-polarised light.

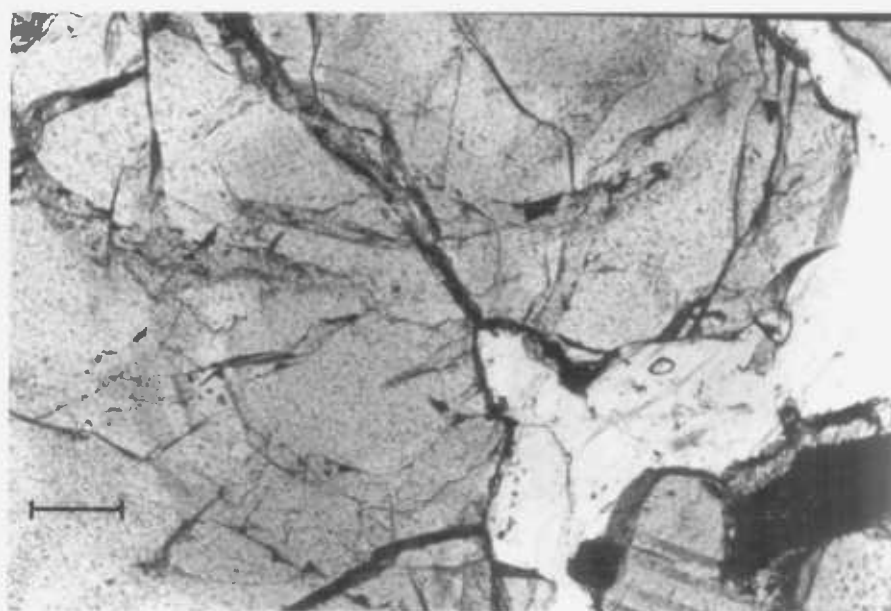


Plate 2.19 Colour zoning in a garnet porphyroblast from a mafic gabbro. The porphyroblast has embayed and scalloped margins against a plagioclase 'collar'. Scale bar is 0.25mm. Sample 81, plane-polarised light.

(labradorite-bytownite). Twin lamellae are strained and may be bent.

Amphiboles are sparse in occurrence and are paragenetic. They are pleochroic from light to dark brown, the colour being emphasised by dusty inclusions of iron oxide.

Magnetite and ilmenite are the dominant opaque phases, commonly associated with hercynite.

The mafic gabbros with higher $\text{FeO}/(\text{FeO} + \text{MgO})$ ratios (assemblage 2). above) show features similar to the above. Colour indices are lower as plagioclase becomes more evident. Some recrystallisation of relict clinopyroxenes has taken place to produce fine grained (0.4 mm.) clinopyroxene-plagioclase areas (Plate 2.20). Garnet porphyroblasts (2 - 3mm. in diameter) are better preserved, although still declining in abundance. Garnets have pronounced 'collars' of plagioclase, with an outer rim of hypersthene and magnetite against relict clinopyroxenes. (Plate 2.21). Green spinel is absent and there is also a lack of symplectite about garnets.

Clinopyroxenes are apple-green salites of composition $\text{En}_{34}\text{Wo}_{46}$ - $\text{En}_{35}\text{Wo}_{47}$. Relict grains contain exsolution lamellae of orthopyroxene, whereas recrystallised grains do not.

Garnet occurs in two distinct forms. Garnet porphyroblasts are of the composition $\text{Am}_{53}\text{Py}_{29}\text{Gr}_{17}$ - $\text{Am}_{54}\text{Py}_{26}\text{Gr}_{18}$. Garnet coronas form fringes (60 mm. wide) on opaques (Plate 2.22) and are of composition $\text{Am}_{58}\text{Py}_{22}\text{Gr}_{19}$. An intervening rim of biotite and/or actinolite is common. The porphyroblastic type is embayed against plagioclase.

Orthopyroxenes are strongly pleochroic from pink to red-brown and are hypersthene (En₅₇₋₅₈). They form an outer ring on plagioclase collars and armour relict salites.

Plagioclase compositions are in the range An₄₉₋₅₅ (andesine-

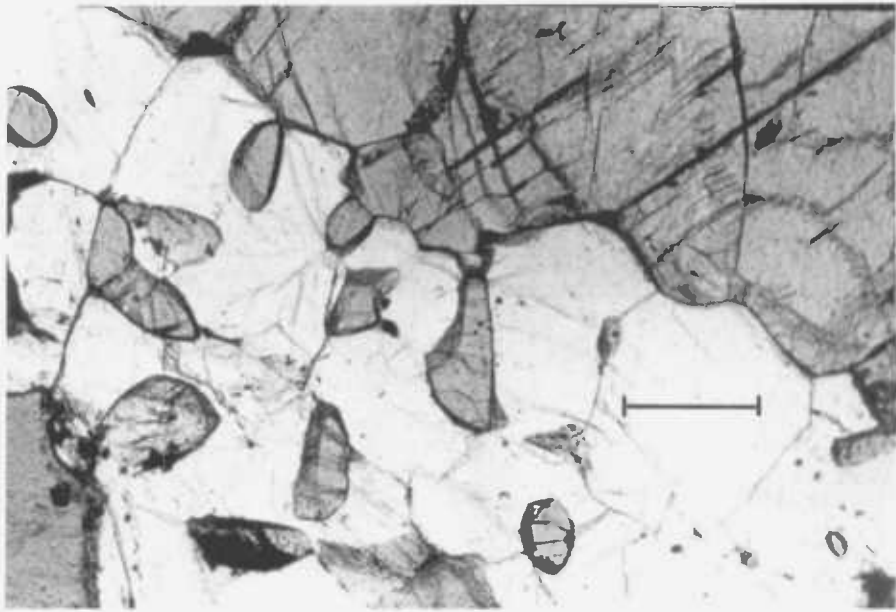


Plate 2.20 Recrystallisation of coarse relict clinopyroxenes (upper field) in a mafic gabbro. The clinopyroxenes are embayed against plagioclase. The recrystallised clinopyroxenes are fine grained in a plagioclase mosaic. Scale bar is 0.25mm. Sample 81, plane polarised light.

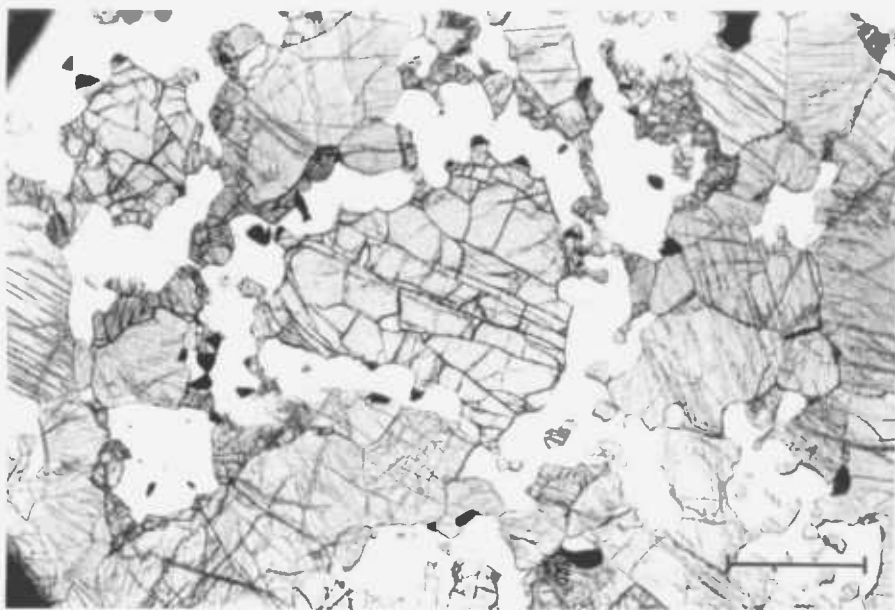


Plate 2.21 Garnet porphyroblast (centre) surrounded by a plagioclase 'collar' (white) and fine grained hypersthene (dark grey) and titanomagnetite (black). Coarse relict clinopyroxene is external to the corona area. Scale bar is 1mm. Sample 81, plane-polarised light.

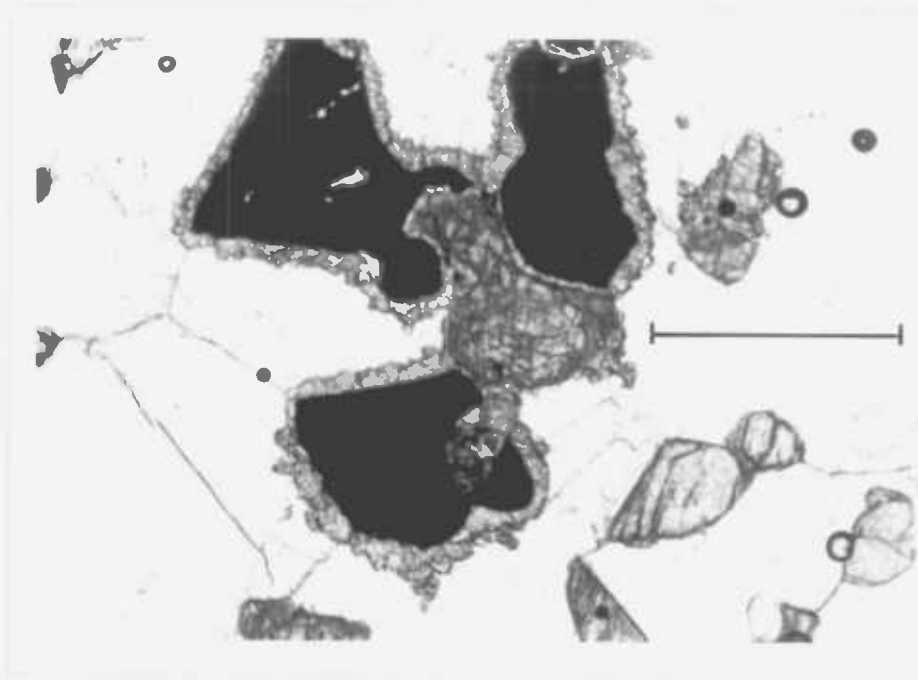


Plate 2.22 Fine grained garnet corona on titanomagnetite (black) and biotite (grey). Scale bar is 0.5mm. Sample 9, plane-polarised light.

labradorite). It is present as well-crystallised rims around garnet porphyroblasts.

Magnetite and ilmenite form associated grains with a necked junction. The occasional bar exsolution lamella may be present.

The sulphide minerals also form associated grains. Their annealed triple-point junctions with the silicate phases indicate that the sulphides are primary, rather than secondary. No exsolution features have been observed. The proportion of sulphides increases markedly in the sulphide-rich mafic gabbros. Quartz may be present as a minor phase in the sulphide-rich rocks.

All the above petrographic descriptions refer to unretrogressed granulites. However, the effects of retrogression are present in all samples. This is generally slight, but may be advanced in some samples.

2.1.4 SUMMARY

The layered mafic-ultramafic sheets represent small sill-like bodies within the host gneisses of the Lewisian Complex. The intrusions have a compositional banding which is founded on an original igneous layering. Gravity stratified units have been identified from field observation and a geochemical traverse. These probably indicate an original multiple injection of magma for the intrusions. Way-up directions established from the rhythmic units suggest that the initial base of the intrusions is now central to the outcrop as a result of isoclinal folding. This folding has transposed the original igneous layering to produce the banding now visible in the intrusions (S_1). This is demonstrably older than the planar fabric present within the acid gneisses and granulites (S_2). From the similarity of the rock types present, and from the distribution of rock types within each intrusion in

the Scourie district, it would not be unreasonable to assume that the bodies may have originally formed one continuous intrusion.

The sheets lack any intrusive, cross-cutting contacts with the host gneisses and granulites. However, there is also a lack of evidence to suggest that the parents of the surrounding gneisses intruded and disrupted the layered sheets as is envisaged to have happened to the layered anorthosites of West Greenland. (Kalsbeek and Myers, 1973; Myers, 1976). Only a minor phase of granitic intrusion (the granite sheets, described below) is clearly younger than the layered mafic-ultramafic sheets themselves.

From the petrographic descriptions and the mineral chemistry (chapter five) of the ultramafites and mafic gabbros, it is apparent that the minerals and textures present are metamorphic, rather than igneous. For example, the high alumina contents of the clinopyroxenes (up to 7%) and orthopyroxenes (up to 5%), the low calcium content of the orthopyroxenes, the aluminous nature of the spinels, Fe/Fe+Mg ratios of co-existing minerals and the presence of garnet. The ubiquitous presence of annealed, granoblastic textures in the ultramafites and mafic gabbros, the early coarse clinopyroxene-garnet mineralogy of the mafic gabbros and complete lack of any ophitic, sub-ophitic or cumulate textures are also consistent with a metamorphic origin.

At this point, it may be relevant to summarise possible modes of origin for the layered sheets compatible with the above observations:-

- 1). High-level intrusion and differentiation of the sheets in a supracrustal sequence of sediments and lavas, possibly in a proto-oceanic environment. Subduction (?) of the intrusions and sediments, followed by tectonic thrusting into an island arc/cordilleran environment. Invasion and disruption by tonalitic/

granodioritic material derived from partial melting of subducted oceanic lithosphere (Ringwood, 1974), accompanied by deformation and high grade metamorphism.

This general model (not specifically for Scourie) has been proposed by Tarney (1976). Problems are that lead isotope studies (Moorbath et al., 1969) indicate a short time period (100 - 200 million years) between derivation of the Scourie gneiss parents from an upper mantle source and subsequent granulite facies metamorphism. This obviously places limits on the time available for the subduction and metamorphism of the layered sheet material after differentiation from the upper mantle.

2). Crystallisation of the intrusions from basaltic magmas parental to the calc-alkaline association of diorites, granodiorites and tonalites in a Cordilleran-type environment. Penecontemporaneous deformation and metamorphism at depth.

This again, is a general model suggested by Windley and Smith (1976), Windley (1976), and Tarney and Windley (1978). This hypothesis is dependent on the sheets being calc-alkaline in character. Fractionation trends might also be expected to be controlled by a high pressure mineralogy (aluminous pyroxenes, spinel, hornblende).

3). Intrusion of the sheets into supracrustal sediments and volcanics followed by tectonic interleaving with quartzofeldspathic gneisses. Invasion and disruption of the older gneisses, supracrustals and layered sheets by granodioritic and tonalitic magmas, followed by metamorphism and deformation. (Davies, 1975).

Isotopic evidence (Chapman and Moorbath, 1977) suggests that the 'older' gneisses of Davies (1975) are not significantly older than the 2,900 million year age assigned to the rest of the complex.

The above models will be reviewed later in light of the geochemical information to be presented in chapter four.

2.2 THE GRANITE SHEETS

These rocks have been briefly mentioned by other workers in the Lewisian. Some idea of the terminology used for these rocks and hypotheses of origin may be gained from the following:-

(i) 'Bands of unusually acid character' (Peach et al., 1907, p.132)

The early survey workers noted the almost exclusive occurrence of the granites within the mafic-ultramafic sheets (p.132-134).

(ii) 'Leucocratic gneisses' (O'Hara, 1961a; Bowdidge, 1969).

Thought to represent partial melts derived from the regional acid gneisses during ultra-metamorphism (Bowdidge, 1969).

(iii) 'Gneissose pegmatites'. (Sheraton et al., 1973). Considered to represent the 'sweat-out' pegmatites of the Badcallian granulite facies metamorphism, rather than the ' π_1 ' pegmatites of Sutton and Watson (1951).

(iv) 'The Clach Boga Gneisses' (Davies, 1975). A pre-Badcallian set of intrusive acid gneisses.

(v) 'Leucocratic gneisses' (O'Hara, 1977a, 1977b; O'Hara and Yarwood, 1978). Regarded as dry partial melts or granulite-facies pegmatites.

All authors note the common location of these rocks within, and at the margins of, the layered mafic-ultramafic sheets. Davies (1975) has suggested that these granite sheets may comprise a significant proportion of the Lewisian Complex as a whole.

2.2.1 FIELD RELATIONS AND PETROGRAPHY

Within the area mapped, granite sheets occur solely as layers, or masses within the layered mafic-ultramafic sheets and as integral

components of the net-veined mafic-ultramafic sheets (described below). Granite sheet material has not been observed in the surrounding acid gneisses (cf. Davies, 1975).

The granite sheets form conspicuous, salmon-pink weathering bands within the mafic and ultramafic rocks, quite often forming prominent crags (Plate 2.24). Layers are sub-concordant to the banding in the mafic and ultramafic rocks (S_1), but are locally discordant (e.g. Geodh' Eanruig, fig. 2.3). Despite being discordant, they are themselves foliated as a result of the D_2 deformation. Granite thicknesses vary from 20cms to 10 metres (individual layers may show such a range in thickness when traced along strike). Layers occasionally branch. They occur in either ultramafite or gabbro lithologies, but not apparently at junctions between these.

The granites are medium to coarse grained in hand-specimen and have a strong tectonic fabric of elongated quartz lenticles, which may be 1 cm. in length. This foliation is parallel to that of the surrounding acid granulites.

The mineral assemblage is:-

$QZ + ANTIPERTH/MESOPERTH + MAG \pm OPX \pm CPX \pm AP$

The fabric is dominated by elongate quartz grains and feldspar is equigranular and interstitial.

Quartz is heavily strained, exhibiting b δ hm lamellae normal to the length of the grains. Recrystallisation of lenticles into a fine grained mosaic is common.

Grains are blue in hand-specimen, a generally accepted feature of granulite facies gneisses.

Feldspar may be a mesoperthite or an antiperthite. Grains are sub-idioblastic. The antiperthite has a potash-rich feldspar exsolved as irregular blebs (0.15mm.) in two crystallographically



Plate 2.23 Sharp contact between granite sheet and pyroxenite. Camas nam Buth intrusion, east of Camas an Lochain.



Plate 2.24 Steep cliff of granite sheet, Camas nam Buth intrusion.

oriented directions. The mesoperthite has a ribbon form of exsolution, ribbons being approximately 10 μ in width. The two feldspar varieties may be present in the same sheet, giving rise to chemical inhomogeneities. The feldspar is commonly sericitised, emphasising the low colour index of the rocks.

Clots of mafic minerals are rare and are generally elongate in the direction of foliation. Both orthopyroxene and clinopyroxene may occur only as relics replaced by biotite and actinolite.

Contacts with the mafic and ultramafic rocks are generally sharp. At Geodh' Eanruig, large clinopyroxene megacrysts are contained in a fine grained, quartzo-feldspathic matrix. The gabbros surrounding this mass of granite are amphibolitised for 1-2 cms. at the contacts, suggesting that the granitic melt was not entirely dry (cf. O'Hara, 1977a; O'Hara and Yarwood, 1978).

The field evidence suggests that the granite sheets are intrusive into the layered mafic-ultramafic rocks, cross-cutting the S_1 banding, but are themselves affected by the D_2 deformation. They are also clearly pre- Π_1 pegmatite (Sutton and Watson, 1951), as the pegmatite at Geodh' Eanruig cuts across the foliation within the granite sheet there. The unusual quartz lenticles may have been produced by the action of tectonic stresses on a still molten granitic melt.

2.3 THE NET-VEINED MAFIC-ULTRAMAFIC SHEETS

These units have not been previously described as such and have in the past been grouped together with the layered mafic-ultramafic sheets. Gross similarities exist with the layered bodies, such as the presence of garnetiferous mafic rocks and amphibole pyroxenites as the main lithological types, but the descriptions

below will help to illustrate the major differences between the two types of intrusions.

2.3.1 FIELD RELATIONS AND PETROGRAPHY

Two outcrops of net-veined bodies occur on Scouriemore (fig. 2.1): on the south side of Camas nam Buth ; and west of the graveyard. Both are well exposed in coastal cliffs, but crop out poorly inland. They are similar in size and form to the layered mafic-ultramafic sheets.

The units are characterised by angular blocks of mafic or ultramafic material (blocks range from tens of centimetres to tens of metres across) set in a matrix of coarse, quartzofeldspathic granulite. This intrusive granitic fraction is identical in macroscopic and chemical characteristics to the granites found within the layered mafic-ultramafic sheets. Granite veins are 10-20 cms. thick on average, although at the margin of the units, the thickness may be 1-2 metres. An idea of the nature of this net-veining may be gained from plates 2.26 - 2.31. The engulfed blocks of mafic and ultramafic material tend to be smaller at the margins of the unit, the size of the blocks increasing inwards, with a reciprocal decrease in the amount of veining granite.

The mineral assemblage in the ultramafites is:-

CPX + OPX + AMPH + PLAG + BI + OP

They are coarse grained, granoblastic aggregates of clinopyroxene, with minor amounts of amphibole, orthopyroxene and plagioclase. No banding or layering is evident within any individual block. The Cleit Mhor unit has ultramafite blocks concentrated along its lower (structural) contact for a width of 10-20 metres, rather than anywhere else in the mass. This is the only evidence of any large-scale layering within the net-veined



Plate 2.25 Granite sheet material veining mafic gabbro,
Geódh' Eanruig.



Plate 2.26 Lower portion of a net-veined unit, Cleit Mhor.



Plate 2.27 Close-up of Plate 2.26.



Plate 2.28 Close-up of Plate 2.27.



Plate 2.29 Thick granite sheet at the base of a net-veined unit, Cleit Mhor.



Plate 2.30 Net-veining of gabbros by granite sheet material, west of the graveyard, Scouriemore.



Plate 2.31 Net-veining of gabbros, west of the graveyard,
Scouriemore.

mafic-ultramafic sheets.

The mineral assemblage in the mafic rocks:-

CPX + OPX + PLAG + AMPH $\pm$ GT + OP

These rocks show textures similar to those of the mafic rocks from the layered mafic-ultramafic sheets. Plagioclase is strongly sericitised. Orientation of the poorly defined S_1 foliation varies from block to block, suggesting a 'jumbling' of the blocks within the acid matrix.

No sulphide bearing gabbroic rocks have been observed.

As mentioned previously, the invading granite material is macroscopically and chemically similar to the granite sheets present within the layered mafic-ultramafic sheets. The granite veins have a foliation (S_2) defined by quartz lenticles, parallel to the foliation within the external acid gneisses. This is only poorly developed where the veins are at an angle to the external foliation. The quartz fabric is discordant to the foliation present within the engulfed mafic and ultramafic blocks.

Similar disrupted mafic-ultramafic intrusions have been described from West Greenland by Myers (1976).

Veins and lenses of presumed metasedimentary material (garnet-biotite-feldspar rocks) are present within these units (fig. 2.1). They have an irregular occurrence, but seem to be intercalated with the mafic rocks. The lenses are thin (1-2 metres in thickness) and pinch out along the strike. They weather to a greyish surface, spotted with rusty coloured garnet porphyroblasts. This metasedimentary material appears to be confined to the net-veined mafic-ultramafic sheets within the region mapped. However, a larger mass of metasediments crops out north of Scourie Bay, apparently unrelated to any net-veined intrusion.

CHAPTER 3

STRUCTURE AND METAMORPHISM

Structural mapping of the layered sheets and surrounding gneisses has been undertaken to aid geochemical sampling and also with an aim of integrating mineralogical data with phases of deformation.

3.1 THE REGIONAL STRUCTURE

The gneisses in the vicinity of Scourie have gently dipping (30° - 40°) foliations trending in a north-east south-west direction. Scourie Dykes are un-deformed and near-vertical, truncating the foliation sharply. Major tectonothermal activity ceased about 2700 million years ago. This Scourian Complex abuts northwards against the Laxfordian Complex (fig. 3.1). The boundary trends north west - south east from Ben Stack in the south to Rubha Ruadh in the north. The Laxfordian Complex consists of granitic (s.l.) gneisses of amphibolite facies grade showing evidence of intermittent tectonothermal activity from 2900 to 1700 million years ago. A region approximately four kilometres wide, south of the junction marks a transition zone (Sutton and Watson, 1951; Beach et al., 1974; Coward, 1974; Davies, 1976). Here, Scourian gneisses are progressively modified north-eastwards so that amphibolite to greenschist facies mineral assemblages result and the structural trend is re-oriented north west - south east. This is accomplished by pre-Scourie Dyke (Inverian or Late Scourian) upright folding and by Laxfordian ductile shearing, both along north west - south east axes. Scourie Dykes are deformed and disrupted in this zone. Davies (1976) has interpreted this zone as a large scale overturned synform, initially developed in Scourian times.

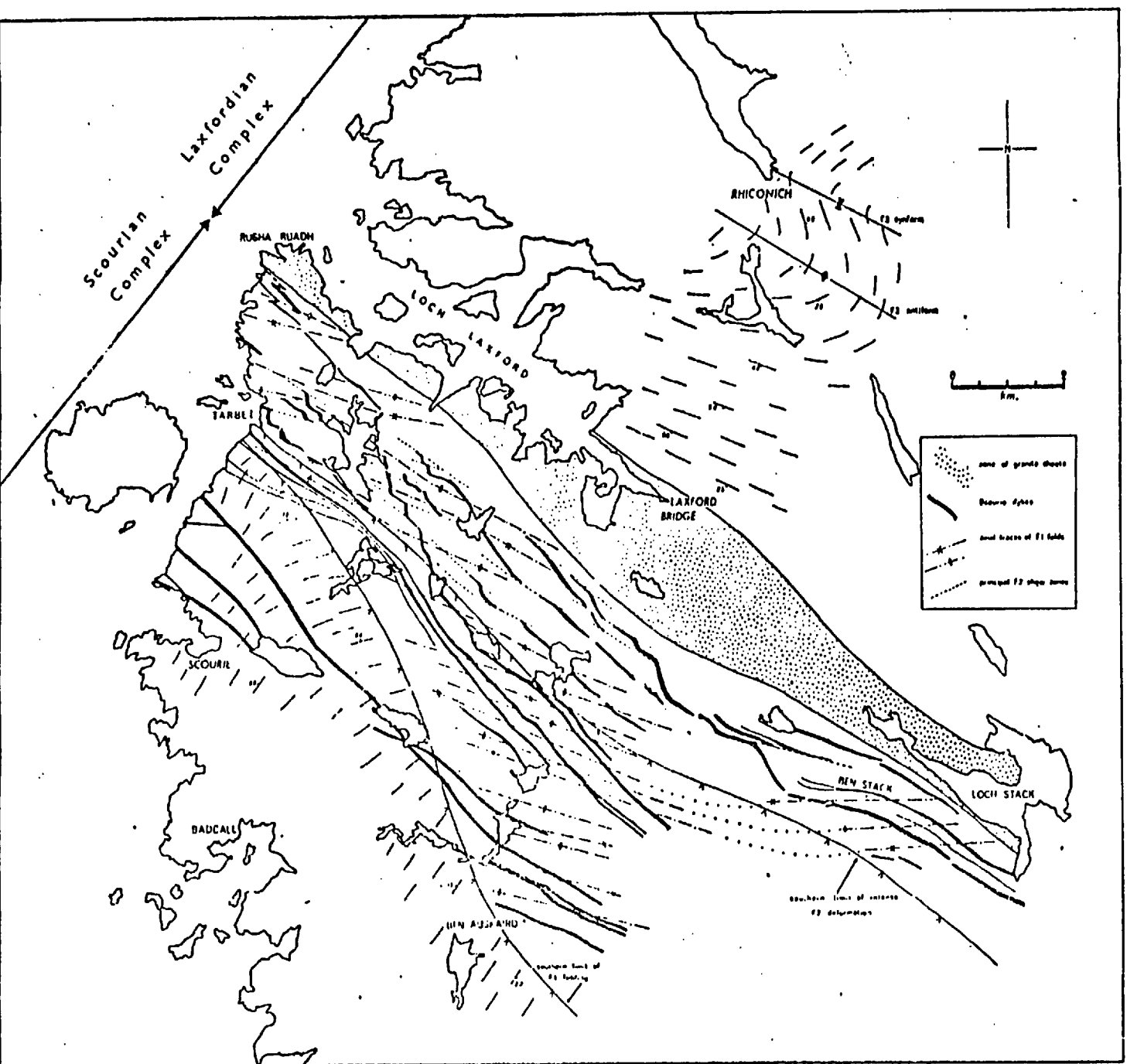


Fig. 3.1 The geological structure of the area north-east of Scourie, from Beach et al. (1974). F_1 and F_2 refer to F_3 and F_4 respectively, of this work.

3.1.1 THE D₁ DEFORMATION

The evidence for an early fold phase prior to the major foliation forming event recognisable in the acid gneisses has been mentioned in chapter two. A summary of this evidence is as follows:-

(i) The geochemical trends and the juxtaposition of sulphide and ultramafite layers observed from the traverse of the Camas nam Buth sheet suggest that the lower portion of this sheet has been inverted relative to the top portion. The central outcrop of the thickest ultramafite layer in all the intrusions studied indicates that the original base now occurs in the middle of the intrusions rather than at one margin. It is necessary to invoke a period of anticlinal isoclinal folding to account for the above observations.

(ii) A diffuse layering may be observed in the acid gneisses and granulites which is folded by the F_2 isoclines (Plates 3.1 and 3.2) This layering is obviously pre- S_2 (which develops as an axial planar structure to the F_2 isoclines) and may be termed S_1 .

(iii) The weak foliation in the mafic and ultramafic blocks of the net-veined mafic-ultramafic sheets shows a considerable discordance with that of the surrounding acid gneisses and granulites. This again, is probably an S_1 foliation and pre-dates the S_2 fabric of the acid gneisses.

(iv) O'Hara (1965) has described small isoclinal folds from ultramafites in layered sheets north-east of Scourie Bay. These plunge to the north-west, parallel to minor folds observed in the surrounding acid gneisses (F_2 of this work). The folded banding in the ultramafites is therefore S_1 .

It is clear therefore, that an early banding or layering may be seen to pre-date the main S_2 foliation. It is believed that

this develops parallel to the isoclinal folding observed from the geochemical trends and distribution of rock types in the layered sheets.

3.1.2 THE D₂ DEFORMATION

This phase of deformation was responsible for the major fabric-forming event of the acid granulites and gneisses. The S₂ foliation is consistent in orientation (fig. 3.2) trending north east - south west and dipping gently (25° - 35°) to the north-west.

The layered mafic-ultramafic sheets may be used as marker horizons to define large-scale isoclinal folds within the Lewisian Complex (Davies, 1974b, 1976; Josey and Shaw, 1974). Two such folds may be identified in the Scourie area (fig. 3.2). These are large recumbent isoclines with shallow dipping axial planes, plunging gently to the north or north-west. The large isoclines refold early F₁ isoclines, as defined by the lithological banding within the layered sheets.

Evidence for the large scale F₂ isoclinal folding within the acid gneisses is restricted to the occurrence of small intrafolial folds and isoclines (Plates 3.1 and 3.2).

3.1.3 THE D₃ DEFORMATION

This is probably Inverian (Late Scourian) in age. Effects of deformation are minor around Scourie, although more pronounced north-eastwards (Park, 1973; Beach et al., 1974; Davies, 1976). They are shown by structures of two different types:

(1) The D₃¹ Deformation

This takes the form of open upright warps which deform the S₂ foliation into gently curving arcs (fig. 3.2). The folding



Plate 3.1 Intrafolial fold in acid granulites, Rubha Shios.



Plate 3.2 Isoclinal fold defined by a mafic band in acid granulites, east of Geodh' nan Cliabh.

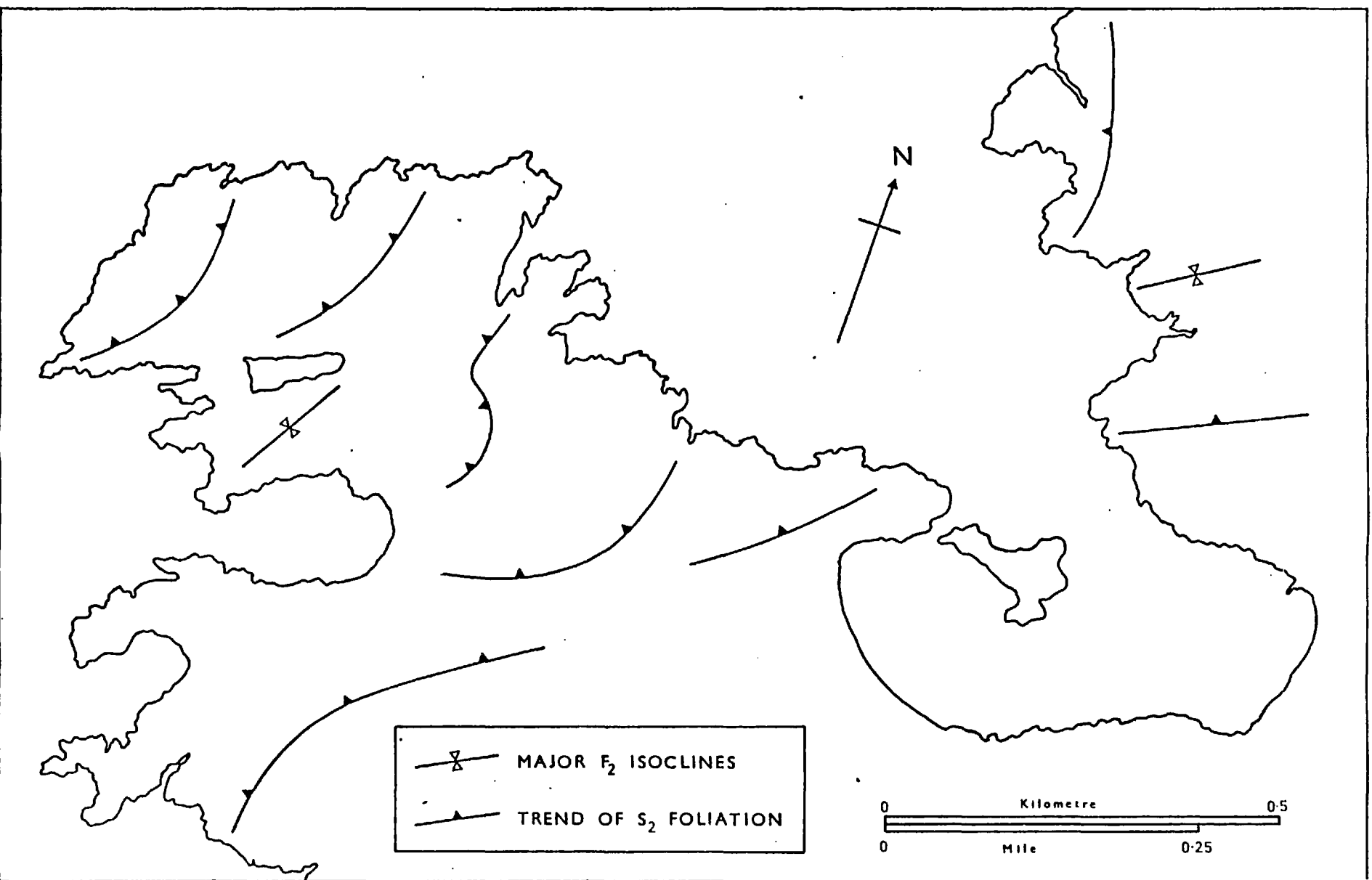


Fig. 3.2 The effects of the D_2 deformation in the Secourie district. Note the gentle warping of the S_2 foliation by F_3 folds.

trends WNW-ESE and is cut sharply by the Scourie Dykes. It may be seen to refold F_2 isoclines eg. the eastern extremities of the Camas nam Buth sheet (Map 1; fig. 2.1; see also Davies, 1976).

(ii) The D_3^2 Deformation

This is concentrated into zones of small-scale tight folds or monoclines, trending NE-SW. These structures have a gentle ten or twenty degree plunge to the south-west. A steep foliation may be developed parallel to the axial planes as is seen in the nose of the F_2 synform at Camas nam Buth (Plate 2.9). Serpentinisation of olivines and amphibolitisation of pyroxenes in tight folds and monoclines suggest that an influx of volatiles was associated with folding. The folding also seems to have been roughly synchronous with the intrusion of the Π_1 pegmatites. At least one such pegmatite lies along the axial plane of an F_3^2 fold at Geodh' Eanruig (fig. 2.5).

3.1.4 THE D_4 DEFORMATION

This is localised and of the ductile shear-zone type, tentatively correlated with the production of Laxfordian shear-zones described from north of Scourie Bay by Beach (1973, 1974a, 1974b). Shear zones trend NNW - SSE, but a minor set also exists trending NE -SW (fig. 3.3). Shears are vertical and have a dextral sense of lateral movement (Plate 3.3). Minor parasitic folding may accompany shearing (Plate 3.4). Beach et al., (1974) noted a preponderance of sinistral Laxfordian shear zones north of Scourie. These shear-zones are discordant to all previous structures. A fine grained fabric associated with amphibolitisation of pyroxenes is developed.

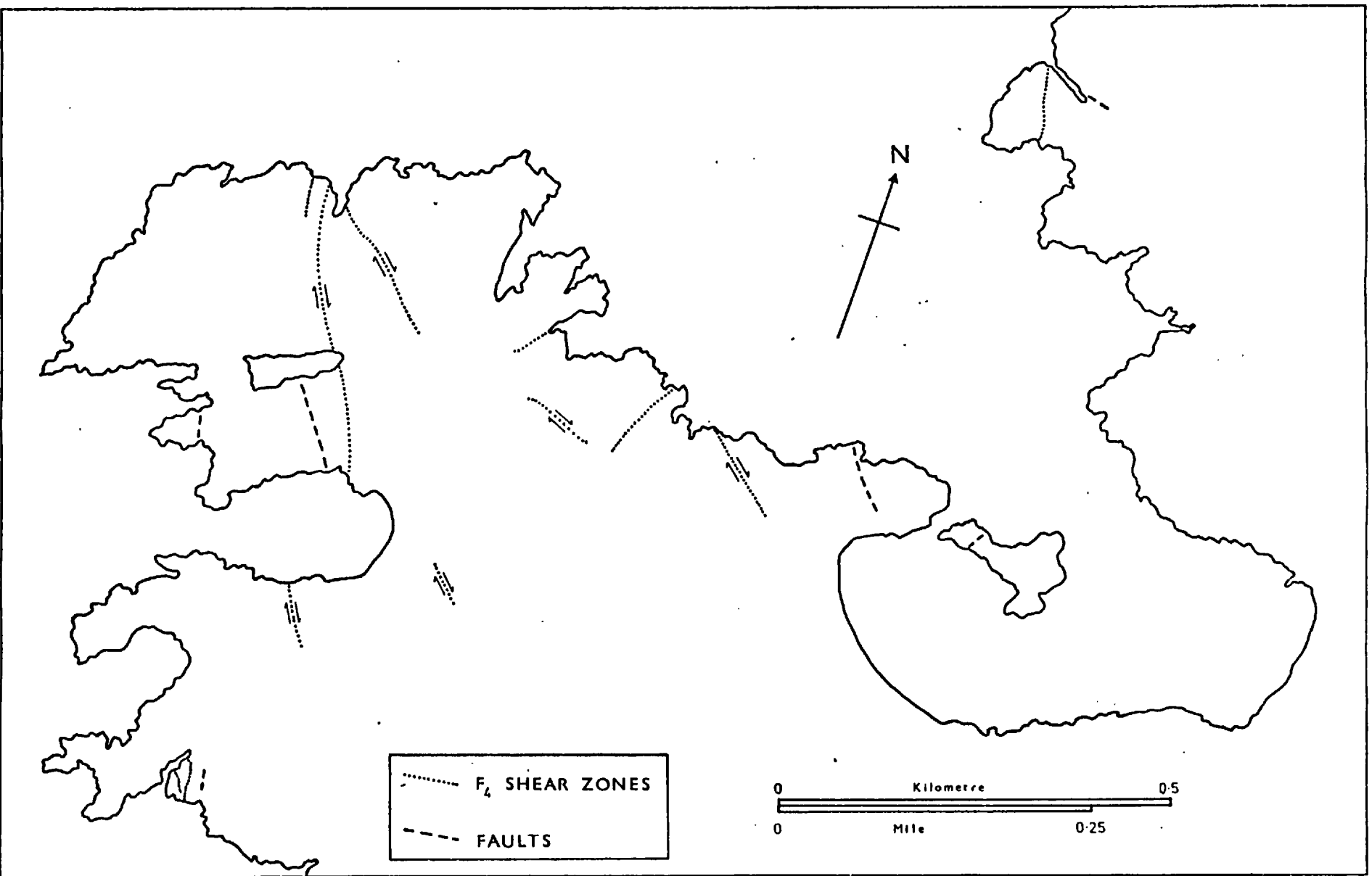


Fig. 3.3 Shear-zones and faults in the Scourie district.



Plate 3.3

D₄ shear-zone in acid
granulites, west of Geodh'
nan Cliabh.

Plate 3.4

Parasitic D₄ synform
adjacent to a shear-zone
east of Geodh' nan Cliabh.



3.1.5 THE D₅ DEFORMATION

This is represented by a few minor faults of uncertain age (fig. 3.3). Orientations are NW-SE to N-S with minor displacements.

3.2 METAMORPHISM

This section discusses information derived from the layered mafic-ultramafic rocks, although reference to other lithological types will be made where relevant. It will become apparent that the mafic rocks in particular are the most useful in tracing the metamorphic history of the Scourie area. Their mineral assemblages are sensitive to changes in intensive parameters (P , P_{H_2O} , T , f_{O_2}) within the range of metamorphic conditions experienced in the area.

Four periods of metamorphic mineral growth have been identified (M_{1-4}), three being of granulite facies grade, the other of upper amphibolite facies grade. Each metamorphic event will be discussed in detail, below.

First, it is necessary to consider the nature of the rocks prior to metamorphism and the type of mineral reactions involved in converting rocks of a 'basaltic' chemistry to the high grade metamorphic rocks around Scourie.

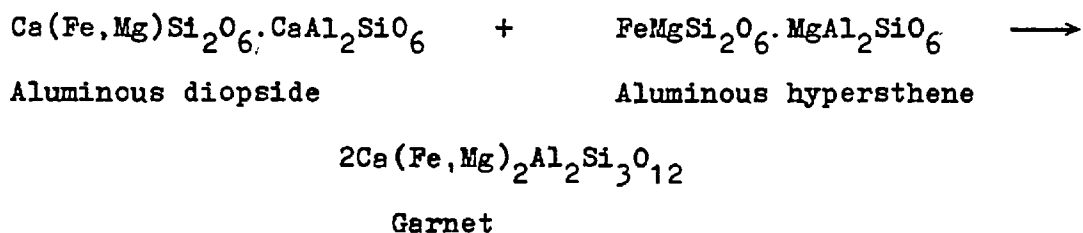
3.2.1. THE GABBRO-ECLOGITE TRANSFORMATION AND ITS RELATION TO THE SCOURIE AREA

The conversion of rock of basaltic composition from a gabbroic to eclogite mineral assemblage by high grade metamorphism is of considerable petrological importance. Models of basalt and andesite genesis, geophysical models of the upper mantle and Moho, and plate tectonic subduction processes depend on the mineralogical constitution of basic rocks in the lower crust and mantle (Ringwood and Green, 1966; Joyner, 1967; Ringwood, 1974). Consequently,

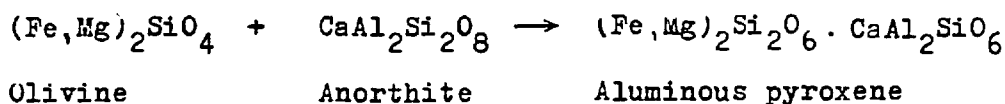
great attention has been paid to the nature and physical conditions of the gabbro-eclogite transition both experimentally (Kushiro and Yoder, 1966; Green and Ringwood, 1967a; Ito and Kennedy, 1970; Green and Ringwood, 1972; Kennedy and Ito, 1972) and on natural samples (Miller, 1970; Griffin, 1971 and 1972; Mysen and Heier, 1972; Griffin and Heier, 1973; Whitney and McLelland, 1973; Loomis, 1977).

These studies indicate that the gabbro-eclogite transition may be divided into a high pressure eclogite-garnet granulite reaction and a lower pressure garnet granulite-gabbro reaction. Although the mafic rocks of the Scourie area exhibit textural evidence of the retrograde eclogite gabbro type reactions (discussed later), direct petrographical evidence of the prograde gabbro-eclogite reactions is lacking. However, high-pressure igneous fractionation trends (chapter four) for the layered sheets suggest that prograde reactions have indeed taken place to effect the transition from gabbro with aluminous pyroxenes to garnet granulite.

The mafic gabbros of the layered complexes are low-alkali, olivine-tholeiites in geochemistry (chapter four). Garnet-forming reactions are probably:-



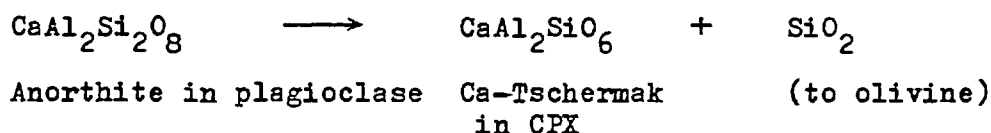
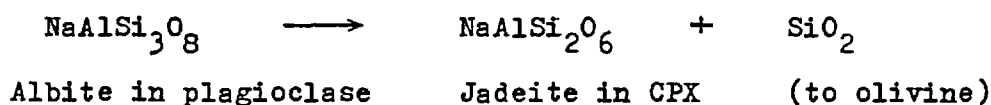
Liquidus phases during the differentiation of the layered sheets are thought to have been aluminous pyroxenes rather than olivine or plagioclase (chapter four). This may be expressed in terms of the normative components:-



One would expect the modal plagioclase to be more albitic than the normative variety as a consequence. This is illustrated in fig.3.4.

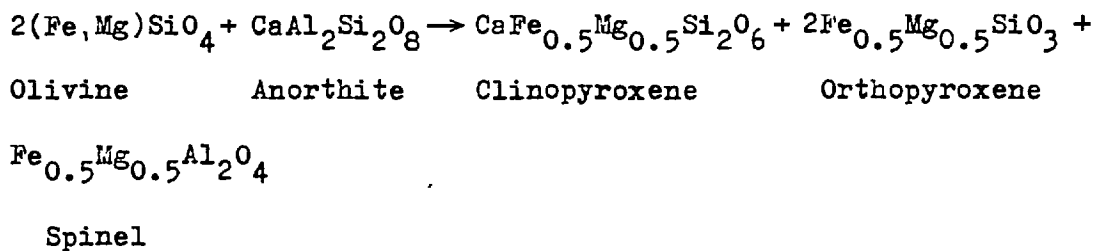
This decrease in anorthite is paralleled by a decrease in total plagioclase content between the norm and the mode (fig. 3.5). Thus it may be assumed that the original igneous rocks had a lower colour index than that indicated by the metamorphic assemblages.

The high alumina content (5-7%) of the analysed clinopyroxenes may be attributed to two reactions:-



Analysed clinopyroxenes show significantly higher Ca-Tschermak than jadeite, suggesting that the latter reaction is more important in producing aluminous pyroxenes in the Scourie samples. This is illustrated in fig. 3.6. Accordingly, decreasing anorthite content of the modal plagioclase should decrease the Ca-Tschermak's content of the co-existing clinopyroxene (Wood, 1975). A reasonable positive correlation is shown between these two components in fig. 3.7.

It may be considered that the ultramafites have also undergone plagioclase and olivine consuming reactions to produce aluminous pyroxenes and spinel. Normative olivine contents range from 19-31 percent, yet the modal range is only 0-9 percent. Similarly normative plagioclase ranges from 19-25 percent and the modal range is 0-14 percent. The reaction is probably:-



Garnet forming reactions are strongly dependent on bulk rock FeO/FeO + MgO, garnet appearing in rocks of low FeO/FeO + MgO at a higher pressure than those with a high ratio (Green and Ringwood, 1967a). The equilibrium pressure for the formation of garnet in the high-MgO (approximately 25%) ultramafites was not reached in the Scourie area. However, Livingstone (1967, 1976a, 1976b) has reported the co-existence of olivine and garnet in ultramafites from the Lewisian of South Harris.

The mafic and ultramafic rocks of the layered sheets provide an insight into the mineral reactions involved in the gabbro-eclogite transition. Although conditions of metamorphism in the Scourie region are thought not to have reached eclogite facies grade, mineral assemblages are consistent with a lower crustal origin (Tarney and Windley, 1978). The main effects of high grade metamorphism have been to:-

- 1). Increase the colour index of the mafic gabbros and ultramafites by a figure of approximately 10-30 percent.
- 2). Increase the density of the rocks by the production of garnet and spinel in the mineral assemblages.
- 3). Produce pyroxenes with high alumina contents, generally concentrated as Ca-Tschermak's molecule.
- 4). Decrease olivine contents in the ultramafites and to effect complete removal of olivine from plagioclase bearing rocks.

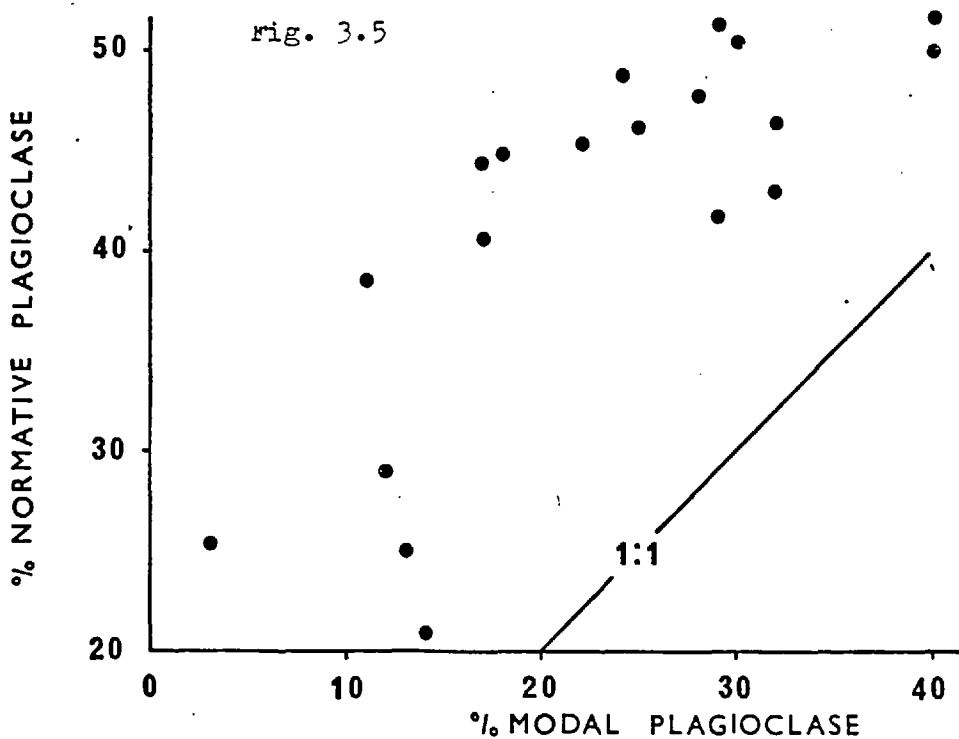
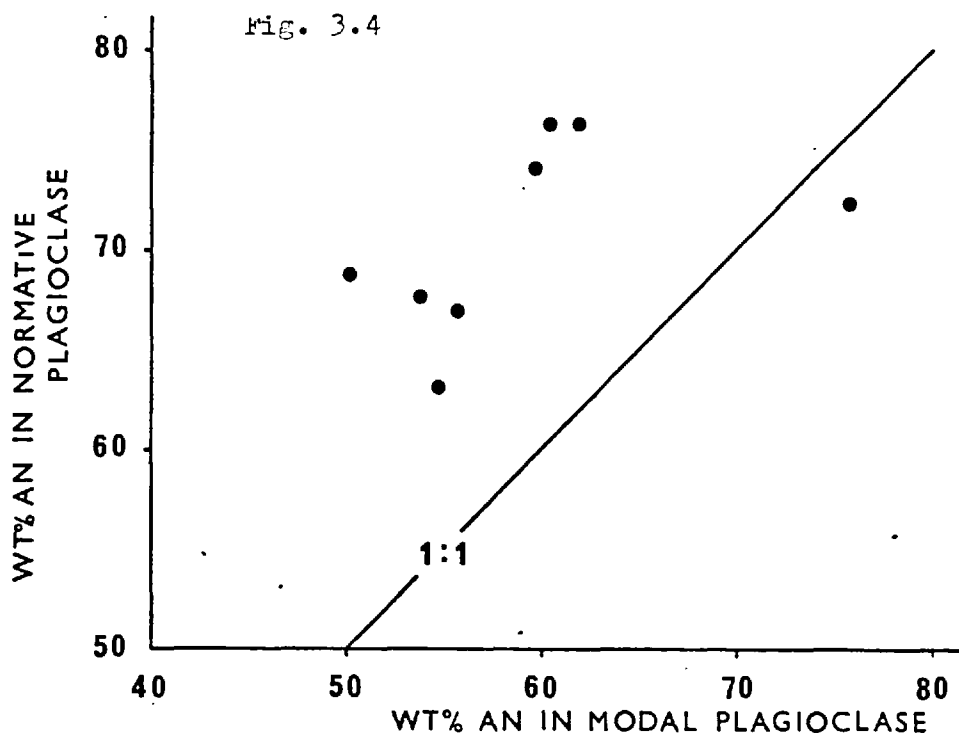


Fig. 3.4 Normative and modal plagioclase compositions of mafic gabbros.

Fig. 3.5 Normative and modal abundance of plagioclase in mafic gabbros.

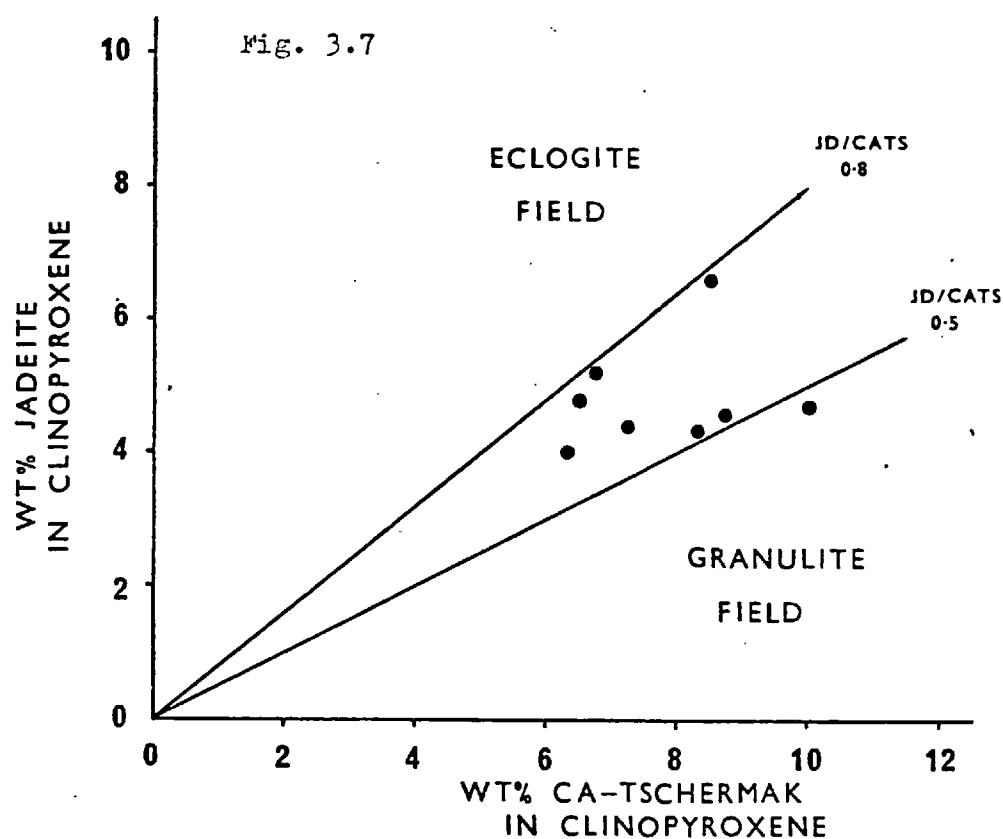
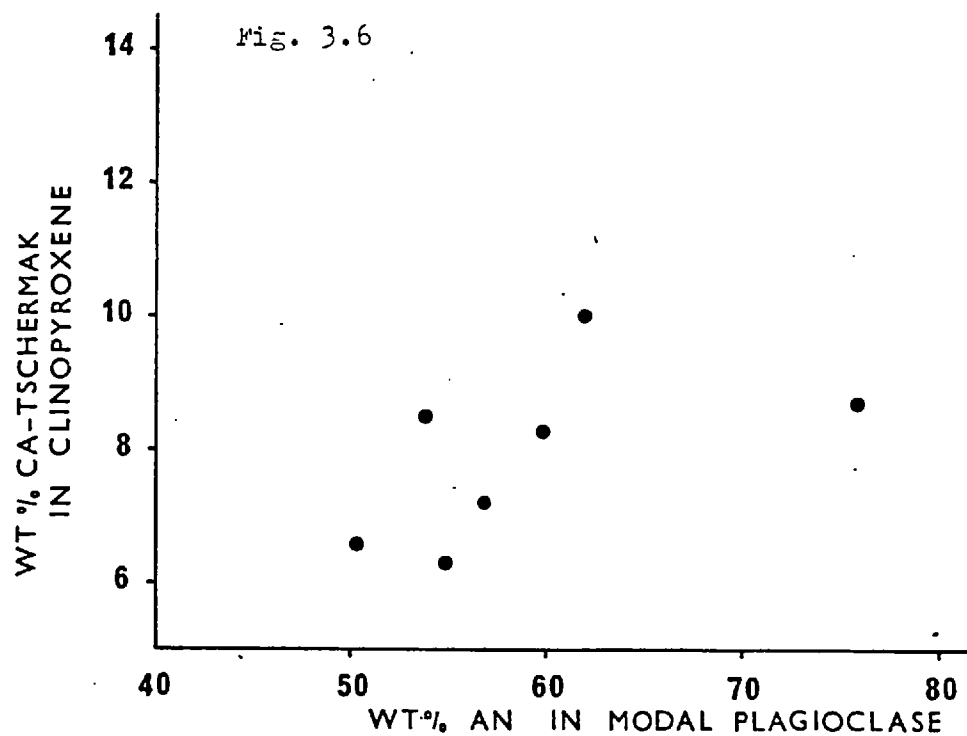


Fig. 3.6 Correlation of Ca-Ts component of clinopyroxene with An content of co-existing plagioclase.

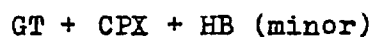
Fig. 3.7 Jadeite and Ca-Tschermak's content of clinopyroxenes of the mafic gabbros. Eclogite and granulite fields from White (1964).

3.2.2 THE M₁ METAMORPHISM

This is characterised by the earliest recognisable mineral assemblages and it may, in many cases, be represented only by relict phases superceded by later minerals. It is presumed to be early Scourian in age (c. 2.9 billion years old).

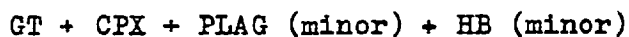
(i) Mafic Gabbros

The mineral assemblage in the magnesian mafic gabbros (77, 78, 84, 91 and also ultramafite 85) is:-



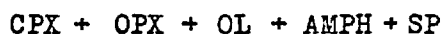
Plagioclase and orthopyroxene are believed to have been totally absent during this metamorphic event. Textural relationships which suggest this have been discussed in chapter two. Clinopyroxenes are assumed to have been unexsolved.

The mineral assemblage in the MgO-poor mafic gabbros is:-



(ii) Ultramafites

Low FeO/FeO + MgO ultramafites (74,75,86-90) have the following M₁ assemblage:-



This assemblage seems to have been little affected by subsequent metamorphic episodes and uplift. Pyroxenes are again assumed to have been homogeneous. Magnetite and pleonaste may possibly have formed a homogeneous solid solution.

Plagioclase-bearing ultramafites (85,104,108) have the assemblage:-



The presence of garnet is indicated by pseudomorphs of plagioclase, orthopyroxene and hercynite.

(iii) Summary

The above mineral assemblages indicate that the M₁ metamor-

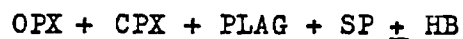
phism was of high-pressure granulite facies grade (Green and Ringwood, 1967a).

3.2.3 THE M₂ METAMORPHISM

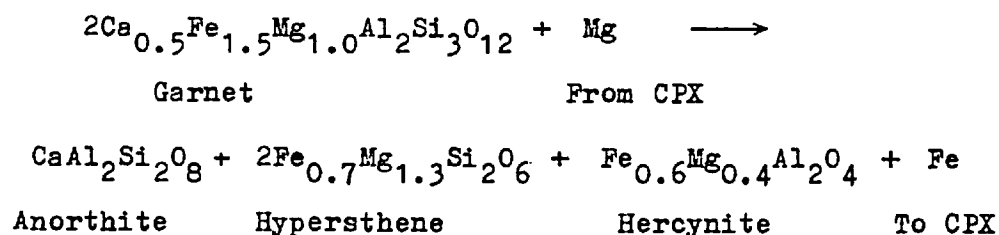
The M₂ mineral assemblage defines the S₂ foliation and is again early Scourian in age.

(i) Mafic Gabbros

Magnesian mafic gabbros have the assemblage:-



The garnet porphyroblasts of the M₁ metamorphism are corroded to a vermiform intergrowth of orthopyroxene, plagioclase and hercynite. The reaction was probably:-



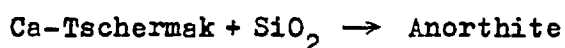
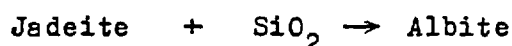
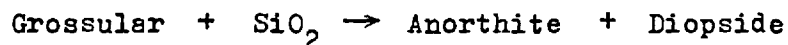
The involvement of clinopyroxene in cation exchange may not be real if any diffusion of Mg in garnet has taken place during re-equilibration with cooling of the complex. However, slight enrichment of clinopyroxene rims in FeO has been detected by microprobe analyses. Areas of larger garnet porphyroblasts show colour zoning (Plate 2.19). This has not been detected by microprobe analysis. Colour zoning has been observed in garnets from metapelites by Yardley (1977). He has suggested that clouded areas are due to the presence of minute cracks and inclusions, and correspond to areas that have changed composition through diffusion.

The exsolution of fine lamellae of orthopyroxene and platelets of iron oxide from clinopyroxene is probably a consequence of a small drop in temperature from M₁ to M₂ conditions.

MgO-poor mafic gabbros have the assemblage:-

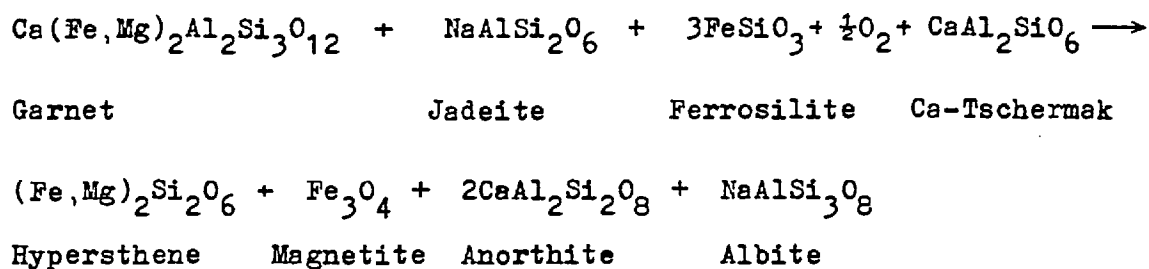


The embayed outlines of the garnet porphyroblasts suggest they were decreasing in abundance. This is to be expected with decreasing pressure and temperature (Green and Ringwood, 1967a). Garnet is replaced by orthopyroxene, plagioclase and magnetite. Pronounced plagioclase collars surround garnet porphyroblasts, and these in turn are rimmed by hypersthene grains against relict clinopyroxenes (Plate 2.21). These corona structures have been interpreted by O'Hara (1961, 1977a; O'Hara and Yarwood, 1978) as due to the exsolution of excess jadeite, calcium tschermak's molecule, hypersthene and iron-titanium spinel from an original high temperature pyroxene during the slow cooling of the complex. However, this does not account for the facts that the bulk of the plagioclase and all the hypersthene only occur adjacent to garnet porphyroblasts and that large areas of monomineralic clinopyroxene occur without plagioclase or hypersthene. Griffin and Heier (1973) Griffin (1972) and Griffin and Raheim (1973) have ascribed plagioclase 'collars' between clinopyroxene and garnet in eclogites to the following reactions:-



decreasing Pressure $\longrightarrow$

The Scourie garnets do not show zoning of grossular as would be expected from the above reactions and there is also the presence of hypersthene to be accounted for. The reaction is best approximated by:-



This accounts for the petrographic features described, and also the presence of magnetite with hypersthene in the coronas. It is apparent from the geometry of the coronas that nucleation and growth were not random, thus plagioclase selectively rims garnet. This may be controlled by intergranular diffusion of Al between garnet and plagioclase. Some authors have suggested that diffusion of aluminium is much slower than that of other major components in silicates and have noted the effect this has on textures (Anderson and Buckley, 1973; Carmichael, 1969; Fisher, 1973; Loomis, 1975, 1977). The exsolution of ferrosilite from clinopyroxene may also help nucleate hypersthene against clinopyroxene.

The consumption of garnet in garnet granulite-gabbro reactions is generally slower than its formation in the reverse situation. This is because its corrosion depends on solid state diffusion within garnet, a notoriously slow process (Loomis, 1977), whereas prograde reactions are not controlled by intragranular diffusion in garnet. Solid state diffusion in garnet may therefore be rate-limiting for the garnet granulite-gabbro transition (Loomis, 1977), resulting in pronounced zoning of the garnet shell. However, no zoning is observed in the Scourie garnets, a lack of which is probably a function of high temperature metamorphism (greater than 700°C) and slow cooling (Yardley, 1977). This has enabled the garnets to re-equilibrate by diffusion and it is therefore unlikely that intragranular diffusion in garnet was the rate-limiting step in this case. Diffusion in garnet has a high activation energy

(Yardley, 1977) and small decreases in temperature lower diffusion coefficients dramatically (diffusion coefficients vary exponentially as a function of reciprocal temperature). Decreasing temperature therefore greatly increases the time for re-equilibration of the garnet composition. A small decrease in temperature may therefore be enough to 'freeze' in chemical components and end reaction (Wood, 1977). Thus the supply of heat may have been rate-limiting in this case ('armouring' of reactant garnet and clinopyroxene by product plagioclase and hypersthene respectively, may also have been significant).

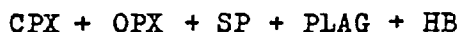
Recrystallised mosaics of fine grained clinopyroxene and plagioclase also occur in these rock types. As no chemical distinction can be made between the two textural types of pyroxene (chapter two), the recrystallisation may have occurred solely due to tectonic processes. Chemical distinctions may have been blurred by subsequent diffusion of components however, as pyroxenes are known to be poor indicators of relict metamorphism (Loomis, 1977). Minor exsolution of jadeite and ca-tschermak molecules from clinopyroxene could have taken place to produce plagioclase and a more diopsidic pyroxene. Relict clinopyroxenes contain orthopyroxene exsolution lamellae, which reflect a drop in temperature during the change from M_1 to M_2 conditions of metamorphism.

Composite grains of ilmenite and magnetite appear to have been homogeneous under M_1 conditions and their separation is coupled with cooling to M_2 conditions.

(ii) Ultramafites

The plagioclase-free ultramafites have the same mineral assemblage as for the M_1 metamorphism. The only visible differences are the presence of exsolved oxides in the pyroxenes and the exsolution of a homogeneous magnetite-spinel phase.

The assemblage for the plagioclase bearing ultramafites is:-



The stable garnets from the M_1 metamorphism broke down to orthopyroxene, plagioclase and spinel. Some orthopyroxene was also exsolved from the clinopyroxenes.

(iii) Summary

The stability of $\text{GT} + \text{OPX} + \text{CPX} + \text{PLAG}$ assemblages in MgO -poor undersaturated basaltic rocks and spinel-lherzolite assemblages in the ultramafic rocks indicate intermediate pressure granulite facies metamorphism (Green and Ringwood, 1967a).

3.2.4 THE M_3 METAMORPHISM

This period of mineral growth reflected the decreasing pressure and temperature and increasing PH_2O experienced by the Lewisian Complex. The appearance of hydrous minerals indicates a retrogression from granulite facies to amphibolite facies conditions. The hydrated assemblages are best developed in association with F_3^2 folding and is tentatively correlated with the Inverian event. The influx of volatiles associated with this event has been noted elsewhere (Sheraton et al., 1973; Sheraton et al., 1973; Holland and Lambert, 1973).

The M_3 metamorphism is represented by the growth of fibrous blue-green amphibole as fringes on clinopyroxenes and, occasionally orthopyroxenes. This is accompanied by minor growth of biotite between oxides, sulphides and plagioclase in the gabbroic rocks. Some serpentinisation of olivines also takes place.

The effects of this metamorphism are localised and vary from sample to sample.

3.2.5 THE M₄ METAMORPHISM

The effects are only visible in the MgO-poor mafic gabbros within the rock suite examined.

Small garnet fringes mantle oxides, sulphides, orthopyroxene and also biotite and amphibole formed during the M₃ metamorphism. These garnets are more iron-rich (5-6 mol. percent greater almandine) than the porphyroblastic type which indicates growth at a slightly lower pressure than the latter (Green and Ringwood, 1967a).

The M₄ metamorphism appears to have been at a higher grade than M₃, showing a return to granulite facies conditions within the area. A rise in temperature from the ambient P-T path would have been necessary to re-initiate garnet growth and overcome kinetic barriers associated with this growth.

3.2.6 CONDITIONS OF METAMORPHISM

This section records an attempt to quantify metamorphic conditions as characterised by mineral assemblages present within the mafic and ultramafic rock types and has involved the use of published mineral geothermometers and geobarometers, as well as experimental work on simple and complex systems.

O'Hara has recently published (O'Hara, 1977a, 1977b; O'Hara and Yarwood, 1978) pressure-temperature estimates of the Scourian granulite facies metamorphism and a review of selected geobarometers and geothermometers as applied to mineral equilibria in rocks from the Scourie area. This work to some extent overlaps that of O'Hara's although interpretations differ in certain areas.

The M₁ metamorphism probably took place under approximately 15 kilobars total pressure and 1000°C (fig. 3.8). There is considerable uncertainty attached to these figures due to the possibility of modification of M₁ mineral assemblages by subsequent

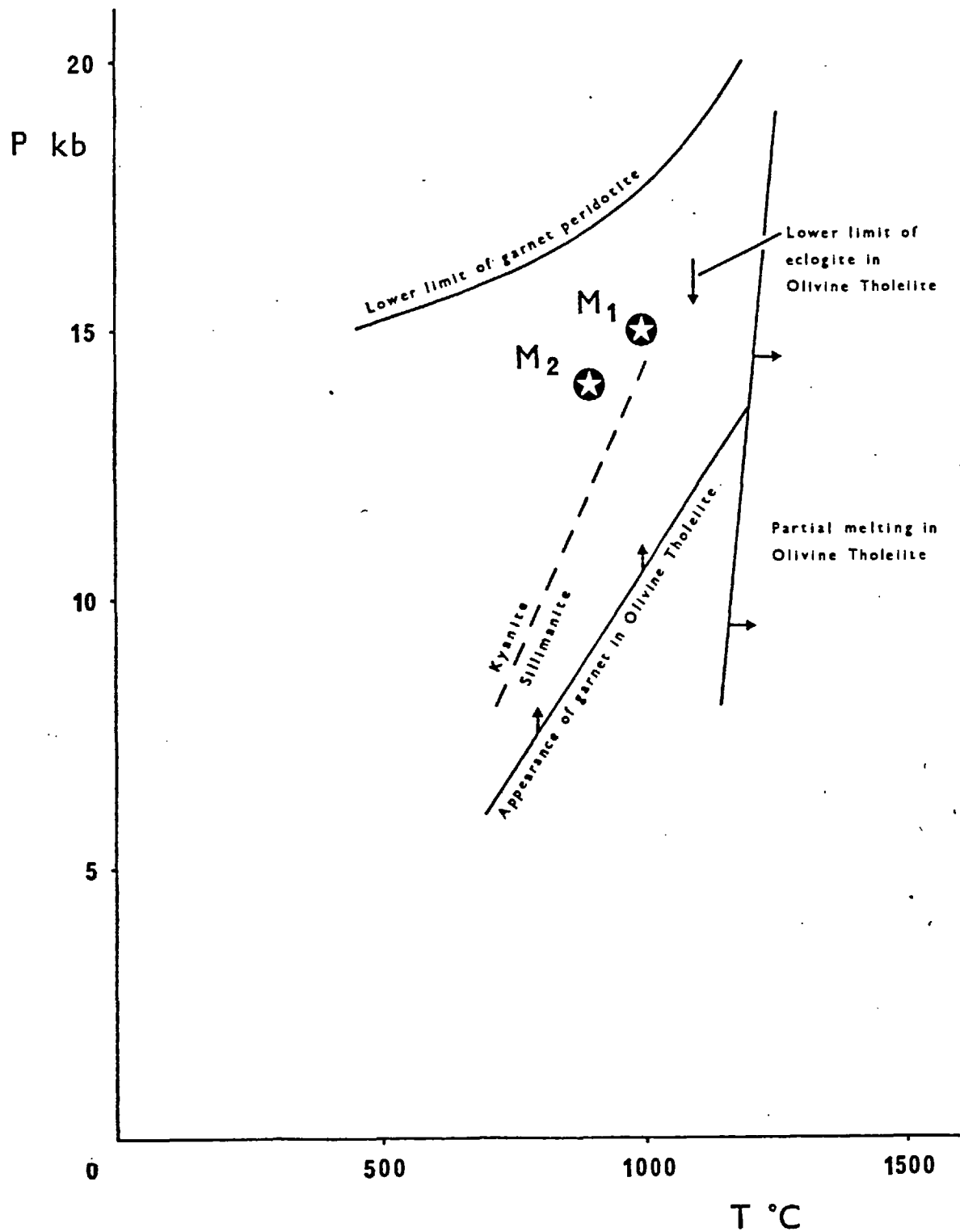


Fig. 3.8 Inferred metamorphic conditions of the M₁ metamorphism. Data from O'Hara (1971), Green and Hibberson (1970), Herzberg (1975), Holdaway (1971), and Green and Ringwood (1967a).

diffusion and recrystallisation through recovery from depth.

Although it may be possible from textural evidence to infer the mineral assemblages present during the metamorphism, compositions of minerals have probably been subsequently modified. This considerably limits the use of mineral equilibria to estimate metamorphic pressures and temperatures (O'Hara, 1977a; O'Hara and Yarwood, 1978). However, certain constraints may be placed on the conditions necessary for the metamorphism. The M_2 P-T estimate (fig. 3.9) is based on published geobarometers and geothermometers (see below). Slightly higher temperatures and pressures for the M_1 metamorphism are needed to explain the exsolution of orthopyroxene, calcium-tschermak's molecule and jadeite from clinopyroxenes in the mafic gabbros and the breakdown of garnet in the magnesian mafic gabbros. This places the conditions of M_1 metamorphism to the right of those of the M_2 metamorphism in fig. 3.9. The plagioclase-free ultramafites show no evidence of ever having been garnetiferous, thus limiting pressure to an upper level of 18 kilobars (O'Hara et al., 1971). The mafic gabbros show no textural evidence that plagioclase has been totally consumed in clinopyroxene and garnet (i.e. eclogite), although plagioclase content was probably quite low (Alderman (1936); O'Hara (1960) and Sanders (1978) have reported eclogites from the Lewisian east of the Moine Thrust at Glenelg. These are probably Archaean in age (Watson, 1978) and whole-rock chemical similarities with mafic gabbros from the Scourie area suggest that they may represent fragments of layered mafic-ultramafic sheets and metamorphosed at a slightly higher pressure). This limits pressure to approximately 16 kilobars (Green and Ringwood, 1967a). No partial melt textures have been observed in the mafic gabbros in the field, putting 1150-1200°C. as an absolute maximum for temperature (Green and

Ringwood, 1967a). Both kyanite and sillimanite have been reported from metasedimentary rock types in the Scourie-Laxford region (Beach, 1973; Davies, 1974a) although petrographic evidence suggests that kyanite was the earlier polymorph (J. Watson, pers. comm.). This places M_1 conditions to the left of the kyanite-sillimanite inversion (fig. 3.8).

O'Hara (1977a, 1977b; O'Hara and Yarwood, 1978) has suggested pressures of 15 kilobars and 1250°C . for this early metamorphic event. The pressure estimate coincides with this work, although temperature differs considerably. O'Hara's temperature estimate is based largely on the interpretation of feldspar textures in the granite sheets and on experimental dry melting of these rock types. Field evidence (chapter two) would suggest that the granite magmas were not intruded in a dry state, and that they post-date D_1 . Clearly, these temperatures would have caused partial melting within the mafic gabbros at pressures of fifteen kilobars (fig. 3.8). As already cited, field evidence suggests otherwise. Thus a temperature of 1000°C . is preferred for the M_1 metamorphism.

Mineral geothermometers and geobarometers have been used to estimate P-T conditions for the M_2 metamorphism. Four independent methods have been utilised, suggesting temperatures of $900 \pm 50^{\circ}\text{C}$. and pressures of 14 ± 3 kilobars (fig. 3.9). Temperatures have been derived by the methods of Wood and Banno (1973), Wells (1977) and Raheim and Green (1974). The distribution coefficient for iron and magnesium (K_D) between garnet and clinopyroxene has been calculated for the Scourie rocks and plotted in fig. 3.9. Only the 30 kilobar experimental data of Raheim and Green (1974) have been used, extrapolating to lower pressures as suggested by Wood (1977). K_D is almost independent of pressure and the Scourie data concentrate in the range $850-950^{\circ}\text{C}$. The empirical geotherm-

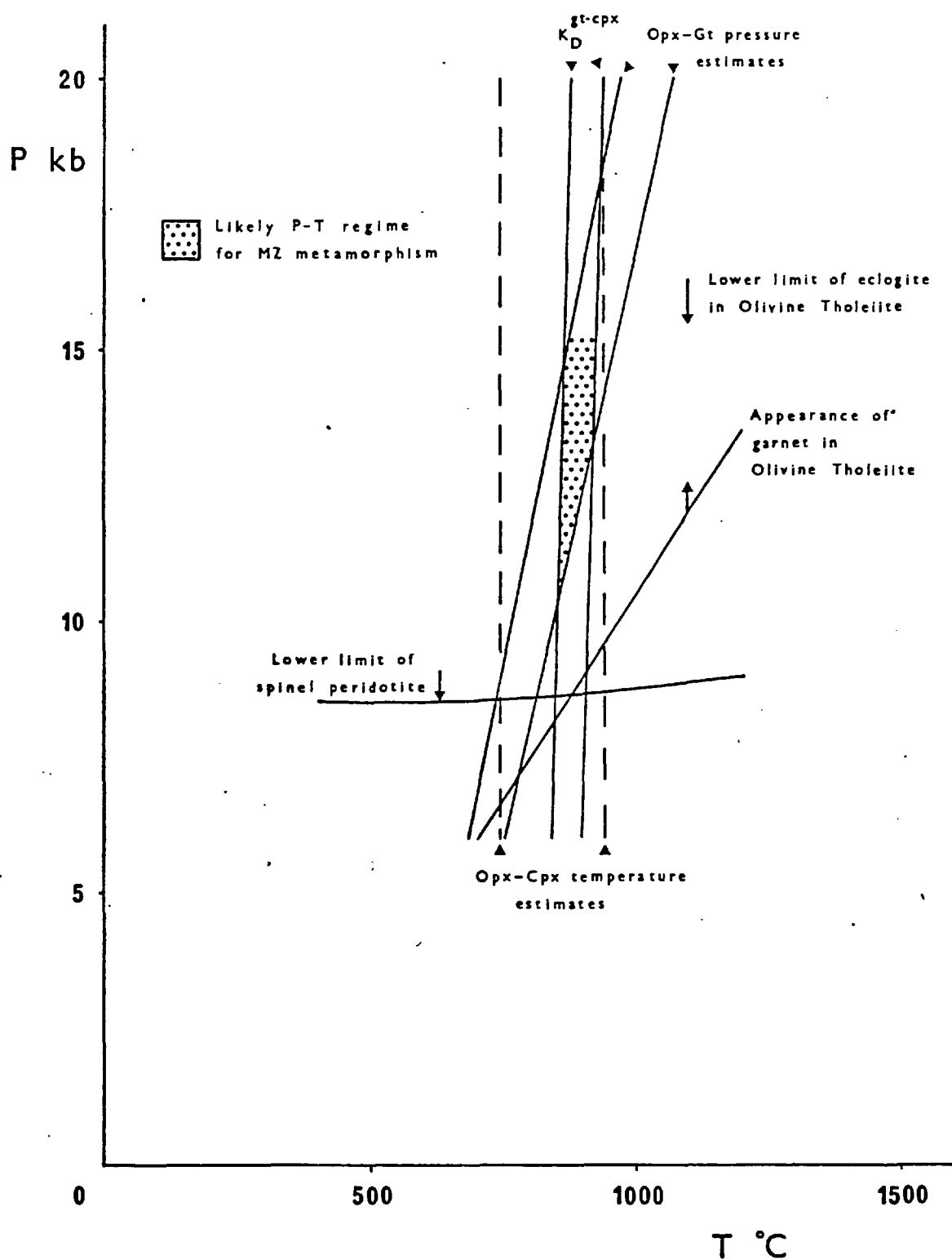


Fig. 3.9 Conditions of M₂ metamorphism. Two-pyroxene and pyroxene-garnet temperature bands include the range of mafic gabbros from Scourie. The orthopyroxene-garnet pressure band is defined by symplectite minerals from samples 84 and 85. Other data is from O'Hara et al. (1971), Green and Hibberson (1970), Herzberg (1975), and Green and Ringwood (1967a).

ometers founded on the enstatite-diopside miscibility gap of Wood and Banno (1973) and Wells (1977) have also been used to calculate temperatures for the Scourie rocks (Table 3.1). These range from approximately 750-940°C. No differences in temperature estimates are apparent between the exsolved pyroxenes of the mafic gabbros and the unexsolved pyroxenes of the ultramafites. The close agreement of the three methods would suggest that the temperatures determined represent a reasonable estimate of M_2 metamorphic conditions. Pressure has been derived by the geobarometer of Wood (equation 12, 1974). This depends on the alumina content of orthopyroxene, co-existing with garnet. As can be seen from fig. 3.9, the geobarometer is strongly dependent on temperature and a range of values result for the Scourie rocks (Table 3.2). This is probably due to re-equilibration of the orthopyroxenes and garnets with uplift (see below). However, the best estimates of pressure have been derived from the symplectite intergrowths of garnet and orthopyroxene from the magnesian mafic gabbros (Samples 84, 85). If any of the garnet and orthopyroxene pairs in the mafic gabbros are in equilibrium, then the symplectite minerals are the best candidates. Two textural types of orthopyroxene are present in the rocks, and the differences in composition are listed in Table 3.3. Alumina content is the significant difference, the symplectite orthopyroxene being poorer in alumina than the relict variety. These symplectite orthopyroxenes give significantly higher pressure estimates from the Wood (1974) geobarometer than the relict orthopyroxenes.

Fraser and Lawless (1978) have recently investigated temperatures and pressures of equilibration of garnet-lherzolite samples using the two-pyroxene and pyroxene-garnet methods quoted above. They observed that two-pyroxene and pyroxene-garnet temperatures

	<u>SAMPLE NO.</u>	<u>TEMPERATURE (°C)</u>	<u>TEMPERATURE (°C)</u>
		<u>WOOD & BANNO (1973)</u>	<u>WELLS (1977)</u>
	74	822	744
	86	885	814
	88	937	865
	104	879	854
	77	861	851
	78	800	865
(Opx-relict Cpx)	81	841	872
(Opx-recrystallised Cpx)	81	799	811
(Opx-Cpx)	84	845	851
(Symplectite Opx-Cpx)	84	856	823
(Opx-Cpx)	85	842	828
(Symplectite Opx-Cpx)	85	851	821
	91	857	871
	101	815	831
	106	813	832
(Retrogressed)	129	790	771

Table 3.1 Temperatures derived from the empirical two-pyroxene geothermometers of Wood & Banno (1973), and Wells (1977). Samples 74, 86, 88, and 104 are ultramafites, the rest are mafic gabbros.

	<u>SAMPLE NO.</u>	<u>PRESSURE (Kb)</u>
		<u>WOOD (1974)</u>
	77	5.9
	78	4.6
	81	7.3
(Symplectite Opx-Gt)	84	14.9
(Opx-Gt)	84	12.7
	85	9.0
	91	5.6
	101	8.0
	106	7.9
(Retrogressed)	129	5.3

Table 3.2 Pressures from the orthopyroxene-garnet geobarometer of Wood (1974) for the mafic gabbros and ultramafites. Temperatures are from the geothermometer of Wells (1977).

	84	84	85	85
	Relict Opx	Symplectite Opx	Opx in Gt-free Symplectite	Opx in Gt- Symplectite
Na ₂ O	-	-	-	0.12
MgO	22.78	24.74	23.94	26.15
Al ₂ O ₃	3.14	1.77	4.56	2.69
SiO ₂	52.63	53.41	52.06	53.41
CaO	0.27	0.18	0.41	0.37
TiO ₂	0.15	0.13	-	-
Cr ₂ O ₃	-	-	-	-
MnO	0.28	0.23	0.38	0.14
FeO (T)	21.51	19.12	19.04	16.99
NiO	-	-	-	-
Total	<u>100.76</u>	<u>99.58</u>	<u>100.39</u>	<u>99.87</u>

Table 3.3 Compositions of orthopyroxenes from samples 84 and 85.
The low-alumina pyroxene in each pair co-exists with garnet.

are coherent at high temperatures, but that discrepancies occur below 1100°C. The two pyroxenes monitor re-adjustment to lower temperatures better than the pyroxene-garnet pairs because of lower 'blocking' temperatures for Ca-Mg exchange between two pyroxenes than for Fe-Mg exchange between pyroxenes and garnets. This is a consequence of lower rates of diffusion in garnets as compared with pyroxenes. It would explain the spread of two-pyroxene temperatures below the range of pyroxene-garnet temperatures shown in fig. 3.9. Similarly, pressures derived by the orthopyroxene-garnet method will be low when coupled with two-pyroxene temperatures (fig. 3.9). Thus the best estimates for the M₂ metamorphism have been derived from pyroxene-garnet temperatures coupled with pyroxene-garnet pressures (fig. 3.9). It should be emphasised that these are only best estimates and still suffer some of the closure problems quoted in Fraser and Lawless (1978).

The M₃ metamorphism reflects an increase in P_{H2O} throughout the complex evidenced by the production of minor amounts of biotite and amphibole.

The M₄ metamorphism saw an increase in the amount of garnet in the MgO-poor mafic gabbros. An increase in temperature relative to the M₁-M₂ recovery path is required to reach the activation energy necessary for the process and to effect the re-growth of garnet. Temperatures from the garnet-clinopyroxene geothermometer (Raheim and Green, 1974) suggest values in the range 800-850°C. The garnet coronas are richer in the almandine molecule than the earlier garnet porphyroblasts, suggesting slightly lower pressures (1-2 kb?) than for the M₂ metamorphism (Green and Ringwood, 1967a). Thus the M₄ metamorphism probably represents a significant thermal event, late in the Scourian metamorphism at approximately 825 ± 50°C.

and 12 ± 2 kb (fig.3.10).

Temperatures from iron-titanium oxides range from 530°C to 740°C . with $\log_{10} f\text{O}_2$ ranging from -28 to -35 bars, depending on the method of recalculation of the mineral formulae (Buddington and Lindsley, 1974; Anderson, 1968; Powell and Powell, 1977a). These temperatures and oxygen fugacities probably represent conditions of metamorphism at the final equilibration of ilmenite-magnetite pairs and need not indicate peak temperatures of metamorphism. This is a common phenomenon in slowly cooled plutonic rocks e.g. Buchanan, 1976b. However, low oxygen fugacities are also indicated from positive europium anomalies in rare-earth element patterns from the granite sheets (chapter four, cf. Drake, 1975).

Fig. 3.11 shows the likely P-T path for the Scourie area, in comparison with various geothermal gradients and the modern cratonic crustal geotherm. The conditions for the intrusion of the Scourie Dykes are based on garnet-clinopyroxene temperatures (Raheim and Green, 1974), using the data of O'Hara (1961b) and pressure estimates are those necessary to preserve olivine and plagioclase, and also to produce garnet coronas between these two phases (O'Hara, 1977b). These are approximately 600°C . and 7 kilobars. This gives the Scourie region a geothermal gradient between $20\text{--}30^{\circ}\text{C}/\text{km}$ depth. Most Archaean granulite facies regions fall in a field bounded by these geothermal gradients (Watson, 1978). The M_1 and M_2 pressure estimates quoted above suggest Archaean crustal thicknesses of forty-fifty kilometers (consistent with present crustal thicknesses) in the area (if it may be assumed that the Scourian granulites represent lower crustal material). Depths of metamorphism of this order seem to be the exception rather than the rule for Archaean granulite facies regions (Watson, 1978). However,

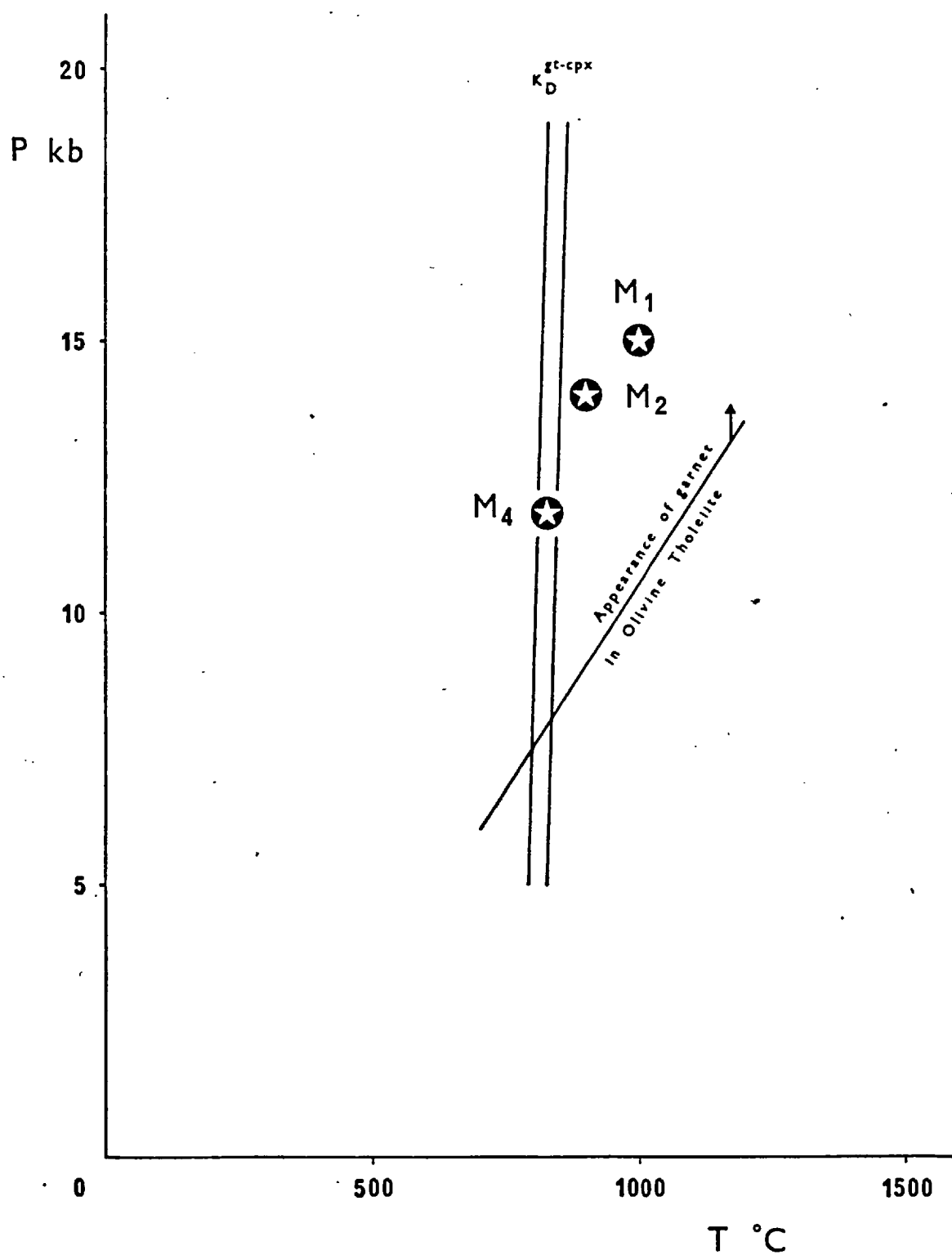


Fig. 3.10 Conditions of M_4 metamorphism. The K_D^{gt-cpx} band defines the P-T range of all mafic gabbros with m_4 garnet. Olivine-tholeiite data is from Green and Kingwood (1967a).

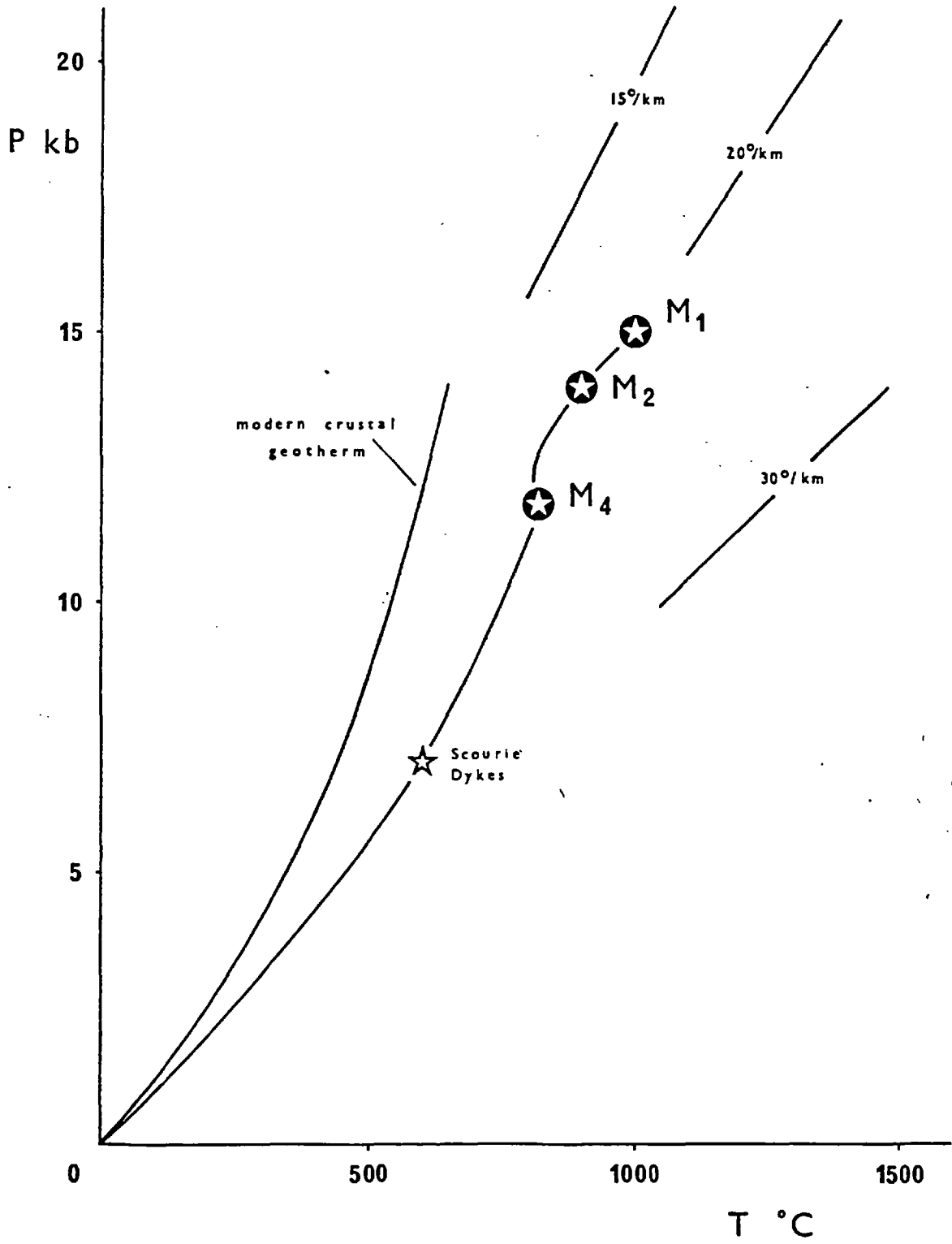


Fig. 3.11 P-T path for the Scourie area. Crustal (cratonic) geotherm from Bott (1971).

these estimates of Scourian crustal thicknesses put in question the validity of assumptions of a universally 'thin' Archaean crust (Fyfe, 1973; Anhaeusser et al., 1969; Lambert, 1976) and non-uniformitarian conceptions of Archaean geological processes (Anhaeusser et al., 1969; Glikson, 1971; Lambert, 1976). A proto-plate tectonic mechanism has recently been proposed for the Archaean (Tarney and Windley, 1978) involving crustal generation at island arc/cordilleran type plate boundaries, and is probably a more realistic approach.

CHAPTER 4

ROCK CHEMISTRY

Major and trace-element data for forty-five rock samples from the layered mafic-ultramafic sheets, granite sheets and net-veined mafic-ultramafic sheets are presented in Tables 5 - 11. A description of analytical techniques is included in Appendix I.

The bulk of the geochemical data comes from a traverse of the Camas nam Ruth layered mafic-ultramafic sheet. This intrusion was selected for sampling because of its good preservation of granulite facies mineral assemblages, its well developed layering and its relative lack of later retrograde metamorphic effects and subsequent deformation. The investigation of the original nature and igneous fractionation trends of the layered sheets is critically dependent on the above factors and on the degree to which the granulite facies metamorphism was isochemical. A great deal of the controversy between O'Hara (1961a, 1965, 1966), Bowes et al. (1964, 1966), and Bowdidge (1969) concerning the original nature of the layered sheets can be ascribed to the fact that each protagonist sampled different intrusions in varying states of metamorphism and deformation, and that attempts have been made to compare the diverse resulting chemical trends.

A small amount of geochemical work has been performed on the granite sheets and the net-veined intrusions and the data from this work are presented in sections 4.2 and 4.3.

4.1 THE LAYERED MAFIC-ULTRAMAFIC SHEETS

As Bowes et al. (1964) pointed out, the mafic and ultramafic rocks of the layered sheets closely resemble un-metamorphosed

igneous mafic and ultramafic rock-types in their chemical characteristics. This similarity is illustrated in Tables 4.1 and 4.2.

The mafic gabbros are iron-rich, alkali-poor olivine-tholeiites, similar to modern mid-ocean ridge basalts (Engel and Fisher, 1975) and island arc tholeiites (Jakes and White, 1972). The ultramafites are iron-rich and reasonably MgO poor (cf. 'Alpine-type' ultramafics, Wyllie, 1967), with appreciable chromium (3000 ppm) and nickel (1000 ppm) contents.

The resemblance between the Camas nam Buth rocks and un-metamorphosed igneous rocks in major and trace element compositions is excellent, considering the degree of metamorphism experienced by the Scourian rocks. The only exception to this generalisation concerns the abundance of rubidium (and thorium), which seems depleted in Camas nam Buth relative to the other rocks presented in Table 4.1. This is probably a consequence of granulite-facies metamorphism and is a well-documented feature of many high grade metamorphic regions including the Lewisian complex itself (Lambert and Heier, 1968; Shaw, 1968; Signinolfi, 1971; Tarney et al., 1972; Sherraton et al., 1973; Heier, 1973). Generally associated with Rb-depletion is a loss of K, Cs, Th and U. Consequently, granulites are characterised by high K/Rb, Ba/Rb and low Rb/Sr, the K/Sr and K/Ba ratios are not anomalous, suggesting that depletion of K, Sr and Ba has been relatively minor in comparison with Rb. The anomalous ratios cited above appear to be accounted for primarily by a loss of rubidium. The ultramafites, on the other hand, show no such depletion, probably because of the abundance of a pargasitic amphibole capable of retaining the large rubidium ion within its structure. Apart from rubidium (and probably thorium) therefore, there is no reason to suggest that the Camas nam Buth sheet behaved in any other way than as a geochemically closed system during metamorphism.

	1	2	3	4	5	6
SiO ₂	46.73	54.02	46.59	47.41	44.00	45.10
TiO ₂	0.35	0.09	1.83	0.89	0.58	0.76
Al ₂ O ₃	7.37	3.90	7.69	8.66	8.30	10.80
Fe ₂ O ₃	3.38	0.32	2.20	2.81	2.20	2.40
FeO	8.91	4.22	10.46	11.15	8.80	8.10
MnO	0.21	0.03	0.18	0.20	0.19	0.18
MgO	22.64	31.76	21.79	19.29	26.00	19.70
CaO	7.09	1.92	7.41	6.76	7.30	9.20
Na ₂ O	0.67	0.32	1.33	1.35	0.90	1.04
K ₂ O	0.34	0.03	0.28	0.43	0.06	0.08
P ₂ O ₅	0.03	0.01	0.11	0.10	0.07	0.09
H ₂ O ⁻	0.10	0.40	0.04	0.11	-	-
L.O.I.	1.89	2.01	0.37	1.45	0.97	2.14
Total	99.71	99.53	100.53	100.61	99.37	99.59

V	141	59		155		
Cr	2991	2300		290	2463	1847
Ni	986	1300		500	1493	943
Cu	19	<2		90	81	109
Zn	118				60	58
Zr	23	<20		90	45	52
Rb	7				1	2
Sr	51	<3		110	87	186
Y	8	<20		24	13	18
Ba	23	2		150	44	55

K/Rb	403				498	332
Ba/Rb	3				44	27
Rb/Sr	.137				.011	.011
K/Ba	123	125		24	11	12
K/Sr	55			32	6	4

Table 4.1 Comparison of the Camas nam Ruth ultramafites with other Mg-rich basalts. 2=Orthopyroxenite, Central Indian Ridge (Engel & Fisher, 1975); 3=Picrite, Hawaii (Wyllie, 1967); 4=Olivine Dolerite, Palisades Sill (Walker, 1969); 5=Picritic Basalt, Baffin Island (Clarke, 1970); 6=Olivine Basalt, Baffin Island (Clarke, 1970). 1 is the average Camas nam Ruth ultramafite.

	<u>1</u>	<u>2</u>	<u>3</u>	<u>4</u>	<u>5</u>	<u>6</u>	<u>7</u>	<u>8</u>
SiO ₂	47.31	50.60	50.67	51.57	46.37	48.56	47.09	46.62
TiO ₂	0.86	0.36	1.24	0.80	0.79	0.24	1.12	0.96
Al ₂ O ₃	14.13	16.29	16.60	15.91	15.82	18.69	14.92	16.05
Fe ₂ O ₃	2.42	1.87	1.33	2.74	1.52	2.27	3.82	11.11
FeO	10.82	4.12	7.37	7.04	10.44	4.30	7.30	-
MnO	0.25	0.09	0.16	0.17	0.09	0.11	0.20	0.17
MgO	9.16	9.59	8.02	6.73	9.61	9.26	9.05	8.15
CaO	11.66	14.33	11.58	11.74	11.29	12.67	12.67	12.66
Na ₂ O	1.87	2.15	2.75	2.41	2.45	1.88	1.87	2.05
K ₂ O	0.27	0.07	0.16	0.44	0.20	0.07	0.10	0.10
P ₂ O ₅	0.09	0.04	0.11	0.11	0.06	0.02	0.10	0.09
H ₂ O ⁻	0.12	0.15	-	0.45	0.09	0.17	-	-
L.O.I.	1.40	0.52	-		0.29	1.72	1.43	-
Total	<u>100.36</u>	<u>100.18</u>		<u>100.11</u>	<u>100.02</u>	<u>99.96</u>	<u>99.67</u>	

V	273	160	250	270	225	110	315	
Cr	461	670	360	50	175	900	310	
Ni	167	240	109	30	135	200	116	
Cu	182	60	93		80	85	156	
Zn	162					45	73	
Zr	37	19	71	70	35	<10	58	47
Rb	1			5			2	2
Sr	130	130	108	200	700	110	133	125
Y	26	<10	35			25	28	23
Ba	45	10	17	75	25	10	148	
K/Rb	2241			730			415	415
Ba/Rb	45			15			74	
Rb/Sr	.0077			.025			.015	
K/Ba	50	58	78	49	66	58	6	
K/Sr	17	4	12	18	2	5	6	7

Table 4.2 Comparison of the Camas nam Butte mafic gabbros with other basalts.

1 = Average Mafic Gabbro; 2 = Average Olivine Gabbro, Central Indian Ridge (Engel & Fisher, 1975); 3 = Average Low-K Tholeiite, Central Indian Ridge (Engel & Fisher, 1975); 4 = Average Island Arc Tholeiite (Jakes & White, 1972); 5 = Average Gabbro from LZ4; Skaergaard (Jager & Brown, 1967); 6 = Gabbro, Romanche Fracture (Melson & Thompson, 1970); 7 = Olivine Tholeiite, Skye (Esson et al., 1975); 8 = Average Low-alkali Tholeiite, Skye (Lattey et al., 1977).

This is not surprising when one considers the evidence of lack of diffusion between minerals on the micron scale (e.g. preservation of symplectite and relict orthopyroxenes with differing compositions in samples 84 and 85). Similar isochemical metamorphism has been observed in other granulite facies terrains with regard to strontium isotope exchange (Moorbath, 1978). The apparent element immobility implicit in these observations casts severe doubts on the scale of diffusion required (on the scale of metres) to produce the gabbroic rock types as 'aureoles' around the ultramafites as suggested by O'Hara (1961a, 1965, 1966).

Retrograde metamorphism has been found to have a considerable effect on the rock chemistry of the ultramafites and mafic gabbros. Analyses 109, 129 and 130, presented in Table 9 as examples, are from the margin of the Camas nam Buth Sheet (109) and the region of F_3^2 folding in the large scale F_2 isocline fold nose at Camas nam Buth (samples 129, 130). The mafic gabbros show significant increases in potash (up to 1.8 wt. percent) and rubidium (up to 24 ppm), both well above the average presented in Table 4.1. There is a concurrent lowering of K/Rb ratios. The ultramarite shows a considerable increase in volatile content (L.O.I. over 9%) and a decrease in K_2O and Rb, presumably as a result of the breakdown of the Rb retaining phase, a pargasitic amphibole.

Analyses of the sulphide-rich mafic gabbros are also presented in Table 7. These should be regarded as qualitative analyses only, as it was impossible to obtain totally fresh samples of these rock types. However, they provide the basis for estimates of the degree of enrichment in iron and copper relative to the other mafic gabbros due to the occurrence of sulphides. The chemistry and differentiation of the layered sheets will therefore be discussed in terms of the ultramafites and mafic gabbros.

4.1.1 GENERAL CHEMISTRY

Geochemical variation in the ultramafites and mafic gabbros of the Camas nam Buth sheet are plotted on an AFM diagram in fig. 4.1. This diagram reveals an alkali-poor 'tholeiitic' trend with a moderate degree of iron enrichment which is clearly distinct from the calc-alkaline trends defined by Irvine and Baragar (1971). The field of compositions analysed by Bowes et al. (1966) from layered mafic-ultramafic intrusions in the Drumbeg area is also plotted on fig. 4.1. for comparison with the Camas nam Buth data. The Drumbeg data define a different trend from the Camas nam Buth rocks, being calc-alkaline in character. It is difficult to reconcile such divergent trends with a common origin for the intrusions in the two areas. It may be that the layered mafic-ultramafic sheets of the mainland Lewisian belong to two (or more) suites. This point will be returned to in the final chapter. The tholeiitic nature of the mafic gabbros and ultramafites from Camas nam Buth is further emphasised in fig. 4.2, a plot devised by Irvine and Baragar (1971) to distinguish tholeiitic from calc-alkaline types of basaltic rocks.

The mafic gabbros and ultramafites plot in the undersaturated, normative olivine and hypersthene tholeiite field of the 'basalt tetrahedron' (Yoder and Tilley, 1962), as shown in fig. 4.3. Only one analysis falls in the nepheline-normative field, and in this the normative nepheline is only 0.1 wt. percent. The ultramafites cluster along the olivine-enstatite join, the mafic gabbros plotting further towards plagioclase. There is no crossing into the quartz-normative field.

Normative feldspar compositions (fig. 4.4) fall in a narrow range between labradorite and bytownite, and are generally poor in the orthoclase component. Two of the ultramafites are anomalous in this respect, showing unusually high orthoclase contents for the

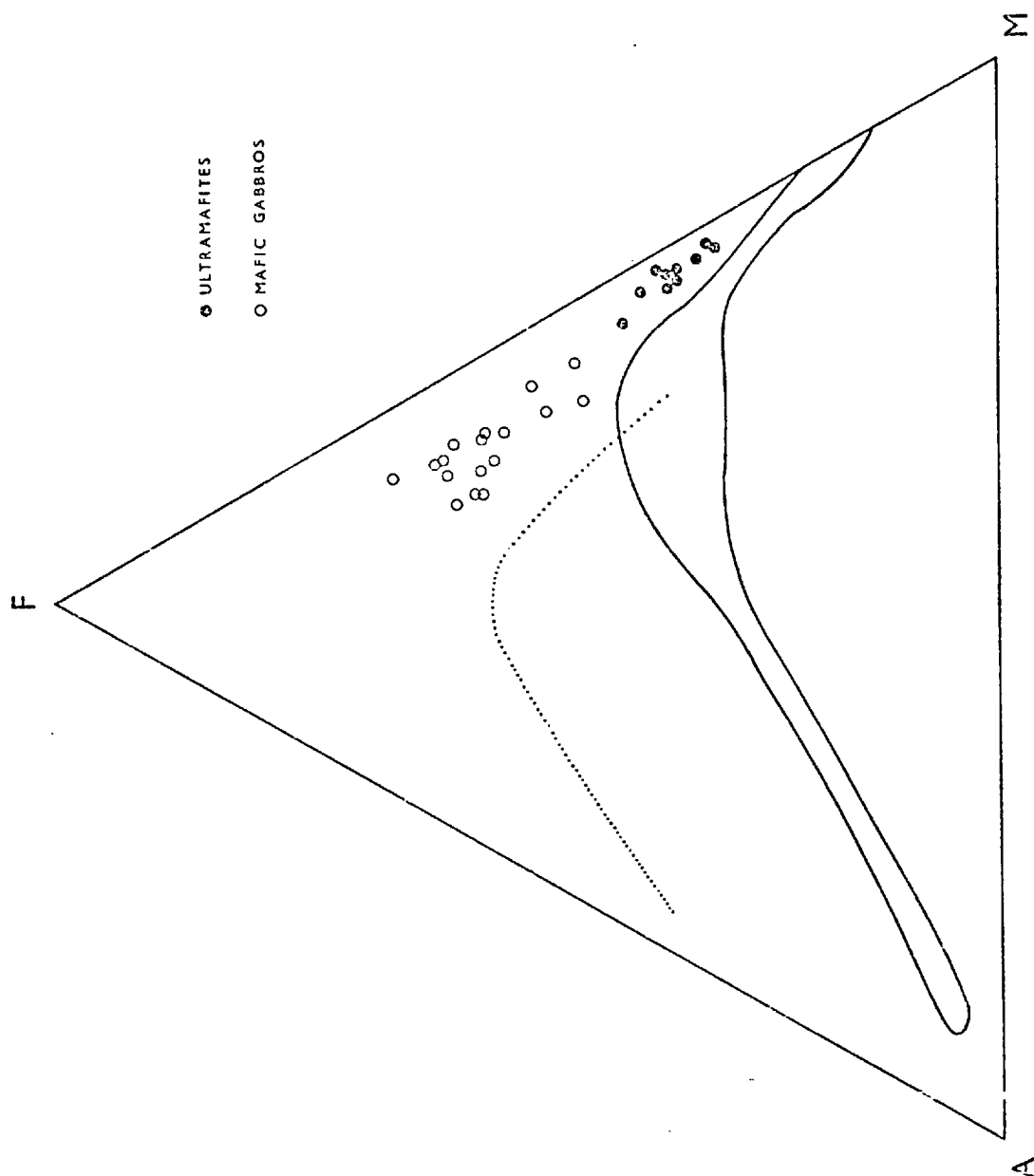


Fig. 4.1 AFM ($\text{Na}_2\text{O} + \text{K}_2\text{O} - \text{FeO(T)} - \text{H}_2\text{O}$) diagram of the Camas bath samples. The dotted line separates tholeiitic (above) and calc-alkaline (below) rock series (Irvine and Baragar, 1971). The field enclosed by the solid line is of rocks from layered mafic-ultramafic sheets from Kamourie to Achiltibuie analysed by Bowes et al. (1965).

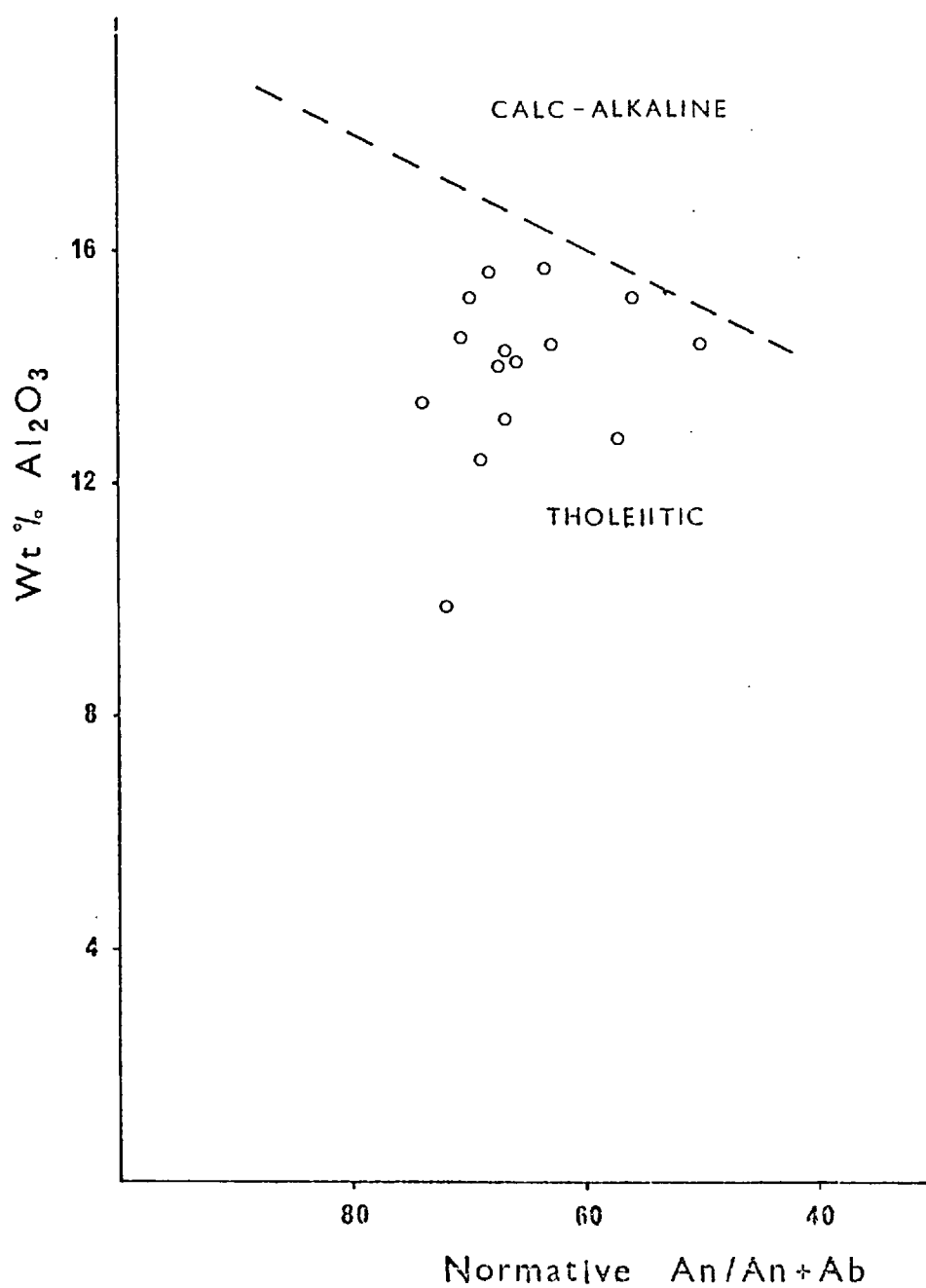


Fig. 4.2 The mafic gabbros plotted on the discriminant diagram of Irvine and Saragamo (1971).

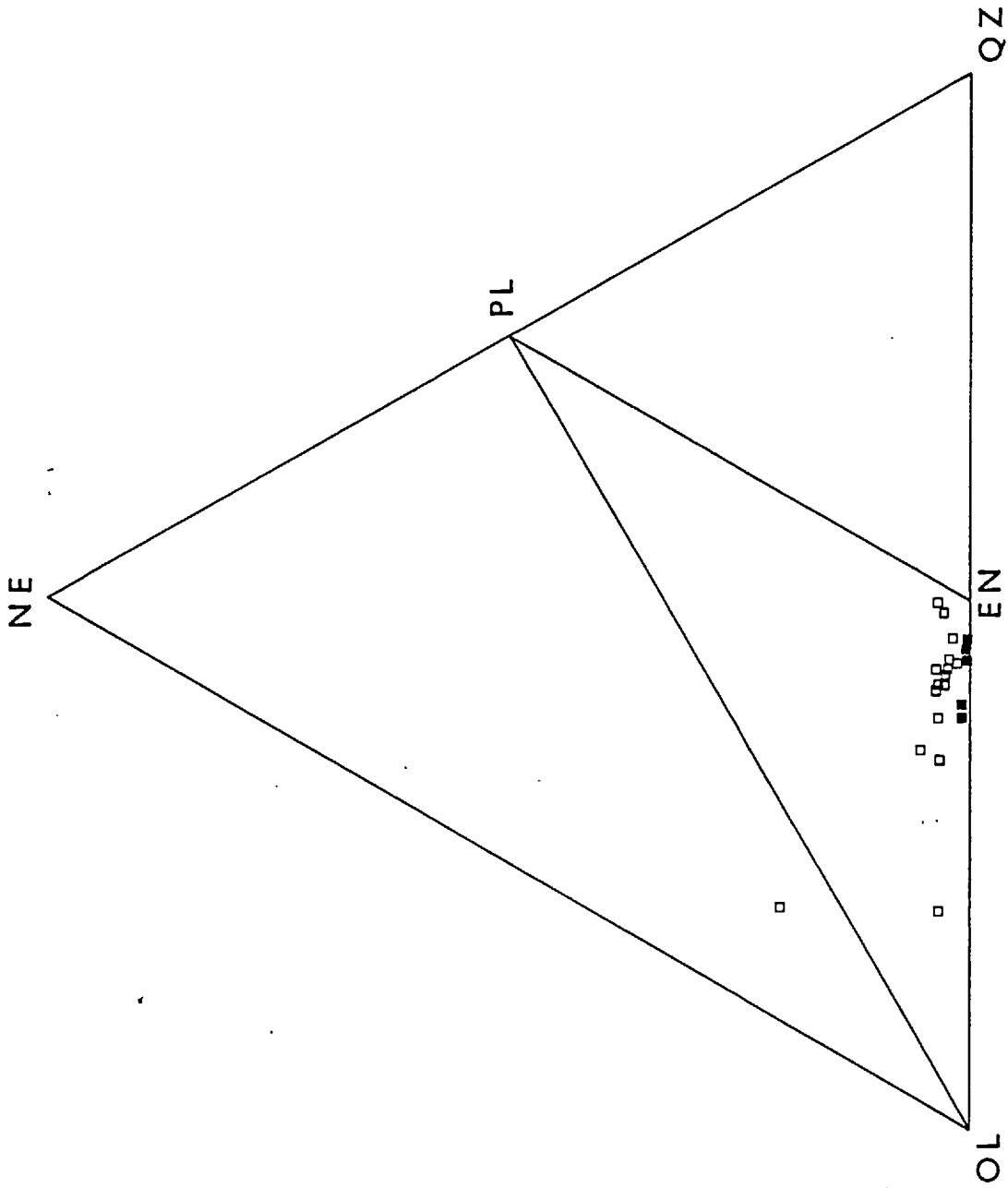


Fig. 4.3 The Camas nam Buth mafic gabbros (open squares) and ultramafites (filled squares) plotted in the basalt tetrahedron of Yoder and Tilley (1962).

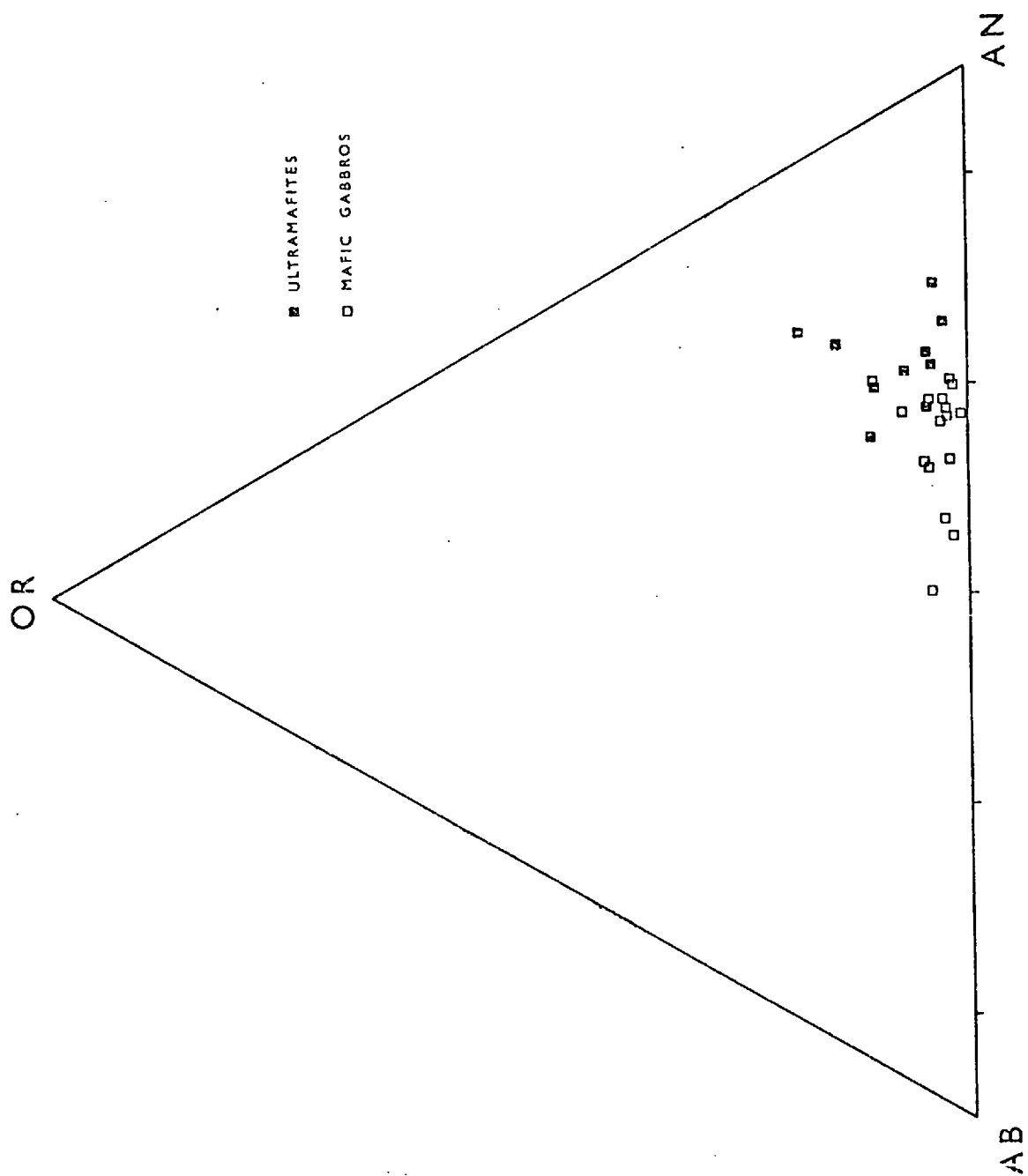


Fig. 4.4 Normative feldspar compositions of the Camas nam Buth samples.

calcic nature of the plagioclase. These both contain modal plagioclase (104, 108) which is considerably clouded and sericitised, probably due to an unfiltration of potash.

4.1.2 GEOCHEMICAL DIFFERENTIATION OF THE SHEETS

It was mentioned in chapter two that gravity-stratified units and igneous differentiation trends may be recognised from the geochemical traverse of the Camas nam Ruth sheet. A hypothesis of multiple injection of magma was invoked to account for the development of the units. It is now necessary to investigate this further to examine the physical conditions of crystallisation and the phases controlling fractionation within the sheet. The data from the geochemical traverse have been plotted on variation diagrams (see below). No distinction between different cyclic units has been made on the diagrams, as each unit appears to have differentiated in the same manner. This would suggest that the periodic influxes of new magma were from a common source.

Major-element concentrations for the ultramafites and mafic gabbros are plotted against Mafic Index ($100 \times \text{FeO} + \text{Fe}_2\text{O}_3 / \text{FeO} + \text{Fe}_2\text{O}_3 + \text{MgO}$) in fig. 4.5 a-h and trace elements against Mafic Index in fig. 4.6 a-h. All the diagrams show a bi-modal distribution of data points, with the ultramafites plotting in a range from 30.0 - 43.0 Mafic Index, and the mafic gabbros plotting in a range from 47.0 - 70.0 Mafic Index. This bi-modality is emphasised for certain elements by scattered distributions and irregular trends (Al_2O_3 , MnO, CaO, K_2O , Cu, Zr, Zn), although other elements show well-defined trends (TiO_2 , Na_2O , P_2O_5 , Ni, Cr, V, Y). Thus, although it seems as if the Camas nam Ruth sheet as a whole has remained as a closed system during metamorphism (section 4.1.1), minor mobilisation of certain elements within the sheet may have taken place.

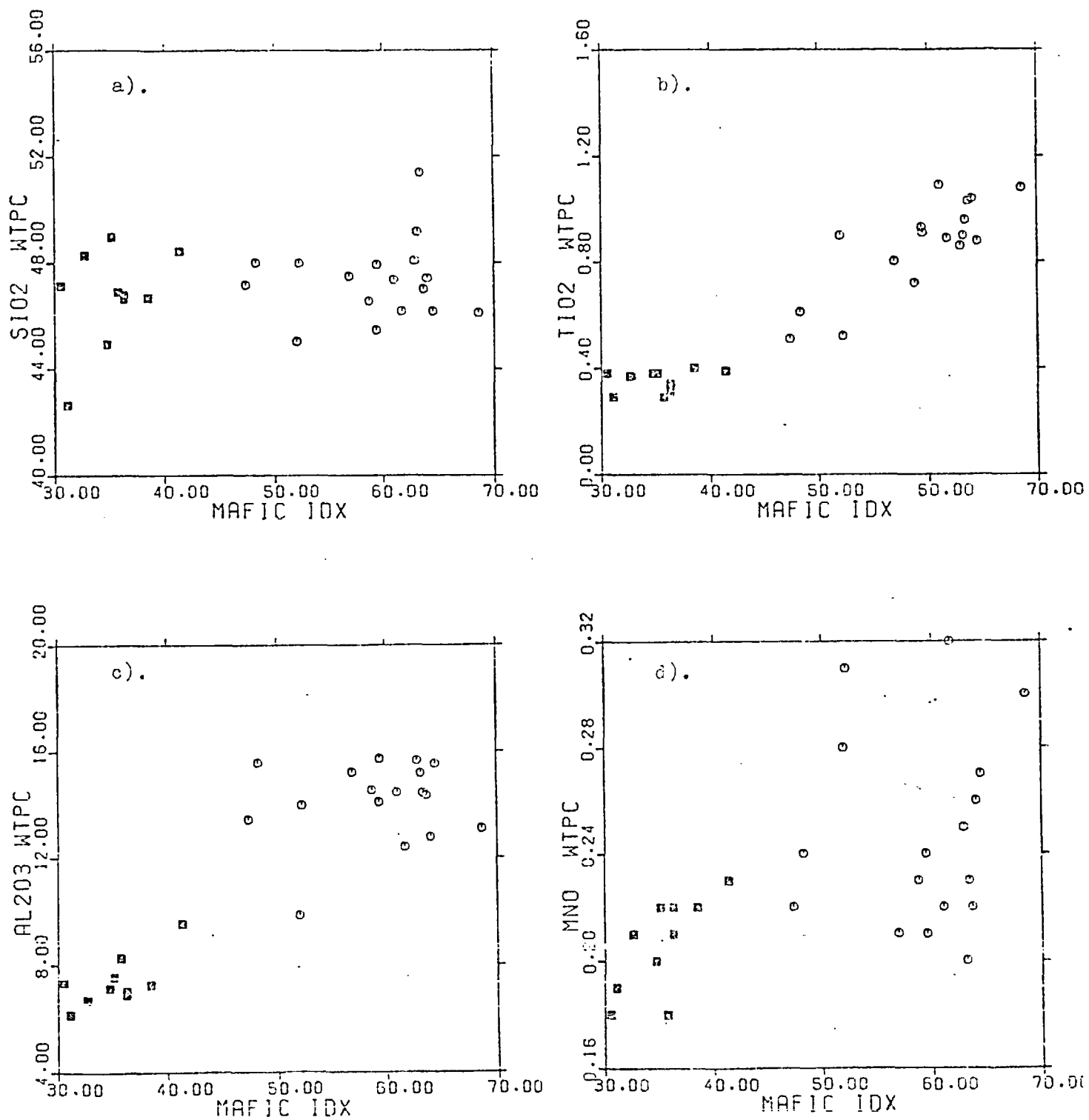


Fig. 4.5 a-h. Major element concentrations plotted against Mafic Index ($100 \times \text{FeO} + \text{Fe}_2\text{O}_3 / \text{FeO} + \text{Fe}_2\text{O}_3 + \text{MgO}$) for the Camas nam bath samples. Ultramafites are plotted as filled symbols, mafic gabbros as open symbols.

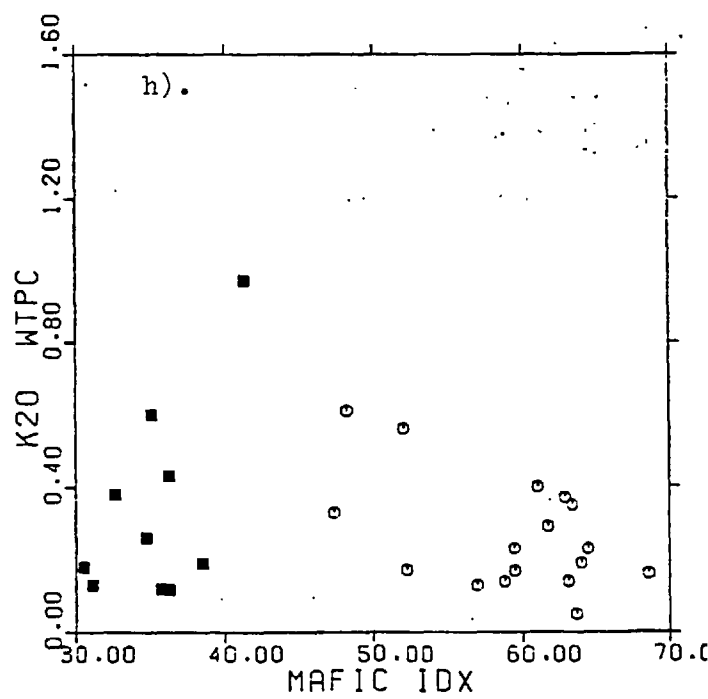
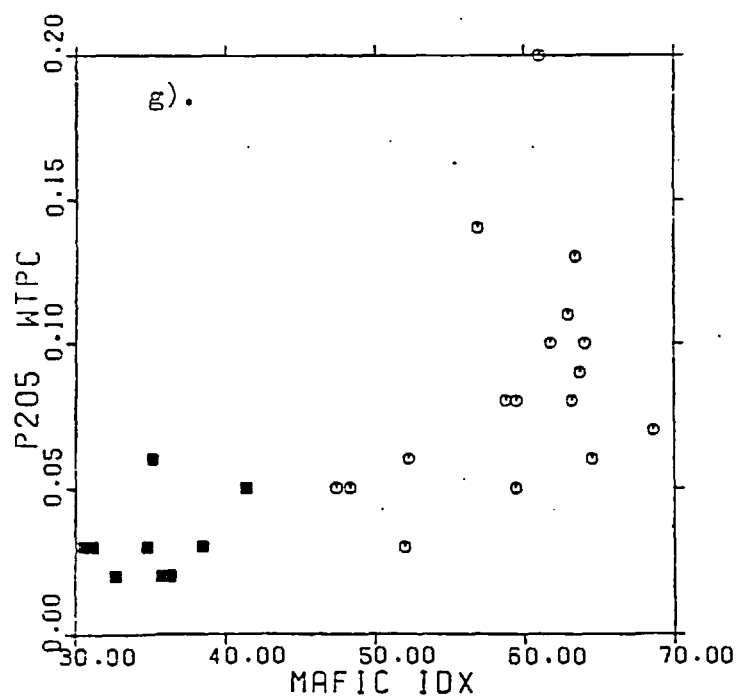
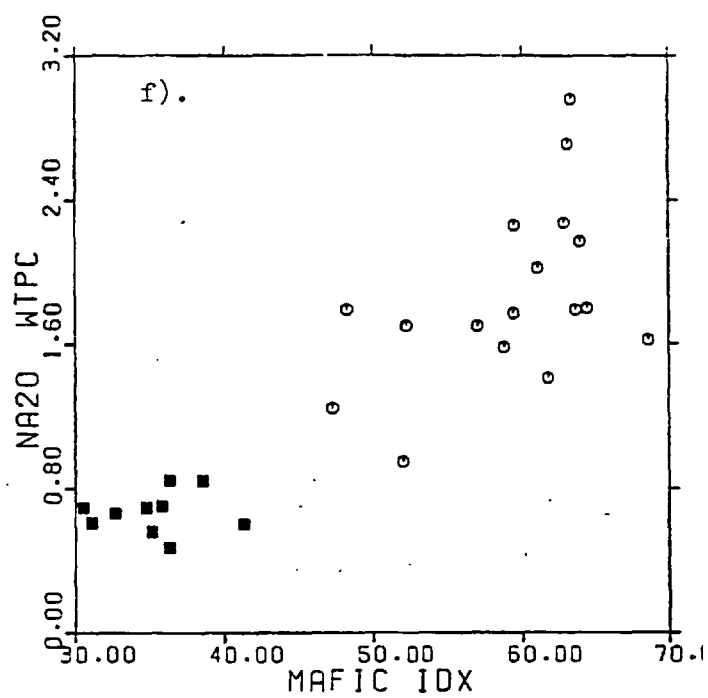
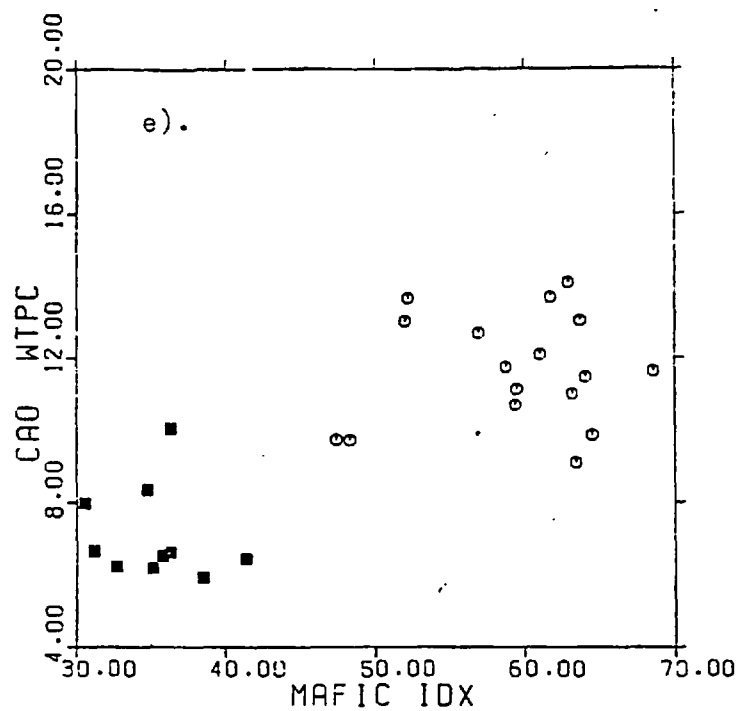


Fig. 4.5'cont'd.

SiO_2 ranges from approximately 40% to 52%, although this tends to be concentrated into a narrow range from 45-48%.

TiO_2 shows a marked increase with differentiation within the mafic gabbros, although remains approximately constant in the ultramafites. In silicate rocks FeO and TiO_2 are highly concentrated in magnetite and ilmenite, and typical tholeiitic series' show increases of both elements during fractional crystallisation (Greenland and Lovering, 1966; Miyashiro, 1974; Miyashiro and Shido, 1975). This is generally taken to indicate delayed crystallisation of magnetite and ilmenite due to a low ambient oxygen fugacity in the magma (Osborn 1962, 1969; Miyashiro and Shido, 1975). The ultramafites have minor amounts of modal magnetite, whereas the mafic gabbros have both ilmenite and magnetite in their mineral assemblages. The well-defined trend for TiO_2 is probably also a consequence of its relative 'immobility' during metamorphism (Pearce and Cann, 1973; Floyd and Winchester, 1975; Morrison, 1978).

Al_2O_3 concentrations are higher in the mafic gabbros than the ultramafites, although there is no trend linking the two rock types. CaO shows a similar distribution.

Na_2O increases with increasing Mafic Index in a broad band in the mafic gabbros, although remaining approximately constant in the ultramafites.

Phosphorus is another element considered to be immobile in basaltic systems (Pearce et al., 1975). It increases with increasing Mafic Index in the Cameas nam Buti samples, the slope increasing sharply at the more fractionated end. This may be a consequence of the inability of phosphorus to enter any likely original cumulus phase (olivine, pyroxene, plagioclase) and hence its tendency to increase in any residual liquid.

MnO and K_2O show scattered distributions with no recognisable

trend. MnO is strongly partitioned into garnet in the mafic gabbros (chapter five). Because garnet porphyroblasts may be very large (10 cms. in diameter), this causes problems for gaining a representative sample for geochemical analysis. This problem is highlighted when a trace element with a large partition coefficient for garnet such as Mn is being determined, and it may explain the scatter of points in the mafic gabbro portion of fig. 4.5,d. The random distribution of points in the K₂O diagram may be due to the mobility of K during metamorphism, as suggested in section 2.1.1.

Trace element concentrations are plotted in fig. 4.6, a-h.

Nickel decreases sharply with fractionation, from approximately 1500 ppm in the ultramafites and curves to 100-200 ppm in the mafic gabbros. This suggests the initial liquid was depleted rapidly in nickel by a silicate phase, as the ultramafites are impoverished in sulphide minerals. Nickel receives a high crystal field stabilisation energy in mineral octahedral sites as opposed to silicate liquid tetrahedral sites (Burns and Fyfe, 1964; Whittaker, 1967) and partitions strongly into early forming olivines in basaltic liquids (Wager and Mitchell, 1950; Walker, 1969; Gunn, 1971; Maclean and Shimazaki, 1976; Campbell, 1977). It would not be unreasonable to assume therefore that the nickel distribution was controlled by olivine. The modal, metamorphic olivines, also have high nickel contents (chapter five) and it is suggested that they may have crystallised on the sites of original igneous olivines.

Chromium shows a trend similar to nickel, although less curved. Concentrations may be as high as 3500 ppm in the ultramafites. Chromium also receives a high crystal field stabilisation energy in octahedral sites in silicate minerals (Burns and Fyfe, 1964) and partitions into early forming pyroxenes (Wager and Mitchell, 1950; Campbell, 1977), although it requires a coupled substitution

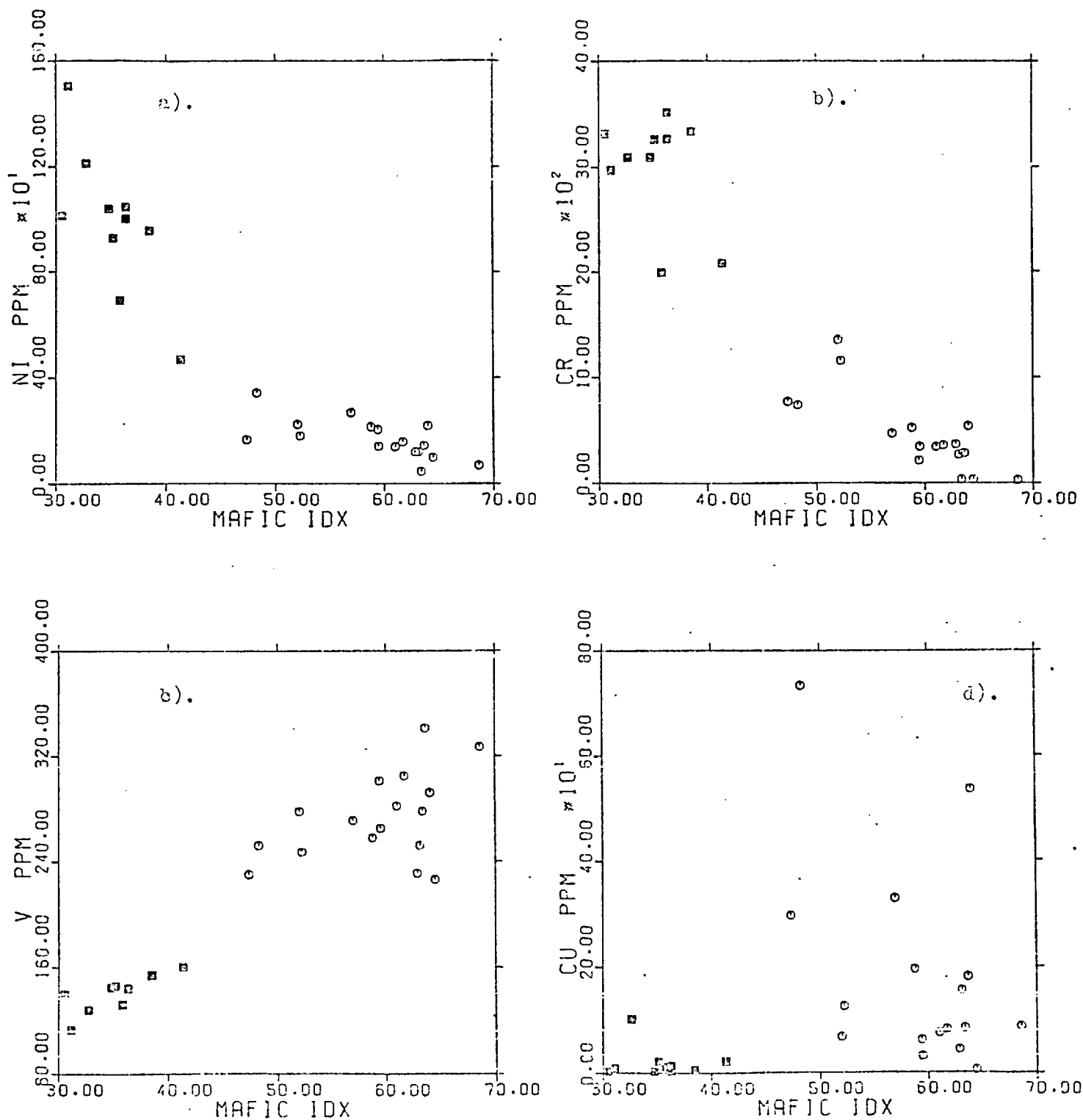


Fig. 4.6a-h. Trace element concentrations plotted against Mafic Index for the Cemas namuth samples. Legend as for the major element diagrams.

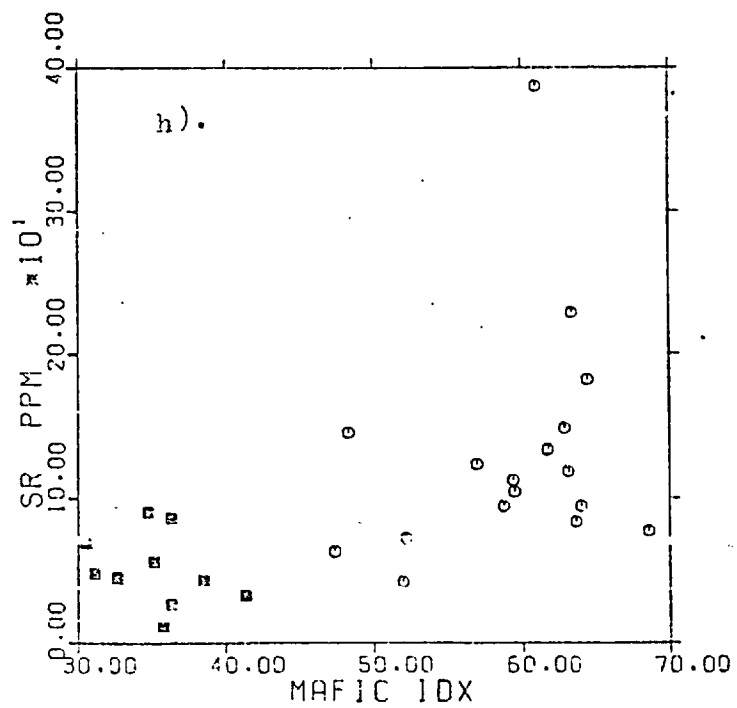
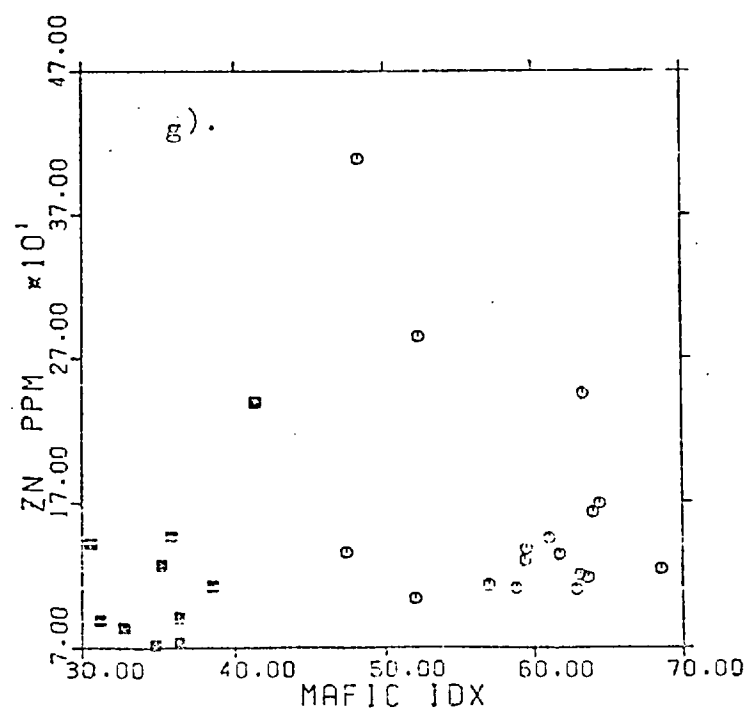
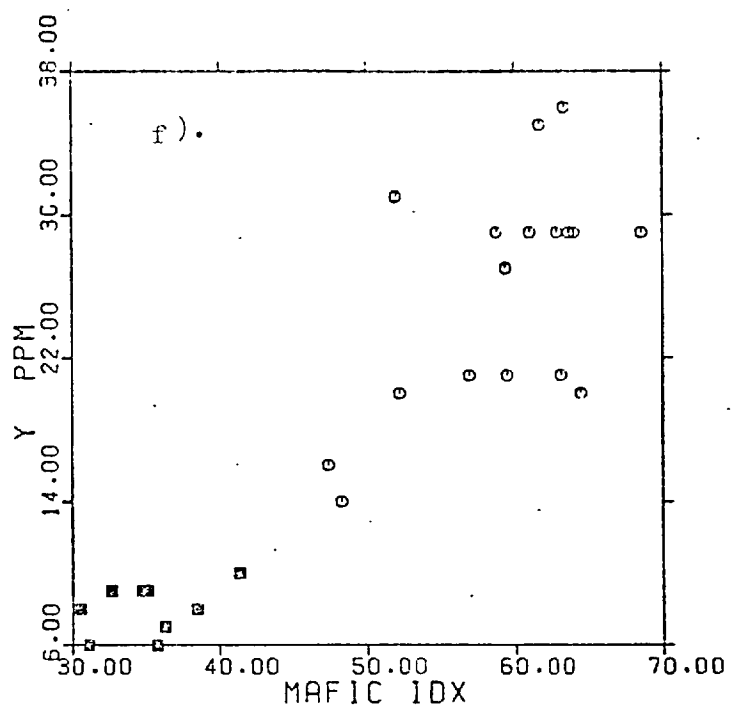
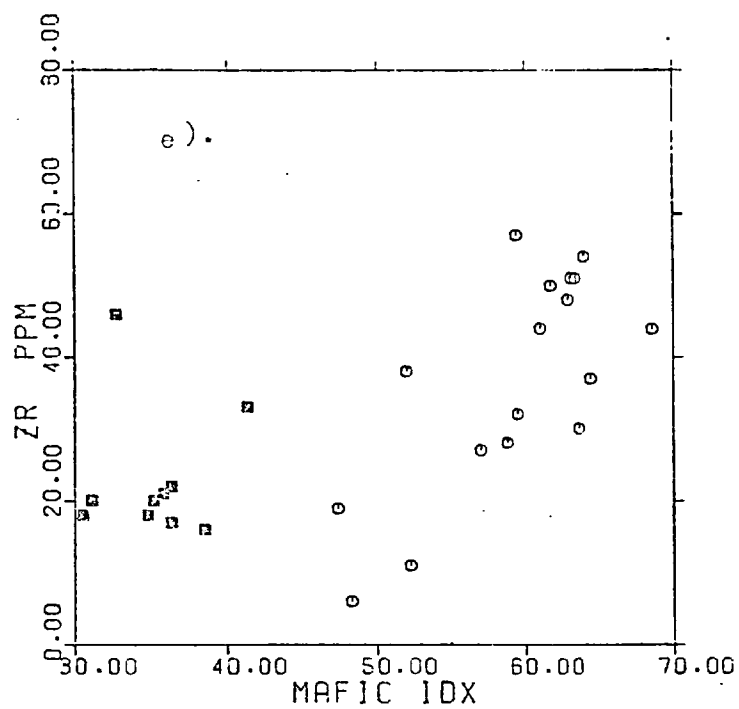


Fig. 4.6 cont'd.

to enter the pyroxene phase. Early fractionation of olivine was probably accompanied by pyroxene in the Camas nam Butth sheet.

Vanadium increases systematically with increasing mafic index from approximately 120 ppm in the ultramafites to 320 ppm in the mafic gabbros. There is a slight break in trend between the two rock types. Vanadium is usually concentrated in magnetite and ilmenite, and to a lesser extent, in pyroxenes (Wager and Mitchell, 1950; Greenland and Lovering, 1966; Duncan and Taylor, 1968; Leeman et al., 1978). Vanadium decreases monotonically with differentiation in calc-alkaline suites (Miyashiro and Shido, 1975). This again, depends on the stage of crystallisation of magnetite and ilmenite. The distribution of vanadium correlates quite closely with that of FeO and TiO_2 . The break between the ultramafites and mafic gabbros is probably due to the presence of modal ilmenite in the latter rock type.

Copper shows no clear trend with differentiation. Although copper may replace sodium and divalent iron in silicates and oxides (Greenland and Lovering, 1966), it will selectively partition into sulphide rather than silicate liquid, if possible (Maclean and Shimazaki, 1976). Copper is indeed partitioned into the dispersed sulphide grains in the mafic gabbros (chapter five) which would explain the lack of any trend in fig. 4.6, d.

Zirconium increases with differentiation in the mafic gabbros, but the ultramafites show a close grouping without any clear trend. Wager and Mitchell (1950) observed that Zr entered pyroxenes in the Skaergaard intrusion, whilst Walker (1969) noticed its concentration in the opaque minerals of the Palisades Sill. Despite this, Zr is widely recognised as an 'incompatible' element (Floyd and Winchester, 1975), its solid/liquid partition coefficient being much less than unity and shows a strong positive correlation with other incompatible

elements such as R and Y. These elements increase in the residual liquid with fractional crystallisation (Beccaluva et al., 1977). Jamieson and Clarke (1970) also noted a strong coherence in these elements from tholeiitic basalts.

Yttrium increases steadily from the ultramafites to the mafic gabbros, defined by one curve. As noted above, this probably indicates its residual character. Lambert and Holland (1974) have suggested its incorporation into calcium-bearing minerals because of the similarity in ionic radius between calcium and yttrium. They noted the general decrease in Ca/Y with differentiation in igneous rocks. The Ca/Y ratio (fig. 4.7, d) for the Camas nam Buth rocks shows such a decrease with increasing mafic index.

Zinc shows considerable scatter, although this may be due to its incorporation into the sulphides (chapter five), however Maclean and Shimazaki (1976) have determined a sulphide/silicate partition coefficient of 0.5 for this element, which suggests a preferential accumulation in the silicate liquid.

Strontium increases gradually with differentiation, although it generally remains below 200 ppm throughout. Sr has an ionic radius between those of potassium and calcium and may be expected to be concentrated in the feldspar fraction of silicate rocks (Taylor, 1966; Walker, 1969). The generally flat trend for this element may therefore indicate a suppression of plagioclase crystallisation.

Rubidium and thorium contents are low, generally 1-4 ppm each or below detection limits in the mafic gabbros. Although these may be reasonable abundances in low-K tholeiites (Jakes and White, 1972; Engel and Fisher, 1975), they may have been higher prior to granulite facies metamorphism. Barium is irregular in concentration, but generally below 100 ppm.

Fig. 4.7, a-d, illustrates a number of trace element ratios

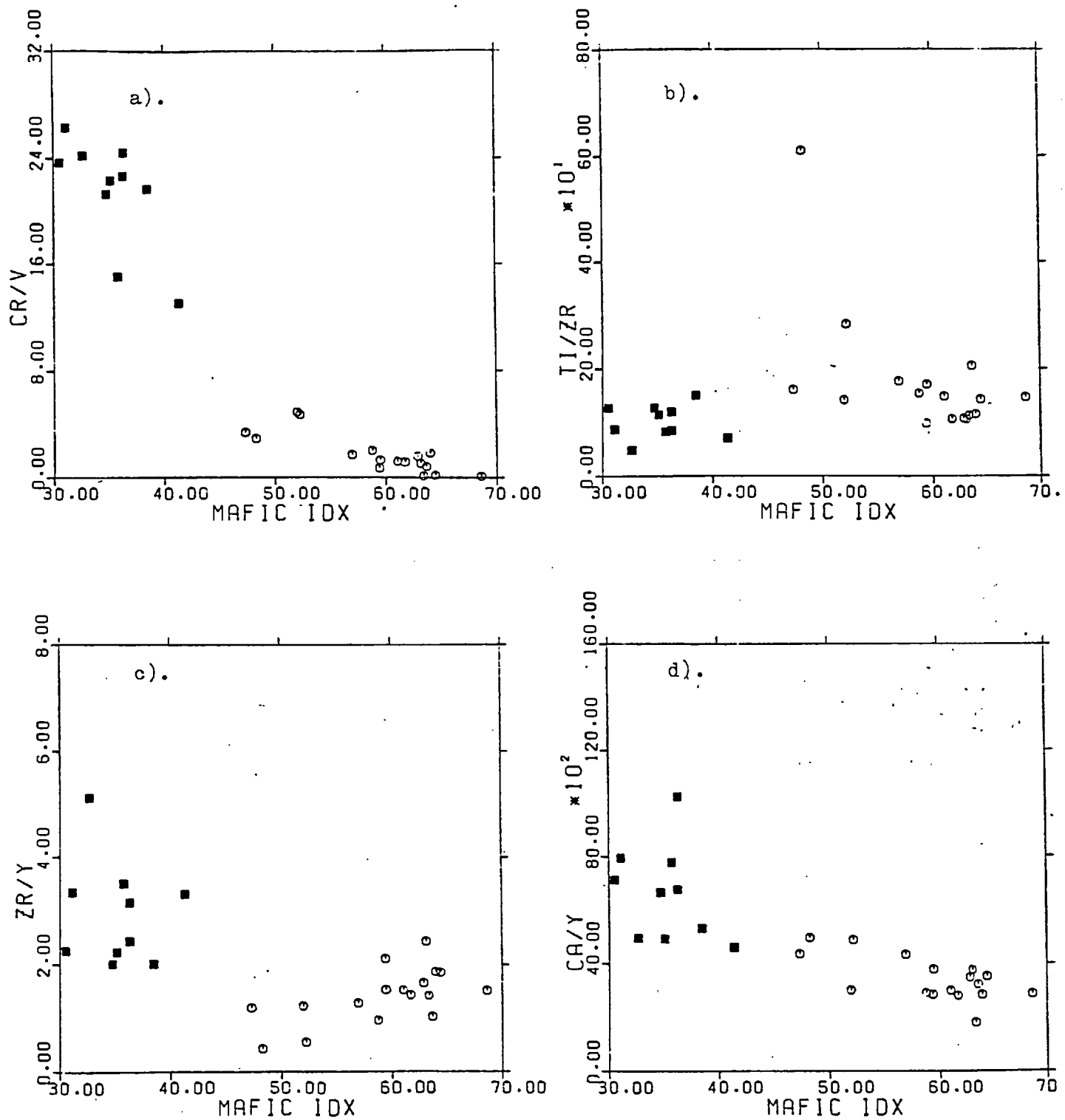


Fig. 4.7 a-d. Trace element ratios plotted against Mafic Index for the Camas nam Buth samples. Legend as for the major element diagrams.

sensitive to igneous fractionation processes (Taylor, 1966; Holland and Lambert, 1974; Nesbitt and Sun, 1976). Cr/V and Ca/Y decrease whilst Ti/Zr increases with differentiation. Zr/Y shows a bimodal grouping of the two rock-types with little connecting trend.

The differentiation of the layered mafic-ultramafic sheets may be investigated further using Pearce variation diagrams (Pearce, 1968a, 1968b, 1970).

Harker-type variation diagrams, in which simple oxides are plotted against each other, distort genetic trends between variables by changing their slopes. Pearce (1968a) showed that dividing two variables by a third parameter does not change the slope of the relationship between them if the third parameter is a constant. For example, if olivine is being fractionated from a basic magma, the relationship between absolute amounts of MgO and SiO₂ is indicated by a trend-line with the same slope as that which expresses the relationship between MgO/Al₂O₃ and SiO₂/Al₂O₃, if Al₂O₃ is not being removed in the fractionation process (the slope of this line will be controlled by the MgO/SiO₂ ratio of the precipitating olivine). Thus, although the weight percent of Al₂O₃ would increase in the magma as olivine was removed, the absolute amounts of Al₂O₃ would remain constant in terms of equivalent amounts of MgO and SiO₂. Simple weight percent comparisons do not take this residual concentration by fractionation of other components into account, hence the necessity to compare equivalent amounts rather than equal weights. Trends become more difficult to interpret if more than one phase is fractionating, or if precipitating phases change composition. In the latter case, trends become curved.

This method of demonstrating trends has been successfully applied to the investigation of the differentiation of the Palisades

Sill (Pearce, 1970), the Dunderberg Sill, Ontario (Pearce, 1968b), the Losberg Intrusion, Transvaal (Ferguson and Fripp, 1971) and also to metamorphosed igneous intrusions of the Lewisian of the Outer Hebrides (Witty, 1975; Soldin, 1978; Okeke, 1978).

TiO_2 has been chosen as a constant parameter for the Camas nam Ruth intrusion. Although variation diagrams suggest incorporation of this element into magnetite and ilmenite, it is an element known to be relatively immobile during metamorphism. As long as the major fractionating phases contain only minor amounts of the 'constant element', the effects on the slopes of the Pearce diagrams is minimal (Pearce, 1968b).

The $\text{FeO} + \text{MgO} : \text{SiO}_2$ diagram (fig. 4.8, a) produces trends indicative of the fractionating phase, independent of any solid solution within that phase. Two distinct, intersecting trends are present. The ultramafites are best resolved as products of the fractionation of orthopyroxene with minor amounts of a mineral with a higher $\text{Fe} + \text{Mg} : \text{Si}$ ratio, possibly olivine or spinel. The fractionating phase of the mafic gabbros appears to be dominantly clinopyroxene, again with minor amounts of a higher $\text{Fe} + \text{Mg} : \text{Si}$ mineral, possibly orthopyroxene.

The $\text{FeO} : \text{SiO}_2$ and $\text{MgO} : \text{SiO}_2$ diagrams (figs. 4.8b and 4.9a) show features similar to the above, but because of the precipitation of more than one ferro-magnesian phase it is only possible to give an estimate of the combined $\text{Fe}:\text{Si}$ and $\text{Mg}:\text{Si}$ ratios, rather than the ratios for each mineral.

The $\text{Al}_2\text{O}_3 : \text{SiO}_2$ diagram (fig. 4.9b) shows slopes for both rock-types greater than would be expected if low-alumina pyroxenes were the sole fractionating phases. Igneous pyroxenes from basic intrusions such as the Bushveld complex which crystallised at low pressures generally have $\text{Al}_2\text{O}_3 : \text{SiO}_2$ ratios lower than 0.05 (Deer,

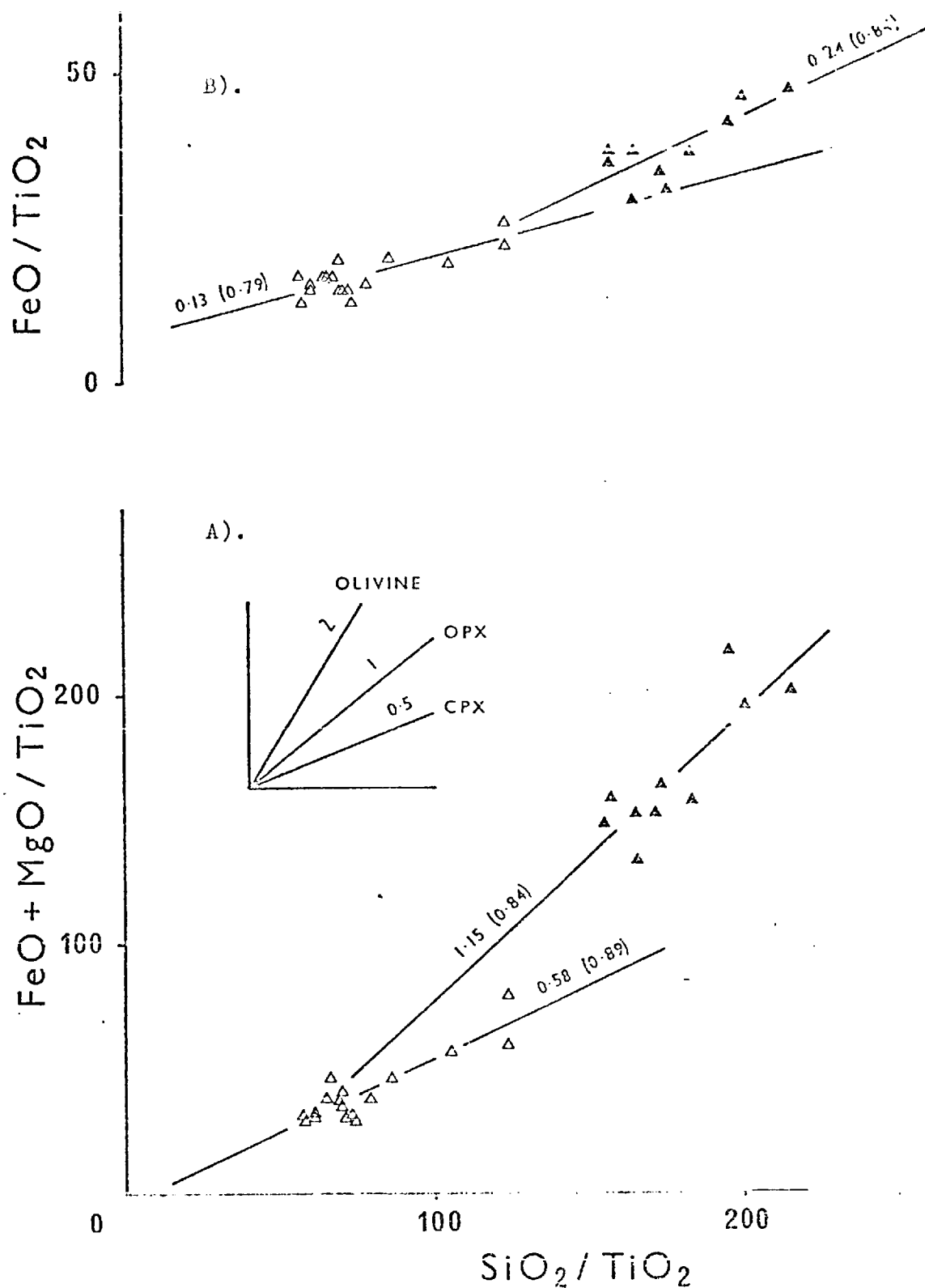


Fig. 4.8 Pearce variation diagrams for the Camas nam Ruth samples. Filled triangles are ultramafites, open triangles are mafic gabbros. Slopes and correlation coefficients (in parentheses) are indicated.

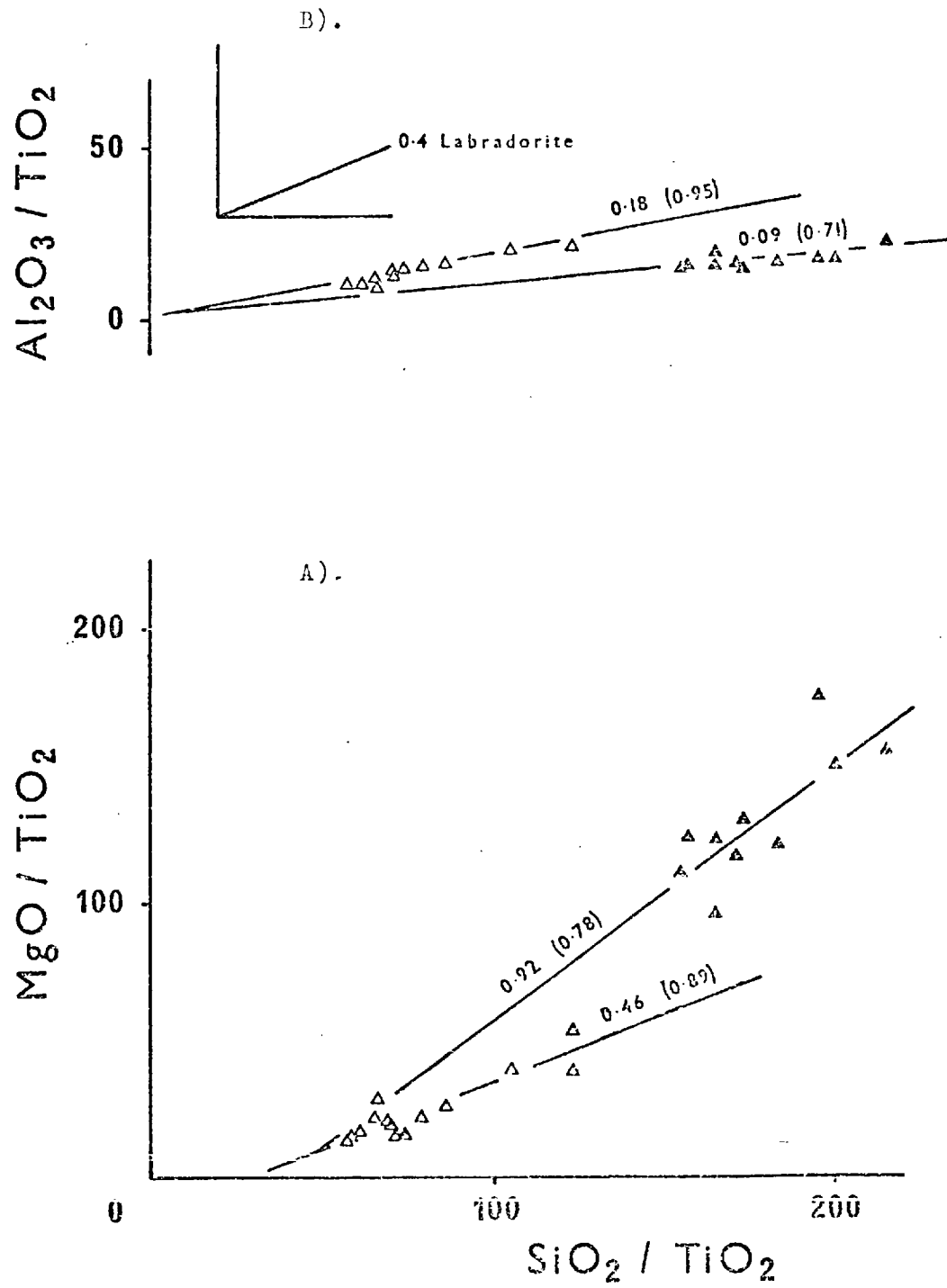


Fig. 4.9 Pearce variation diagrams of the Camas nam Ruth samples.

Legend as for fig. 4.8.

Howie and Zussman, 1972). A plagioclase of labradoritic composition would seem too aluminous to explain the trends. Interpretation of the $\text{CaO} : \text{SiO}_2$ diagram (fig. 4.10a) is also ambiguous. The ultramafite and mafic gabbro slopes are too low for either clinopyroxene (diopside) or labradorite fractionation. Thus both the Al_2O_3 and CaO diagrams cannot easily be interpreted in terms of likely phases in low-pressure crystallisation. This point will be returned to subsequently.

The $\text{NiO} : \text{SiO}_2$ and $\text{Cr}_2\text{O}_3 : \text{SiO}_2$ diagrams (figs. 4.10, b, c) help one to identify the minor ferromagnesian phase precipitating with orthopyroxene in the ultramafites. If spinel was the contributing phase, then the slopes in each case would be infinitely large. The slopes are in fact consistent with olivine controlling the NiO distribution and orthopyroxene controlling the Cr_2O_3 contents, from available analyses of olivines and pyroxenes from basic intrusions (Deer, Howie and Zussman, 1972). Similarly, both the NiO and Cr_2O_3 slopes in the mafic gabbros may be controlled by clinopyroxene.

The two intersecting trends present in all the diagrams may be explained by considering 4.11. As an example, the $\text{FeO} + \text{MgO} : \text{SiO}_2$ diagram is used. P_1 is a hypothetical composition of the initial magma from which the Camas nam Buth sheet crystallised. However, it is necessary to invoke two waves of crystallisation to achieve the two intersecting trends. From the initial magma P_1 , orthopyroxene (plus minor olivine) crystallised, producing essentially a series of cumulates and cumulate derived liquids. This is the ultramafite trend and the slope $P_1 - P_2$ is controlled by the $\text{FeO} + \text{MgO} : \text{SiO}_2$ ratio of the precipitating orthopyroxene. During this crystallisation, the liquid composition moves from P_1 to P_2 , as indicated by the arrows. At P_2 , a clinopyroxene (plus minor orthopyroxene) crystallisation commences. Again, this produces a series of cumulates and cumulate

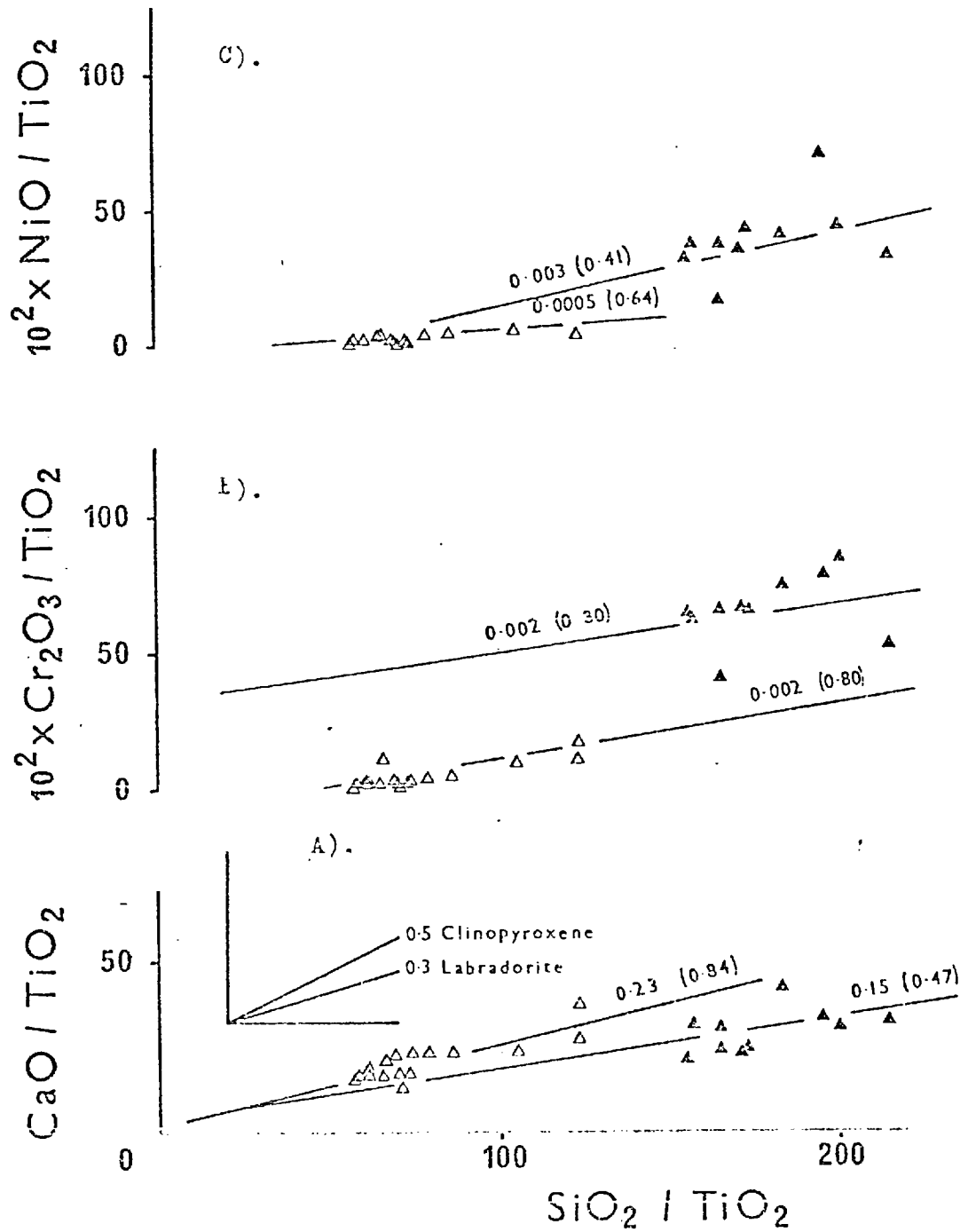


Fig. 4.10 Pearce variation diagrams for the Camas nam Ruth samples. Filled triangles are ultramafites, open triangles are mafic gabbros. Slopes and correlation coefficients (in parentheses) are indicated.

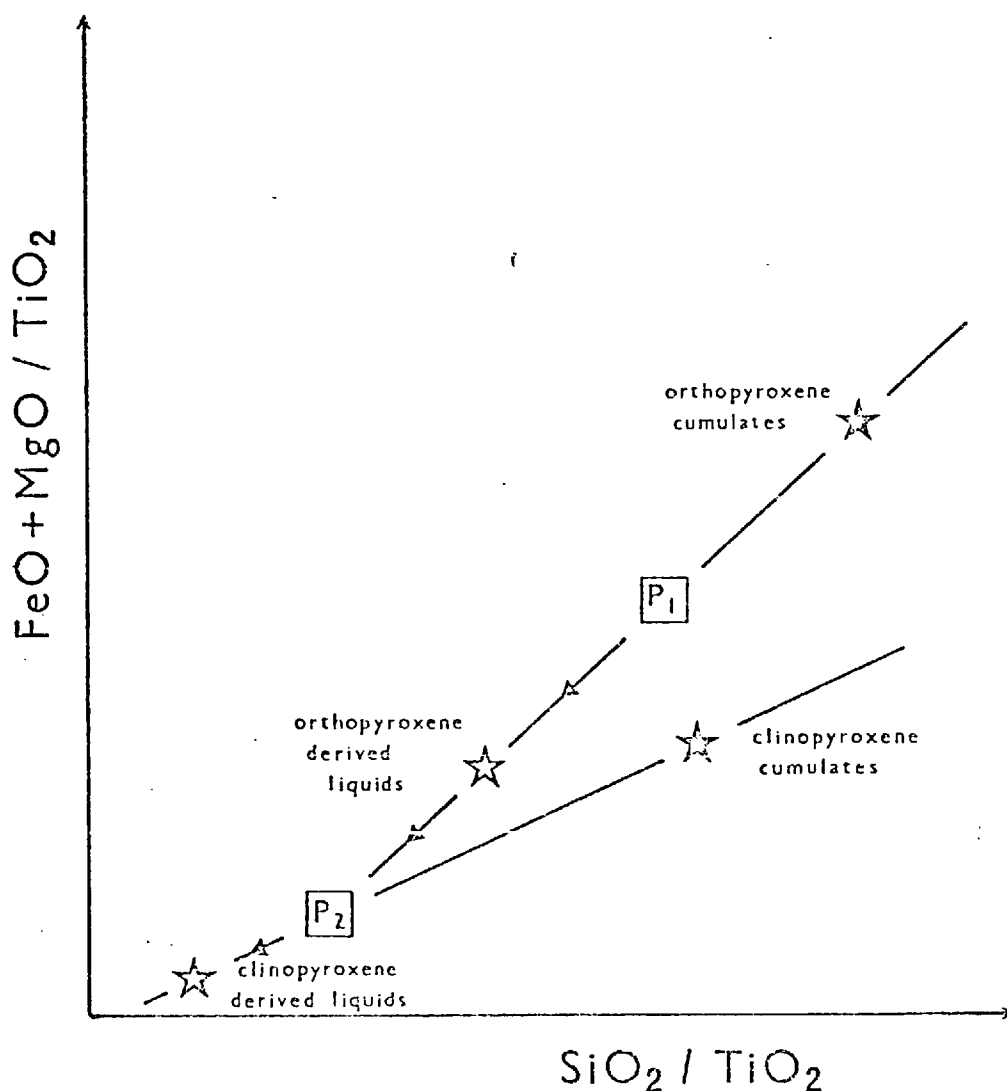


Fig. 4.11 Hypothetical interpretation of the $\text{FeO} + \text{MgO} : \text{SiO}_2$ Pearce variation diagram. (see text). P_1 is the initial liquid of the intrusion. The slope of the upper line is controlled by the composition of the precipitating orthopyroxene (the ultramafite trend). The slope of the lower line is controlled by the composition of the precipitating clinopyroxene (the mafic gabbro trend). The liquid line of descent is indicated by the arrows. P_2 is the composition of the liquid after the crystallization of the ultramafites, and is the liquid from which the mafic gabbros crystallised.

derived liquids. This is the mafic gabbro trend, the slope of the line being controlled by the $\text{FeO} + \text{MgO} : \text{SiO}_2$ ratio of the precipitating clinopyroxene. The liquid composition continues to evolve as indicated by the arrows, until crystallisation is complete.

The variation diagrams emphasise the minor importance of olivine and the possible absence of plagioclase as crystallising phases in the pre-metamorphic differentiation of the Camas nam Buth sheet. These are the dominant liquidus phases during fractional crystallisation of olivine tholeiite magma at atmospheric pressure (Tilley et al., 1963, 1964, 1965; Green and Ringwood, 1967b). Suppression of crystallisation of these phases occurs at high pressures (Green and Ringwood, 1967b; Thompson, 1974). Green and Ringwood (1967b) noticed the following at 9Kb during fractional crystallisation of olivine tholeiite:-

- 1). Initial crystallisation of olivine, followed by olivine plus aluminous orthopyroxene, then aluminous orthopyroxene and aluminous, sub-calcic clinopyroxene. Plagioclase crystallisation is suppressed to a late and minor stage.

- 2). Derived liquids show little change in SiO_2 content, although Al_2O_3 , CaO and Na_2O are steadily enriched in the residual liquid.

- 3). Fractionation trends show decreasing normative hypersthene and increasing normative plagioclase, but with no tendency for movement into the quartz nepheline normative fields of the basalt tetrahedron (Yoder and Tilley, 1962).

Green and Ringwood's (1968b) analysis of orthopyroxene and clinopyroxene crystallising at 9Kb from an olivine tholeiite are presented in Table 4.3, accompanied by their $\text{Al}_2\text{O}_3/\text{SiO}_2$ and CaO/SiO_2 ratios (Green and Ringwood did not analyse the 9Kb clinopyroxenes, but suggest they are similar to those co-existing with aluminous orthopyroxene in the olivine tholeiite at 13.5 Kb. This composition is presented in Table 4.3, instead). There is an extremely close comparison between these figures and those derived from the Pearce variation diagrams for the Camas nam Buth sheet. It is suggested therefore that the Camas nam Buth sheet differentiated at high pressures, probably in the range 9-14 Kb. The original high pressure igneous mineral assemblage subsequently recrystallised at sub-solidus temperatures ($\approx 1000^\circ\text{C}$) and higher pressures ($\approx 15\text{Kb}$) to give the current metamorphic assemblage, as described in chapters three and five.

The geochemistry and differentiation of the Camas nam Buth sheet is similar to that of a mafic-ultramafic granulite inclusion suite in a Carboniferous analcinite sill, NSW, Australia, described by Wilkinson (1975)

The essential features of the pre-metamorphic differentiation of the Camas nam Buth sheet may now be summarised:-

- 1). Multiple injection of a parental tholeiitic magma at a depth of 25-45 km (9-14 Kb) under a low oxygen fugacity and low $f\text{H}_2\text{O}$ (apparently the liquid had a low $f\text{H}_2\text{O}$ because the appearance of a sub-silicic amphibole as a liquidus phase would increase the saturation of derived liquids in a manner not shown by the analyses (Green and Ringwood, 1968; Holloway and Burnham, 1972).

- 2). Crystallisation of olivine and aluminous orthopyroxene, giving rise to the ultramafites in each cyclic unit. This depleted

	<u>CPX</u> ¹	<u>OPX</u> ²		
SiO ₂	49.3	54.5		
TiO ₂				
Al ₂ O ₃	11.3	4.9		
Cr ₂ O ₃				
Fe ₂ O ₃				
FeO	8.3	7.8		
MnO				
MgO	21.2	30.0		
NiO				
CaO	9.9	2.9	Ratios derived from Pearce variation diagrams	
Na ₂ O				
K ₂ O				
Total	<u>100.0</u>	<u>100.1</u>	<u>Mafic Gabbro</u>	<u>Ultramafite</u>
			<u>CPX</u>	<u>OPX</u>
Al ₂ O ₃ /SiO ₂	0.13	0.05	0.18	0.09
CaO/SiO ₂	0.22	0.06	0.23	0.15

CPX¹, OPX² = compositions of pyroxenes from Green & Ringwood, 1967b.

CPX¹ - T = 1310°C, P = 13.5 Kb (coexists with aluminous OPX).

OPX² - T = 1290°C, P = 9 Kb (coexists with olivine).

Table 4.3 Comparison of the compositions of pyroxenes (crystallising at 9 Kb pressure) from the experimental work of Green and Ringwood (1967b) with those estimated from the Pearce variation diagrams of the Camas nam Bush samples.

the magma in nickel, chromium and magnesium.

3). Crystallisation of aluminous, sub-calcic clinopyroxene and aluminous orthopyroxene to give the mafic gabbro trend. Delayed crystallisation of titanomagnetite led to increases in FeO, TiO_2 and V.

4). Plagioclase crystallisation was suppressed, enriching residual liquids in alumina.

5). Crystallisation of sulphides, enriched in Fe and Cu, but depleted in Ni (see below).

4.1.3 THE ORIGIN OF THE SULPHIDE LAYERS

It has been briefly mentioned that the position of the sulphide and oxide concentrations within the Camas nam Buth sheet may be used to define a primary 'way-up' in the intrusion and indicate junctions between new influxes of magma.

Sulphide minerals in these concentrations are pyrite, pyrrhotite and chalcopyrite, analyses (Table 22) revealing low nickel contents. This is unusual considering the high nickel concentration in the ultramafites and large sulphide/silicate liquid partition coefficient for Ni ($K_D = 150$; Maclean and Shimazaki, 1976). Therefore, it would be difficult to envisage the Camas nam Buth sulphides precipitating in a supercooled zone at the base of the magma chamber after each magma pulse, as is envisaged to have occurred in the Jimberlana Intrusion, Western Australia (Campbell, 1977). Quite clearly, the Camas nam Buth liquid was depleted in nickel by early precipitation of olivine before crystallisation of the sulphides. This enriches the sulphides in iron and copper (Maclean and Shimazaki, 1976; Bowles, 1978).

The sulphide precipitation is obviously related to the influx of new magma as evidenced by the occurrence of sulphides beneath the

ultramafite layers which form the base of the rhythmic units. Solubility of sulphur in a basaltic melt is dependent on temperature, f_{O_2} and f_{S_2} (Haughton et al., 1974). Increasing temperature and f_{S_2} would increase sulphur solubility whereas increasing f_{O_2} would decrease sulphur solubility. It is unlikely that the temperature of the incoming magma would be lower than that already in the magma chamber, and so changes in f_{S_2} and f_{O_2} must have caused sulphide precipitation. Buchanan (1976a, 1976b) has suggested that the oxide and sulphide concentrations in the Bushveld Intrusion of South Africa were caused by an increase in f_{O_2} accompanying new influxes of magma. This may well have been the case for the Camas nam Bush sulphides.

4.1.4 RARE-EARTH ELEMENTS

Seven ultramafites and eight mafic gabbros have been analysed for ten rare-earth elements and scandium (Table 12).

The rare-earth elements (REE) are generally considered to be immobile elements, even at the highest grades of metamorphism (Green et al., 1969, 1972). Modification of REE patterns may occur however, if movement of fluid through the rocks has taken place (Frey and Haskin, 1964). This is not thought to be significant for the Camas nam Bush samples, although sample 130, recrystallised during the high P_{H_2O} conditions of the M_3 metamorphism shows large discrepancies compared with other ultramafites unaffected by this metamorphism (figs. 4.12 and 4.13). There is a general decrease in total REE content coupled with a modification of the pattern. With this specific exception, the regularity of the observed patterns are thought to represent original igneous features.

The REE patterns normalised to chondrite (Haskin et al., 1966) for the ultramafites are presented in fig. 4.13. These are lightly

Fig. 4.12 REE pattern of a retrogressed ultramafite, sample 130.

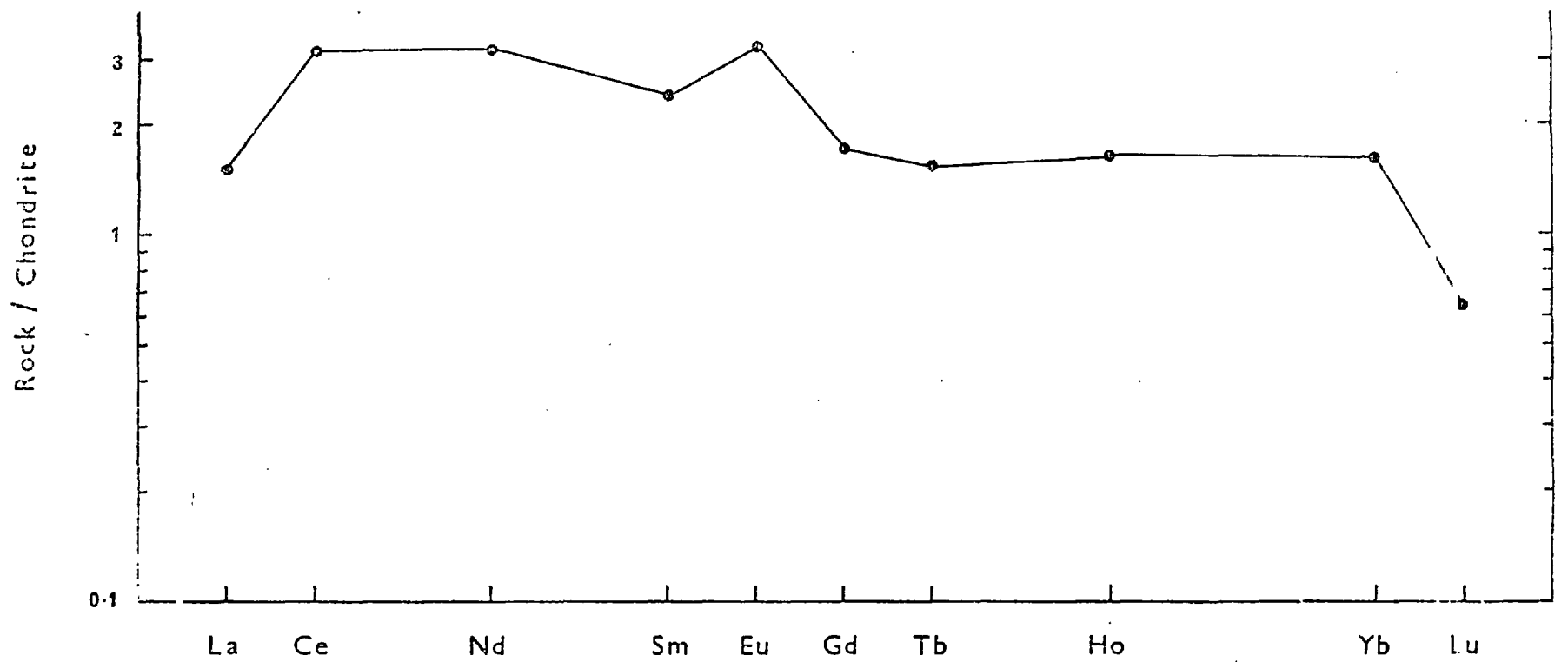
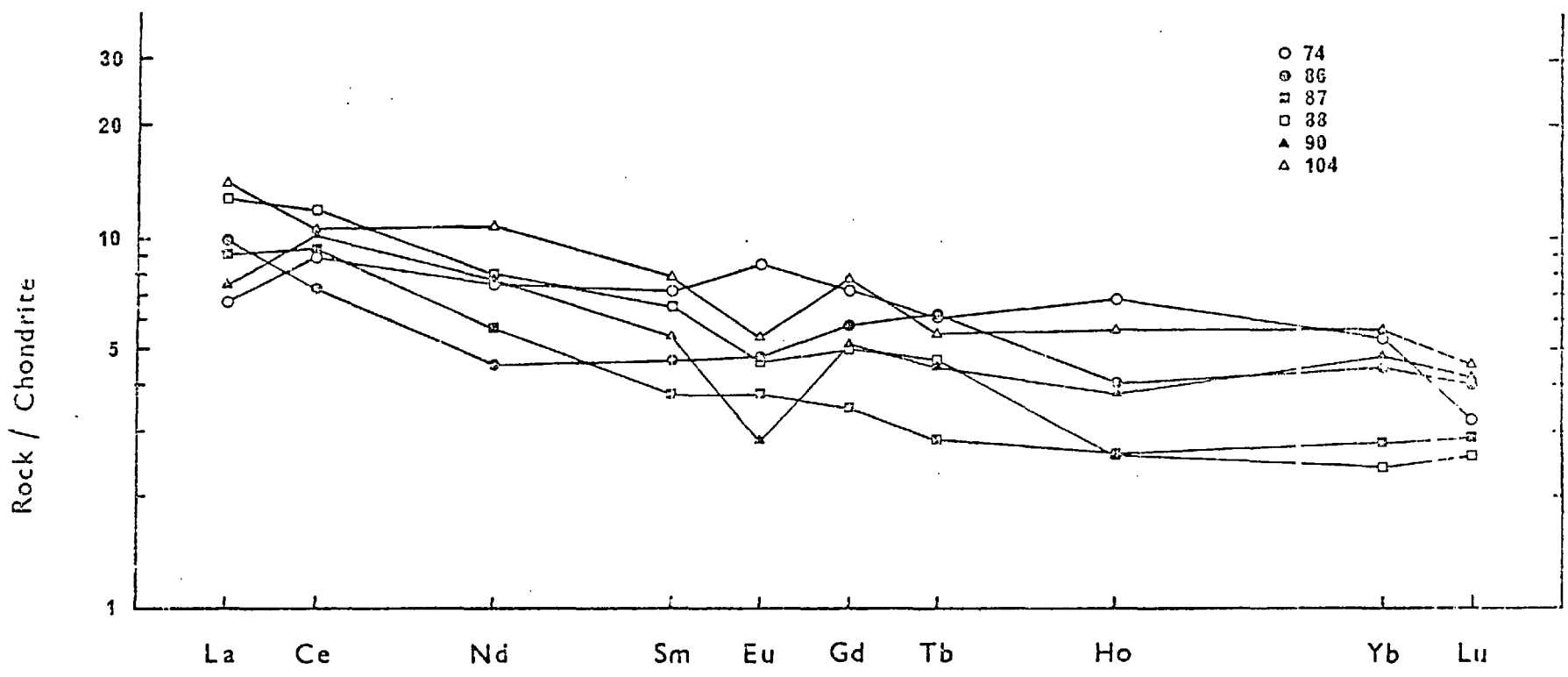


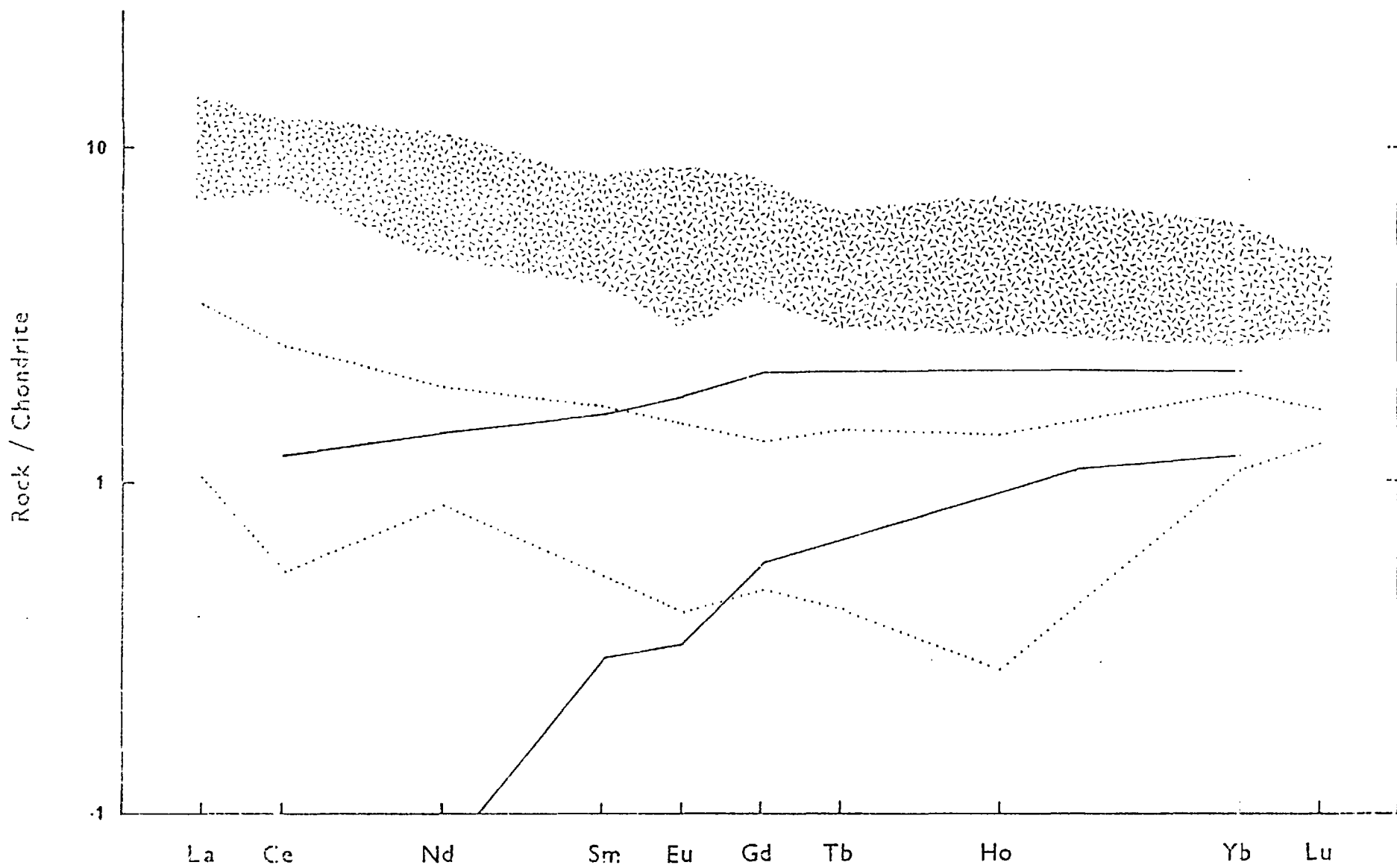
FIG. 4.13 REE patterns of the Cameroonian Yb₂ ultramylonites.



fractionated relative to chondrite with Ce_N/Yb_N ratios varying from 1.6 to 5.0. Total abundances range from 2 to 15 times that of chondrite. Four of the samples show slight negative europium anomalies, the other two having slight positive anomalies. The Scourie ultramafites are enriched in REE by a factor of between 2 and 5 relative to 'Alpine-type' ultramafics (Loubet et al., 1975) and typical ultramafics from stratiform intrusives (Frey et al., 1974). This is illustrated in fig. 4.14. This may be a consequence of the similarity Scourie ultramafites show to basaltic rather than ultrabasic geochemistry. This is reflected in the close similarity of their REE distribution to picritic gabbros from the Skaergaard intrusion (Haskin and Haskin, 1968).

The REE patterns for the mafic gabbros are plotted in figs. 4.15 and 4.16. Five of these (78,79,80,81,105) show flat, unfractionated patterns with Ce_N/Yb_N between 0.6 and 1.2 and three show a slightly more light REE enriched pattern (91,106,107) with Ce_N/Yb_N between 1.7 and 3.7. La_N/Sm_N ratios (a measurement of light REE enrichment or depletion) range from 0.93 to 1.95. These are quite distinct from the light REE depleted patterns of modern mid-ocean ridge basalts, having a similar major and trace element chemistry to the mafic gabbros, which have La_N/Sm_N ratios in the range 0.35 to 1.7 (Kay et al., 1970; Schilling, 1971, 1975; Frey et al., 1974; Shih, 1972). There are no significant europium anomalies. The mafic gabbro REE patterns are compared with the REE patterns of island arc tholeiites, and calc-alkaline rocks of island arcs in fig. 4.17 (after Jakes and Gill, 1970). The mafic gabbros lack the extreme fractionated patterns exhibited by the calc-alkaline rocks, and, as previously mentioned, do not have the light REE depleted patterns characteristic of the mid-ocean ridge basalts. In this respect, the mafic gabbros show greatest similarity

FIG. 4.14 The benthic ultramafites (stippled) and geochemistry as compared with that of 'Alpine-type' peridotites (area between solid lines) and ultramafites from stratiform intrusions (area between dotted lines).



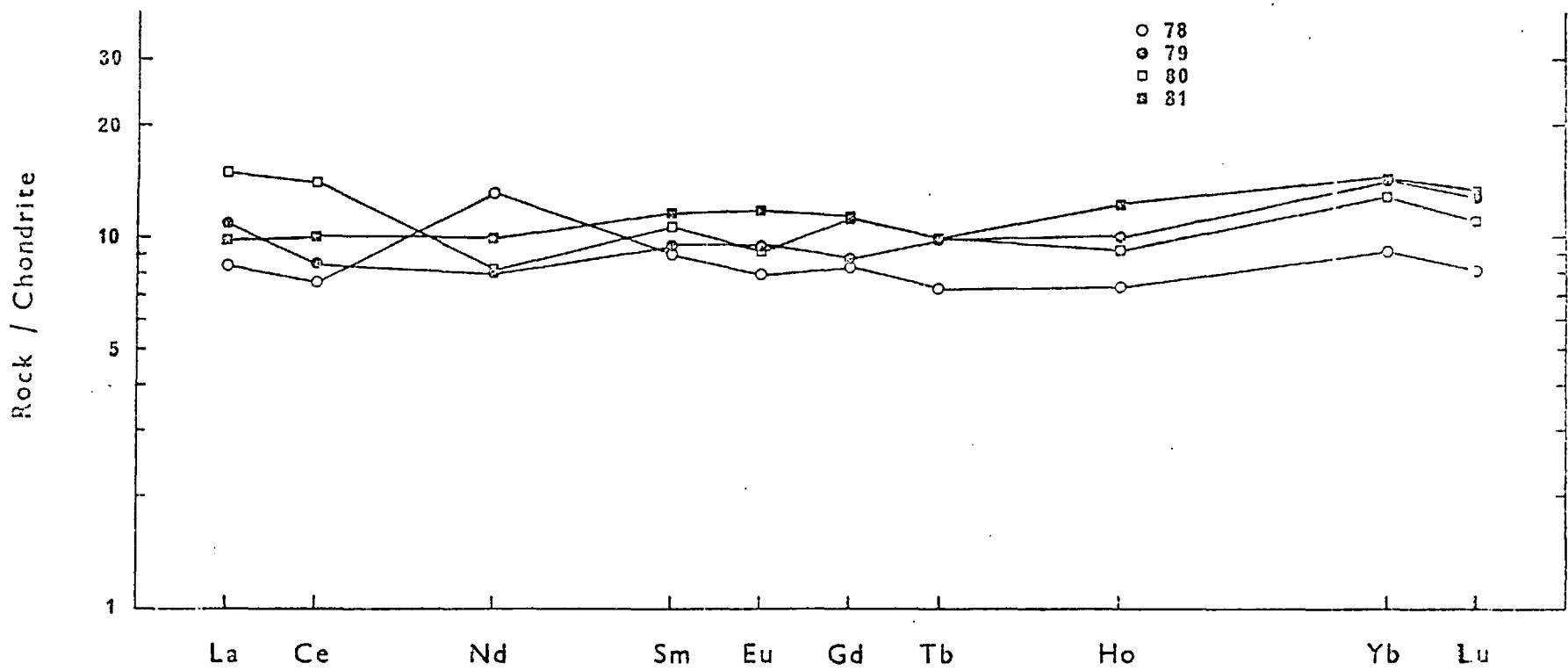


Fig. 4.15 REE patterns of four mafic gabbros from the Camas nan bath intrusion.

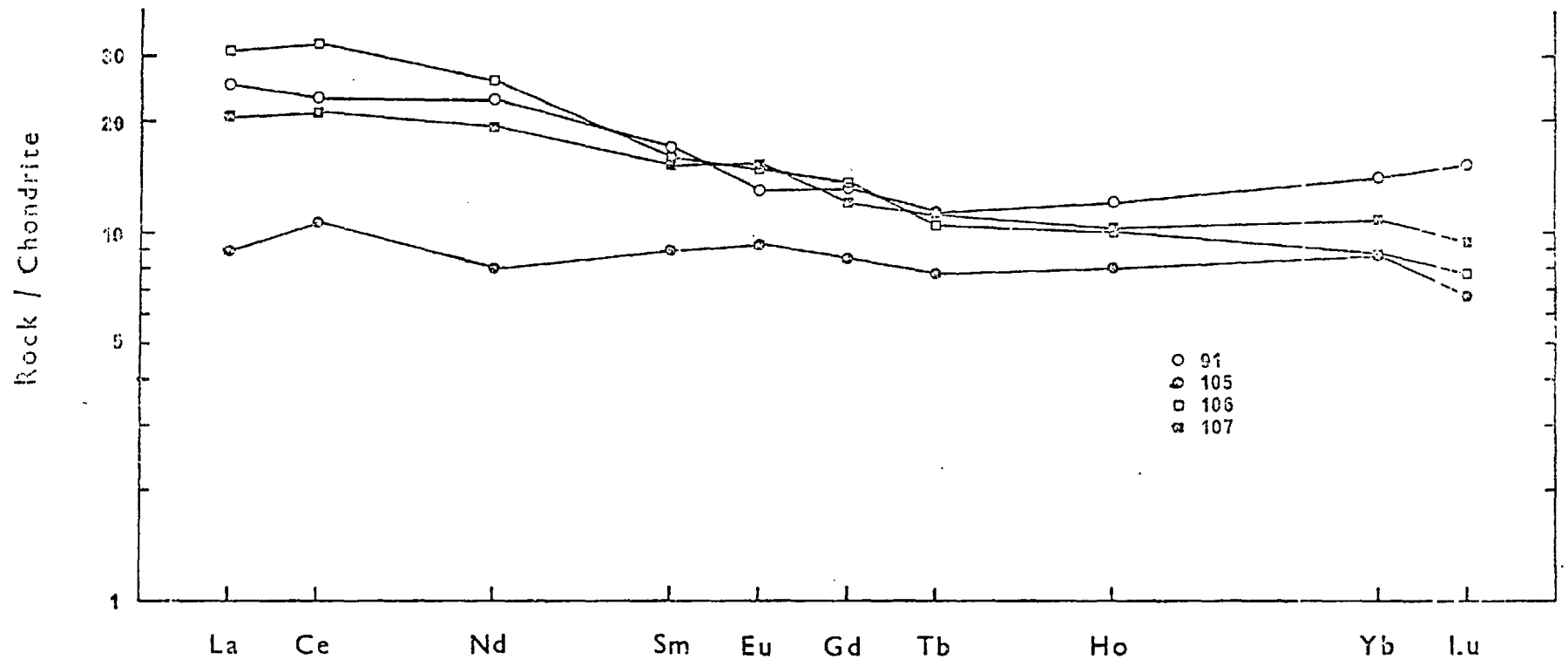


Fig. 4.16 REE patterns of four mafic gabbros from the Cames nan bath intrusion.

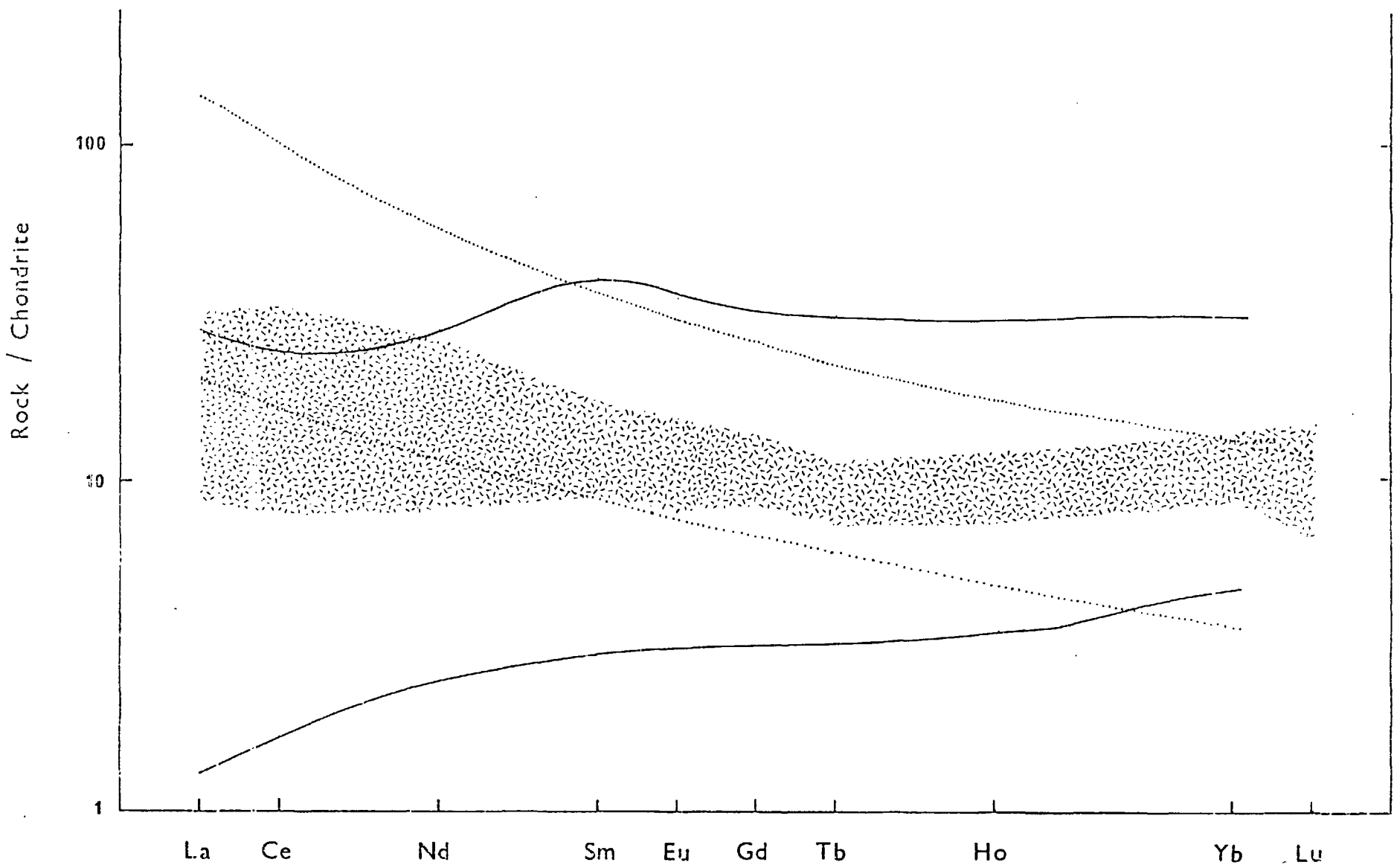


FIG. 4.1/ The Scourie mafic gabbros (stippled) REE geochemistry as compared with that of calc-alkaline volcanics from island arcs (area between dotted lines) and that of island arc tholeiites (area between solid lines).

to the island arc tholeiites.

Europium anomalies are generally thought to result from the accumulation or separation of feldspar from a melt under a low oxygen fugacity (Drake, 1975). Certainly, feldspar is the only common rock forming mineral with a prominent positive europium anomaly in the mineral/melt distribution coefficients for the REE (Hanson, 1978). As the mafic gabbros show no positive europium anomaly, the parent magma presumably had a slight negative europium anomaly.

Scandium values increase from approximately 20 ppm in the ultramafites to 60 ppm in the mafic gabbros. Norman and Haskin (1968) have noted the geochemical similarity between Sc and Fe^{2+} . The increase in Sc concentration with differentiation may be due to the incorporation of Sc into magnetite, and thus parallels the behaviour of FeO , TiO_2 and V.

4.1.5 THE TECTONIC ENVIRONMENT AND ORIGIN OF THE LAYERED MAFIC-ULTRAMAFIC SHEETS.

As was pointed out in chapter two, numerous theories exist for the mechanism of generation of Archaean crust and the tectonic environment of the 'high-grade gneiss' regions. An examination of the trace element geochemistry of the Camas nam Buth samples may help to shed some light on these theories.

Table 4.4 presents average analyses of basalts from various tectonic environments. In terms of major elements, the mafic gabbros show greatest similarity with the mid-ocean ridge basalts (MORB) and Island arc tholeiites (IAT), particularly in terms of TiO_2 and K_2O . Ba and Sr values in the mafic gabbros are intermediate between the IAT and MORB, although K/Ba and Ba/Sr ratios are akin to the IAT values. Zr is lower than either the IAT or MORB. Ca and Ni are both

	<u>1</u>	<u>2</u>	<u>3</u>	<u>4</u>	<u>5</u>	<u>6</u>
SiO ₂	46.73	47.31	51.57	50.59	49.20	51.90
TiO ₂	0.35	0.86	0.80	1.05	1.39	1.01
Al ₂ O ₃	7.37	14.13	15.91	16.29	15.80	14.10
Fe ₂ O ₃	3.38	2.42	2.74	3.66	2.20	2.30
FeO	8.91	10.82	7.04	5.08	7.20	3.90
MnO	0.21	0.25	0.17	0.17	0.16	0.18
MgO	22.64	9.16	6.73	8.96	8.50	7.10
CaO	7.09	11.66	11.74	9.50	11.10	10.30
Na ₂ O	0.67	1.87	2.41	2.89	2.70	2.30
K ₂ O	0.34	0.27	0.44	1.07	0.26	0.14
P ₂ O ₅	0.03	0.09	0.11	0.21	-	-
H ₂ O	0.10	0.12	0.45	0.81	-	0.10
L.O.I.	1.89	1.40		-	-	1.10
Total	99.71	100.36	100.11	100.28	98.51	99.43
V	141	273	270	255		
Cr	2991	461	50	40	297	407
Ni	986	167	30	25	97	150
Cu	19	182			87	41
Zn	118	162				
Zr	23	37	70	100	100	53
Rb	7	1	5	10		3
Sr	51	130	200	330	123	97
Th	0	0	.5	1.1		
Y	8	26		20	43	17
Ba	53	45	75	115	12	
K/Rb	403	2241	730	888		387
Ba/Rb	3	45	15	11		
Rb/Sr	.137	.0077	.025	.030		.031
K/Ba	123	50	49	77	180	
K/Sr	55	17	18	27	18	12
Ba/Sr	1.0	.3	.4	.3	.1	

Table 4.4 Comparison of the Camas nam Butte rocks with other basaltic rocks
1=Average Ultramafite; 2=Average Mafic Gabbro; 3=Average
Island Arc Tholeiite (Jakes & White, 1972); 4=Average Calc-
alkaline Basalt (Jakes & White, 1972); 5=Average Mid-Atlantic
Ridge Tholeiite (Nelson & Thompson, 1971); 6=Average Archean
Tholeiite, Coolgardie (Hallberg, 1970).

high in the mafic gabbros and show greatest similarity with the Archaean tholeiite. High Cr and Ni values are typical features of Archaean basaltic rocks and this is widely attributed to a greater degree of partial melting in the mantle to generate these magmas (Nesbitt and Sun, 1976).

REE distributions are similar to the patterns of island arc tholeiites (previous section) rather than the fractionated, light REE (La-Sm) enriched patterns of calc-alkaline basalts (Jakes and Gill, 1970), or the light REE depleted patterns of MORB.

Thus, taking the major, trace and REE data together, although there are some similarities with MORB, the nearest analogue to the mafic gabbros are the IAT's.

The time-period spanned by the Scourian (3000-2500 m.y. ago) was a particularly active one for the generation of crustal material (Moorbath, 1978). In this connexion, it is pertinent to consider a model for continental growth proposed by Jakes (1973):-

- 1). Generation of oceanic crust at ocean ridges by partial melting of upper mantle materials and transported by sea-floor spreading to subduction zones.

- 2). Dehydration of this oceanic crust leads to partial melting in the overlying mantle producing an island arc base of tholeiitic rocks. This is followed by melting of the subducted oceanic lithosphere itself producing a capping of calc-alkaline andesitic and dacitic lavas.

- 3). The crust is then vertically and laterally zoned and zoning may be subsequently accentuated by Andean volcanism and plutonic activity. The lower parts of the island arc may be partially melted to produce granulite facies mineral assemblages and enrich the upper crust in lithophilic elements.

This model has also been suggested by Ringwood (1969, 1974)

and Taylor (1968):

With this, and the geochemical evidence presented in this chapter, in mind, it may be possible to suggest origins for the layered mafic-ultramafic sheets and associated Scourie gneisses:-

1). Partial melting of mantle pyrolite over a primitive subducted oceanic lithosphere to produce a tholeiitic melt, LIL-depleted with an unfractionated REE pattern.

2). Crystallisation and differentiation of this magma at depth (30-35 Kms) in the initial stages of island arc development. The layered mafic-ultramafic sheets are thought to represent fragments of these intrusions.

3). Partial melting of the subducted lithosphere at greater depth produces later, calc-alkaline magmas (tonalites, granodiorites) typically LIL and light REE enriched, which would have invaded and disrupted the early tholeiitic base. These are represented by the acid gneisses and granulites of the Scourian, which, although Rb, Th and U depleted (see 4 below) have a granodioritic major-element chemistry, are Ba Sr rich, and if heavy rare-earth behaviour parallels that of Y (Frey et al., 1968), REE patterns must be fractionated relative to chondrite as suggested by XRF determinations of La, Ce and Y (Tarney, 1976).

4). Finally, granulite facies metamorphism (and deformation) in the lower part of the island arc pile would deplete the base in elements such as Rb, Th and V and would assist in the vertical stratification and 'ripening' (Jakes, 1973) of the early Scourian crust.

This hypothesis is consistent with the isotopic evidence of a short period of time between mantle derivation and granulite facies metamorphism of the Scourian (Mooreath et al., 1969), which is a prime requirement of any proposed model, as outlined in

chapter two.

An island arc environment has been invoked for the origin of the Lewisian gneisses of South Harris by Carson and Livingstone (1973) and Witty (1975). The relationships between the layered intrusions of the mainland and Hebridean Lewisian will be examined in chapter six.

4.2 THE GRANITE SHEETS

Seven granite sheet samples have been analysed for major and trace elements, five of these for the REE (Tables 10 and 12). Four of the samples come from the Camas nam Buth sheet (97,98,99,100), two from the Cleit Mhor net-veined complex (144, 145) and one from Geodh' Eanruig (148). Iron is expressed as FeO in the analyses, which affects the computed norm. The content is low enough, however, for the error to be relatively insignificant. Repeated wet-chemical analyses produced FeO contents higher than total iron contents from XRF analyses. This may be caused by the oxidation of elements other than Fe²⁺ in the metavanadate method used.

The rocks are granodioritic to adamellite, containing approximately thirty percent modal and normative quartz. Normative feldspar compositions are variable (fig. 4.18) even for samples from the same sheet (e.g. samples 97 and 98). This is due to variations in potash content. The anorthite content of the normative feldspar is also considerable. Silica content is high, ranging from 69 to 76 wt. percent, although generally concentrated in the range from 74 to 76 wt. percent. The rocks are slightly peraluminous with normative corundum ranging from 0 to 1.4 wt. percent. Both FeO and MgO are low, due to the scarcity of ferromagnesian minerals. The granites (s.l.) plot near the alkali apex of the AFM triangle (fig. 4.19). There is a considerable gap between the granitic and layered mafic-ultramafic rocks (c.f. fig. 4.1). It is clearly inconceivable that the granites

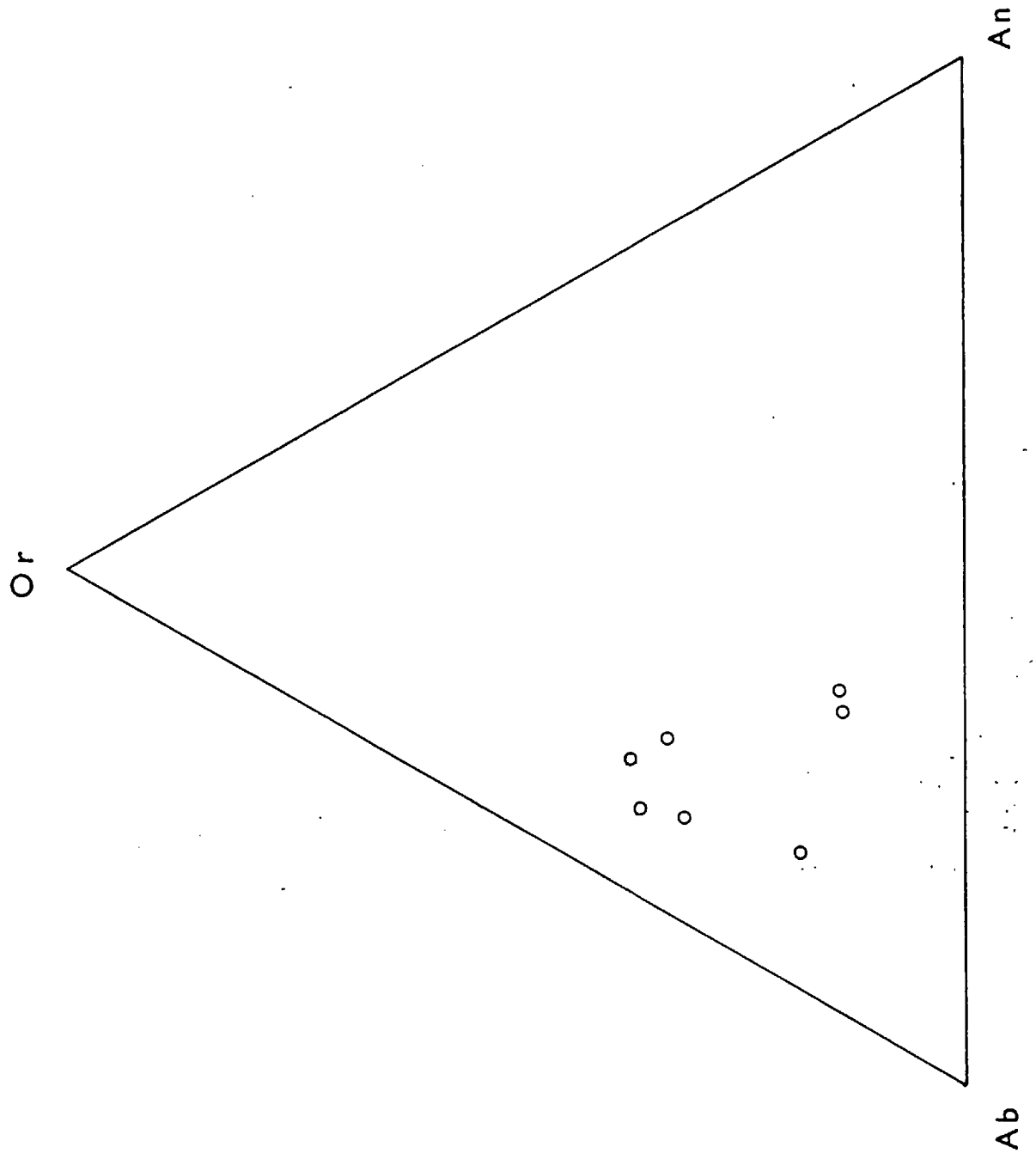


Fig. 4.18 Normative feldspar compositions of the granite sheets.

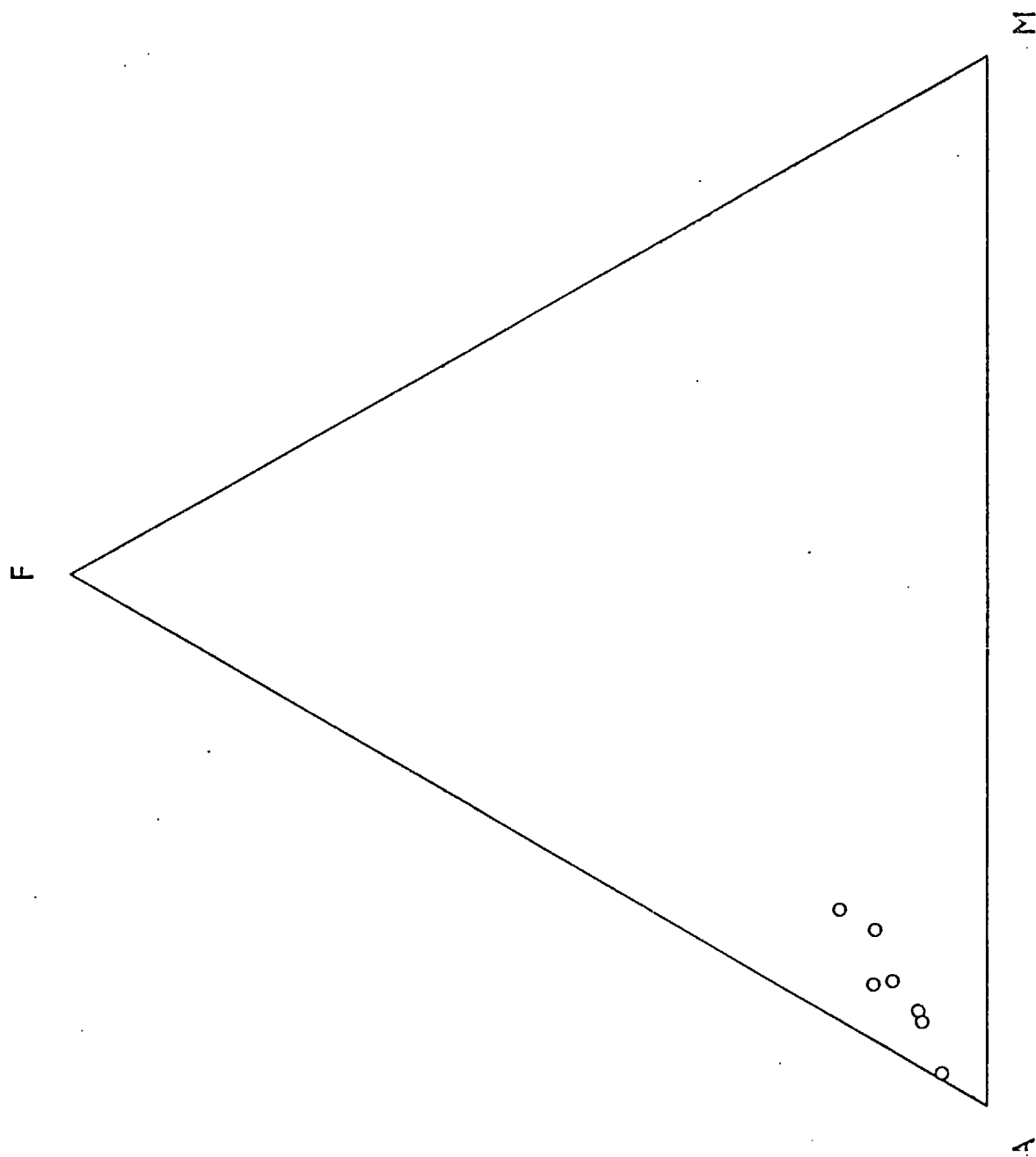


Fig. 4.19 AFM ($\text{Na}_2\text{O} + \text{K}_2\text{O} - \text{FeO(T)} - \text{MgO}$) diagram of the granite sheets.

represent some late residual liquid from the differentiation of the layered sheets. The granites fall relatively close to the quartz-albite cotectic in the residual system, quartz-albite-orthoclase (fig. 4.20), assuming a P_{H_2O} of 2Kb and an Ab/An ratio of 3.8 (Winkler, 1967).

The chemical composition of the granite sheets is compared with an 'average' granite, G-2, in Table 4.5. Barium in the granite sheets is half that of G-2, whilst strontium is approximately the same. Rubidium, however is anomalously low (24 ppm) and K/Rb ratios are consequently high (960). The granite sheets also have a high Ba/Rb ratio (38), whereas Rb/Sr is low (0.06). Zr and Th concentrations are low (20 and 0, respectively) bearing in mind the acidity of the rock types and Yttrium is also below the detection of the XRF (2 ppm). As might be expected, Cr, Ni, V, Cu and Zn all average below 20 ppm, reflecting the low ferromagnesian content of the granite sheets.

The REE patterns for five of the granitic rocks (97, 100, 144, 145, 148) are presented in fig. 4.21. All the patterns are strongly fractionated with Ce_N/Yb_N ratios of between 23 and 238. La concentrations may be 100 times that of chondrites, whereas Lu may be a tenth of the chondrite value. Although all the samples cluster reasonably closely for the light REE, divergence occurs at the heavy REE end (c.f. samples 144 and 148). All samples have a marked positive europium anomaly, suggesting a low oxygen fugacity during their formation (Drake, 1975). These depleted heavy-REE and fractionated patterns suggest that garnet was a residual mineral during the partial melting process to derive the granites, as garnet is the only common mineral to retain the heavy REE (Hanson, 1978). Yttrium's geochemical behaviour is thought to parallel that of the heavy REE, and is also partitioned strongly into garnet

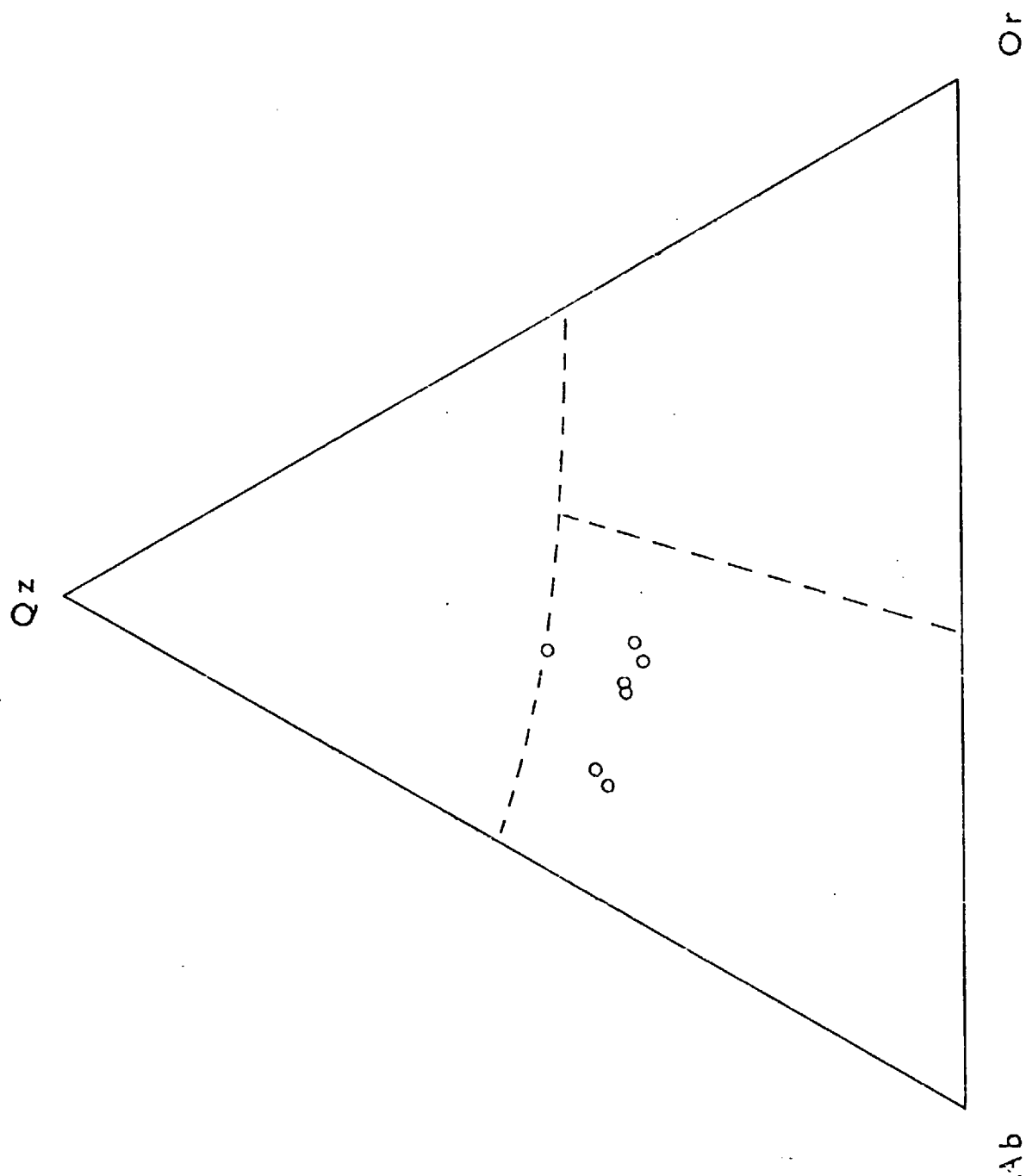
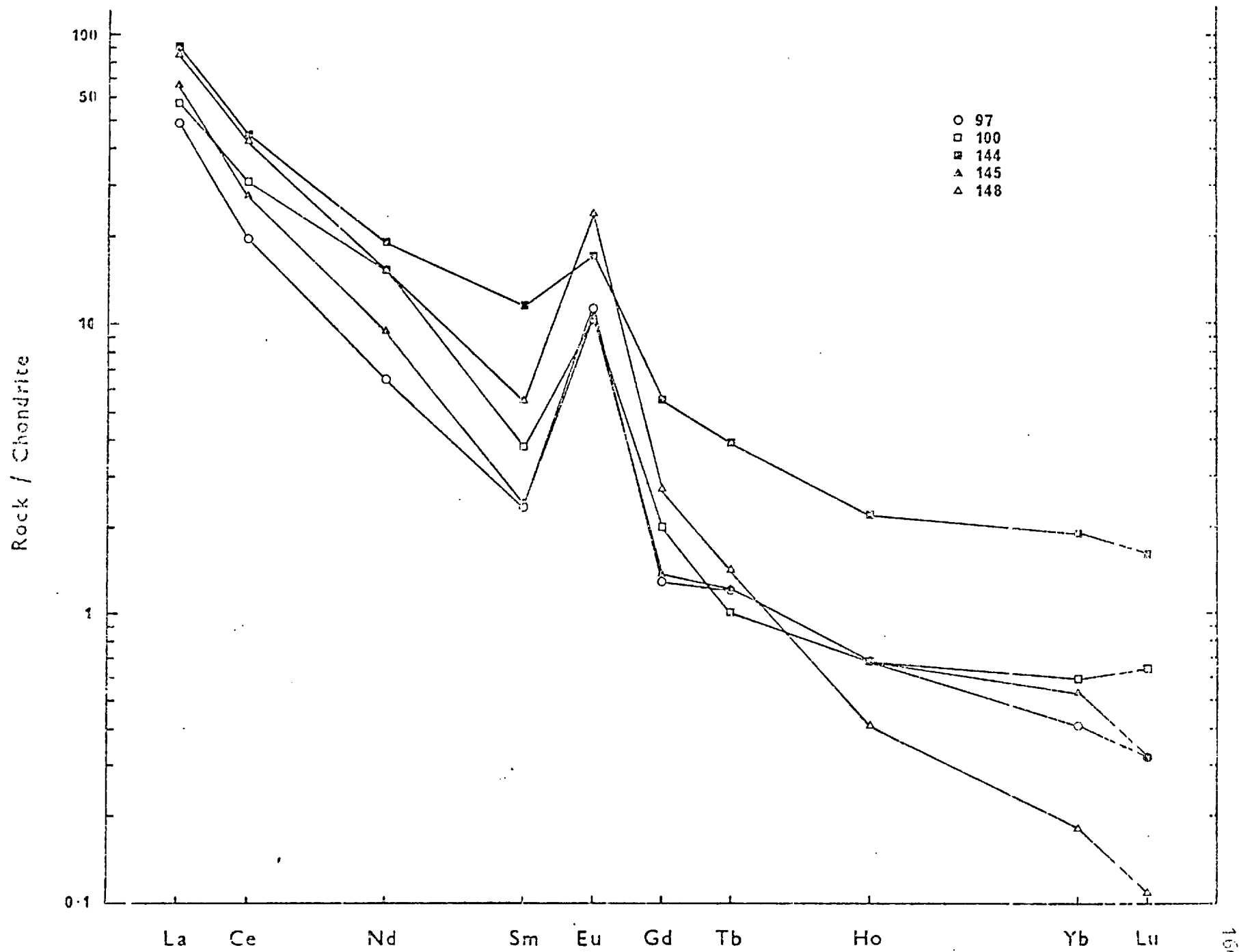


Fig. 4.20 The granite sheets plotted in the quartz - albite - orthoclase residual system. The cotectic lines are for $P_{H_2O} = 2 \text{ Kb}$ and $Ab/An = 3.6$ (Winkler, 1967).

	<u>1</u>	<u>2</u>	<u>3</u>
SiO ₂	63.56	74.05	69.69
TiO ₂	0.46	0.08	0.50
Al ₂ O ₃	15.83	14.43	15.46
Fe ₂ O ₃	2.69	-	2.67
FeO	1.83	0.80	-
MnO	0.06	0.02	0.04
MgO	2.57	0.51	0.78
CaO	4.92	2.25	1.99
Na ₂ O	4.69	4.14	4.09
K ₂ O	0.91	2.77	4.55
P ₂ O ₅	0.17	0.04	0.14
Cr	52	17	9
Ni	44	12	6
Rb	9	24	171
Sr	631	415	483
Y	5	0	12
Zr	191	20	302
Nb	5		
Ba	728	903	1863
Ce	44	28	
La	20	21	
Pb	12		
Th	1	0	24
K/Rb	883	958	221
Ba/Rb	86	38	11
Ba/Sr	1.2	2.2	3.9
Rb/Sr	.013	.053	.354

Table 4.5 Comparison of the average granite sheet (2) with the average hornblende-biotite gneiss from Assynt (Tarney, 1976), and G-2, the U.S.G.S. standard granite (Planagan, 1973; Abbey, 1973).

Fig. 4.21 REE patterns of five of the granite sheets.



between mineral/melt (Frey et al., 1968). The low concentrations of this element in the granites is also consistent with a garnet-bearing residue.

Similar highly fractionated REE patterns (also believed by the respective authors to be due to residual garnet) have been described by Onions and Pankhurst (1974) for archaean Amitsoq gneisses from Godthaab, West Greenland and by Compton (1978) for grey gneisses and trondhjemites from the archaean of Buksefjorden, West Greenland. These patterns are presented in fig. 4.22.

Following the presentation of the field evidence (chapter two) and the geochemistry (above) it is now possible to comment on the origins of the granite sheets.

The sheets are localised in occurrence, are intrusive, small in volume and were injected at a late stage in the Scourian metamorphism (chapter two). As suggested by other workers (Bowdidge, 1969; Sheraton et al., 1973), they clearly bear the hallmark of a local partial melt derived by high grade metamorphism. If this were the case, and, bearing in mind the high silica content of the granites, the only feasible residuum would be the acid granulites present within the area. Table 4.5 compares the analyses of Lewisian hornblende-biotite gneisses (Terney, 1976) with the average granite sheet from this study. The modal mineralogy of the acid gneisses around Scourie is quartz-andesine-hornblende-augite-hypersthene-biotite. Published mineral/melt distribution coefficients for granitic rocks are such that this assemblage would decrease Sr in a partial melt, leave Ba, Rb, K/Rb, K/Ba more or less constant, and increase Ba/Sr and Rb/Sr (Hanson, 1978). This is consistent with the analyses present in Table 4.5. However, REE patterns are somewhat contradictory to this simple interpretation. If plagioclase were a residual phase in the gneisses, one

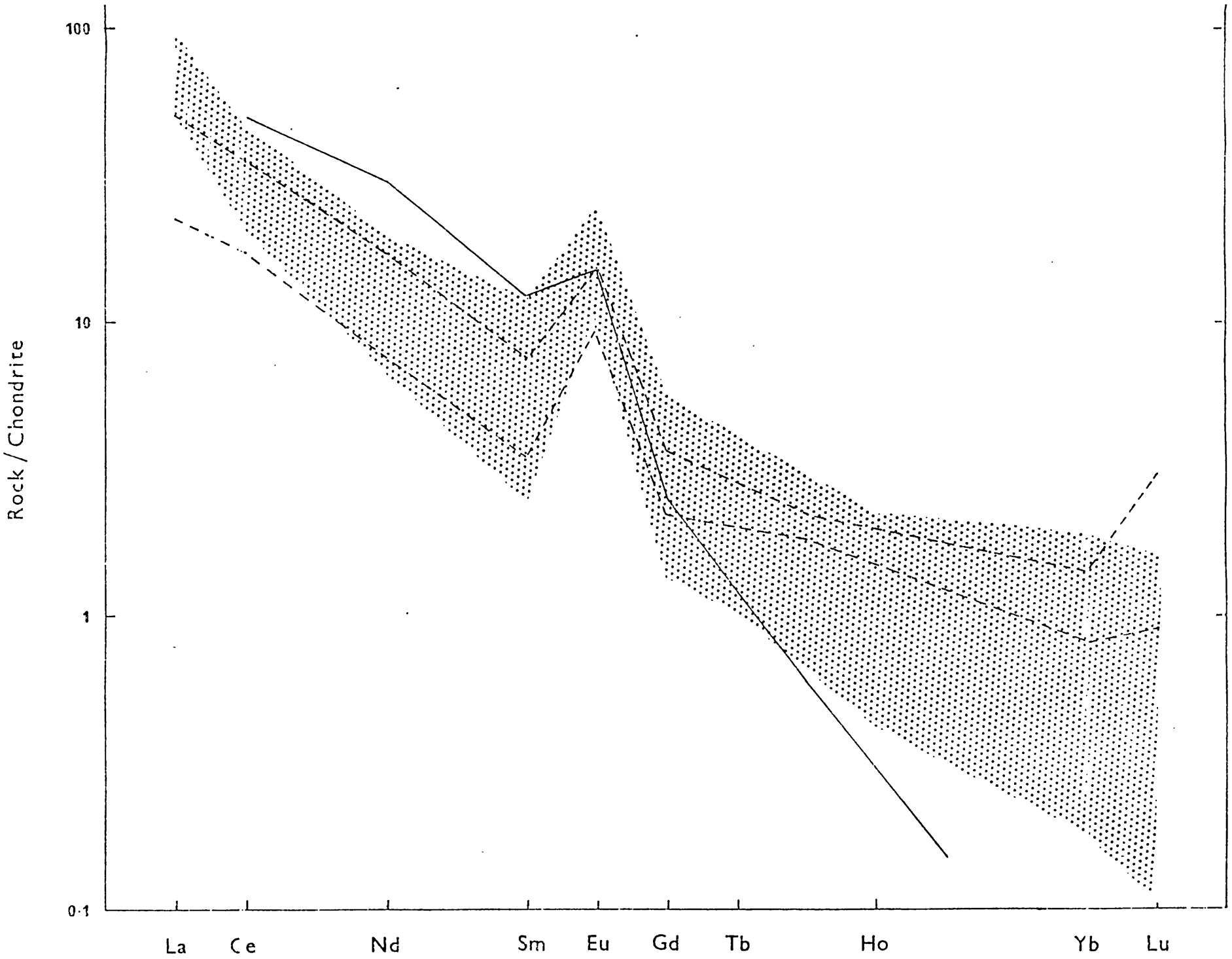


Fig. 4.22 REE patterns for the Granite sheets (stippled) compared with those of gneiss (dashed lines) from the same area. The shaded area represents the range between the two patterns.

would expect a negative europium anomaly in the melt (Hanson, 1978). This is clearly not the case. Such a large positive europium anomaly in the granites is probably the result of a low ambient oxygen fugacity. As previously mentioned, the heavy-REE depletion is probably due to residual garnet. So although Ba, Sr and Rb concentrations suggest that the host gneisses are possibly parental to the granite sheets, the REE patterns are not consistent with the present mineralogy. Since suitable garnet-granulites are not exposed at the surface in the Scourie area, it is tentatively suggested that these occur at a greater depth. As metamorphic isograds are more or less horizontal in the Scourian, it is assumed that garnet is likely to be developed at depth in the intermediate rock-types. Thus the granite sheets are interpreted as being minor, late-stage partial melts of a garnet-granulite assemblage with Scourie Gneiss geochemistry.

4.3 THE MAFIC AND ULTRAMAFIC COMPONENTS OF THE NET-VEINED SHEETS.

Only two ultramafites (samples 146, 147) and one mafic gneiss (sample 158) have been analysed for major, trace and rare-earth elements, so only a brief comment on the nature of the chemistry of these rocks will be made.

Table 4.6 compares the analyses of the net-veined rocks with the analogous components of the layered sheets. Broad similarities exist, although the net-veined ultramafites have lower MgO (14%) and higher CaO (18%), lower Cr and Ni (499, 281, respectively) and much higher Rb, Sr and Ba (20, 107 and 369 respectively) than their layered counterparts. The norm has a high diopside content (65%) and no hypersthene, a marked contrast with the layered ultramafites. The basic rocks are broadly comparable, although the concentration of Ba is 5 times higher in the net-veined mafics than in the

	<u>1</u>	<u>2</u>	<u>3</u>	<u>4</u>
SiO ₂	46.73	51.04	47.31	46.20
TiO ₂	0.35	0.45	0.86	1.27
Al ₂ O ₃	7.37	5.22	14.13	14.64
Fe ₂ O ₃	5.58	0.31	2.42	4.36
FeO	8.91	7.21	10.82	9.54
MnO	0.21	0.15	0.25	0.11
MgO	22.64	14.37	9.16	7.50
CaO	7.09	18.21	11.66	12.12
Na ₂ O	0.67	0.93	1.87	1.60
K ₂ O	0.34	0.53	0.27	0.92
P ₂ O ₅	0.03	0.07	0.09	0.12
H ₂ O ⁻	0.10	0.28	0.12	0.32
L.O.I.	1.89	1.83	1.40	1.74
Total	<u>99.71</u>	<u>100.60</u>	<u>100.36</u>	<u>100.44</u>
V	141	179	273	
Cr	2991	499	461	171
Ni	936	281	167	
Cu	19	80	182	
Zn	118	68	162	
Zr	23	29	37	
Rb	7	20	1	
Sr	51	107	130	
Th	0	0	0	
Y	8	14	26	
Ba	23	369	45	221
K/Rb	403	220	2241	
Rb/Sr	.137	.187	.008	
K/Ba	123	119	50	35
K/Sr	55	41	17	
Ba/Rb	3.3	18.4	45	
Ba/Sr	.4	3.4	.3	

Table 4.6 Comparison of the rock-types of the layered and net-veined mafic-ultramafic sheets. 1 = Average Layered Ultramafite; 2 = Average Net-veined Ultramafite; 3 = Average Layered Mafic Gabbro; 4 = Net-veined Mafic Gneiss.

layered mafics.

REE patterns for one ultramafite (146) and a basic gneiss (158) are presented in 4.23. The ultramafite has a fractionated pattern with a Ce_N/Yb_N ratio of 6.0 with concentrations approximately 30 times chondrite for La and twice chondrite for Lu. There is no appreciable europium anomaly. The basic gneiss has a flat, unfractionated pattern with Ce_N/Yb_N of 1.6. Concentrations are approximately fifteen times that of chondrite, with no europium anomaly.

Although the geochemical data presented here are scant, comparisons with rocks from the layered sheets reveal some differences. This is particularly evident in the LIL elements which are enriched in the net-veined rocks - Ba, Sr, Rb and the light REE. This is accompanied by lower K/Rb, K/Ba, K/Sr and higher Rb/Sr, Ba/Rb and Ba/Sr. These values are more typical of continental, calc-alkaline tholeiites (Jakes and White, 1970) and are in marked contrast with the layered sheets. The net-veined complexes seem to have more geochemical similarities with the calc-alkaline acid granulites and gneisses of the Scourie district. Thus the net-veined rocks may be derived from the parental calc-alkaline magmas of the surrounding acid gneisses, as envisaged by Windley and Smith (1976) and Tarney and Windley (1978) for the layered mafic complexes of the Archaean high-grade regions.

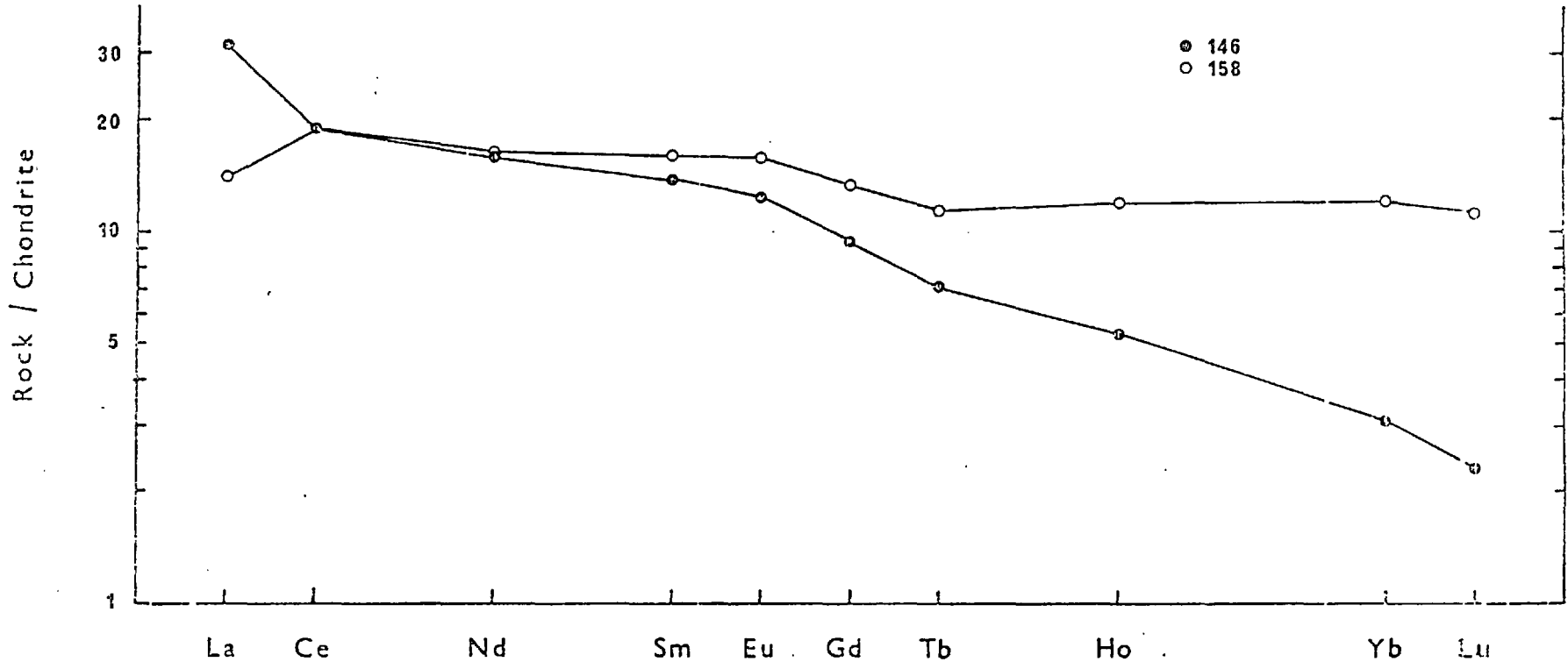


Fig. 4.23 REE patterns of an ultramylonite (146) and a mafic gneiss (158) from a mel-veined intrusion.

CHAPTER 5

MINERAL CHEMISTRY

5.1 GENERAL INTRODUCTION

A total of 78 mineral analyses have been made from the mafic and ultramafic rocks of the Camas nam Buth intrusion (Tables 13-22). Analyses were obtained on an energy-dispersive microprobe (EDS) and a computer-automated 'Geoscan'. Details of the analytical techniques may be found in Appendix II. Fe is generally expressed as Fe^{2+} (Total), although a theoretical ferric iron content has been calculated for some mineral species (clinopyroxenes, garnets, spinels, magnetites and ilmenites) and has been plotted in certain figures. This calculation assumes a stoichiometric cation total. Jadeite and calcium-tschermak's molecules of clinopyroxenes have been calculated according to the scheme of Cawthorn and Collerson (1974).

Each mineral species is dealt with in turn and the final sections cover element distribution between co-existing minerals, and a summary of mineral chemistry.

5.2 PYROXENES

Pyroxenes are generally preserved in every sample from the Camas nam Buth intrusion in varying stages of replacement by fibrous green amphibole. This fact has provided a range of compositions for analysis, and twelve pairs of co-existing calcium-rich and calcium-poor pyroxenes have been analysed (fig. 5.1), Tables 13-14.

Tie-lines between co-existing pyroxenes are sub-parallel, and the projections of these tie-lines intersect the enstatite-wollastonite join between Wo_{80} and Wo_{92} . These intersections are similar to those described by Hess (1941) for co-existing pyroxenes from igneous rocks. Deer et al. (1972) suggest that it is not possible

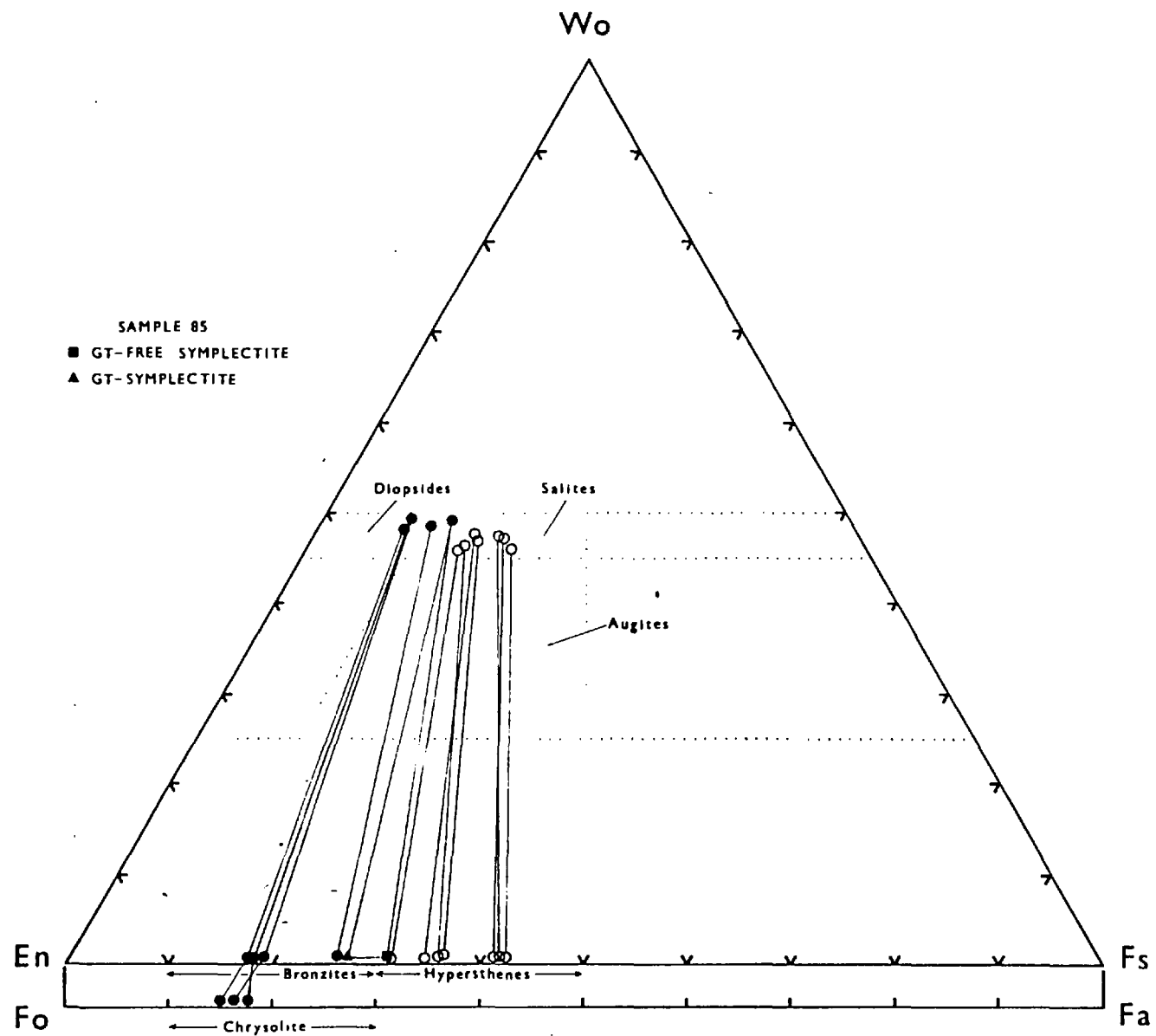


Fig. 5.1 Compositions of pyroxenes in the enstatite-wollastonite-ferrosilite triangle. Co-existing olivines are also plotted in terms of forsterite-fayalite. Filled symbols refer to an ultramafite host rock, open symbols a mafic gabbro host rock.

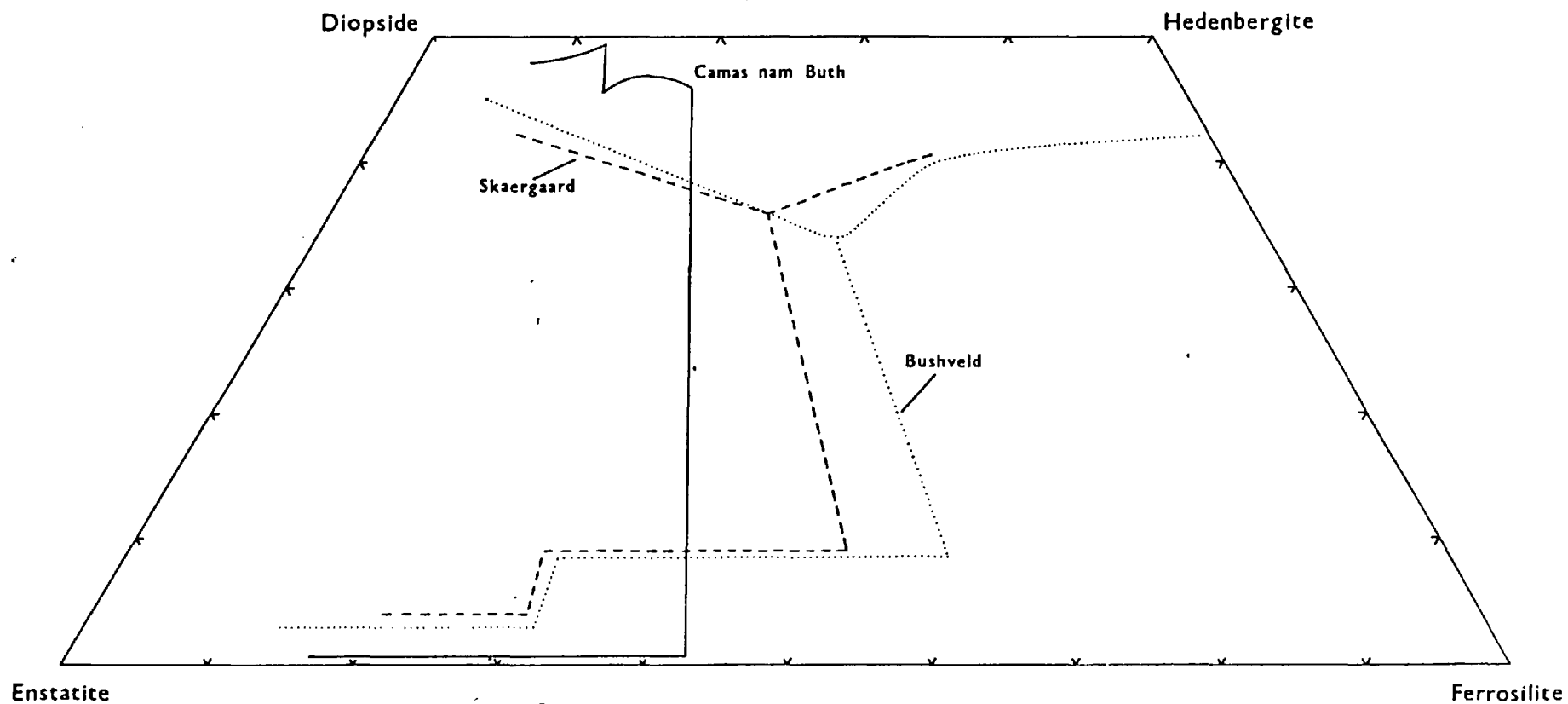
to distinguish igneous from metamorphic pyroxenes using these intersections, although they suggest that sub-parallel tie-lines do indicate crystallisation under equilibrium conditions.

A comparison with pyroxene trends from un-metamorphosed layered basic intrusions is presented in fig. 5.2. Although this diagram illustrates the broad similarity in iron-enrichment trends for the clinopyroxenes and orthopyroxenes in the intrusions, it highlights the calcium-poor nature of the orthopyroxenes and calcium-rich nature of the clinopyroxenes in the Camas nam Buth intrusion as compared with those of the un-metamorphosed intrusions.

A marked correlation between host rock $\text{FeO}/\text{FeO} + \text{MgO}$ and pyroxene $\text{FeO}/\text{FeO} + \text{MgO}$ is illustrated in fig. 5.3. The bulk of both the clinopyroxenes and orthopyroxenes have lower $\text{FeO}/\text{FeO} + \text{MgO}$ ratios than their corresponding host rocks.

The ultramafite clinopyroxenes are aluminous diopsides, with low Ti contents. Clinopyroxenes from the plagioclase-free ultramafites have jadeite contents between 1.4 and 3.7 mol.percent and ca-tschermak contents between 6.4 and 7.6 mol.percent. Those from plagioclase-bearing ultramafites have higher jadeite and ca-tschermak contents of 4.6 - 5.2 and 8.8 - 10.6 mol.percent, respectively. The mafic gabbro clinopyroxenes are salitic with jadeite contents between 4.5 and 7.3 mol.percent and ca-tschermak contents between 6.5 and 8.9 mol.percent. Taking the clinopyroxenes as a whole, the highest jadeite and ca-tschermak contents are in the plagioclase-bearing ultramafites and the low $\text{FeO}/\text{FeO} + \text{MgO}$ mafic gabbros. This distribution is in accord with the metamorphic reactions described in chapter three. It was suggested that the plagioclase in these rock types was due solely to the breakdown of garnet (producing orthopyroxene, plagioclase and spinel) without the exsolution of jadeite and ca-tschermak components from the

Fig. 5.2 Trends of co-existing pyroxenes from the Camas nam Buth sheet and some igneous layered intrusions.



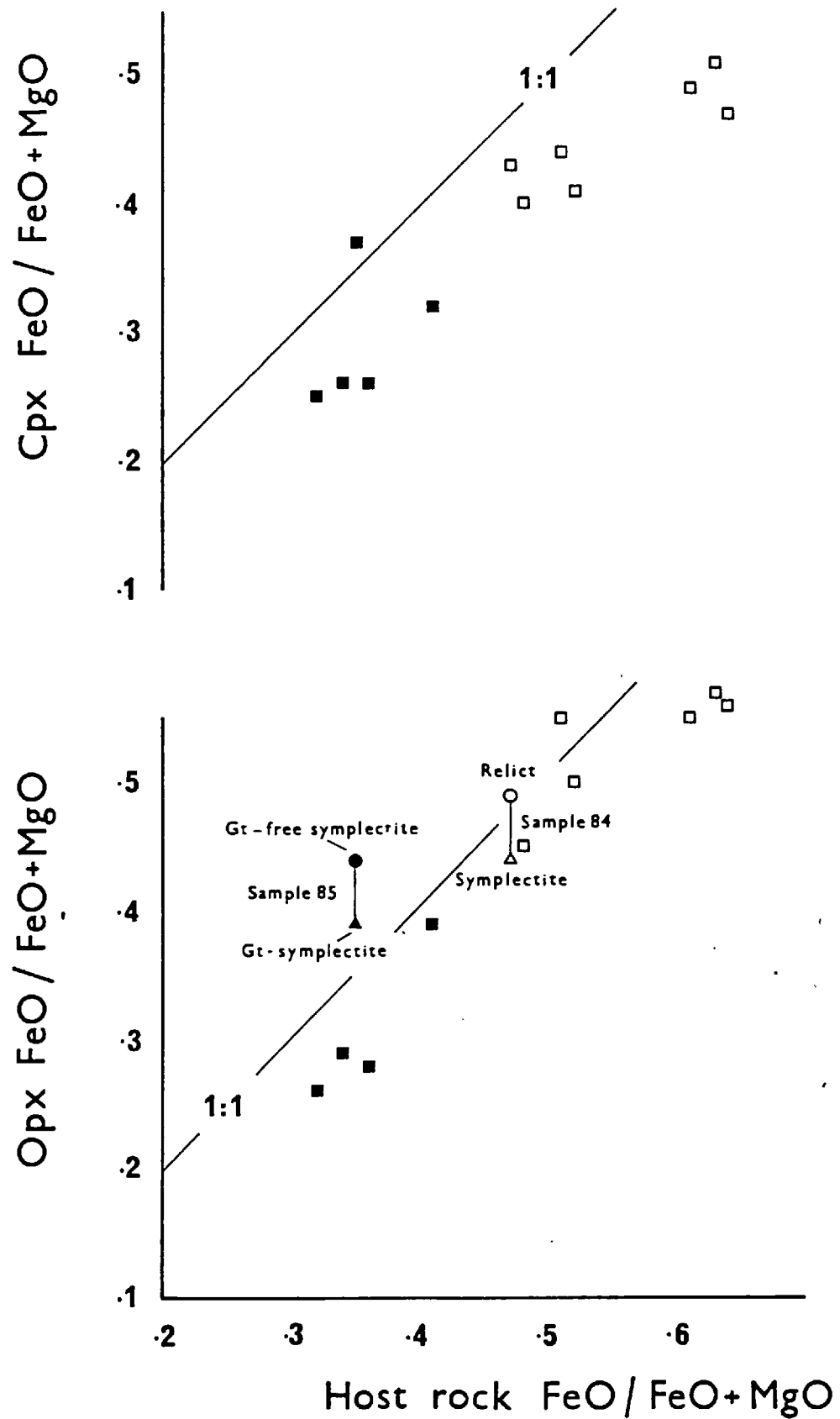


Fig. 5.3 The relation between pyroxene FeO/FeO+MgO and host rock FeO/FeO+MgO. Filled symbols refer to an ultramafite host rock, open symbols a mafic gabbro host rock.

clinopyroxenes, whereas the bulk of the plagioclase in the high FeO/FeO + MgO mafic gabbros was due to the partial breakdown of garnet, coupled with minor exsolution of ca-tschermak and jadeite from the clinopyroxenes.

Minor element concentrations are plotted against Fe/Fe + Mg ratios of the clinopyroxenes in fig. 5.4. All elements show gradations between ultramafic and mafic clinopyroxenes, except for alumina which is probably strongly affected by co-existing aluminous minerals (spinel, amphibole, garnet and plagioclase in the ultramafites, garnet and plagioclase in the mafic gabbros). Cr_2O_3 contents range up to 0.3 wt.percent in the ultramafite clinopyroxenes.

Orthopyroxenes in the ultramafites are bronzites, whilst those of the mafic gabbros are hypersthene. All are aluminous and calcium-poor (wollastonite < 1 mol.percent). Although alumina contents are lower than for the clinopyroxenes, contents still reach five wt.percent. MnO, CaO and Al_2O_3 are more or less constant with increasing Fe/Fe + Mg (fig. 5.5). As mentioned in chapter three, the orthopyroxenes in the garnet-symplectites of sample 85 have lower Fe/Fe + Mg and alumina contents than those in the garnet-free symplectites. This due to the selective partitioning of alumina and iron into the garnet.

5.3 OLIVINES

Three olivines from the ultramafites have been analysed (Table 15). These are chrysolites with forsterite contents ranging from 82-85 mol.percent (fig. 5.1). They are relatively iron-rich when compared with olivines from 'Alpine-type' ultrabasic rocks e.g. Fo_{89-91} , Josephine Peridotite, Oregon (Dick, 1977); Fo_{90-91} , Finero Peridotite, Italy (Brodie and White, 1978).

Nickel contents are high (up to 0.5 wt.percent) and are within the range described by Simkin and Smith (1970), who noted

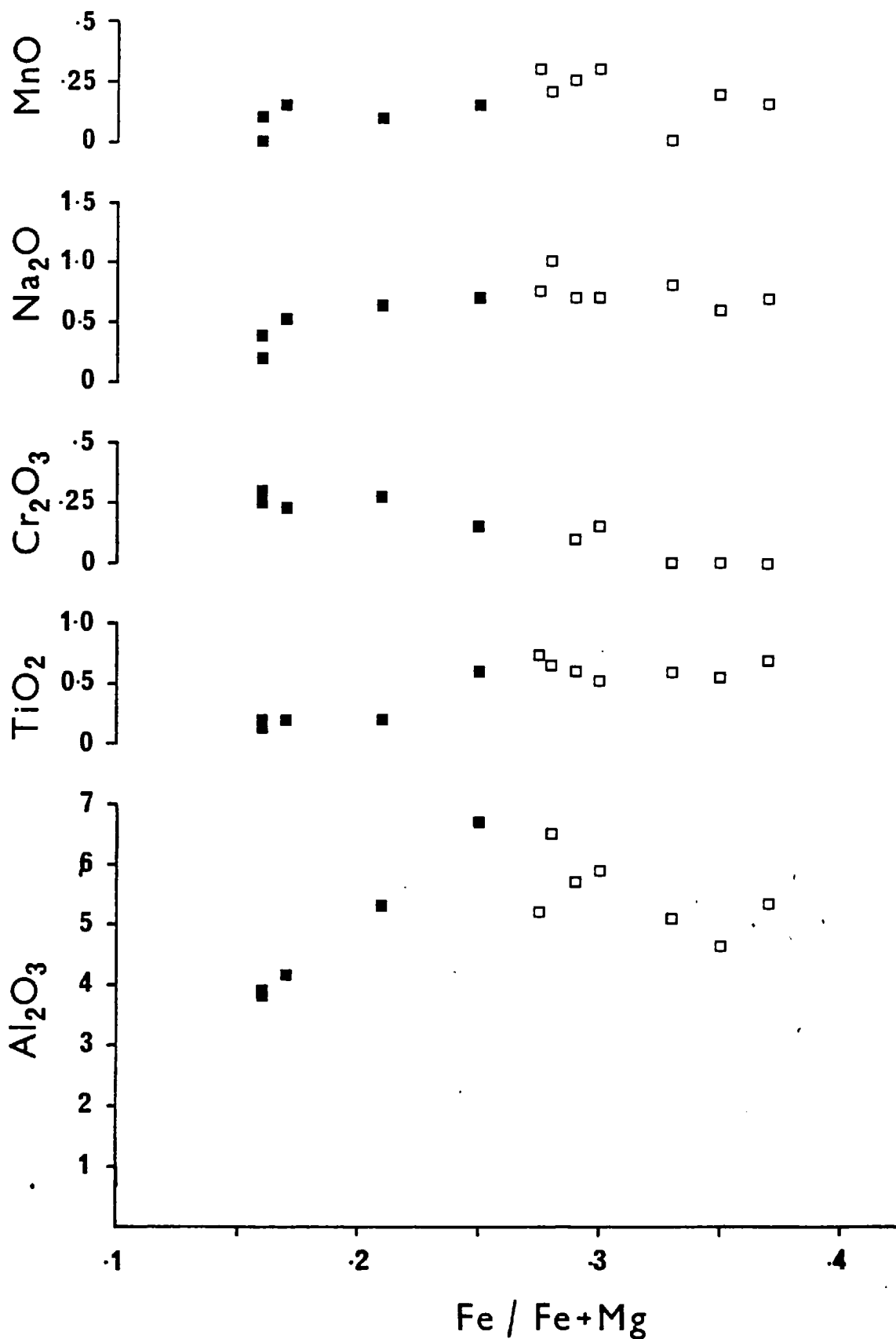


Fig. 5.4 Minor element variation versus Fe/Fe+Mg ratio of clinopyroxenes. Filled symbols refer to an ultramafite host rock, open symbols a mafic gabbro host rock.

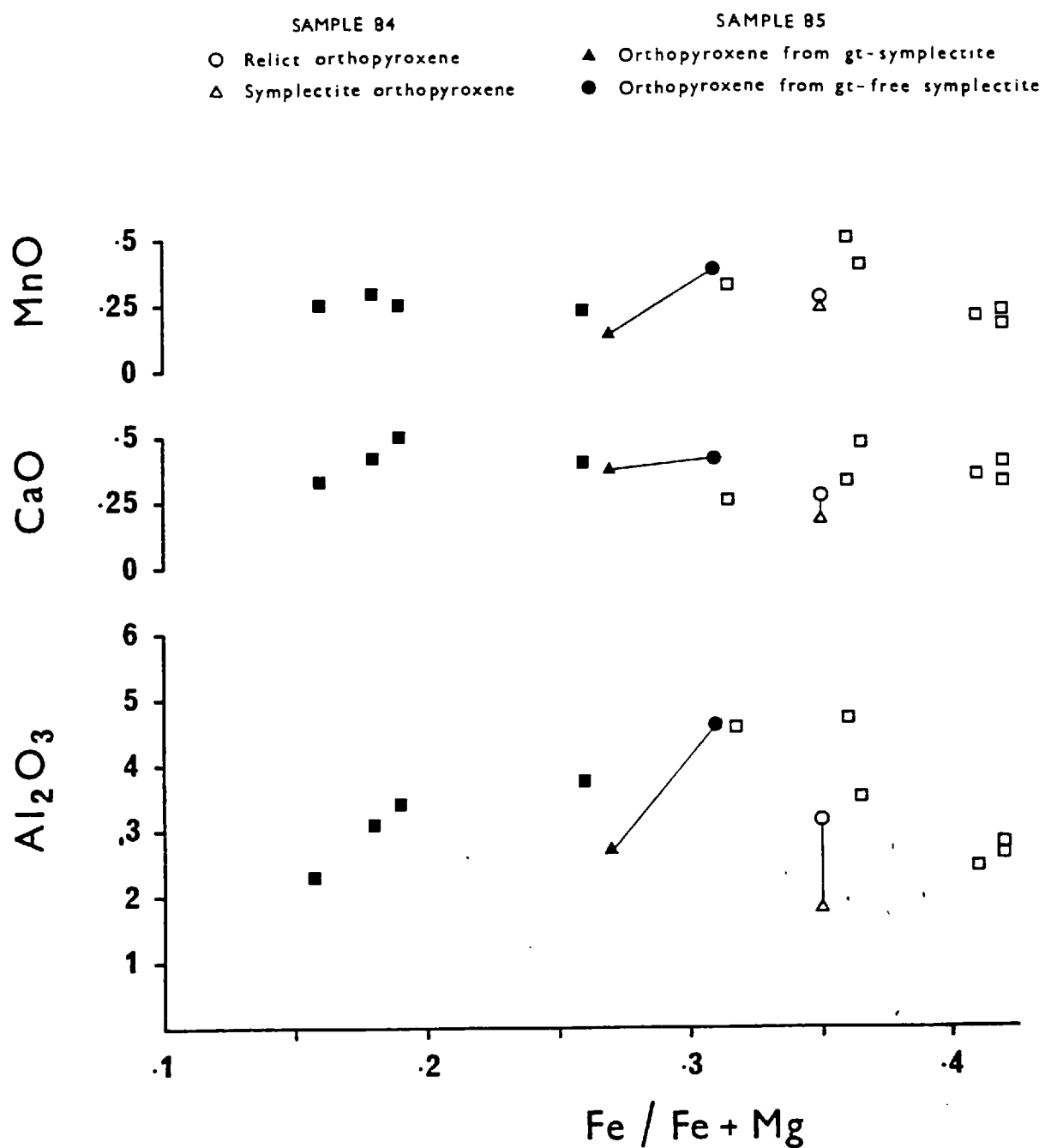


Fig. 5.5 Minor element variation versus Fe/Fe+Mg ratio of orthopyroxenes. Filled symbols refer to an ultramafite host rock, open symbols a mafic gabbro host rock.

a positive correlation between nickel and forsterite contents of terrestrial olivines. Fleet et al. (1977) decided that nickel content was an exponential function of forsterite content and were able to distinguish separate distributions for olivine from different rock associations (fig. 5.6). The Camas nam Buth olivines show greatest similarity with those from igneous layered intrusions.

CaO has not been detected in the olivines. Simkin and Smith (1970) noted that CaO contents of olivines from plutonic rocks are generally less than 0.1 wt.percent. MnO contents are about 0.2 wt.percent, well within the range described by Simkin and Smith (1970) in their survey.

5.4 GARNETS

Garnet porphyroblasts and coronas have been analysed in a number of samples from the Camas nam Buth intrusion (Table 16). All are almandine-rich with pyrope content varying from 21 to 36 mol. percent. Grossular contents are approximately constant at about 17 mol.percent, spessartine forms a minor constituent, not more than 3.1 mol.percent. The corona garnets are richer by 5-6 mol.percent in almandine than their co-existing porphyroblasts, a change accompanied by a corresponding decrease in the pyrope molecule (fig. 5.7). The field of garnet compositions is encompassed by the range of garnet compositions from granulites and amphibolites described by Chappell and White (1970) (fig. 5.8). Calculated theoretical ferric oxide contents for the garnets do not exceed 2.1 wt.percent, with no difference between porphyroblastic and corona types.

Cr_2O_3 and NiO have not been detected, and the Scourie garnets show no preference for these elements.

The $\text{FeO}/\text{FeO} + \text{MgO}$ ratios of the garnets have a direct relation-

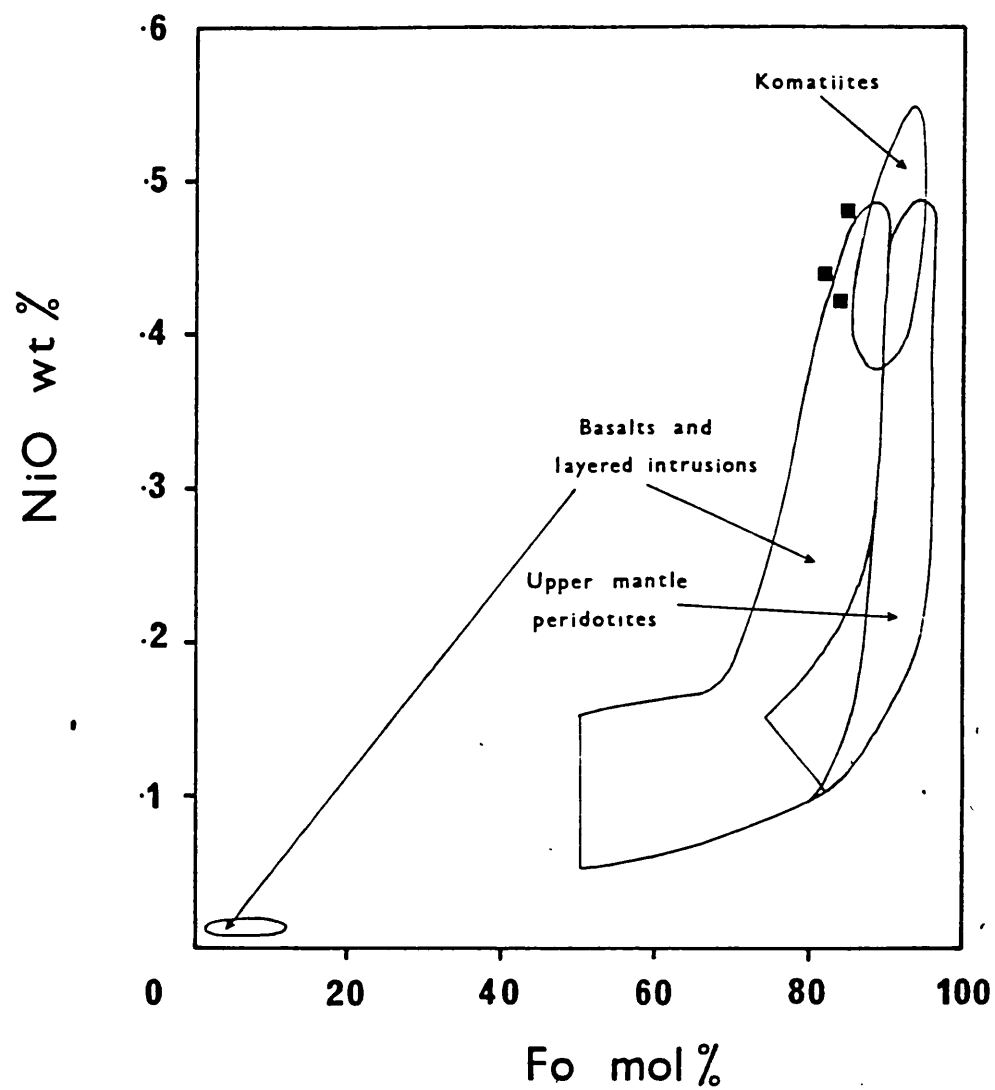


Fig. 5.6 Variation of nickel content with forsterite content of analysed olivines (filled squares). Compositional fields after Fleet et al. (1977).

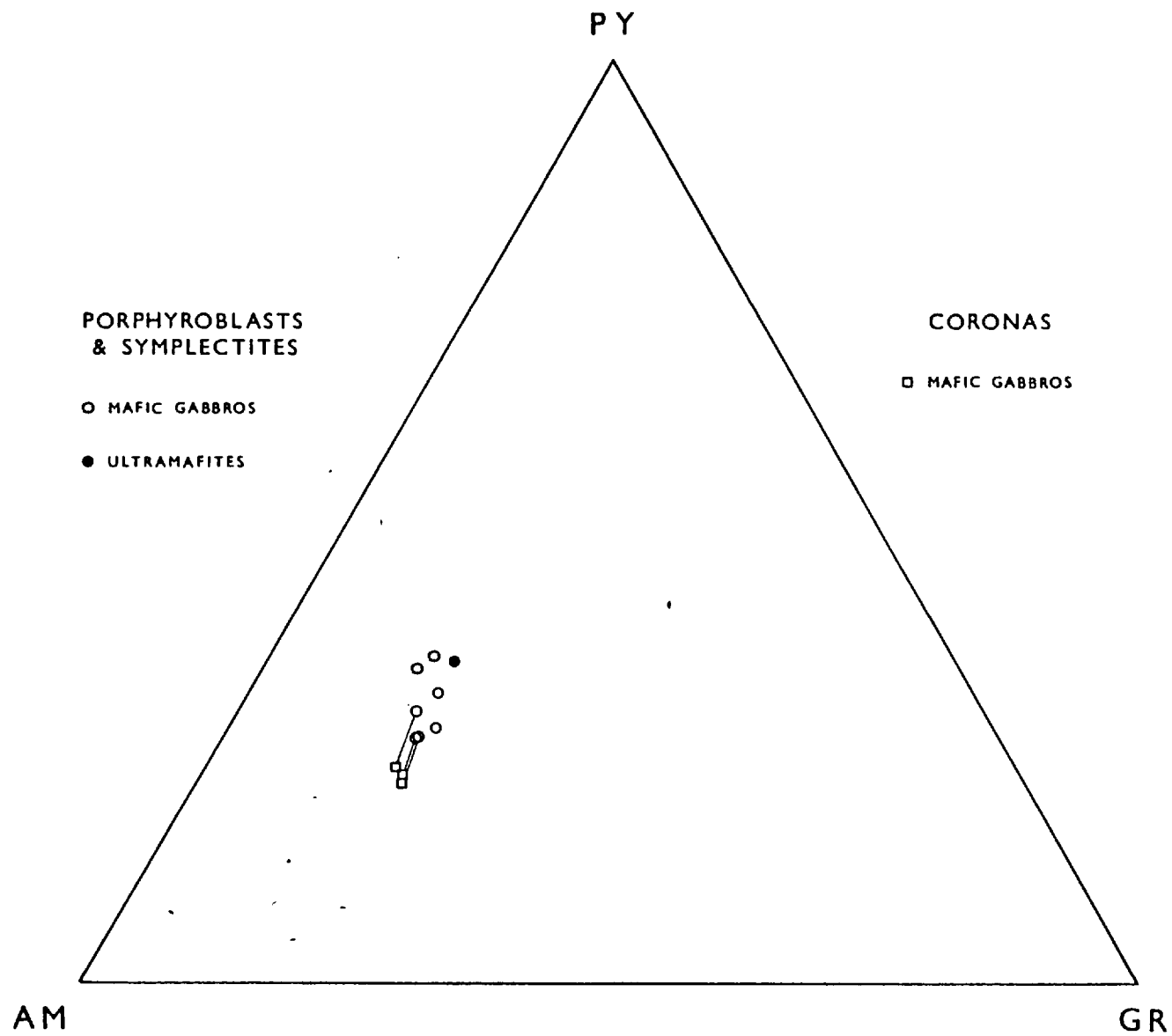


Fig. 5.7 Compositions of garnets in terms of pyrope ($\text{Mg}_3\text{Al}_2\text{Si}_3\text{O}_{12}$), almandine ($\text{Fe}_3\text{Al}_2\text{Si}_3\text{O}_{12}$), and grossular ($\text{Ca}_3\text{Al}_2\text{Si}_3\text{O}_{12}$).

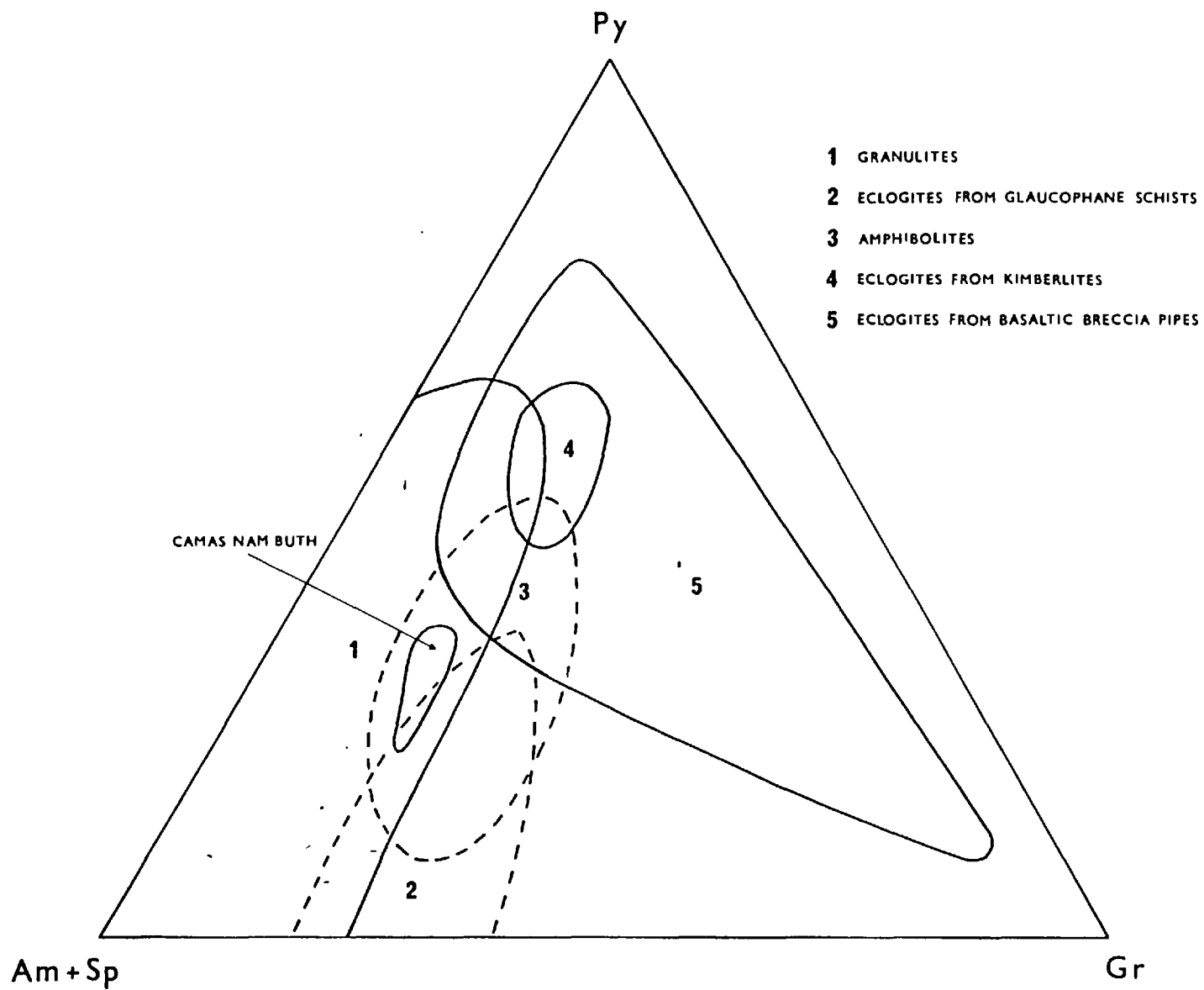


Fig. 5.8 Field of Camas nam Buth garnet compositions, compared with garnets from other metamorphic terrains. Data from Chappell and White (1970).

ship to the $\text{FeO}/\text{FeO} + \text{MgO}$ ratios of their host rocks (fig. 5.9). The ratio is greater in the garnets, but approaches a 1:1 relationship as the ratio in the host rock increases. This feature has been observed in garnets from granulite facies rocks from Lofoten, Norway by Griffin and Heier (1969). A host rock $\text{FeO}/\text{FeO} + \text{MgO}$ ratio of 0.35 - 0.40 appears to be the minimum for M1-M2 garnet growth (chapter three) in the Scourie area. Similarly, a ratio of 0.55 appears to be the minimum for corona garnet growth during the M_4 metamorphism (chapter three).

5.5 SPINELS

The green spinels in the mafic and ultramafic rock types are members of the MgAl_2O_4 - FeAl_2O_4 solid solution series and are pleonastes or hercynites (fig. 5.10, Table 17). Cr_2O_3 contents are low, below 7 wt.percent. There is very little substitution on the trivalent cation site, which is dominated by Al (fig. 5.11). Calculated theoretical ferric iron contents are also low, not exceeding 5.2 wt.percent.

NiO may reach 0.5 wt.percent and MnO 0.25 wt.percent, although there is no systematic variation with $\text{Fe}/\text{Fe} + \text{Mg}$ ratio (fig. 5.12). TiO_2 has not been detected in the spinels. One of the spinels (sample 104) was analysed for ZnO and found to contain a significant amount (1.13 wt.percent).

The spinels in the garnet-symplectites of sample 85 have lower $\text{Fe}/\text{Fe} + \text{Mg}$ ratios than those of the garnet-free symplectites, because of the preferential entry of iron into garnet.

The Camas nam Buth spinels show no similarity to those of stratiform or Alpine-type intrusions (fig. 5.13), mainly because of the low Cr and high Al contents. High $\text{Al}/\text{Al} + \text{Cr}$ ratios are typical of spinels from metamorphosed basaltic rocks (Griffin, 1971; Griffin

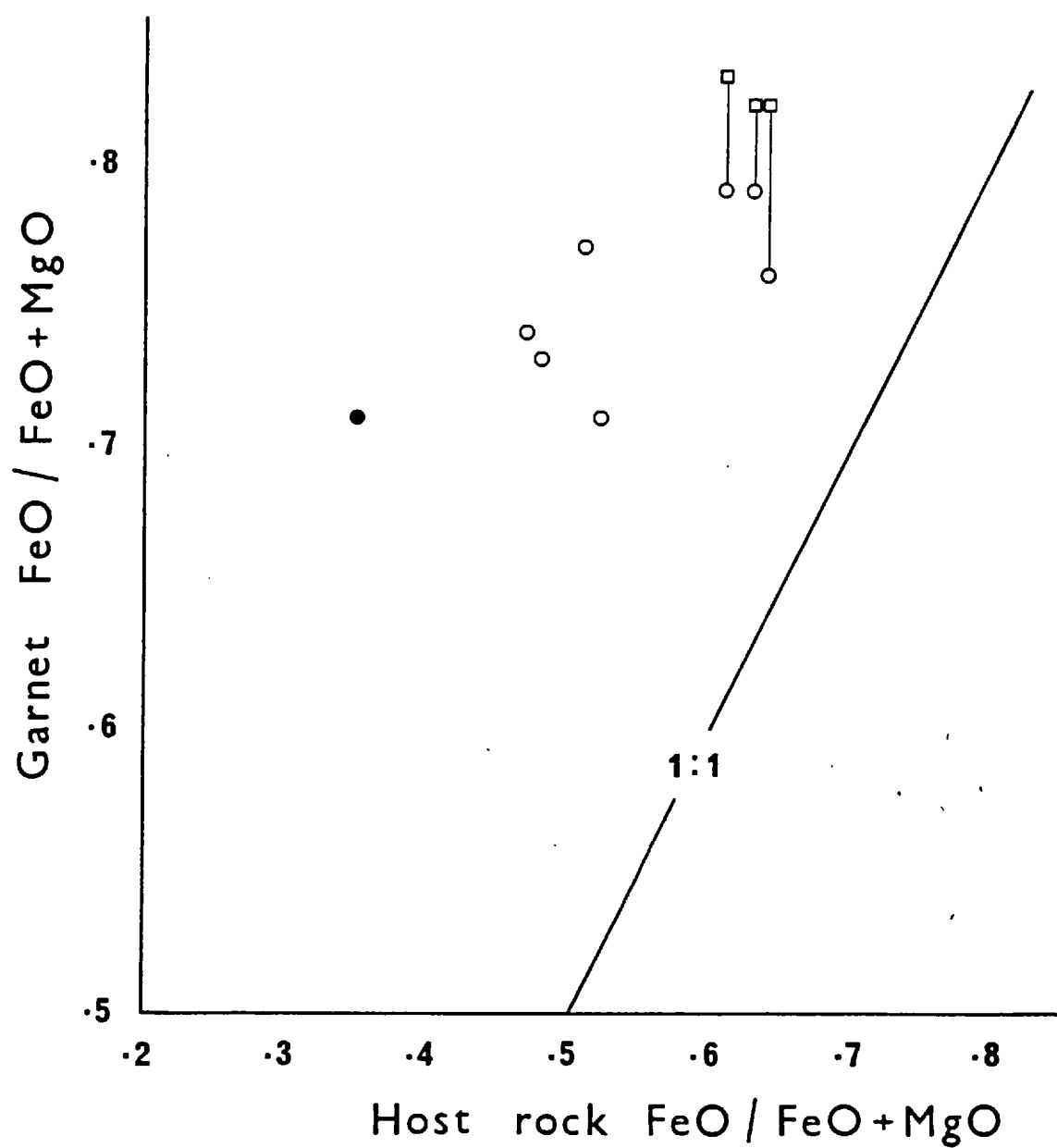


Fig. 5.9 Variation of garnet $\text{FeO} / (\text{FeO} + \text{MgO})$ with host rock $\text{FeO} / (\text{FeO} + \text{MgO})$. Symbols are as for fig. 5.7.

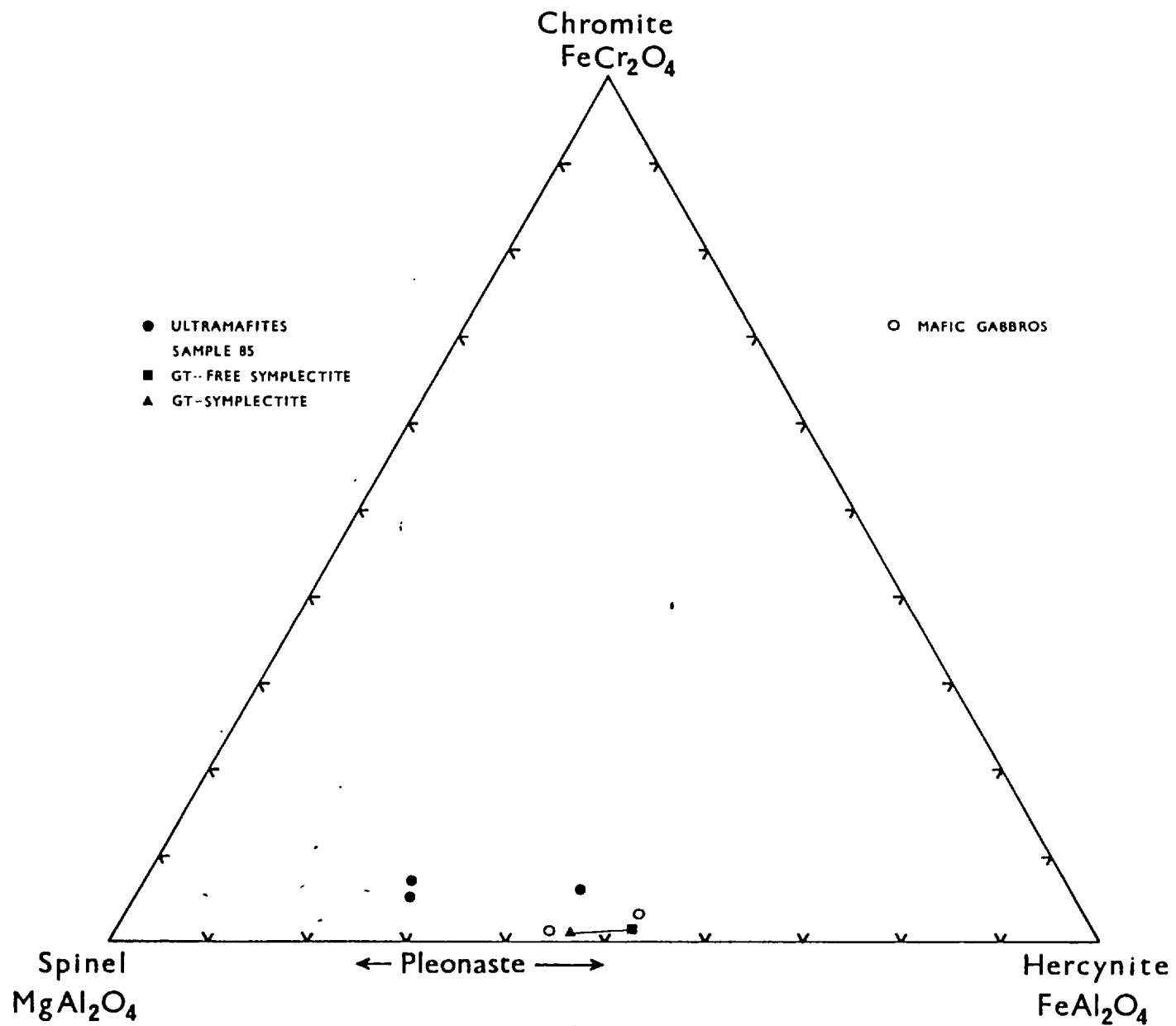


Fig. 5.10 Compositions of spinels in terms of the spinel-chromite-hercynite triangle.

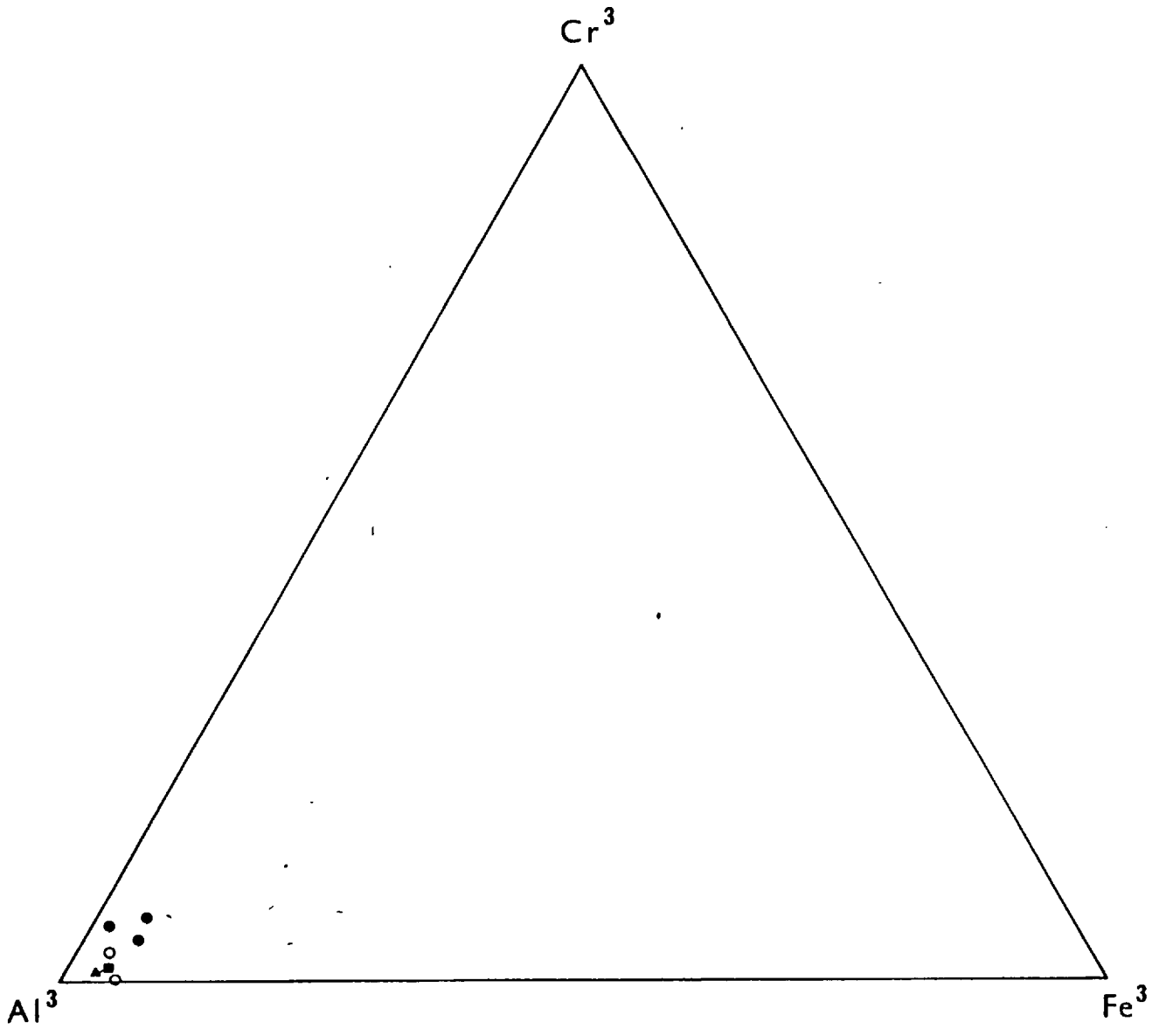


Fig. 5.11 Compositions of spinels in terms of the trivalent cations. Ferric iron concentrations were calculated from mineral stoichiometry. Symbols are as for fig. 5.10.

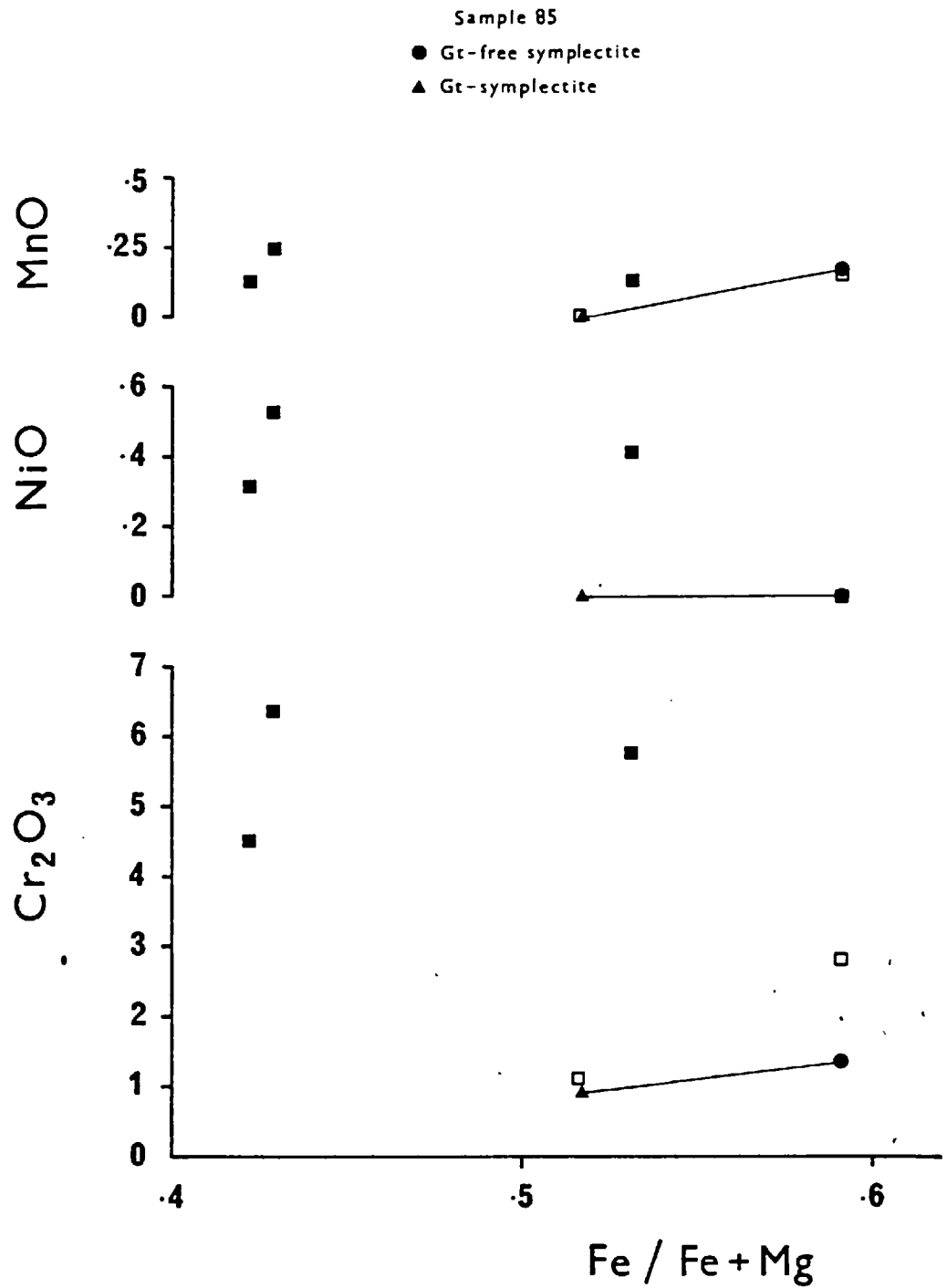


Fig. 5.12 Minor element variation versus Fe/Fe+Mg ratio of spinels. Filled symbols refer to an ultramafite host rock, open symbols a mafic gabbro host rock.

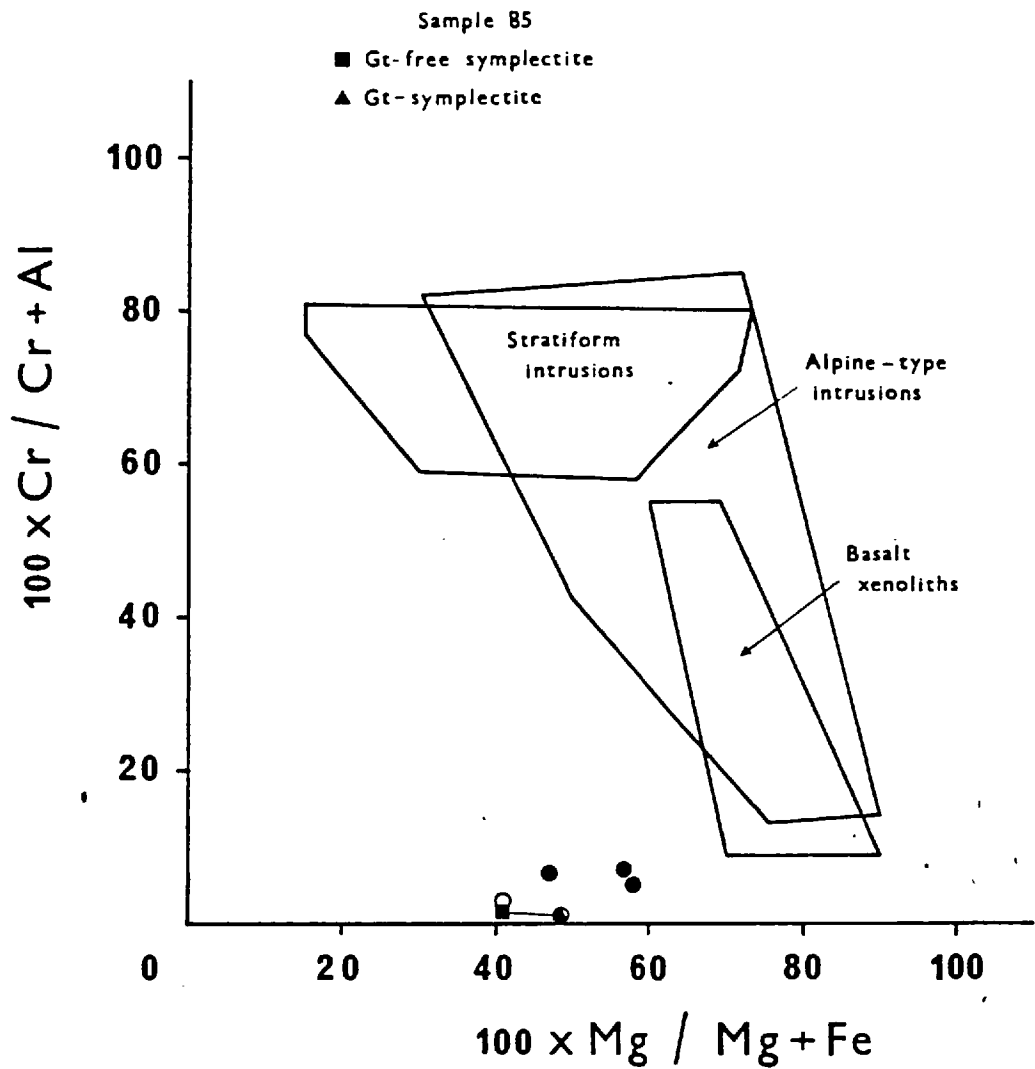


Fig. 5.13 Comparison of the Camas nam Buth spinels with those of stratiform intrusions, alpine-type intrusions, and basalt xenoliths. Diagram after Irvine (1967) and Evans and Frost (1975). Symbols are as for fig. 5.10.

and Heier, 1973; Wilkinson, 1975), whereas primary spinels in basalts have low $Al/(Al + Cr)$ ratios (Evans and Wright, 1972; Jackson, 1969; Irvine, 1965, 1967). Hence, the Camas nam Butth spinels are consistent with equilibration during metamorphism.

5.6 AMPHIBOLES

Only the high-grade, granulite facies amphiboles (chapter 2) have been analysed (Table 18). These are dark green-brown in colour, as opposed to the emerald-green colour of the amphibole that replaces pyroxene during retrogression.

All the amphiboles analysed may be described as pargasites. However, amphibole classification is complex and three schemes of classification are utilised here.

Leake (1968) presents one such classification, using Si atoms plotted against niggli mg ($Mg/(Mg + Fe + Mn)$) as illustrated in fig. 5.14. This is designed for amphiboles with $Ca + Na + K$ atoms greater than 2.5 and Ti atoms less than 0.5. The amphiboles from the ultramafites plot as pargasites, whilst those from the mafic gabbros plot as ferroan pargasites. This scheme of classification will be adopted here.

Fig. 5.15 shows the amphiboles plotted on a tetrahedral Al atoms versus $Na + K$ atoms classification (Deer et al., 1963). There is no discrimination of the amphibole types in this diagram and all plot towards the composition of pargasite.

The third classification is as described by Ernst (1968) and modified by Windley and Smith (1974). The amphiboles from the ultramafites are separated from those from the mafic gabbros and plot towards a pargasitic composition (fig. 5.16). The more iron-rich amphiboles from the mafic gabbros, although still plotting in the pargasite field, tend towards a ferro-pargasite composition.

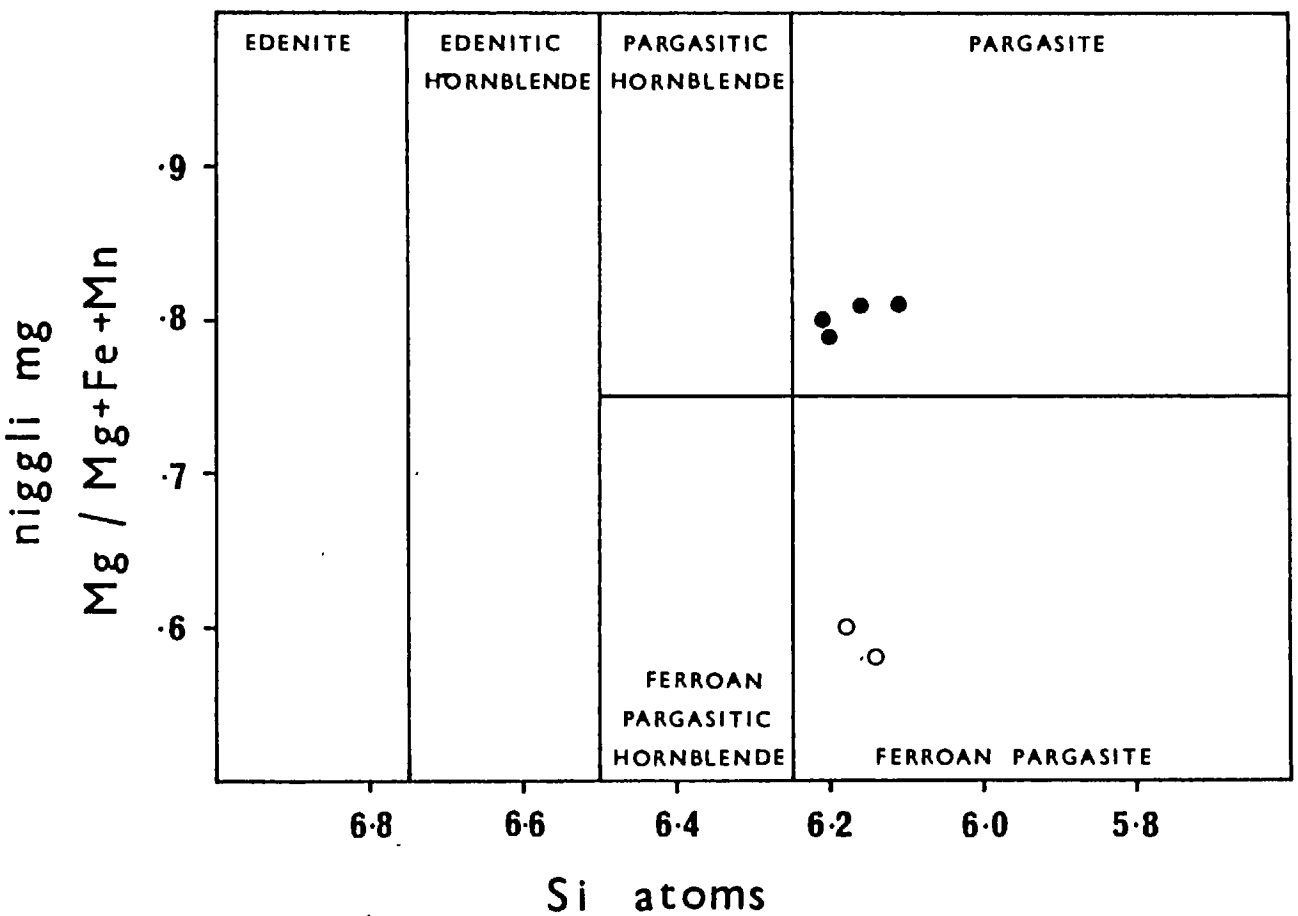


Fig. 5.14 Classification of amphiboles after Leake (1968). All the amphiboles have Na Ca K atoms greater than 2.5 and Ti atoms greater than 0.5. Filled symbols refer to an ultramafite host rock, open symbols a mafic gabbro host rock.

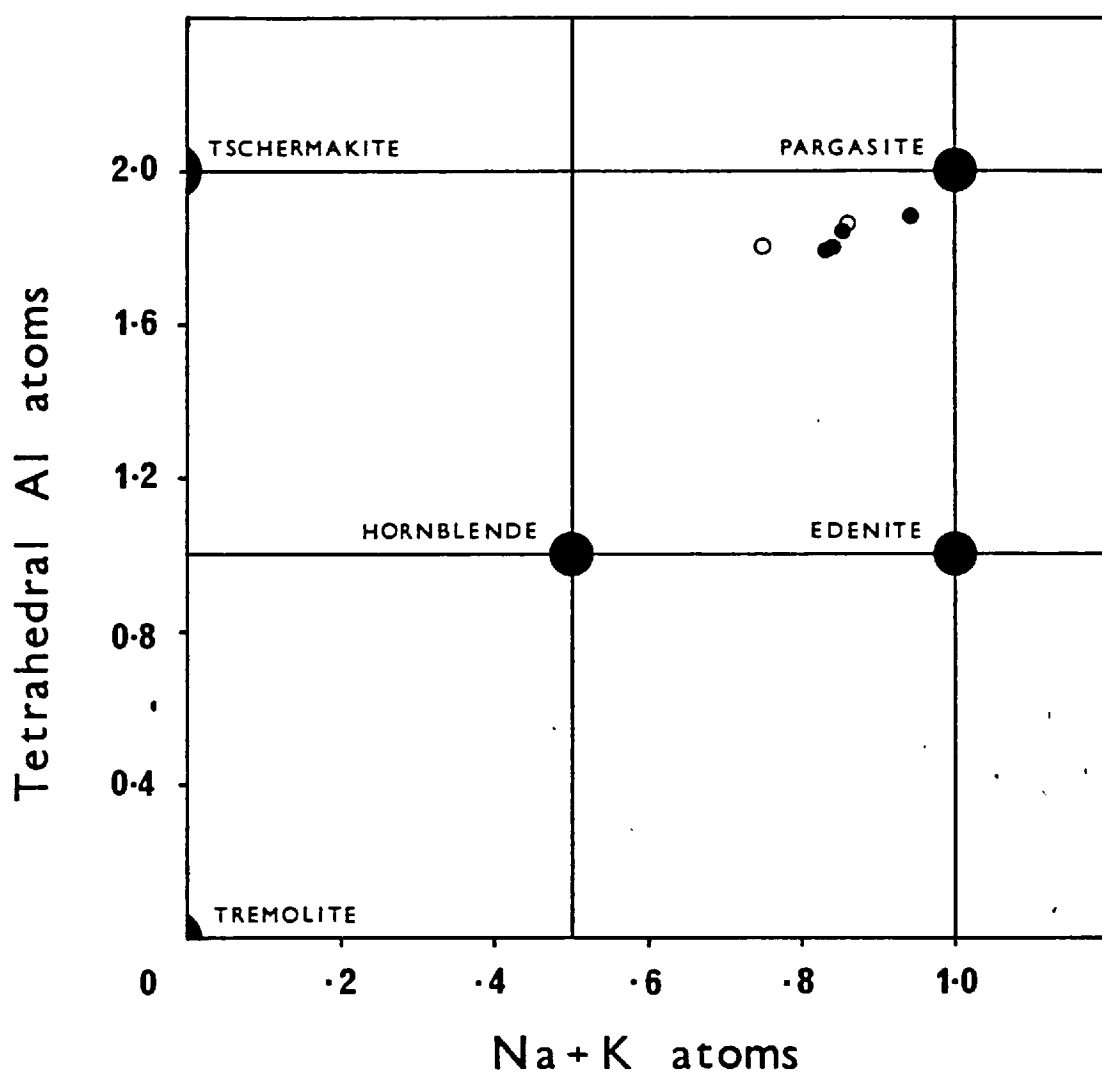


Fig. 5.15 Classification of amphiboles, after Deer et al. (1963).
Symbols are as for fig. 5.14.

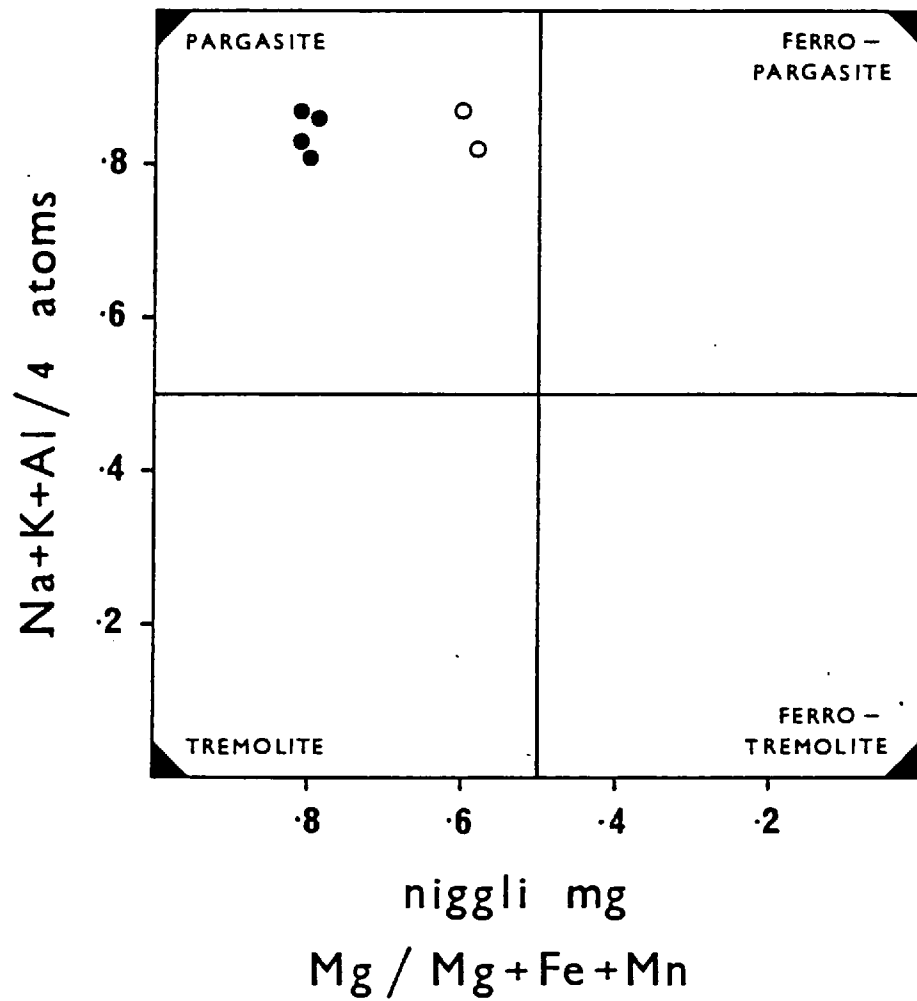


Fig. 5.16 Classification of amphiboles after Windley and Smith (1974), and Ernst (1968). Symbols are as for fig. 5.14.

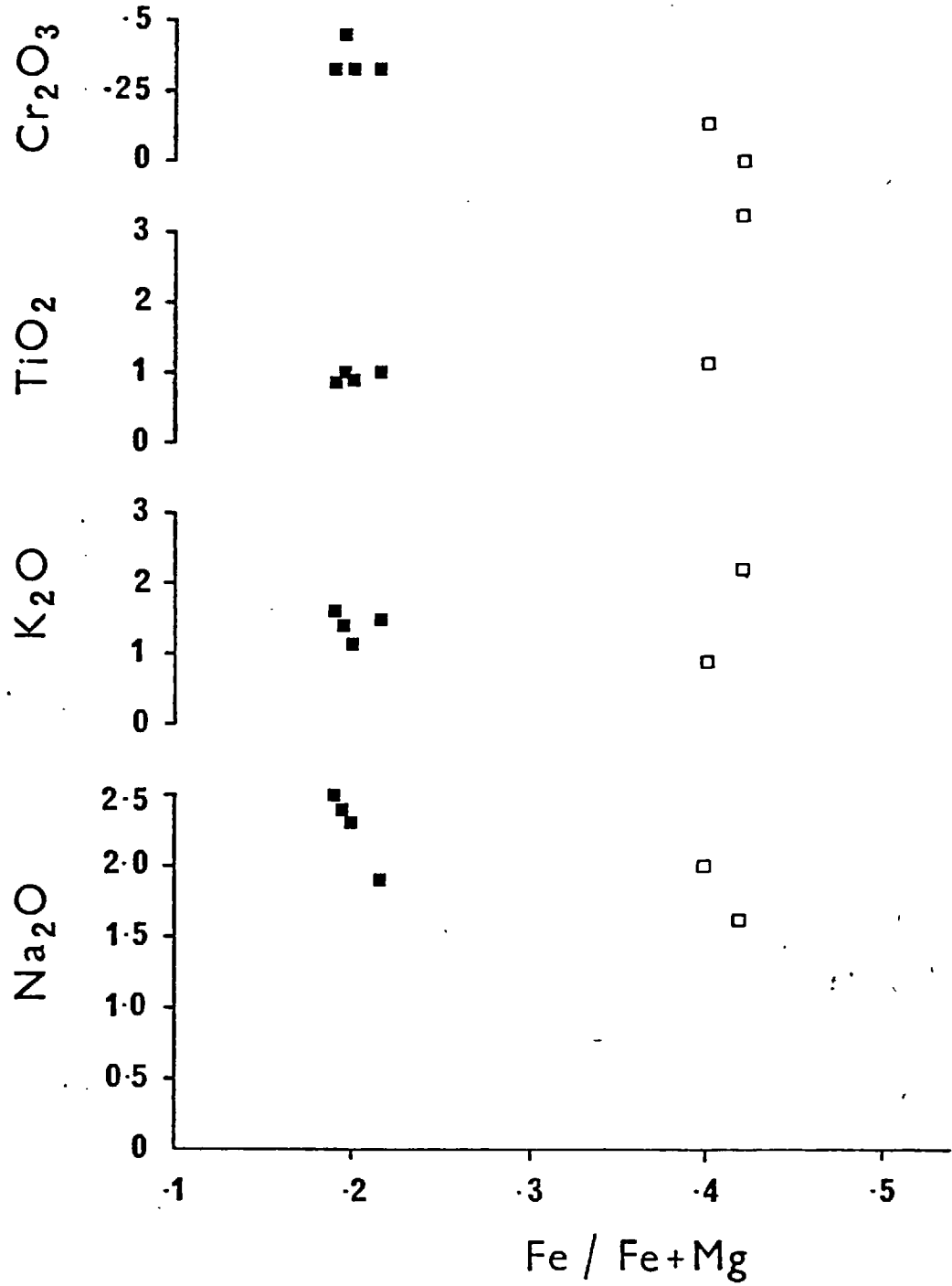


Fig. 5.17 Minor element variation versus Fe/Fe+Mg ratio of amphiboles. Filled symbols refer to an ultramafite host rock, open symbols a mafic gabbro host rock.

Cr_2O_3 contents are highest in the pargasites, in which they reach 0.4 wt.percent. TiO_2 is approximately 1 wt.percent, although one sample has a high value of 3.2 wt.percent. K_2O is high throughout, in the range 1-2 wt.percent. Considering the low concentration of potassium in the rocks as a whole, the amphiboles quite clearly control the distribution of this element. The Na_2O contents decrease from approximately 2.5 to 1.6 wt.percent with increasing $\text{Fe}/\text{Fe} + \text{Mg}$.

Raase (1974), in a survey of calciferous amphiboles from a variety of metamorphic terrains noted a correlation between octahedrally coordinated Al and pressure of metamorphism. The Camas nam Buth amphiboles are plotted on a Si atoms versus Al atoms diagram in fig. 5.18. The dashed line should separate amphiboles of low grade metamorphism (below) from those of high-grade metamorphism (above) as suggested by Raase (1974). The analysed amphiboles are thought to have crystallised under the conditions of the M1 metamorphism, at an estimated pressure of 15 kb (chapter three). This quite clearly conflicts with fig. 5.18. Obviously, this diagram is an over-simplification of the correlation of Al contents with pressure of formation. Raase also noted increasing Ti content of amphiboles with increasing metamorphic grade, and quotes a modal value of 0.27 Ti atoms per formula unit for granulite facies rocks. The Camas nam Buth amphiboles have a mean value of 0.15 Ti atoms per formula unit.

5.7 BIOTITES

Only one biotite has been analysed, from a mafic gabbro (Table 19). It occurs as a retrogressive fringe on a garnet porphyroblast. The biotite is potash-rich (9.52), with Na_2O below detection limits, also with high TiO_2 (3.14 wt.percent).

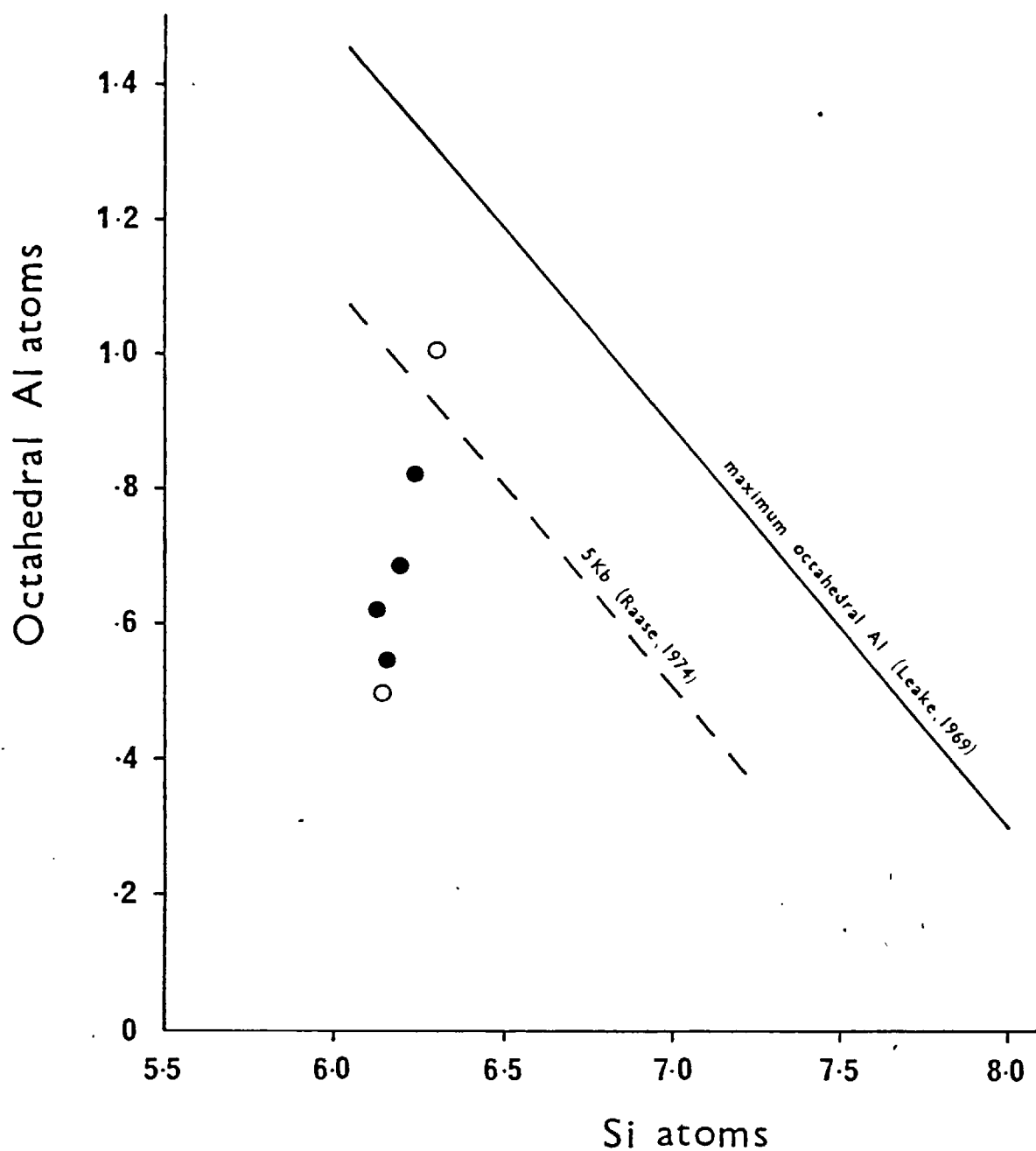


Fig. 5.18 Plot of octahedral Al atoms versus Si atoms for the Camas nam Buth amphiboles, after Raase (1974). The dashed line separates amphiboles metamorphosed at pressures greater than 5 Kb (above) and less than 5Kb (below). Symbols are as for fig. 5.14.

5.8 PLAGIOCLASES

Plagioclase compositions range from An_{48} to An_{74} with the orthoclase molecule contributing less than 1 mol.percent (fig. 5.19, Table 20). The most calcic plagioclase is found in the ultramafic rock types, apart from one sample (91). This mafic gabbro has a plagioclase composition of An_{74} , well outside the range of other mafic gabbro plagioclases (An_{48-58}). This analysis may be suspect as the feldspar in this rock is highly clouded, and the high An content may be a feature of this alteration.

Zoning of the feldspars is minimal and there may be only 1 mol.percent difference in a particular end-member between rims and cores.

Feldspar stoichiometry is such that the concentration of each element is strictly controlled by the overall composition of the mineral. Thus the An content will control the amounts of Si and Al in the feldspar. Fig. 5.20 compares the SiO_2 and Al_2O_3 concentrations for the modal feldspar with a theoretical stoichiometric feldspar of given An content. Although the silica distribution appears random about the 1:1 ratio, Al_2O_3 seems to be consistently deficient in the modal plagioclases.

FeO is higher than K_2O contents in all the plagioclases analysed, and significant FeO contents of granulite facies plagioclases have been observed by Sen (1960) and Ribbe and Smith (1966). Sen (1960) noted the increase in FeO in plagioclase on passing from amphibolite facies to granulite facies grade and attributed this to increasing temperature. Similarly, Ribbe and Smith (1966) observed a positive correlation between FeO and the An content of plagioclase. This is exhibited weakly by the Camas nam Buth plagioclases (fig. 5.21). Both sets of authors (above) suggest that Fe^{2+} substitutes as a divalent ion in Ca^{2+} sites, or as a

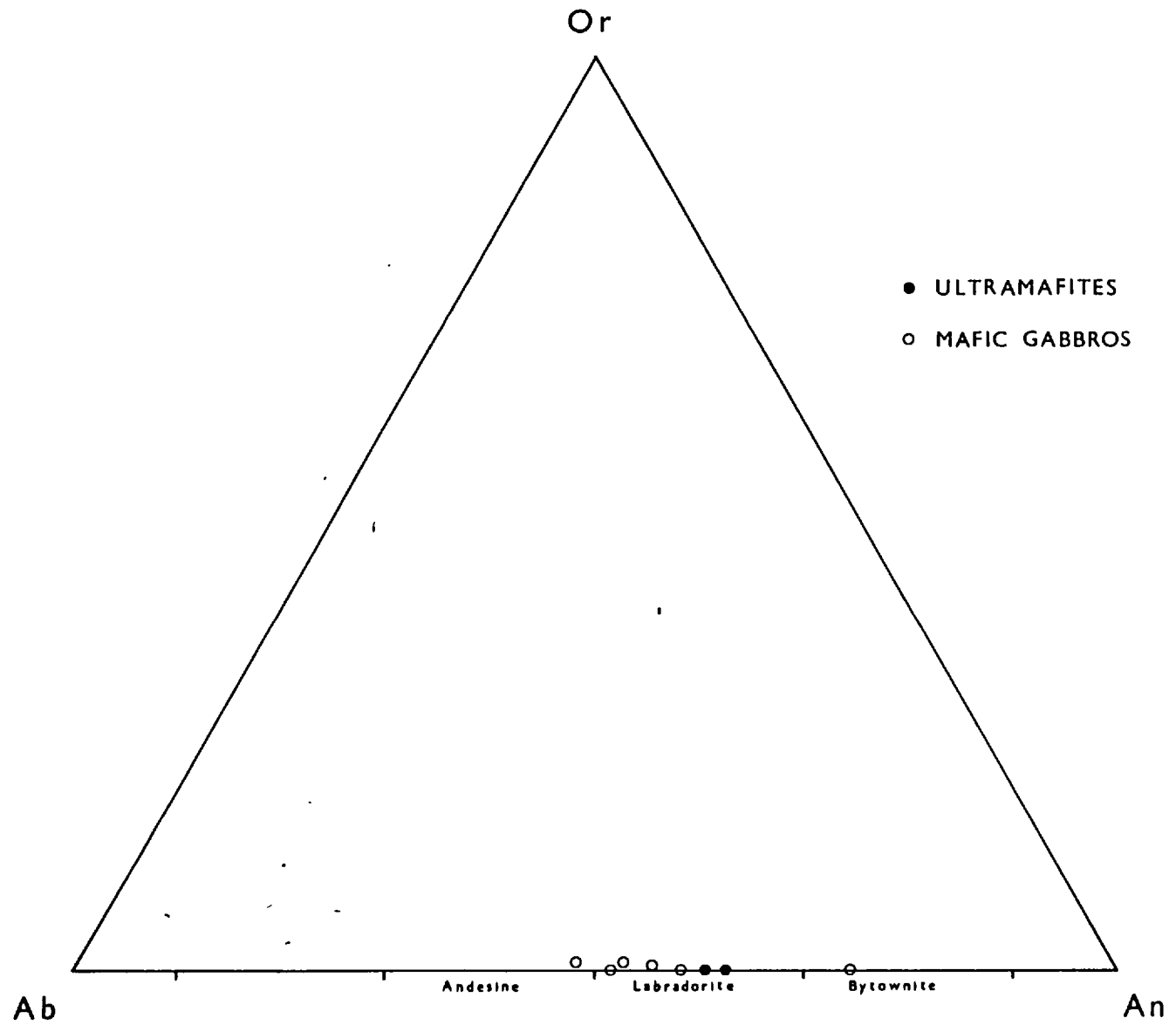


Fig. 5.19 Plagioclase compositions in terms of the albite-orthoclase-anorthite triangle.

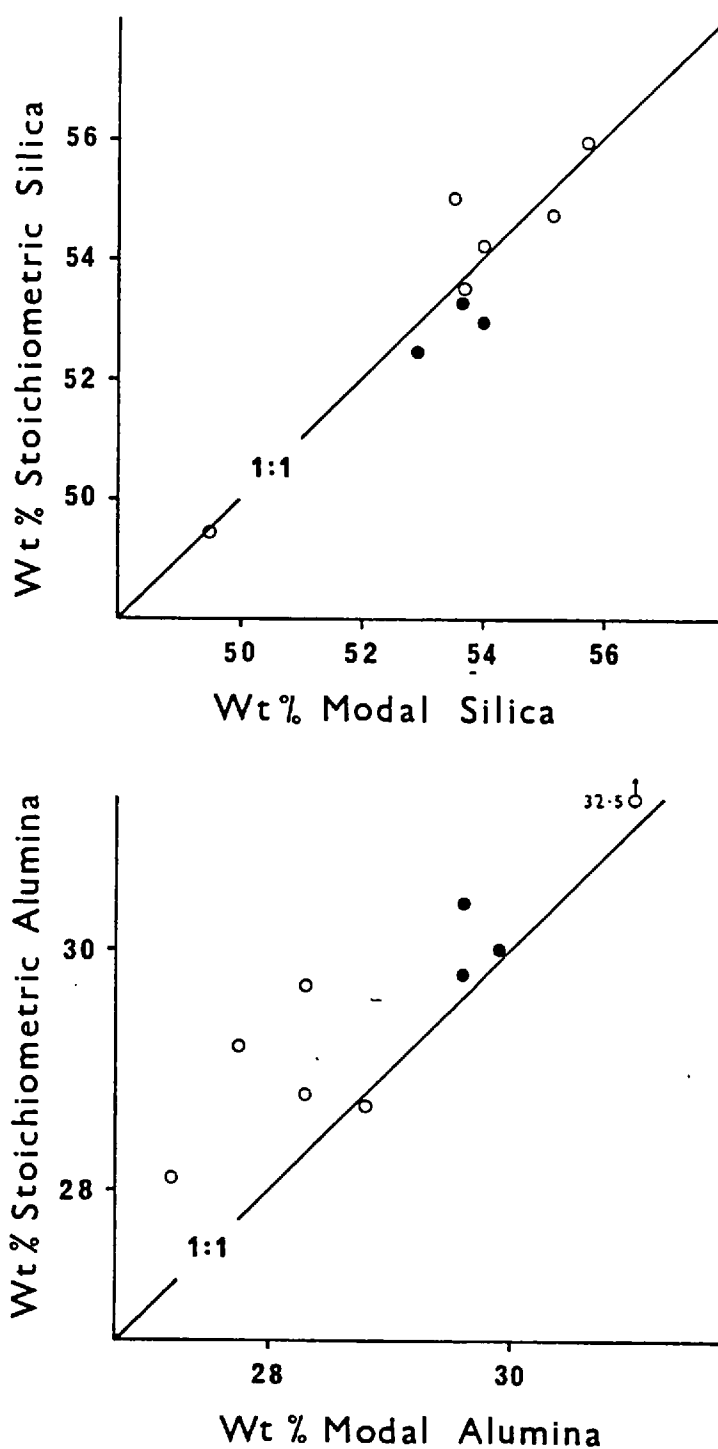


Fig. 5.20 Comparison of modal silica and alumina contents with a stoichiometric plagioclase of given An:Ab:Or ratio. Symbols are as for fig. 5.19.

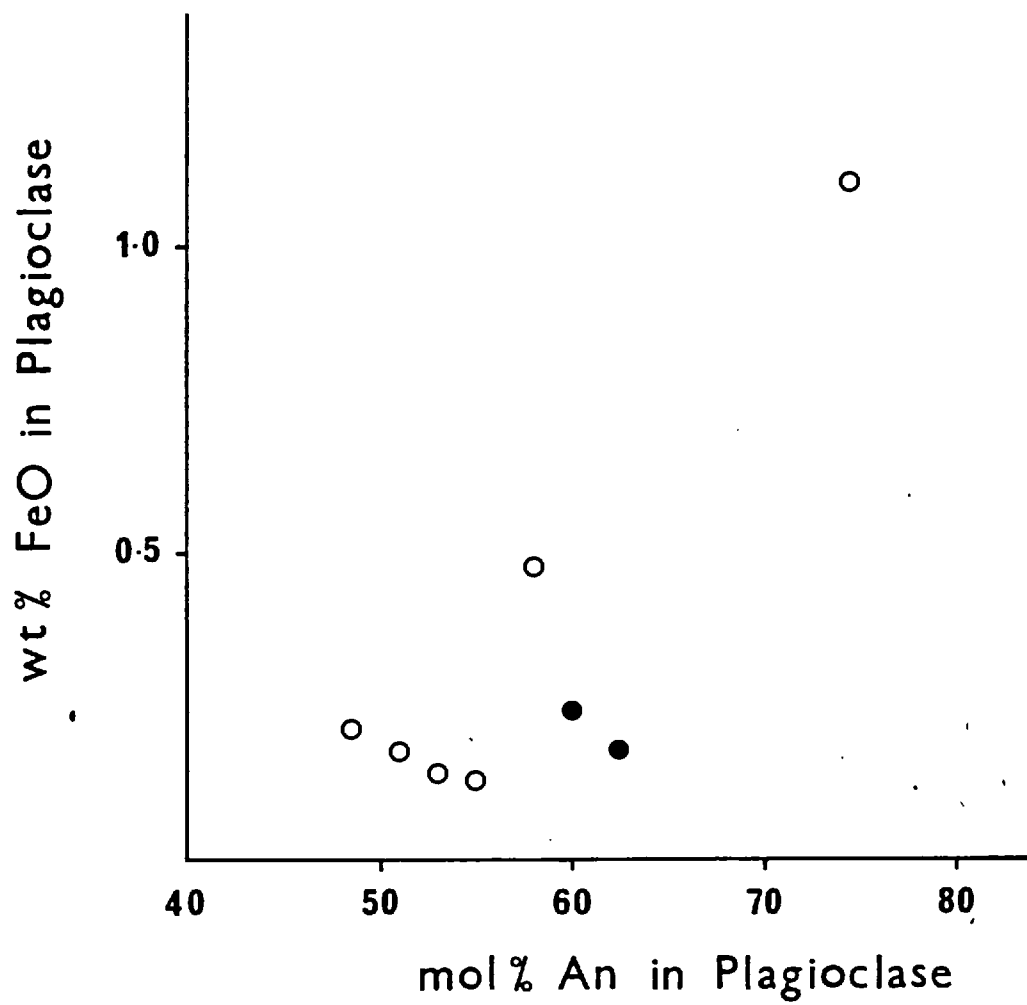


Fig. 5.21 Variation of FeO content with anorthite content of plagioclase. Symbols are as for fig. 5.19.

trivalent ion in the tetrahedral Al sites.

Many granulite feldspars in the Scourie area contain small rods which look microscopically like those in the granulite quartz grains (J. Watson, pers. comm.). The rods in quartz are conventionally believed to be rutile. TiO_2 has not been detected in the Camas nam Buth plagioclases, which indicates that the rods in the plagioclases are unlikely to be rutile.

5.9 IRON-TITANIUM OXIDES AND SULPHIDES

Six magnetites and two ilmenites have been analysed from the mafic and ultramafic rocks. (Table 21)

The magnetites from ultramafic rocks have less than 5.5 mol. percent ulvospinel in solid solution with chromite contents ranging up to 10 mol.percent. Spinel may also reach 4 mol.percent. Nickel contents may be as high as 0.63 wt.percent oxide. These magnetites have no co-existing ilmenite and appear to have formed a solid solution (from textural evidence described in section 2) with the green pleonastes at high temperatures. The significant amounts of molecules other than magnetite and ulvospinel also points to this conclusion.

The magnetites co-existing with ilmenite in the mafic gabbros have less than 1 mol.percent ulvospinel in solid solution and chromite contents of less than 0.8 mol.percent. Vanadium contents are up to 1.4 weight percent.

The ilmenites in the mafic gabbros have possible haematite contents less than 2.7 mol.percent. Vanadium concentrations may be up to 0.3 wt.percent.

Pyrrhotite, chalcopyrite and pyrite have been analysed from a sulphide-rich mafic gabbro and the analyses are presented in Table 22. Minor elements present are cobalt (up to 0.7 wt.percent),

nickel (up to 0.8 wt.percent) and zinc (up to 0.2 wt. percent).

5.10 ELEMENT DISTRIBUTION BETWEEN CO-EXISTING MINERALS

The inter-element distributions between certain co-existing ferromagnesian minerals are investigated in this section. Partitioning data may be used to estimate the degree of equilibration between the various mineral pairs.

Figs. 5.22 and 5.23 show Fe-Mg partitioning between the more important mineral phases together with published partitioning data from charnockites and granulites.

It is apparent from fig. 5.22a that orthopyroxene and clinopyroxene show probable equilibrium partitioning of Fe and Mg. Although the Camas nam Buth data fall off the general granulite trend, the curve defined by the data points parallels the general trend.

Kretz (1963) noted that the Mg-Fe distribution coefficient (K_D) between co-existing orthopyroxene and clinopyroxene appears to be constant for a given set of equilibrium conditions, where K_D is defined as :-

$$K_D \text{ (Mg-Fe)} = \frac{X_{\text{Mg}}^{\text{Opx}}}{(1 - X_{\text{Mg}}^{\text{Opx}})} \cdot \frac{(1 - X_{\text{Mg}}^{\text{CPx}})}{X_{\text{Mg}}^{\text{CPx}}}$$

$$X_{\text{Mg}}^{\text{Opx}} = \text{Mg/Mg + Fe (Opx)}$$

$$X_{\text{Mg}}^{\text{CPx}} = \text{Mg/Mg + Fe (CPx)}$$

K_D is a function of pressure and temperature of equilibration. Thus Kretz found a set of granulite facies pyroxenes to have a value of 0.54 and Skaergaard pyroxenes a value of 0.73. The Camas nam Buth pyroxenes have an average K_D of 0.8, well above both these values. This may be a function of the high temperatures ($T=900^\circ\text{C}$) and pressures ($P=14\text{Kb}$) estimated for the M_2 metamorphism (chapter three).

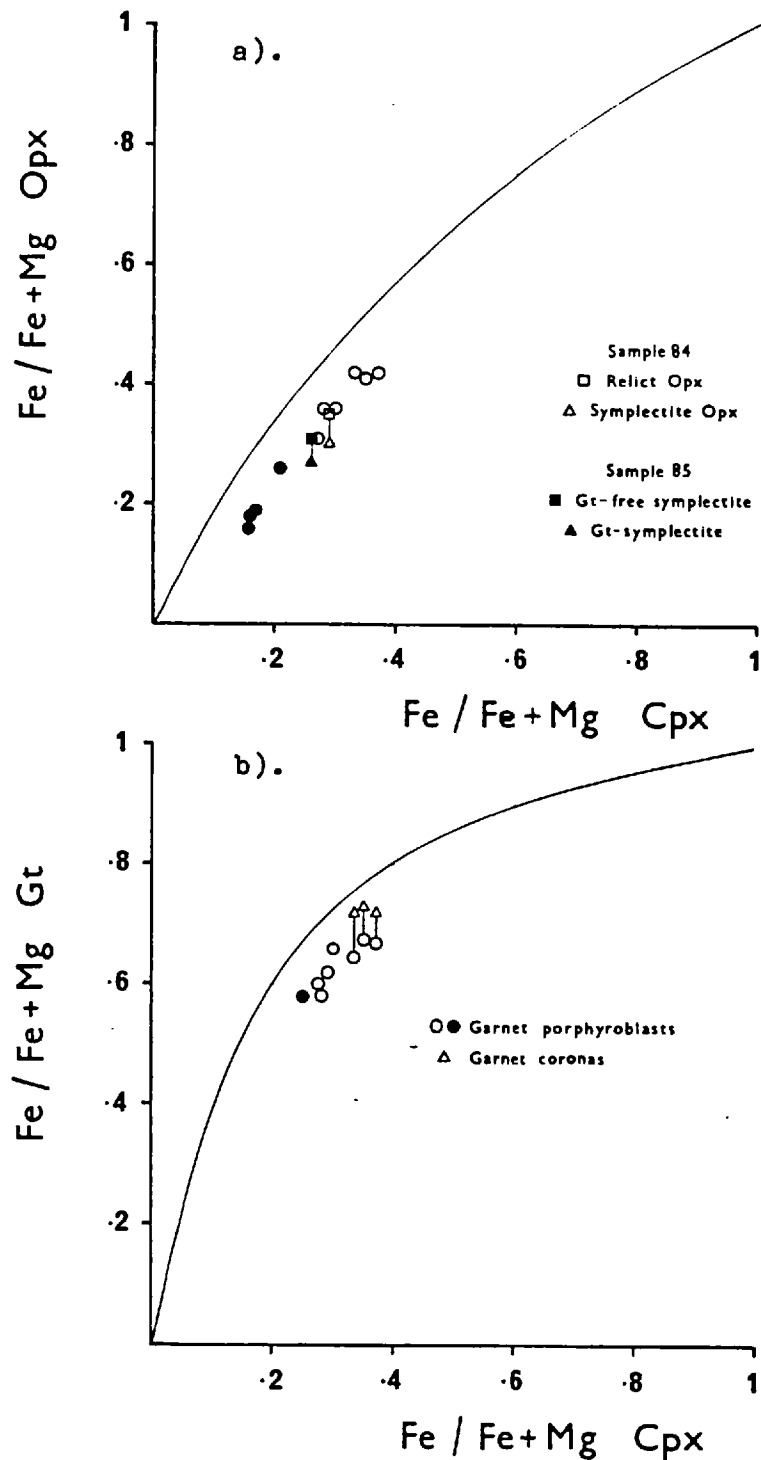


Fig. 5.22 Distribution of Fe and Mg between orthopyroxene and clinopyroxene (a), and garnet and clinopyroxene (b). The reference curves are from minerals in granulite facies rocks from the literature, after Loomis (1977). Data sources are Davidson (1968), Griffin (1971), Groves (1935), Himmelberg and Phinney (1967), Howie and Subramaniam (1957), and Saxena (1968a, 1968b; 1969).

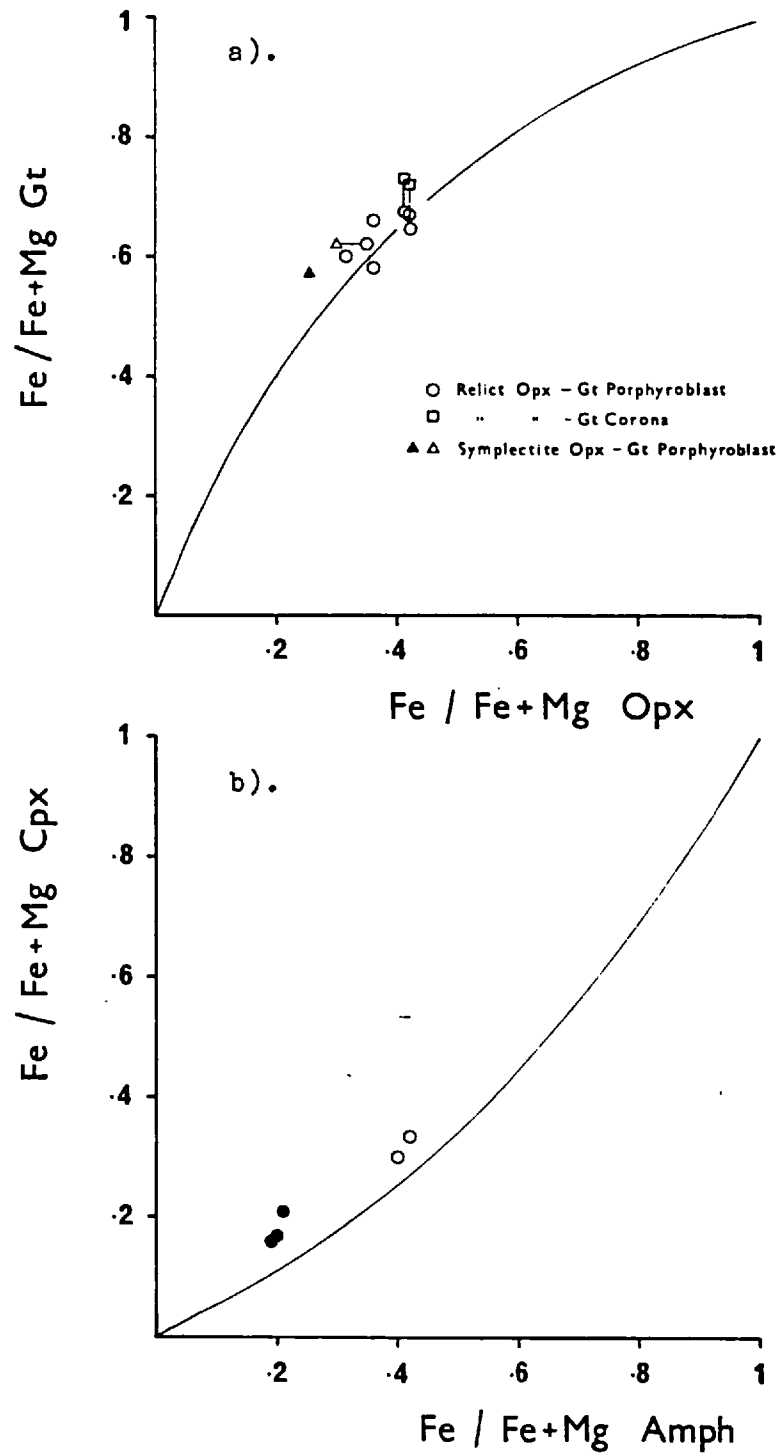


Fig. 5.23 Distribution of Fe and Mg between garnet and orthopyroxene (a), and clinopyroxene and amphibole (b). The reference data are as for fig. 5.22 and also from Onuki (1966).

Fe-Mg partitioning between garnet and clinopyroxene (fig. 5.22b) also plots off the main granulite trend. As increasing temperature tends to partition Fe more strongly into clinopyroxene (Banno, 1970; Raheim and Green, 1974), it may be seen that the Camas nam Buth data plot on the high-temperature side of the main granulite curve. Similarly, the garnet coronas appear to have formed at a lower temperature than the garnet porphyroblasts in the same rock, as outlined in chapter three.

Fe-Mg partitioning between orthopyroxene and garnet is less straightforward, although the data spread about the main granulite trend.

Fe-Mg partitioning for clinopyroxene and amphibole parallels the general trend for granulite facies rocks but falls on the low temperature side (amphibolite facies) of this curve, as suggested by Onuki (1966).

Some minor element distributions (Cr, Mn, Ti) are plotted for co-existing minerals in fig. 5.24. Although trends are erratic, the order of preference for these elements is:-

Cr	spinel	>>	amphibole	>	clinopyroxene
Mn	garnet	>>	orthopyroxene	>	clinopyroxene
Ti	amphibole	>>	clinopyroxene	>	orthopyroxene

5.11 SUMMARY

The salient points arising from the description of the mineral chemistry are summarised below:-

(1) Pyroxenes show iron-enrichment trends, broadly similar to those of igneous layered intrusions, although controlled by host-rock $\text{FeO}/\text{FeO} + \text{MgO}$. The high alumina contents are concentrated as jadeite and ca-tschermak molecules, being particularly high in the mafic gabbros with low $\text{FeO}/\text{FeO} + \text{MgO}$ ratios. The jadeite/ca-tschermak

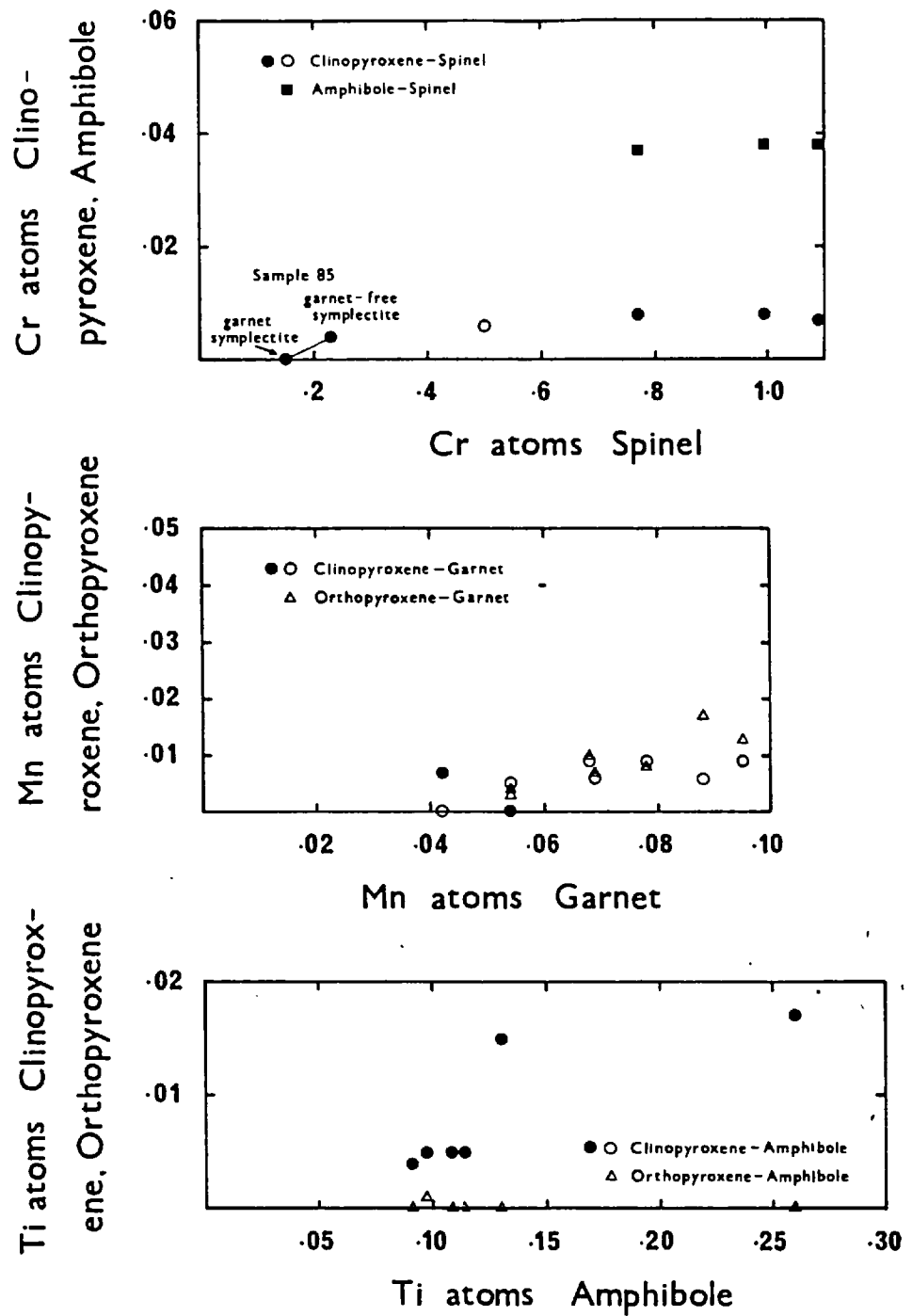


Fig. 5.24 The relation between Cr, Mn, and Ti contents of co-existing garnets, spinels, amphiboles, clinopyroxenes, and orthopyroxenes. Filled symbols refer to an ultramafite host rock, open symbols a mafic gabbro host rock.

ratios are still typical of granulite facies rocks, rather than eclogites (chapter three). Pyroxene Mg-Fe K_D 's are abnormally high and may be a consequence of high pressure and temperature during metamorphism.

(2) Olivines have high nickel contents, and are the only phase with any appreciable amount of this element in the ultramafites. It was suggested in chapter four that early crystallising olivines may have controlled nickel distribution during the igneous evolution of the Camas nam Buth rocks. The metamorphic olivines may have crystallised on the sites of these original igneous olivines.

(3) Garnet porphyroblasts are almandine-rich, and garnet coronas have even higher FeO/FeO+MgO ratios. Fe-Mg K_D 's between garnet and clinopyroxene suggest that the Scourie granulites formed under higher temperature conditions than are usual for granulite terrains.

(4) Spinel has low Cr/Cr+Al ratios, typical of spinels from metamorphosed basaltic rocks. Textural evidence (chapter two) also supports the inference that these equilibrated under metamorphic conditions.

(5) Amphiboles which appear, from petrographic evidence (chapter two), to be part of the initial metamorphic assemblage are pergasitic. The scarcity of the amphiboles in the mafic gabbros and abundance in the ultramafites is taken to indicate a stability dependence on host rock composition, rather than any spatial variation in PH_2O .

(6) Plagioclases are labradoritic or bytownitic in composition generally depending on host-rock composition. The appreciable FeO contents and low K_2O contents are typical of feldspars in basic granulites.

(7) Little solid solution between magnetite-ulvospinel and ilmenite-hematite is shown by the oxide phases, and this is probably a function of slow cooling rate and accompanying re-equilibration of compositions.

CHAPTER 6

CONCLUSIONS AND DISCUSSION

It is not intended to use this chapter to summarise the other five. It is hoped that the salient features have been summarised well enough already, at the relevant stage of each chapter. Instead, issues raised by this work, of significance to the geology of the Lewisian Complex, and others of wider significance will be discussed.

6.1 THE ORIGINAL NATURE OF THE LAYERED MAFIC-ULTRAMAFIC SHEETS

The evidence presented in this thesis suggests that the layered mafic-ultramafic sheets observed around Scourie were once parts of a stratified igneous sill or sheet, differentiated at depth, and that the currently observed lithological layering is essentially an igneous layering which has suffered little chemical modification. Implicit in this statement is the inference that the ultramafites were related to the mafic gabbros by a process of fractional crystallisation, albeit that this was by a two-stage process, and that the mafic gabbros were originally igneous rocks. This inference is in marked contrast to suggestions made by O'Hara that the sheets represent original igneous peridotites surrounded by aureoles of mafic gneiss (mafic gabbros of this work) developed by diffusional mass transfer between the peridotites and the acid gneisses (O'Hara, 1961a, 1965, 1966). It has been shown that the major and trace element compositions of the mafic gabbros closely resemble those of modern basaltic rocks. It is highly unlikely that these basaltic compositions could have been derived by diffusional mass transfer of components from the peridotites into the acid gneisses (Bowdidge, 1969). Furthermore, the layered sheets have clearly defined physical and chemical boundaries

against the acid gneisses. In his discussion, O'Hara highlighted the distinct chemical compositions of the ultramafites and the mafic gabbros, and the small thicknesses of the rock units compared with those usually recorded where similar units are formed by fractional crystallisation of a common magma. These are important points, but it is difficult to account for the observed sequence of layers - ultramafite, mafic gabbro, sulphide-rich mafic gabbro, the geochemical differentiation in each cyclic unit, and the almost ubiquitous occurrence of the two rock types together in all sheets observed in the Scourie-Laxford region without invoking a cosanguineous origin for the two rock types in a layered igneous intrusion. Accepting this, it is necessary to account for O'Hara's observations. Firstly, the ultramafites and mafic gabbros must have behaved as separate systems during metamorphism, with elemental re-distribution taking place within each unit to maintain the separate compositions, but without migration from ultramafite to mafic gabbro. The second point may be resolved by noting that the layered sheets need not have been intrusive into the acid gneisses, but that the converse relationship is more likely considering the evidence produced from other high-grade gneiss terrains of the world, particularly West Greenland. For example, Myers, among others, has observed granitic sheets intruding and disrupting amphibolites and stratified anorthosites in the Fiskenaesset region (Myers, 1976). Typically, the amphibolites and anorthosites have been subsequently veined and fragmented by younger granite sheets which, when followed along strike, show increasing amounts of veining and dilation. This intrusive origin for granitic gneisses in Greenland was first established by Macgregor (1973) at Godthaab, and a similar mechanism has been proposed for the Scourie gneisses by Davies (1975). Intrusion of large volumes of acid magma would cause disruption of the layered

intrusions, which, coupled with tectonic thinning associated with the two isoclinal fold phases would produce the observed distribution of rock types.

Thus, in conclusion, it is believed that the best explanation of the origin of the layered mafic-ultramafic sheets is as a layered igneous sheet which has been subsequently modified by metamorphism, deformation and intrusion.

This reinforces much of the argument put forward by Bowes, Wright and Park (1964, 1966) that the layered sheets are derived from igneous intrusions. However, a major difference from Bowes et al.'s results lies in the chemical affinity of the layered sheets. The Camas nam Buth rocks are clearly tholeiitic, whilst the rocks analysed by Bowes et al. (1966) from Scourie-Achiltibuie are clearly calc-alkaline (fig. 4.1). Although some variation is feasible along the FeO-MgO side of the AFM diagram, it is difficult to explain why one set of data shows no alkali-enrichment, the other extreme alkali-enrichment. As Bowes et al. only published four of their thirty-seven analyses represented in the field of fig. 4.1, it is difficult to comment on their data. The only rocks analysed by the author from the Scourie layered sheets which show such alkali-enrichment are the granite sheets. Bowes et al. do not mention these granitic rock types in any of their descriptions of the layered masses. Occasionally, the contacts between the granites and the mafic gabbros are diffuse (e.g. Geodh' Eanruig), giving rise to 'hybrid' rocks. These plot towards the alkali apex on an AFM diagram. Without any other suitable explanation, it is tentatively suggested that Bowes et al. sampled these rocks for analysis, producing the alkali-enrichment in the AFM diagram.

6.2 LEWISIAN CHRONOLOGY

A few words are necessary concerning the role of the layered mafic-ultramafic sheets in the framework of the geological history of the Lewisian Complex.

Age relationships between the layered sheets and the acid gneisses and granulites of the Lewisian are uncertain. For example, it has been suggested that the layered sheets were intruded into an older Pre-Scourian complex of gneisses, and later disrupted by younger invading tonalites and granodiorites, all subsequently metamorphosed in the Scourian (Watson, 1975; Davies, 1975), or that the layered sheets were intruded into a supracrustal sequence of andesitic, dacitic and basaltic lavas and volcanoclastic sediments, all derived during the Scourian, representing juvenile additions from the mantle (Bowes, 1976). Evidence from this present work and elsewhere pertinent to this discussion is :-

1). The layered sheets pre-date the acid gneisses and granulites surrounding them. Although clear intrusive relationships are nowhere visible for the acid gneisses, it is believed that the disruption and fragmentation of the layered sheets was caused by the invasion of younger granitic magmas as is envisaged to have happened between the layered Fiskeraasset intrusion and the granitic Kuk gneisses of W. Greenland. It is difficult to see how tectonic processes alone could have caused the fragmentation of the layered sheets.

2). Strontium and lead isotopic evidence indicates that the Scourian granulite parents had a short history between derivation from the upper mantle (probably by a two-stage melting process) and subsequent conversion to a gneiss complex, probably in the period 3100-2900 m.y. ago (Moorbath, 1975). Proposed older Pre-Scourian gneisses (Davies, 1975) show the same isotopic

characteristics and ages as the bulk of the complex (Chapman and Moorbath, 1977).

With the possible exception of metasedimentary lithologies (Okeke, 1978), it is suggested that the layered sheets are the oldest components of the Scourian Complex, although the age gap between these and the other lithological units may be small (within the 200 m.y. quoted above). It is envisaged that the early Scourian crust was essentially basaltic, the layered sheets representing the deeper portions of this crust. The later acid gneisses (essentially tonalites and granodiorites) are partial melts of this subducted basaltic material, and probably much basaltic material was consumed in this way. As the exposed portion of the Lewisian is not predominantly basaltic, very large volumes of granitic material would have had to have been introduced to explain the observed proportions of lithological types. This is not unreasonable when one considers that the period 2,900-2,700 m.y. was a considerable 'event' in the production of sialic crust (Moorbath, 1973).

It will be outlined below (section 6.4) that the layered sheets of the mainland Lewisian strongly resemble the mafic gabbros rimming the Reneval anorthosite of South Harris (Witty, 1975) in terms of structural, physical and chemical features. It is believed that the mafic gabbros of South Harris are part of the same suite of layered sheets occurring on the mainland Lewisian. Although the mainland lacks any major anorthosite occurrence, this is not crucial to the point in question. It is therefore suggested that these intrusions may be used as time-markers between the mainland and hebridean Lewisian, in much the same way as the Scourie Dykes have been used to separate the Scourian and Laxfordian episodes (Sutton and Watson, 1951).

6.3 METAMORPHISM AND THERMAL GRADIENTS

The evidence presented in chapter three suggests that the Scourie area has experienced three episodes of granulite facies metamorphism, the highest grade being almost eclogitic (15 kb, 1000°C). This estimate is still much lower than that proposed by O'Hara for the area, a figure of 15 kb and 1250-1300°C (O'Hara, 1977a, 1977b; O'Hara and Yarwood, 1978). This higher temperature estimate is based on four pieces of evidence :-

1). The presence of ferrohypersthene lamellae in an augite host in a quartz - ferrohypersthene - augite - magnetite gneiss, has been assumed to indicate the former presence of ferro-pigeonite in the rock. The ferrohypersthene lamellae contain lamellae of augite, presumably exsolved from the suggested original ferro-pigeonite (O'Hara and Yarwood, 1978). Similar pyroxenes appear to be stable at 1040°C at atmospheric pressure.

2). The plagioclase-hypersthene coronas around garnet in the mafic gabbros are taken to be exsolved from a high temperature pyroxene. O'Hara's experimental work (O'Hara, 1977a; O'Hara and Yarwood, 1978) on separated pyroxenes indicates that temperatures of 1100°C at 15 kb are necessary to dissolve the plagioclase, hypersthene and iron ore into the clinopyroxenes.

3). Intergrowths of feldspars in the granite sheets would indicate the original co-existence of a calcic potash feldspar ($\text{Ca}_{19}\text{Na}_{54}\text{K}_{27}$) and a plagioclase ($\text{Ca}_{24}\text{Na}_{70}\text{K}_6$) which yield temperatures of 1145°C using a modified version of the Powell and Powell feldspar geothermometer (Powell and Powell, 1977a).

4). Dry melting experiments on the granite sheets indicate temperatures in excess of 1200°C for fusion.

Points 2). and 4). have previously been discussed and rejected in chapter three. Evidence from this thesis suggests

that the coronas developed as a consequence of coupled garnet-clinopyroxene reaction rather than simple exsolution from clinopyroxene and that the granite sheets contained appreciable amounts of dissolved H_2O during fusion, thus obviating the need for temperatures in excess of $1000^{\circ}C$. Points 1). and 3). are harder to comment upon, due to the lack of any occurrence of these particular textures in the suite of rocks studied by the author. The credibility of either of these temperature estimates is based on the assumptions made in re-tracing the various exsolution stages from a supposed original high temperature mineral assemblage.

The author therefore has no reason to suggest that the M_1 metamorphism took place at temperatures appreciably greater than $1000^{\circ}C$.

The M_2 P-T estimate of this work ($900^{\circ}C$, 14 kb) is in good agreement with the estimate for the metamorphism of the South Harris Complex, derived by Wood (1977) of $825^{\circ}C$ and 13 kb. The metamorphism of the South Harris Complex is believed to be Scourian in age (Witty, 1975), thus there is a good correlation between the Scourie and South Harris areas in terms of metamorphic grade, rock types present and early structural sequence (see below).

These P-T estimates imply a geothermal gradient of 20° - $30^{\circ}C/km$ depth (chapter three), which is typical for Archaean granulite facies terrains (Watson, 1978). However, the Scourie and South Harris granulites are atypical in depths of metamorphism which is of the order of 40-50 kilometres. Although this may have been just a local thickening of the Archaean crust, the geochemical evidence suggesting metamorphism in an island-arc type environment (chapter four; Witty, 1975; Horsley, 1978) also

points to an Archaean crust thick enough for subduction in a proto-plate tectonic situation.

The metamorphic and geochemical data also afford some evidence on the nature of the granulite facies metamorphism. In terms of elements such as U, Th, Rb, Cs, and K the Scourian granulite facies metamorphism was clearly of an open system type, these elements being extremely mobile, due to the lack of stable minerals at this grade of metamorphism capable of accommodating these ions in their structure (micas, amphiboles). However the rock geochemistry suggests that other elements were relatively immobile, re-distributions occurring only on the scale of 1-2 cms. The occurrence of different textural types of orthopyroxenes with correspondingly different compositions in mafic gabbro coronas only millimetres in diameter is also consistent with this conclusion.

6.4 COMPARISON WITH OTHER ARCHAEOAN META-IGNEOUS ROCKS

Layered basic, ultrabasic or anorthositic sheets are a common feature of the 'high-grade' regions and the greenstone belts of the Archaean, and have been described from most of the Archaean provinces of the world. These, naturally vary considerably in thickness, lithological type and metamorphic grade. It would be impossible to compare the Scourie layered sheets with all these different types of metamorphosed intrusions, so the more important conclusions derived from this work will be examined in the light of evidence from one or two other selected intrusions.

As mentioned previously, the Scourie layered sheets show a great similarity to the components of the 'gabbroic rim' surrounding the anorthosite of South Harris. This has been studied in detail by Witty (1975). He observed the following :-

1). The sheets are layered, defined by alternations of mafic gabbros and ultramafites. The layering is essentially an unmodified igneous banding pre-dating the main regional S_1 foliation.

2). The intrusions differentiated under gravity to produce a layered sequence.

3). The rocks were metamorphosed at high pressure granulite facies grade.

4). Although interpreted by Witty as being a nepheline-normative alkalic suite, the bulk of the rocks are not undersaturated and are more akin to an alkali-poor tholeiitic suite.

5). The basaltic magma is interpreted as being generated in an island-arc type environment.

Although the anorthosite post-dates the gabbros in South Harris, there is the observed trend towards high-alumina basaltic liquids being generated at high pressures as suggested by Green and Ringwood (1967b). So although the mainland Lewisian shows no major anorthosite intrusion, the Camas nam Buth sheet shows the same high-pressure fractionation of basaltic magma.

The Fiskensæset Complex, W. Greenland is a much larger Archaean layered intrusion, being 200-1000 m. in thickness (Windley and Smith, 1974). This intrusion is dominated by anorthositic rock types, the basaltic and ultramafic components being of relatively minor importance. Deformation and metamorphism of the sheet is less pervasive than that of the Lewisian layered intrusions, so cumulate structures and relict igneous mineralogy are occasionally preserved. Layering is well developed and rhythmic units provide a means of determining way-upness. The parental magma is interpreted as being a high-alumina basaltic

liquid and the differentiation is interpreted as being controlled by amphibole and plagioclase. Windley (1976) has suggested that these rocks may represent the refractory residue of a basaltic magma parental to the calc-alkaline acid gneisses invading the complex. Although the Camas nam Ruth sheet has no calc-alkaline affinities, there are obvious similarities to the Fiskenaesset Complex in terms of the implied tectonic environment of the intrusions, at destructive plate boundaries. Also the indications are that the magmas differentiated at high pressures and that metamorphism was penecontemporaneous.

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APPENDIX IROCK ANALYSIS: TECHNIQUES AND RESULTS1. SAMPLE PREPARATION

Weathered portions of samples were removed using a mechanical splitter, and the samples cut into approximately one-inch cubes. These fragments were jaw-crushed to fourteen mesh. A representative sample was taken and ground in a 'Tema' mill (tungsten carbide mortar) to approximately two hundred mesh.

2. WET-CHEMICAL ANALYSIS

Na_2O and FeO determinations were made using flame photometry and titration with potassium dichromate solution, respectively.

3. XRF ANALYTICAL TECHNIQUES

Approximately seven grains of rock powder were oven-dried (110°C) overnight and re-weighed to determine H_2O^- . This was then ignited at 900°C for thirty minutes, and re-weighed to determine volatile loss (loss on ignition). The ignited sample and lithium tetraborate flux were accurately weighed in a sample: flux ratio of 1:7. The mixture was fused at 1150°C in a platinum crucible to produce a glass bead. The glass bead was ground to four hundred mesh-size in a 'Tema' mill. The resulting powder was pressed into a briquette using a polyvinyl alcohol binder. These briquettes were prepared in duplicate for major-element analysis. Trace-element analysis was performed on briquettes of five grams of ignited rock powder, bound with polyvinyl alcohol and a boric acid backing.

Major elements (SiO_2 , TiO_2 , Total Fe_2O_3 , MnO , MgO , CaO , K_2O , P_2O_5) and trace elements (Cr , Ni , Cu , Zn , Zr , Rb , Sr , Th , Y , Ba , V) were analysed on a Philips PW 1212 automatic X-ray fluorescence spectrometer, utilising a flexible calibration line technique based

on data for international rock standards.

Further details of the XRF technique employed at Imperial College may be found in a technical report written by R.J. Parker (Parker, 1977).

4. INSTRUMENTAL NEUTRON ACTIVATION ANALYTICAL TECHNIQUES

0.1 grams of rock powder were heat-sealed in a polythene capsule. Eight sample capsules and three standard capsules (international rock standards or REE solution standards) were sealed into a polythene tube. The tubes were irradiated at the University of London nuclear reactor at Ascot for approximately thirty-two hours at a neutron flux of $10^{12} \text{ n cms}^{-2} \text{ sec}^{-1}$. Time was allowed for short-lived ($t_{1/2} < \text{one day}$) nuclides to decay before analysis by high resolution gamma ray spectroscopy using a Ge(Li) detector for La and Sc, and an intrinsic Ge detector for Ce, Nd, Sm, Eu, Gd, Tb, Ho, Yb and Lu. Counting was done on an automatic sample changer, using 'Link Systems' equipment consisting of a 4000 channel analyser and a 'Nova' mini-computer. Conditions of analysis are summarised in Table 1.

5. DATA PROCESSING

Raw major-element XRF data were processed using programs similar to SORT, REORD and MW (Parker and Willis, 1977) and raw trace-element XRF data processed using programs TRANS and NUTRACE (written by R.J. Parker). CIPW norms were calculated with the aid of programs LEMNORM (written by R.W. Le Maitre) and AGNORM (University of California, Berkely). Some of the XRF data were plotted using an interactive graphics computing package, JPLOT (written by R.J. Parker).

The REE data were corrected for line interferences by hand and the absolute concentrations calculated by program RIALTO (University of London Reactor Centre). Corrections were made for

<u>NUCLIDE</u>	<u>HALF-LIFE</u>	<u>ENERGY (KeV)</u>	<u>COUNT-TIME</u>
¹⁴⁰ La	40.2 hours	1595.0	600 secs
¹⁴¹ Ce	33 days	145.6	3600 secs
¹⁴⁷ Nd	11.1 days	91.4	"
¹⁵² Eu	12 days	121.9	"
¹⁵³ Sm	47 hours	103.1	"
¹⁵³ Gd	242 days	97.7	"
¹⁶⁰ Tb	72.1 days	87.0	"
¹⁶⁶ Ho	29.9 hours	80.7	"
¹⁶⁹ Yb	31.8 days	63.4	"
¹⁷⁷ Lu	6.7 days	208.4	"
⁴⁶ Sc		889.4	600 secs

Table 1 Details of nuclides, half-lives, energies, and count-times used in the instrumental neutron activation analysis.

variation in neutron flux across sample tubes and radioactive decay during counting.

6. STATISTICS

42 quantitative and 3 qualitative XRF analyses are presented in Tables 5-11. Of the quantitative analyses 39 oxide totals lie in the range 99.00 - 101.00. Those outside this range are samples 90 (oxide total 98.78, but greater than 99.00 if Cr_2O_3 and NiO are taken into account), 81 (oxide total 101.10), and 94 (oxide total 101.25). An idea of the precision of the XRF major element data may be gained from the analytical results of the seven preparations of one sample, summarised in Table 2. This is far from the ideal size of a statistical sample ($n > 50$), but gives some indication of the quality of the XRF data. The silica range (47.96 - 49.57 wt. percent) is rather large at this concentration level but is probably due to the large range of oxide totals for this particular sample (99.0 - 102.0). Detection limits of trace elements and determined trace element concentrations of a number of international rock standards are presented in Table 3. Table 4 gives an estimate of the average relative errors of rare-earth element determinations.

<u>ELEMENT</u>	<u>RANGE</u>	<u>MEAN</u>	<u>STANDARD DEVIATION</u>
SiO ₂	47.96-49.57	48.65	0.59
TiO ₂	0.52-0.55	0.54	0.01
Al ₂ O ₃	13.95-14.15	14.04	0.06
Fe ₂ O ₃	2.15-2.42	2.33	0.10
MnO	0.31-0.33	0.32	0.01
MgO	9.66-10.37	9.93	0.28
CaO	13.64-13.87	13.76	0.09
K ₂ O	0.12-0.18	0.16	0.02
P ₂ O ₅	0.03-0.06	0.05	0.01

Table 2 Variability of analyses of seven preparations of one mafic gabbro sample.

	Cr (1)	Mn (2)	Cu (2)	Zn (3)	Zr (3)	Rb (1)	Sr (1)	Th (2)	Y (2)	Ba (23)	V (7)
W-1	114 132 120	76 76 73	110 12 110	86 80 86	105 95 105	21 20 21	190 189 190	2 3 2	25 24 25	160 173 160	265 265 240
BCR	18 16 16	16 13 13	18 19 19	120 120 120	191 187 186	47 48 46	331 329 331	6 3 6	37 41 46	677 704 682	400 401 411
AGV	12 13 12	19 18 17	60 56 64	85 83 85	227 216 222	68 68 68	662 663 665	6 6 6	21 22 26	1218 1210 1210	126 127 126
GSP	13 9 13	13 35	34 32 99	99 99 506	506 512 253	257 257 233	236 237 233	81 32	31 32 1314	1314 1313 53 57	50
G-1	20 39 20	25 13	13 13 45	45 50 211	211 230 221	221 213 251	251 255 251	47 13	13 15 1205	1205 980 17 20	17
G-2	7 13 9	7 11	12 14 86	86 86 302	302 321 170	482 475 483	24 23 12 10	1883 1837 1863	36 43 34		
PCC	2864 2956 3158	2667 12	12 9 38	38 47 7	0 0	0 0	2 2	0 0	0 2	0 10 9	31 26 33
DTE	4008 3718 4388	2564 7	7 5 45	45 46 0	0 0	0 0	0 1	0 2	0 110	2 0 225 165	10 9 220
NIM-N	40 128 34	78 114 120	13 11 13	80 57 25	25 3 2	259 4	7 6 90 33	225 165 220			
NIM-P	29900 8167 24700	470 603 568	17 17 14	100 102 100	20 12 20	2 37	4 3	56 14 224 213	249		
NIM-S	13 21 12	8 9 8	23 21 19	21 12 20	30 13 30	531 63	1 <5	6 2452	10		
NIM-L	21 12 15	11 4 10	15 20 15	320 409 287	0 188 189	4472 205	71 51 31 17	451 386 78 75			
NIM-G	12 17 15	6 13	15 7 60	60 48 48	275 331 341	11 62 44	100 179 180 72	2 1 2			
NIM-D	2880 2952	2114	8 9 90 84	6 0	3 0	0 3	0 0	0 65 42	29		

Table 3 Trace-element values determined in this work for various international rock standards as compared with values suggested by Abbey(1973), and Flanagan (1973). For each standard the data are laid out as follows:-

FLANAGAN SAVAGE
ABBEY

<u>ELEMENT</u>	<u>RELATIVE ERROR (%)</u>
La	± 2
Yb	± 5
Ho	± 4
Tb	± 15
Nd	± 1
Gd	± 10
Sm	± 1
Eu	± 5
Ce	± 15
Lu	± 15

Table 4 Precision of rare-earth element determinations.

APPENDIX II

MINERAL ANALYSIS: TECHNIQUES AND RESULTS

The bulk of the mineral analyses (approximately 80%) were performed on an energy-dispersive microprobe in the Department of Mineralogy and Petrology, Cambridge, although a few were obtained on an automated Cambridge Instruments 'Geoscan' at Imperial College. Of the fifty-four analyses of pyroxenes, olivines, garnets and plagioclases, only four analyses lie outside the range 99.00 - 101.00. The phases with likely high ferric iron contents (magnetites, ilmenites, spinels) have similar ideal oxide totals on recalculation for a theoretical ferric iron content. The analysed amphiboles have oxide totals between 96.75 and 97.37. Although analysis for structurally bound volatiles in the amphiboles has not been made in this study, O'Hara (1961a) determined H₂O contents of approximately 1.8 wt.percent in pargasitic amphiboles from the adjacent layered mafic-ultramafic sheet on Scouriemore (Geodh'nam Cliebh). As the amphiboles presented here were analysed at the same time as the other ferromagnesian phases it is believed they are of similar quality as the anhydrous silicates.

Analyses at Cambridge were for 12-14 elements using an electron microprobe with a Harwell Si (Li) detector and pulse processor (Statham, 1976). Operating conditions were 20 Kv accelerating voltage with a specimen current of 3na. Sample count times were 80 live seconds. A cobalt metal standard olivine and jadeite were analysed to assess the quality of the results. Correction procedures followed Sweatman and Long (1969).

Elements at Imperial College were analysed in pairs using a Cambridge 'Geoscan' with automated stage and spectrometers. Operating conditions were 15 Kv accelerating voltage with a specimen

current of up to 1 microamp. Sample count times were 80 seconds.

Standards used were:-

WOLL2 (natural wollastonite) for Ca and Si

FORST (synthetic forsterite) for Mg

TIOX (synthetic TiO_2) for Ti

MGALOX (synthetic spinel, MgAl_2O_4) for Al

RHODON (natural rhodonite) for Mn

JAD2 (natural jadeite) for Na

CR203 (synthetic CR203) for Cr

NIOXID (synthetic NiO_2) for Ni

KTANT (synthetic KTaO_3) for K

ZNS (synthetic ZNS) for Zn and S

PUREFE (metallic iron) for Fe

A standard pyroxene 'FRISCH' was included in the analyses to assess the quality of the results. Corrections to the raw data were made by the methods of Mason et al. (1969), and Frost (1977).

The tables of mineral analyses are grouped according to mineral type and host rock composition.

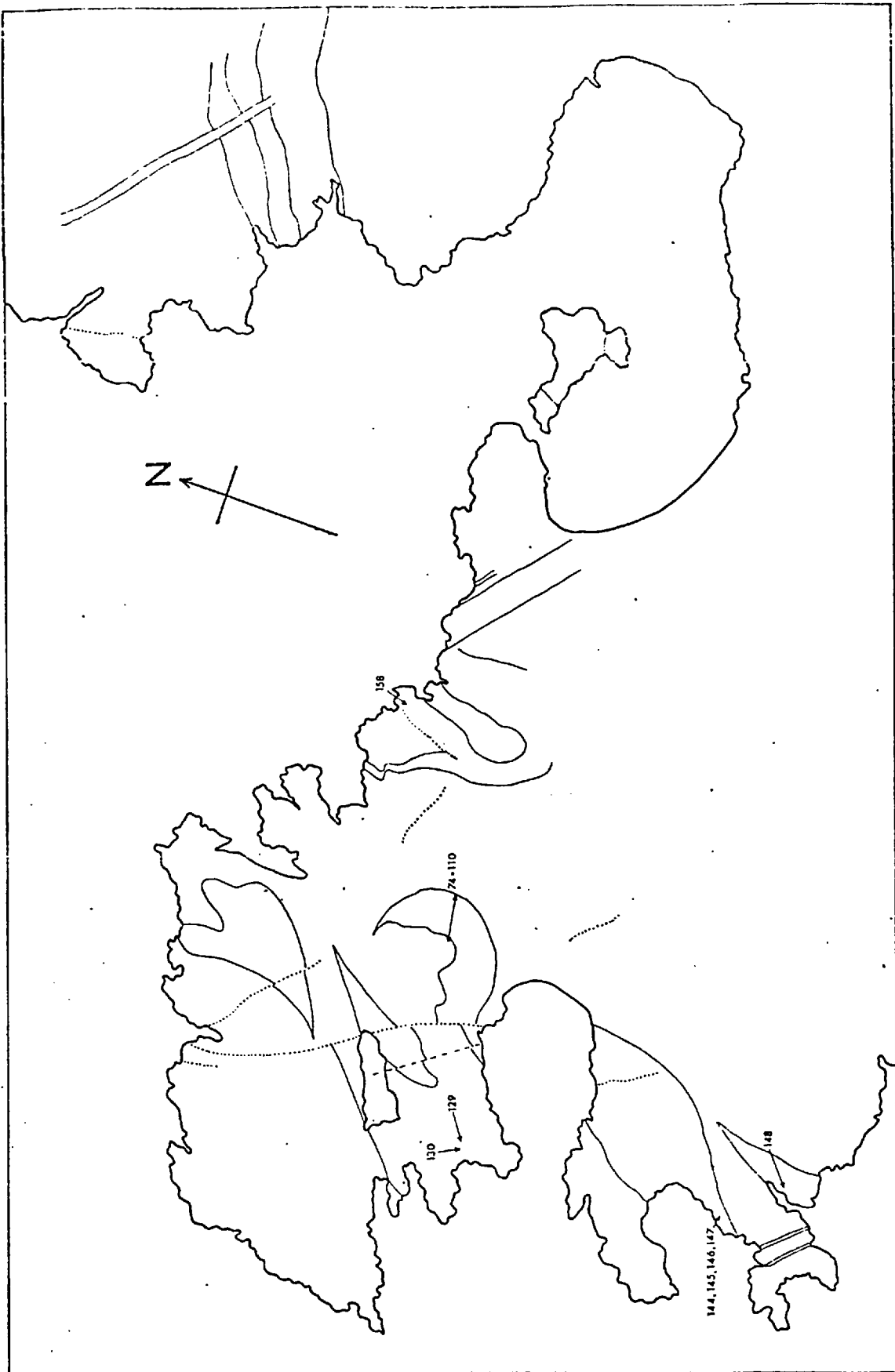


Fig. 1 Sample localities for analysed rocks.

TABLES OF ROCK AND MINERAL ANALYSES

	74	75	85	86	87	88	89	90	104	108
SI02	44.93	47.12	46.88	46.76	42.63	48.26	46.62	46.67	48.42	48.96
TI02	.38	.38	.29	.34	.29	.37	.31	.40	.39	.38
AL203	7.12	7.35	8.26	7.00	6.15	6.69	6.88	7.24	9.52	7.53
FE203	4.85	3.82	2.71	3.67	5.09	2.71	4.45	4.00	1.20	1.33
FEO	7.67	6.50	9.85	8.11	6.45	9.01	8.85	9.87	12.05	10.72
MNO	.20	.18	.18	.21	.19	.21	.22	.22	.23	.22
MGO	23.51	23.47	22.55	20.67	25.53	24.16	23.34	22.15	18.78	22.25
CAO	8.35	7.97	6.52	10.03	6.67	6.22	6.61	5.93	6.44	6.18
NA2O	.69	.69	.70	.84	.61	.66	.47	.84	.60	.56
K2O	.26	.18	.12	.43	.13	.38	.12	.19	.97	.60
P2O5	.03	.03	.02	.02	.03	.02	.02	.03	.05	.06
H2O-	.03	.02	.12	.10	.32	.05	.03	.04	.11	.16
LOI	1.75	2.09	1.27	1.05	5.68	1.21	1.14	1.20	1.83	1.66
TOTAL	99.77	99.80	99.47	99.23	99.77	99.95	99.06	98.78	100.59	100.61
V	145	140	132	144	113	128	144	154	160	146
CR	3087	3314	1992	3260	2970	3092	3513	3333	2087	3259
NI	1040	1012	692	1047	1503	1212	1002	957	467	928
CU	3	3	12	3	9	101	14	6	22	21
ZN	72	142	146	74	89	84	91	112	239	127
ZR	18	18	21	17	20	46	22	16	33	20
RB	7	4	0	14	4	9	4	6	8	13
SR	90	69	12	86	48	45	27	43	33	56
TH	0	0	0	0	0	0	0	0	0	3
Y	9	8	6	7	6	9	7	8	10	9
BA	0	0	0	65	0	0	0	0	79	88
K/RB	308	374	0	255	270	351	249	263	1007	383
RB/SR	.078	.058	0	.163	.083	.200	.148	.140	.242	.232
K/BA	0	0	0	55	0	0	0	0	102	57
K/SR	24	22	83	41	22	70	37	37	244	89
BA/RB	0	0	0	5	0	0	0	0	10	7
BA/SR	0	0	0	0.8	0	0	0	0	2.4	1.6
TI/ZR	126	126	83	120	87	48	84	150	71	114
ZR/Y	2.0	2.2	3.5	2.4	3.3	5.1	3.1	2.0	3.3	2.2
DI	7.38	6.90	6.63	9.65	5.93	7.83	4.69	8.23	10.81	8.29
CI	68.57	68.24	64.11	66.09	67.50	63.73	64.44	60.13	57.60	60.99
QZ	0	0	0	0	0	0	0	0	0	0
CO	0	0	0	0	0	0	0	0	0	0
ZR	0	0	0	0	0	.01	0	0	.01	0
OR	1.54	1.06	.71	2.55	.77	2.25	.71	1.12	5.73	3.55
PL	21.40	22.26	24.96	21.17	18.82	19.75	20.28	22.53	25.49	21.00
(AB)	5.84	5.84	5.92	7.11	5.16	5.58	3.98	7.11	5.08	4.74
(AN)	15.56	16.43	19.04	14.06	13.66	14.17	16.31	15.42	20.42	16.26
NE	0	0	0	0	0	0	0	0	0	0
DI	20.33	18.11	10.52	28.31	15.17	13.18	13.00	11.02	9.07	11.25
(WO)	10.73	9.58	5.50	14.87	8.04	6.92	6.83	5.77	4.69	5.86
(EN)	8.22	7.45	3.92	10.96	6.37	5.08	5.08	4.15	3.02	4.05
(FS)	1.38	1.08	1.10	2.49	.75	1.19	1.09	1.10	1.35	1.34
HY	17.70	29.17	33.40	15.91	19.57	35.92	37.84	35.50	36.07	39.32
(EN)	15.16	25.48	26.10	12.96	17.50	29.10	31.16	28.09	24.92	29.53
(FS)	2.54	3.70	7.30	2.94	2.07	6.81	6.69	7.42	11.15	9.79
OL	29.20	20.75	23.96	24.14	31.45	22.91	18.97	20.74	19.69	20.89
(FO)	24.65	17.89	18.32	19.31	27.82	18.21	15.34	16.07	13.19	15.30
(FA)	4.56	2.86	5.64	4.83	3.62	4.70	3.63	4.68	6.50	5.59
MT	7.03	5.54	3.93	5.32	7.38	3.93	6.45	5.80	1.74	1.93
IL	.72	.72	.55	.65	.55	.70	.59	.76	.74	.72
AP	.07	.07	.05	.05	.07	.05	.05	.07	.12	.14

Table 5 Ultramafite analyses from the Camas nam Buth sheet.

Table 6 Analyses of mafic gabbros from the Camas nam Butth intrusion.

	77	78	79	80	81	82	83	84	91	92	94	95	101	102	105	106	107
SI02	48.00	47.97	46.55	49.16	46.99	47.49	45.45	47.16	45.01	47.40	48.07	46.17	47.35	51.38	47.90	46.19	46.11
TI02	.61	.52	.72	.90	1.03	.80	.93	.51	.90	1.04	.86	.89	1.09	.96	.91	.88	1.08
AL203	15.57	13.98	14.53	15.17	14.33	15.22	14.09	13.44	9.86	12.79	15.65	12.41	14.45	14.43	15.71	15.53	13.12
FE203	.78	2.18	3.20	3.94	2.38	1.67	3.90	1.89	3.06	2.77	2.72	3.04	1.08	1.02	2.18	2.92	2.35
FE0	9.76	8.38	10.05	8.42	11.65	10.30	10.64	10.47	11.13	12.07	8.46	11.28	11.75	11.74	10.43	12.87	14.55
HNO	.24	.31	.23	.20	.22	.21	.24	.22	.28	.26	.25	.32	.22	.23	.21	.27	.30
HGO	11.29	9.67	9.30	7.22	8.01	9.06	9.99	13.73	13.11	8.34	6.60	8.88	8.19	7.37	8.59	8.70	7.73
CA0	9.71	13.65	11.73	11.00	13.04	12.68	10.68	9.73	13.01	11.45	14.09	13.67	12.07	9.10	11.12	9.84	11.63
NA20	1.79	1.70	1.58	2.71	1.79	1.70	1.77	1.24	.95	2.17	2.27	1.41	2.02	2.96	2.26	1.80	1.62
K20	.61	.17	.14	.14	.05	.13	.23	.33	.56	.19	.37	.29	.40	.35	.17	.23	.16
P205	.05	.06	.08	.08	.09	.14	.05	.05	.03	.10	.11	.10	.20	.13	.08	.06	.07
H20-	.02	.08	.07	.05	.14	.08	.13	.12	.06	.25	.33	.14	.12	.10	.11	.11	.08
LOI	1.57	1.07	1.31	1.19	1.38	1.29	1.95	1.21	1.70	1.59	1.47	1.54	1.44	1.16	1.19	1.32	1.38
TOTAL	100.00	99.74	99.49	100.18	101.10	100.77	100.13	100.10	99.66	100.42	101.25	100.14	100.38	100.93	100.86	100.72	100.18
V	252	247	258	252	341	271	301	230	278	292	231	305	282	278	265	226	327
CR	741	1160	522	269	283	469	215	771	1355	533	366	356	341	36	346	37	30
NI	342	180	213	119	143	266	202	165	223	217	118	156	138	46	141	100	67
CU	733	125	195	155	180	328	62	296	68	536	44	82	75	84	31	6	87
ZN	409	285	111	120	118	113	130	135	104	163	110	134	145	245	138	169	124
ZR	6	11	28	51	30	27	57	19	38	54	48	50	44	51	32	37	44
RB	4	0	0	0	0	0	3	0	3	0	0	2	2	0	0	0	0
SR	145	72	94	118	83	123	112	63	42	94	148	133	387	229	104	182	77
TH	0	0	3	0	0	0	0	3	3	0	0	3	4	0	3	0	0
Y	14	20	29	21	29	21	27	16	31	29	29	35	29	36	21	20	29
BA	138	52	0	0	0	0	54	0	46	0	65	49	97	105	0	106	51
K/RB	1266	0	0	0	0	0	636	0	1550	0	0	1204	1660	0	0	0	0
RB/SR	.028	0	0	0	0	0	.027	0	.071	0	0	.015	.005	0	0	0	0
K/DA	37	27	0	0	0	0	35	0	101	0	47	49	34	28	0	18	26
K/SR	35	20	12	10	5	9	17	43	111	17	21	18	9	13	14	10	17
BA/RB	34	0	0	0	0	0	18	0	15	0	45	24	48	0	0	0	0
BA/SR	0.9	0.7	0	0	0	0	0.5	0	1.1	0	0.4	0.4	0.2	0.5	0	0.6	0.7
TI/ZR	609	283	154	106	206	178	98	161	142	115	107	107	148	113	170	143	147
ZR/Y	0.4	0.5	1.0	2.4	1.0	1.3	2.1	1.2	1.2	1.9	1.7	1.4	1.5	1.4	1.5	1.8	1.5
DI	10.75	15.39	14.20	23.76	15.44	15.15	16.34	12.44	11.35	19.48	21.29	13.65	19.46	27.11	20.13	16.59	14.65
CI	57.88	60.71	57.36	50.45	55.21	58.93	55.47	60.87	60.21	49.01	55.59	55.52	52.50	43.53	54.69	53.37	49.50
OZ	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
CO	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
ZR	0	0	.01	.01	.01	.01	.01	0	.01	.01	.01	.01	.01	.01	.01	.01	.01
OR	3.61	1.00	.83	.83	.30	.77	1.36	1.95	3.31	1.12	2.19	1.71	2.36	2.07	1.00	1.36	.95
PL	47.79	44.40	45.51	51.75	46.06	47.90	44.80	40.62	29.02	42.96	50.40	38.61	46.27	50.10	51.34	48.85	41.76
(AB)	15.15	14.38	13.37	22.93	15.15	14.38	14.98	10.49	8.04	18.36	18.98	11.93	17.09	25.05	19.12	15.23	13.71
(AN)	32.65	30.01	32.14	28.81	30.92	33.51	29.82	30.13	20.99	24.60	31.42	26.68	29.18	25.05	32.22	33.62	28.05
NE	0	0	0	0	0	0	0	0	0	0	.12	0	0	0	0	0	0
DI	12.43	30.32	20.85	20.59	27.39	23.36	18.63	14.44	35.16	26.04	31.00	33.27	24.39	16.03	10.41	12.29	24.34
(WO)	6.38	15.59	10.67	10.55	13.87	11.91	9.55	7.45	18.11	13.19	15.79	16.93	12.33	8.07	9.38	6.22	12.20
(EN)	3.79	9.49	6.17	6.21	7.00	6.60	5.44	4.65	11.22	6.77	8.69	9.17	6.20	3.85	5.14	3.14	5.46
(FS)	2.26	5.24	4.01	3.83	6.44	4.85	3.44	2.34	5.82	6.00	6.52	7.18	5.86	4.11	3.90	2.94	6.68
HY	14.61	7.79	15.87	15.80	8.19	10.62	11.31	20.83	2.15	10.29	0	6.11	5.28	22.06	10.47	16.61	12.95
(EN)	9.16	5.02	9.62	9.82	4.29	6.12	7.03	13.85	1.41	5.42	0	3.43	2.72	10.68	5.95	8.58	5.82
(FS)	5.45	2.77	6.25	6.06	3.90	4.50	4.28	6.98	.73	4.07	0	2.68	2.57	11.38	4.52	8.03	7.12
BL	17.58	10.80	8.86	2.30	12.03	12.50	14.30	17.10	22.05	11.95	9.92	12.44	16.44	5.82	13.26	14.16	13.11
(FO)	10.62	6.71	5.17	1.37	6.01	6.90	8.56	10.99	14.02	6.01	5.43	6.68	8.05	2.68	7.22	6.97	5.50
(FA)	6.96	4.09	3.70	.93	6.02	5.60	5.75	6.11	8.02	5.94	4.49	5.76	8.39	3.14	6.04	7.19	7.53
MT	1.13	3.16	4.64	5.71	3.45	2.42	5.77	2.74	4.44	4.02	3.94	4.41	1.57	1.48	3.16	4.23	3.41
IL	1.16	.99	1.37	1.71	1.96	1.52	1.77	.97	1.71	1.98	1.63	1.69	2.07	1.82	1.73	1.67	2.05
AP	.12	.14	.19	.19	.21	.33	.12	.12	.07	.24	.26	.24	.47	.31	.19	.14	.17

	76	93	103
SI02	49.39	46.15	60.72
TI02	.08	3.35	.71
AL203	24.62	5.43	9.06
FE203	0	5.81	8.65
FE0	3.74	10.87	5.55
MNO	.09	.31	.09
MGO	5.53	11.12	3.06
CA0	8.47	16.00	5.25
NA20	3.41	1.04	3.11
K20	1.71	.01	.38
P205	.09	.29	.13
H2O-	.04	.27	.55
LOI	2.51	1.51	3.15
TOTAL	99.68	102.16	100.41
V	30	511	206
CR	266	1604	191
NI	290	108	196
CU	1205	409	808
ZN	90	137	108
ZR	0	126	56
RB	20	0	0
SR	504	51	98
TH	0	0	0
Y	0	29	14
BA	306	0	76
K/RB	710	0	0
RB/SR	.040	0	0
K/BA	46	0	41
K/SR	28	2	32
BA/RB	15	0	0
BA/SR	0.6	0	0.8
TI/ZR	0	159	76
ZR/Y	0	4.3	4.0
DI	38.29	8.86	56.71
CI	51.30	57.18	21.79
RZ	0	0	28.15
CO	1.89	0	0
ZR	0	.03	.01
OR	10.11	.06	2.25
PL	69.03	18.92	35.95
(AB)	27.38	8.80	26.32
(AN)	41.65	10.12	9.64
NE	.80	0	0
DI	0	54.12	12.45
(WO)	0	28.13	6.52
(EN)	0	18.99	4.68
(FS)	0	7.00	1.25
HY	0	11.59	3.73
(EN)	0	8.47	2.94
(FS)	0	3.12	.79
OL	14.98	.23	0
(FO)	9.65	.17	0
(FA)	5.33	.07	0
MT	0	8.42	12.54
IL	.15	6.36	1.35
AP	.21	.69	.31

Table 7 Qualitative analyses of sulphide-rich mafic gabbros from the Camas nam Buth intrusion.

	146	147	158
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SiO ₂	50.49	51.58	46.20
TiO ₂	.49	.41	1.27
Al ₂ O ₃	5.88	4.56	14.64
Fe ₂ O ₃	.31	0	4.36
FeO	6.81	7.61	9.54
MnO	.14	.16	.11
MgO	14.99	13.75	7.50
CaO	17.04	19.38	12.12
Na ₂ O	.88	.97	1.60
K ₂ O	.93	.13	.92
P ₂ O ₅	.06	.08	.12
H ₂ O-	.25	.31	.32
LOI	1.84	1.81	1.74
TOTAL	100.11	100.75	100.44
U	184	174	0
CR	555	442	171
NI	329	233	0
CU	56	104	0
ZN	60	76	0
ZR	19	39	0
RB	37	3	0
SR	129	85	0
Y	10	17	0
BA	662	75	221
K/RB	209	360	0
RB/SR	.287	.035	0
K/BA	12	14	35
K/SR	60	13	0
BA/RB	18	25	0
BA/SR	5.1	0.9	0
TI/ZR	155	63	0
ZR/Y	1.9	2.3	0
DI	12.60	8.98	18.98
CI	67.32	67.20	53.49
QZ	0	0	0
CO	0	0	0
ZR	0	.01	0
OR	5.51	.77	5.44
PL	16.01	15.91	43.59
(AB)	6.67	8.21	13.54
(AN)	9.34	7.70	30.05
NE	.42	0	0
DI	60.00	70.78	23.93
(WO)	31.31	36.73	12.25
(EN)	21.85	24.38	7.11
(FS)	6.85	9.68	4.57
HY	0	.97	9.44
(EN)	0	.70	5.75
(FS)	0	.28	3.70
OL	14.60	9.24	6.98
(FO)	10.85	6.43	4.08
(FA)	3.75	2.81	2.90
MT	.45	0	6.32
IL	.93	.78	2.41
AP	.14	.19	.28

Table 8 Analyses of ultramafites (146, 147) and a mafic gneiss from a net-veined mafic-ultramafic intrusion.

	130	96	109	129
SI02	38.27	64.47	40.33	46.86
TI02	.17	.48	.29	.64
AL203	3.80	10.07	17.79	17.06
FE203	6.55	0	.58	1.48
FE0	6.05	6.31	7.21	9.41
MNO	.16	.13	.18	.19
MGO	32.11	5.57	9.54	10.06
CA0	3.09	8.46	9.33	10.44
NA20	.09	2.82	1.40	1.43
K20	.01	.29	2.56	1.15
P205	.02	.11	.03	.07
H20-	.42	.19	.22	.11
LOI	9.40	.94	2.35	1.76
TOTAL	100.14	99.84	99.81	100.66
V	66	132	139	193
CR	4651	1254	360	472
NI	1735	379	79	283
CU	0	61	41	30
ZN	74	85	101	81
ZR	15	133	0	20
RB	0	0	24	12
SR	4	202	321	277
TH	0	0	0	0
Y	4	22	9	16
BA	0	76	362	167
K/RB	0	0	886	796
RB/SR	0	0	.075	.043
K/BA	0	32	59	57
K/SR	21	12	66	34
BA/RB	0	76	15	14
BA/SR	0	0.4	1.1	0.6
TI/ZR	68	22	0	192
ZR/Y	3.7	6.0	0	1.2
DI	.82	47.25	26.98	18.90
CI	68.60	32.64	55.67	59.57
RZ	0	21.68	0	0
CO	0	0	0	0
ZR	0	.03	0	0
OR	.06	1.71	15.14	6.80
FL	10.70	37.82	46.54	48.83
(AB)	.76	23.86	11.85	12.10
(AN)	9.94	13.96	34.69	36.73
NE	0	0	0	0
DI	4.13	22.47	9.39	11.99
(WO)	2.20	11.43	4.83	6.15
(EN)	1.80	6.15	2.97	3.62
(FS)	.13	4.89	1.58	2.22
HY	22.00	13.86	6.08	9.22
(EN)	20.53	7.72	3.97	5.71
(FS)	1.47	6.14	2.11	3.50
OL	43.57	0	18.68	18.45
(FO)	40.39	0	11.78	11.01
(FA)	3.18	0	6.90	7.44
MT	9.50	0	.84	2.15
IL	.32	.91	.55	1.22
AP	.05	.26	.07	.17

Table 9 Analyses of a retrogressed ultramafite (130), and retrogressed mafic gabbros from the Camas nam Buth intrusion.

	97	98	99	100	144	145	148
SI02	74.92	75.16	75.60	74.89	69.23	71.90	76.67
TI02	.04	.06	.10	.09	.05	.04	.15
AL203	14.02	13.80	13.79	14.06	16.39	16.40	12.88
FE203	0	0	0	0	0	0	0
FE0	.87	.41	.65	.67	1.25	.92	.80
MNO	.02	.01	.02	.02	.02	.02	.02
MGO	.50	.07	.40	.46	.83	.81	.52
CA0	1.59	1.63	1.11	1.28	4.27	3.95	1.89
NA20	4.75	3.81	4.18	4.28	4.27	4.39	3.31
K20	1.83	4.17	3.95	3.28	1.57	1.46	3.15
F205	.02	.02	.03	.05	.08	.05	.04
H2O-	.20	.16	.16	.17	.14	.14	.11
LOI	.80	.35	.58	.80	.93	.63	.45
TOTAL	99.56	99.65	100.57	100.05	99.03	100.71	99.99
V	0	0	8	0	66	0	14
CR	19	11	13	14	35	14	12
NI	14	8	6	11	24	14	5
CU	18	17	5	18	15	10	5
ZN	18	19	18	29	26	20	13
ZR	0	0	19	34	0	0	90
RB	11	32	37	33	10	7	37
SR	397	264	223	429	579	540	474
TH	0	0	0	0	0	0	0
Y	0	0	0	0	0	0	0
BA	564	935	731	846	752	592	1901
K/RB	1381	1082	886	825	1303	1732	707
RB/SR	.028	.121	.166	.077	.017	.013	.078
K/BA	27	37	45	32	16	20	14
K/SR	38	131	147	63	21	22	55
BA/RB	51	29	20	26	75	85	51
BA/SR	1.4	3.5	3.3	2.0	1.3	1.1	4.0
TI/ZR	0	0	31	16	0	0	10
ZR/Y	0	0	0	0	0	0	0
DI	86.38	89.96	91.41	89.11	72.48	76.14	86.47
CJ	8.87	8.35	6.23	7.13	22.39	20.97	10.56
DZ	35.37	33.07	32.69	33.50	27.07	30.36	39.83
CO	1.29	0	.61	1.15	0	.43	.49
ZR	0	0	0	.01	0	0	.02
OR	10.82	24.65	23.35	19.39	9.28	8.63	18.63
FL	48.19	40.47	40.90	42.55	57.05	56.71	37.66
(AB)	40.19	32.24	35.37	36.22	36.13	37.15	28.01
(AN)	8.00	8.23	5.53	6.33	20.92	19.56	9.65
NE	0	0	0	0	0	0	.0
DI	0	0	0	0	.07	0	0
(WO)	0	0	0	0	.03	0	0
(EN)	0	0	0	0	.02	0	0
(FS)	0	0	0	0	.02	0	0
HY	2.81	.85	2.06	2.26	4.28	3.68	2.55
(EN)	1.25	.17	1.00	1.15	2.05	2.02	1.30
(FS)	1.57	.67	1.07	1.12	2.23	1.66	1.26
OL	0	0	0	0	0	0	0
(FO)	0	0	0	0	0	0	0
(FA)	0	0	0	0	0	0	0
MT	0	0	0	0	0	0	0
JL	.08	.11	.19	.17	.09	.08	.28
AP	.05	.05	.07	.12	.19	.12	.09

Table 10 Analyses of the granite sheets.

	110
SI02	53.11
TI02	.31
AL2O3	10.41
FE2O3	4.38
FeO	8.89
MNO	.25
MGO	7.52
CAO	9.42
NA2O	2.62
K2O	.43
F2O5	.07
H2O-	.29
LOI	1.72
TOTAL	99.42
V	184
CR	558
NI	40
CU	99
ZN	162
ZR	81
RB	0
SR	206
TH	0
Y	42
BA	192
K/RB	0
RB/SR	0
K/BA	19
K/SR	17
BA/RB	0
BA/SR	0.9
TI/ZR	23
ZR/Y	1.9
DI	30.62
CJ	39.26
QZ	5.91
CO	0
ZR	.02
OR	2.54
FL	37.54
(AB)	22.17
(AN)	15.37
NE	0
II	25.33
(WO)	12.95
(EN)	7.39
(FS)	4.99
HY	19.01
(EN)	11.34
(FS)	7.67
OL	0
(FO)	0
(FA)	0
MT	6.35
IL	.59
AP	.17

Table 11 Analysis of a mafic gneiss.

ULTRAMAFITES → ← MAFIC GABBROS															
	<u>74</u>	<u>86</u>	<u>88</u>	<u>87</u>	<u>90</u>	<u>104</u>	<u>78</u>	<u>79</u>	<u>80</u>	<u>81</u>	<u>91</u>	<u>105</u>	<u>106</u>	<u>107</u>	
La	2.00	2.96	3.81	2.72	2.24	4.26	2.52	3.27	4.34	2.94	7.53	2.67	9.28	6.15	La
Ce	7.51	6.15	10.03	7.86	8.61	8.91	6.35	7.12	11.76	8.44	19.36	9.02	27.20	17.60	Ce
Nd	4.33	2.66	4.65	3.31	4.56	6.29	7.63	4.62	4.75	5.76	13.20	4.63	14.90	11.20	Nd
Sm	1.72	0.98	1.37	0.80	1.14	1.66	1.90	2.00	2.22	2.42	3.59	1.87	3.34	3.19	Sm
Eu	0.63	0.35	0.34	0.28	0.21	0.40	0.59	0.70	0.68	0.88	0.97	0.69	1.11	1.14	Eu
Gd	2.32	1.86	1.60	1.12	1.66	2.25	2.65	2.81	3.66	3.63	4.28	2.73	4.40	3.87	Gd
Tb	0.30	0.30	0.23	0.14	0.22	0.27	0.36	0.48	0.49	0.49	0.56	0.38	0.51	0.56	Tb
Ho	0.50	0.30	0.19	0.19	0.28	0.41	0.54	0.73	0.67	0.89	0.89	0.58	0.73	0.76	Ho
Yb	0.91	0.76	0.41	0.48	0.81	0.95	1.56	2.41	2.16	2.45	2.37	1.47	1.48	1.84	Yb
Lu	0.10	0.12	0.08	0.09	0.13	0.14	0.25	0.39	0.34	0.41	0.47	0.21	0.24	0.29	Lu
Sc	29	32	26	25	29	34	60	59	49	56	54	40	31	50	Sc
GRANITES →															
	<u>97</u>	<u>100</u>	<u>144</u>	<u>145</u>	<u>148</u>	<u>146</u>	<u>158</u>	<u>130</u>							
La	14.65	17.25	27.19	19.88	25.88	9.44	4.22	0.45							
Ce	16.56	25.97	37.49	23.22	36.00	15.84	15.94	2.65							
Nd	3.75	8.90	11.05	5.47	8.87	9.18	9.50	1.84							
Sm	0.49	0.79	2.43	0.51	1.15	2.86	3.35	0.50							
Eu	0.84	0.76	1.27	0.78	1.78	0.92	1.16	0.24							
Gd	0.41	0.64	1.77	0.44	0.87	3.00	4.28	0.55							
Tb	0.06	0.05	0.19	0.06	0.07	0.35	0.56	0.09							
Ho	0.05	0.05	0.16	0.05	0.03	0.39	0.87	0.12							
Yb	0.07	0.10	0.32	0.09	0.03	0.53	2.04	0.27							
Lu	0.01	0.02	0.05	0.01	0.005	0.07	0.35	0.02							
Sc	1	1	3	1	1	53	39	14							

Table 12 Rare-earth element analyses for the ultramafites and mafic gabbros from the Camas nam Bath intrusion, five granites, one ultramafite and one mafic gneiss from a net-veined intrusion (samples 146 and 158 respectively), and a retrogressed ultramafite from the Camas nam Bath intrusion. (130).

	ULTRAMAFITES						MAFIC GABBROS										
	74 [*] (3)	85 ¹ (8)	85 ² (4)	86(4)	88(3)	104(6)	77 [*] (1)	78 [*] (7)	81(4)	84 ¹ (2)	84 ² (2)	91(6)	101(6)	106(4)	129 [*] (6)		
SiO ₂	53.43	52.06	53.41	53.99	54.57	52.67	49.09	49.89	50.84	52.63	53.41	51.60	51.65	50.98	50.50	SiO ₂	
TiO ₂	0.03	-	-	-	-	-	0.10	0.07	-	0.15	0.13	-	-	-	0.34	TiO ₂	
Al ₂ O ₃	3.38	4.56	2.69	3.08	2.17	3.67	4.57	4.69	2.77	3.14	1.77	3.51	2.44	2.71	4.10	Al ₂ O ₃	
Cr ₂ O ₃	n.d.	-	-	-	0.12	0.13	n.d.	n.d.	-	-	-	-	-	-	n.d.	Cr ₂ O ₃	
FeO	12.30	19.04	16.99	11.60	10.85	16.20	20.87	22.55	25.45	21.51	19.12	21.85	24.88	25.25	21.29	FeO	
MnO	0.24	0.38	0.14	0.28	0.22	0.23	0.32	0.53	0.17	0.28	0.23	0.40	0.21	0.19	0.20	MnO	
NiO	n.d.	-	-	-	-	-	n.d.	n.d.	-	-	-	-	-	-	n.d.	NiO	
MgO	29.51	23.94	26.15	29.62	30.69	25.81	25.57	22.62	19.50	22.78	24.74	21.44	19.96	19.53	23.61	MgO	
CaO	0.49	0.41	0.37	0.43	0.38	0.41	0.25	0.36	0.33	0.27	0.18	0.47	0.35	0.42	0.75	CaO	
Na ₂ O	-	-	-	-	-	-	0.06	0.01	-	-	-	-	-	-	-	Na ₂ O	
Total	99.38	100.39	99.75	99.00	99.00	99.21	100.83	100.72	99.06	100.76	99.58	99.27	99.49	99.13	100.04	Total	
Si	1.910	1.897	1.937	1.926	1.942	1.917	1.812	1.852	1.937	1.930	1.961	1.929	1.952	1.940	1.862	Si	
Al ^{iv}	0.090	0.103	0.063	0.074	0.058	0.083	0.188	0.148	0.063	0.070	0.039	0.071	0.048	0.060	0.138	Al ^{iv}	
Al ^{vi}	0.052	0.090	0.052	0.056	0.033	0.079	0.011	0.058	0.061	0.066	0.038	0.084	0.061	0.061	0.040	Al ^{vi}	
Ti	0.001	-	-	-	-	-	0.003	0.002	-	0.004	0.003	-	-	-	0.009	Ti	
Fe	0.368	0.580	0.515	0.346	0.323	0.493	0.644	0.700	0.810	0.660	0.587	0.683	0.786	0.803	0.657	Fe	
Cr	-	-	-	-	0.003	0.003	-	-	-	-	-	-	-	-	-	Cr	
Mg	1.572	1.300	1.413	1.575	1.627	1.400	1.407	1.252	1.107	1.245	1.353	1.194	1.124	1.110	1.297	Mg	
Ni	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	Ni	
Mn	0.007	0.012	0.004	0.009	0.007	0.007	0.010	0.017	0.003	0.008	0.007	0.013	0.007	0.006	0.006	Mn	
Ca	0.019	0.016	0.014	0.016	0.015	0.015	0.010	0.014	0.013	0.011	0.007	0.019	0.014	0.017	0.030	Ca	
Na	-	-	-	-	-	-	0.005	0.001	-	-	-	-	-	-	-	Na	
Total	4.019	4.001	3.998	4.002	4.008	3.997	4.090	4.044	3.994	3.994	3.995	3.993	3.992	3.997	4.039	Total	
En	80.3	68.6	72.8	81.3	82.8	73.4	68.2	63.7	57.3	65.0	69.5	63.0	58.4	57.5	65.4	En	
Fs	18.8	30.6	26.5	17.9	16.4	25.8	31.3	35.7	42.0	34.4	30.1	36.0	40.8	41.6	33.1	Fs	
Wo	0.9	0.8	0.7	0.8	0.8	0.8	0.5	0.7	0.7	0.6	0.4	1.0	0.7	0.9	1.5	Wo	
Fe/Fe+Mg	.190	.308	.267	.180	.166	.260	.314	.358	.422	.346	.303	.364	.411	.420	.336	Fe/Fe+Mg	

Table 13 Orthopyroxenes from the ultramarites and mafic gabbros. Analyses are by EDS, except those marked * which indicates a crystal-spectrometer analysis. Figures in parentheses after the sample no. indicate the number of points included in the analysis. Fe is expressed as FeO(Total). The formula unit is for six oxygens. 85¹ is from a garnet-free symplectite, 85² from a garnetiferous symplectite. 84¹ is a relict grain, 84² is a symplectite grain.

	ULTRAMAFITES										MAFIC GABBROS					
	<u>74</u> (6)	<u>85</u> ¹ (6)	<u>85</u> ² (6)	<u>86</u> (2)	<u>88</u> (3)	<u>104</u> (6)	<u>77</u> [*] (5)	<u>78</u> [*] (13)	<u>81</u> ¹ (5)	<u>81</u> ² (4)	<u>84</u> (10)	<u>91</u> (19)	<u>101</u> (8)	<u>106</u> (2)	<u>129</u> [*] (12)	
SiO ₂	51.21	50.16	49.66	51.36	51.48	50.96	48.29	48.62	49.99	50.12	50.22	49.30	50.24	49.58	48.95	SiO ₂
TiO ₂	0.19	0.62	0.69	0.15	0.19	0.19	0.72	0.66	0.70	0.72	0.59	0.53	0.56	0.63	0.95	TiO ₂
Al ₂ O ₃	4.14	6.62	6.90	4.01	3.80	5.34	5.20	6.50	5.35	4.75	5.69	5.89	4.65	5.33	6.73	Al ₂ O ₃
Cr ₂ O ₃	0.29	0.15	-	0.29	0.30	0.28	n.d.	n.d.	-	-	0.19	0.15	-	-	n.d.	Cr ₂ O ₃
FeO	5.24	7.30	7.23	5.32	5.26	6.32	9.15	8.87	11.45	10.14	8.89	9.50	10.67	9.85	8.06	FeO
MnO	0.15	0.15	-	0.14	-	-	0.29	0.19	0.16	-	0.26	0.29	0.19	-	0.15	MnO
NiO	-	-	-	-	-	-	n.d.	n.d.	-	-	-	-	-	-	n.d.	NiO
MgO	14.71	12.58	12.32	14.60	15.26	13.40	13.59	12.76	11.10	11.34	11.97	12.19	11.21	11.24	12.43	MgO
CaO	23.88	22.34	22.15	22.97	22.42	22.16	21.36	21.28	20.68	21.65	21.41	21.39	21.43	21.19	22.46	CaO
Na ₂ O	0.53	0.73	0.70	-	0.51	0.64	0.74	1.02	0.67	0.83	0.66	0.71	0.61	0.89	1.03	Na ₂ O
Total	100.34	100.71	99.65	98.84	99.22	99.29	99.84	99.90	100.10	99.52	99.98	99.95	99.56	98.71	100.76	Total
Si	1.836	1.847	1.844	1.910	1.906	1.890	1.820	1.820	1.879	1.889	1.875	1.849	1.895	1.880	1.816	Si
Al ^{iv}	0.114	0.153	0.156	0.090	0.094	0.110	0.180	0.180	0.121	0.111	0.125	0.151	0.105	0.120	0.184	Al ^{iv}
Al ^{vi}	0.066	0.137	0.146	0.086	0.072	0.124	0.052	0.107	0.116	0.100	0.125	0.110	0.102	0.118	0.111	Al ^{vi}
Ti	0.005	0.017	0.019	0.004	0.005	0.005	0.020	0.018	0.020	0.020	0.017	0.015	0.016	0.018	0.027	Ti
Fe	0.161	0.225	0.225	0.166	0.163	0.196	0.288	0.278	0.360	0.319	0.278	0.298	0.337	0.312	0.250	Fe
Cr	0.008	0.004	-	0.008	0.009	0.008	-	-	-	-	0.006	0.005	-	-	-	Cr
Mg	0.808	0.691	0.682	0.809	0.842	0.741	0.764	0.712	0.622	0.637	0.666	0.682	0.631	0.635	0.687	Mg
Ni	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	Ni
Mn	0.004	0.005	-	0.004	-	-	0.009	0.006	0.005	-	0.009	0.009	0.006	-	0.005	Mn
Ca	0.943	0.882	0.881	0.915	0.890	0.881	0.883	0.853	0.833	0.875	0.856	0.860	0.867	0.861	0.893	Ca
Na	0.037	0.052	0.051	-	0.037	0.046	0.054	0.079	0.049	0.061	0.048	0.052	0.045	0.065	0.075	Na
Total	4.032	4.013	4.004	3.992	4.018	4.001	4.070	4.053	4.005	4.012	4.005	4.031	4.004	4.009	4.048	Total
En	42.2	38.4	38.1	42.8	44.4	40.8	39.5	38.6	34.3	34.8	37.0	37.0	34.4	35.1	37.5	En
Fs	8.4	12.5	12.6	8.8	8.6	10.8	14.9	15.1	19.8	17.4	15.4	16.2	18.4	17.2	13.7	Fs
Wo	49.3	49.1	49.3	48.4	47.0	48.4	45.6	46.3	45.9	47.8	47.6	46.8	47.2	47.7	48.8	Wo
Fe/Fe+Mg	.167	.246	.248	.170	.162	.209	.274	.281	.367	.304	.294	.305	.348	.329	.267	Fe/Fe+Mg

Table 14 Clinopyroxenes from the ultramafites and mafic gabbros. Analyses are by EDS, except those marked * which indicates a crystal-spectrometer analysis. Figures in parentheses after the sample no. indicate the number of points included in the analysis. Fe is expressed as FeO(Total). The formula unit is for six oxygens. B1¹ is a relict grain. B1² is a recrystallized grain. B5¹ is from a garnet-free symplectite, B5² is from a garnetiferous symplectite.

	<u>74(4)</u>	<u>86(4)</u>	<u>88(5)</u>
SiO ₂	39.93	39.40	39.72
TiO ₂	-	-	-
Al ₂ O ₃	-	-	-
Cr ₂ O ₃	-	-	-
FeO	15.44	16.45	14.03
MnO	0.17	0.23	0.20
NiO	0.42	0.44	0.48
MgO	44.38	43.26	45.07
CaO	-	-	-
Total	<u>100.39</u>	<u>99.78</u>	<u>99.50</u>
Si	1.003	1.000	1.000
Al	-	-	-
Ti	-	-	-
Fe	0.324	0.349	0.295
Cr	-	-	-
Mg	1.659	1.637	1.691
Ni	0.008	0.009	0.010
Mn	0.004	0.005	0.004
Ca	-	-	-
Total	<u>2.998</u>	<u>3.000</u>	<u>3.000</u>
Fo	83.7	82.4	85.1
Fa	16.3	17.6	14.9

Table 15 Olivines from the ultramafites. All analyses are by EDS. Figures in parentheses after the sample no. indicate the number of points included in the analysis. Fe is expressed as FeO(Total). The formula unit is for four oxygens.

	<u>95</u> (7)	<u>77</u> [*] (9)	<u>78</u> [*] (7)	<u>81</u> ^P (5)	<u>81</u> ^C (2)	<u>84</u> (2)	<u>91</u> (5)	<u>101</u> ^P (4)	<u>101</u> ^C (4)	<u>106</u> ^P (4)	<u>106</u> ^C (4)	<u>129</u> [*] (4)	
SiO ₂	39.30	35.67	39.36	39.19	38.71	39.61	39.20	39.08	38.90	39.02	38.54	39.03	SiO ₂
TiO ₂	-	0.05	0.09	-	-	-	-	0.10	-	-	-	0.08	TiO ₂
Al ₂ O ₃	22.14	21.81	20.57	21.11	21.10	21.51	21.90	21.50	21.46	21.56	21.46	22.01	Al ₂ O ₃
Cr ₂ O ₃	-	n.d.	n.d.	-	-	-	-	-	-	-	-	n.d.	Cr ₂ O ₃
FeO	22.31	24.20	23.02	25.98	27.10	23.76	24.10	25.22	27.11	24.75	26.81	23.44	FeO
MnO	0.85	1.04	1.36	0.72	0.62	1.21	1.47	1.04	0.67	0.64	0.56	0.69	MnO
NiO	-	n.d.	n.d.	-	-	-	-	-	-	-	-	-	NiO
MgO	9.12	9.00	9.30	7.05	5.90	8.28	7.09	6.85	5.58	7.65	5.95	9.54	MgO
CaO	6.51	5.42	5.74	6.69	6.80	6.63	7.00	6.68	6.91	6.20	6.54	6.70	CaO
Na ₂ O	-	n.d.	n.d.	-	-	-	-	-	-	-	-	n.d.	Na ₂ O
Total	100.73	100.19	99.44	100.74	100.23	101.00	100.76	100.45	100.63	99.82	99.86	101.49	Total
Si	3.009	2.968	3.034	3.015	3.016	3.016	3.004	3.011	3.017	3.010	3.006	2.951	Si
Al	1.974	1.974	1.869	1.915	1.938	1.931	1.978	1.953	1.962	1.961	1.973	1.962	Al
Ti	-	0.003	0.005	-	-	-	-	0.006	-	-	-	0.009	Ti
Fe	1.411	1.553	1.484	1.672	1.766	1.513	1.545	1.625	1.758	1.596	1.749	1.482	Fe
Cr	-	-	-	-	-	-	-	-	-	-	-	-	Cr
Mg	1.028	1.029	1.054	0.808	0.685	0.939	0.810	0.786	0.644	0.880	0.691	1.075	Mg
Ni	-	-	-	-	-	-	-	-	-	-	-	-	Ni
Mn	0.054	0.068	0.088	0.054	0.041	0.078	0.095	0.069	0.044	0.044	0.042	0.037	Mn
Ca	0.527	0.447	0.474	0.552	0.568	0.541	0.575	0.551	0.574	0.513	0.546	0.543	Ca
Na	-	-	-	-	-	-	-	-	-	-	-	-	Na
Total	8.003	8.042	8.008	8.016	8.014	8.018	8.007	8.001	7.999	8.002	8.002	8.068	Total
Am	46.7	50.2	47.6	54.2	57.7	49.3	51.2	53.6	58.2	52.7	57.9	47.1	Am
Py	34.0	33.2	34.3	26.2	22.4	30.6	26.8	25.9	21.3	29.0	22.9	34.2	Py
Gr	17.4	14.4	15.2	17.9	18.6	17.6	19.0	18.2	19.0	16.9	18.1	17.2	Gr
Sp	1.8	2.2	2.9	1.7	1.3	2.5	3.1	2.3	1.5	1.4	1.2	1.4	Sp
Fe/Fe+Mg	.578	.602	.581	.674	.720	.617	.656	.674	.732	.645	.717	.580	Fe/Fe+Mg

Table 16 Garnet analyses. 85 is an ultramafite, the rest are mafic gabbros. Analyses are by EDS, except those marked * which indicates a crystal-spectrometer analysis. Figures in parentheses after the sample no. indicate the number of points included in the analysis. Fe is expressed as FeO(Total). The formula unit is for twelve oxygens. P and C as superscripts refer to porphyroblast and corona, respectively.

ULTRAMAFITES → ← MAFIC GABBROS

	74(5)	85 ¹ (4)	85 ² (4)	86(4)	104(3)	77 [*] (3)	84(2)	
SiO ₂	-	0.35	-	-	0.20	-	0.22	SiO ₂
TiO ₂	-	-	-	-	-	-	-	TiO ₂
Al ₂ O ₃	57.28	59.56	60.88	55.61	56.82	60.66	56.72	Al ₂ O ₃
Cr ₂ O ₃	4.52	1.34	0.90	6.34	5.75	1.09	2.81	Cr ₂ O ₃
FeO	20.71	28.39	24.55	20.58	23.57	24.67	26.91	FeO
MnO	-	0.17	-	0.23	-	-	0.15	MnO
NiO	0.31	-	-	-	0.41	n.d.	-	NiO
MgO	15.93	11.03	12.88	15.36	11.67	12.99	10.45	MgO
CaO	-	-	-	-	-	n.d.	-	CaO
ZnO	n.d.	n.d.	n.d.	n.d.	1.13	n.d.	-	ZnO
Na ₂ O	-	-	-	-	-	n.d.	-	Na ₂ O
Total	98.75	100.84	99.21	98.65	99.55	99.41	97.26	Total
Si	-	0.077	-	-	0.043	-	0.048	Si
Al	14.480	15.172	15.450	14.294	14.657	15.405	14.950	Al
Ti	-	-	-	-	-	-	-	Ti
Fe	3.712	5.130	4.419	3.752	4.312	4.444	5.032	Fe
Cr	0.768	0.229	0.152	1.093	0.995	0.185	0.496	Cr
Mg	5.088	3.551	4.133	4.991	3.805	4.170	3.480	Mg
Ni	0.088	-	-	0.092	0.071	-	-	Ni
Mn	-	0.031	-	0.042	-	-	0.024	Mn
Ca	-	-	-	-	-	-	-	Ca
Zn	-	-	-	-	0.183	-	-	Zn
Na	-	-	-	-	-	-	-	Na
Total	24.136	24.190	24.154	24.264	24.066	24.204	23.030	Total
Fe/Fe+Mg	.422	.591	.517	.429	.531	.516	.591	Fe/Fe+Mg

Table 17 Spinel analyses from the ultramafites and mafic gabbros. Analyses are by EDS, except those marked * which indicates a crystal - spectrometer analysis. Figures in parentheses after the sample no. indicate the number of points included in the analysis. Fe is expressed as FeO(Total). The formula unit is for thirty-two oxygens. 85¹ is from a garnet-free symplectite, 85² is from a garnetiferous symplectite.

	ULTRAMAFITES $\longleftrightarrow$			$\longleftarrow$ MAFIC GABBROS		
	<u>74(3)</u>	<u>86(1)</u>	<u>88(1)</u>	<u>104(1)</u>	<u>91(1)</u>	<u>106(1)</u>
SiO ₂	42.53	41.95	42.04	42.76	42.87	40.76
TiO ₂	0.89	0.77	1.17	0.94	0.97	3.23
Al ₂ O ₃	13.96	14.52	13.95	15.02	15.69	13.36
Cr ₂ O ₃	0.32	0.30	0.45	0.35	0.13	-
FeO	7.26	6.88	6.89	7.89	11.93	13.43
MnO	-	-	-	-	0.13	-
NiO	-	-	0.15	-	-	-
MgO	15.92	15.95	15.91	14.26	10.71	10.78
CaO	12.49	12.44	12.60	12.25	12.04	11.80
Na ₂ O	2.31	2.49	2.38	1.88	2.01	1.63
K ₂ O	1.17	1.68	1.45	1.40	0.89	2.23
Total	<u>96.85</u>	<u>96.98</u>	<u>96.99</u>	<u>96.75</u>	<u>97.37</u>	<u>97.22</u>
Si	6.201	6.126	6.145	6.240	6.299	6.131
Al ^{iv}	1.799	1.874	1.855	1.760	1.701	1.869
Al ^{vi}	0.688	0.625	0.549	0.825	1.016	0.498
Ti	0.097	0.084	0.129	0.104	0.107	0.365
Fe	0.885	0.840	0.842	0.963	1.466	1.688
Cr	0.036	0.034	0.051	0.041	0.015	-
Mg	3.458	3.472	3.465	3.100	2.344	2.415
Ni	-	-	0.017	-	-	-
Mn	-	-	-	-	0.016	-
Ca	1.952	1.946	1.974	1.915	1.895	1.900
Na	0.652	0.704	0.674	0.531	0.571	0.476
K	0.216	0.313	0.270	0.260	0.166	0.428
Total	<u>15.984</u>	<u>16.018</u>	<u>15.971</u>	<u>15.738</u>	<u>15.596</u>	<u>15.770</u>
Fe/Fe+Mg	.204	.194	.195	.237	.385	.411

Table 18 Amphiboles from the ultramafites and mafic gabbros. All analyses are by EDS. Figures in parentheses after the sample no. indicate the number of points included in the analysis. Fe is expressed as FeO(Total). The formula unit is for twenty-three oxygens.

	<u>91(2)</u>
SiO ₂	38.70
TiO ₂	3.14
Al ₂ O ₃	16.69
Cr ₂ O ₃	0.20
FeO	12.03
MnO	-
NiO	-
MgO	17.12
CaO	0.37
Na ₂ O	-
K ₂ O	9.52
Total	<u>97.77</u>

Si	5.531
Al ^{iv}	2.469
Al ^{vi}	0.349
Ti	0.337
Fe	1.436
Cr	0.023
Mg	3.650
Ni	-
Mn	-
Ca	0.056
Na	-
K	1.736
Total	<u>15.587</u>

Fe/Fe+Mg	.282
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Table 19 Biotite analysis from a mafic gabbro. The analysis was by EDS. Figures in parentheses after the sample no. indicate the number of points included in the analysis. Fe is expressed as FeO(Total). The formula unit is for twenty-two oxygens.

ULTRAKAFITES					MAFIC GABBROS							
	85 ¹ (4)	85 ² (4)	104(4)	78(2)	81(3)	84(2)	91(3)	101(6)	106(3)			
SiO ₂	54.04	53.65	52.89	53.50	54.00	53.97	49.40	55.18	56.03	SiO ₂		
TiO ₂	-	-	-	-	-	-	-	-	-	TiO ₂		
Al ₂ O ₃	29.86	29.64	29.57	28.83	27.94	28.61	31.03	28.27	27.37	Al ₂ O ₃		
Cr ₂ O ₃	-	-	-	-	-	-	-	-	-	Cr ₂ O ₃		
FeO	0.24	0.21	0.17	0.18	0.20	0.27	1.34	0.14	0.19	FeO		
MnO	-	-	-	-	-	-	-	-	-	MnO		
NiO	-	-	-	-	-	-	-	-	-	NiO		
MgO	-	-	-	-	-	-	0.36	-	-	MgO		
CaO	12.53	12.16	12.69	11.88	11.33	11.80	14.43	10.87	9.94	CaO		
Na ₂ O	4.54	4.67	4.23	6.00	4.91	4.70	2.73	5.18	5.83	Na ₂ O		
K ₂ O	-	-	-	-	0.12	-	-	0.16	0.16	K ₂ O		
Total	101.21	100.33	99.55	100.39	98.50	99.35	99.29	99.80	99.52	Total		
Z	Si	9.661	9.674	9.621	9.689	9.900	9.816	9.111	9.961	10.128	Si	Z
	Al	6.293	6.301	6.340	6.156	6.039	6.135	6.747	6.017	5.833	Al	
	Ti	-	-	-	-	-	-	-	-	-	Ti	
X	Mg	-	-	-	-	-	-	0.100	-	-	Mg	X
	Fe	0.036	0.031	0.026	0.027	0.030	0.040	0.209	0.022	0.029	Fe	
	Ca	2.401	2.349	2.475	2.306	2.225	2.298	2.850	2.103	1.925	Ca	
	Na	1.573	1.632	1.491	2.107	1.746	1.654	0.976	1.813	2.043	Na	
	K	-	-	-	-	0.027	-	-	0.038	0.038	K	
Z	15.954	15.975	15.961	15.845	15.939	15.951	15.858	15.978	15.961	Z		
X	4.010	4.012	3.992	4.440	4.028	3.992	4.135	3.976	4.035	X		
An	60.4	59.0	62.4	51.4	55.6	58.1	74.5	53.2	48.0	An		
Ab	39.6	41.0	37.6	48.6	43.7	41.9	25.5	45.8	51.0	Ab		
Or	0.0	0.0	0.0	0.0	0.7	0.0	0.0	1.0	1.0	Or		

Table 20 Plagioclase analyses from the ultramafites and mafic gabbros. All analyses are by EDS. Figures in parentheses after the sample no. indicate the number of points included in the analysis. Fe is expressed as FeO(Total). The formula unit is for thirty-two oxygens. 85¹ is from a garnet-free symplectite, 85² is from a garnetiferous symplectite.

ULTRAMAFITES				MAFIC GABBROS					
	74 ^{*M} (4)	86 ^M (3)	88 ^M (2)	91 ^M (3)	101 ^M (2)	101 ^I (2)	106 ^M (2)	106 ^I (2)	
SiO ₂	0.14	0.48	0.28	0.69	0.33	0.35	0.80	0.63	SiO ₂
TiO ₂	1.08	1.35	1.88	0.24	0.36	51.06	0.20	52.01	TiO ₂
Al ₂ O ₃	1.23	1.00	1.85	0.60	0.59	-	0.67	0.21	Al ₂ O ₃
Cr ₂ O ₃	3.71	4.89	6.72	3.02	0.52	-	-	-	Cr ₂ O ₃
FeO	84.71	83.06	81.12	86.99	88.42	47.05	90.49	46.78	FeO
MnO	0.06	0.20	-	-	-	0.20	-	-	MnO
NiO	n.d.	0.63	0.55	-	-	-	-	-	NiO
V ₂ O ₅	n.d.	n.d.	n.d.	n.d.	1.43	0.31	0.65	-	V ₂ O ₅
MgO	0.82	0.59	0.61	0.11	-	0.65	0.13	0.82	MgO
CaO	n.d.	-	-	-	-	-	-	-	CaO
ZnO	n.d.	n.d.	n.d.	n.d.	-	-	-	-	ZnO
Na ₂ O	n.d.	-	-	-	-	-	-	-	Na ₂ O
Total	91.75	92.20	93.01	91.65	91.65	99.62	92.94	100.45	Total
Si	0.056	0.188	0.107	0.279	0.137	0.094	0.322	0.169	Si
Al	0.576	0.464	0.835	0.285	0.285	-	0.319	0.067	Al
Ti	0.323	0.397	0.540	0.075	0.110	10.380	0.062	10.390	Ti
Fe	28.131	27.246	25.854	29.353	30.124	10.635	30.322	10.395	Fe
Cr	1.163	1.516	2.025	0.962	0.166	-	-	-	Cr
Mg	0.485	0.345	0.348	0.066	-	0.264	0.079	0.326	Mg
Ni	-	0.198	0.168	-	-	-	-	-	Ni
Mn	0.020	0.067	-	-	-	0.047	-	-	Mn
V	-	-	-	-	0.467	0.068	0.210	-	V
Ca	-	-	-	-	-	-	-	-	Ca
Zn	-	-	-	-	-	-	-	-	Zn
Na	-	-	-	-	-	-	-	-	Na
Total	30.753	30.421	29.157	31.020	31.289	21.488	31.314	21.365	Total

Table 21 Magnetites and Ilmenites from the ultramafites and mafic gabbros.

Analyses are by EDS, except those marked * which indicates a crystal-spectrometer analysis. Figures in parentheses after the sample no. indicate the number of points included in the analysis. Fe is expressed as FeO(Total). The formula unit is for thirty-two oxygens. M and I as superscripts refer to magnetite and ilmenite, respectively.

	<u>PYRRHOTITE(7)</u>	<u>CHALCOPYRITE(1)</u>	<u>PYRITE(1)</u>
S	39.88	35.48	54.88
Fe	58.97	30.30	46.10
Cu	0.22	34.43	0.01
Co	0.28	0.09	0.70
Ni	0.79	-	0.07
Zn	0.20	-	0.07
Total	<u>100.34</u>	<u>100.30</u>	<u>101.84</u>
Fe	1.700	0.980	0.964
Cu	0.005	0.979	0.000
Ni	0.022	-	0.001
Co	0.008	0.003	0.014
Zn	0.005	-	0.001
Total	<u>1.740</u>	<u>1.962</u>	<u>1.980</u>

Table 22 Sulphide analyses from a sulphide-rich mafic gabbro, sample 76. All analyses are by crystal-spectrometer. Fe is expressed as Fe^{II}(Total). The formula unit is for two sulphur atoms. Figures in parentheses indicate the number of points included in the analysis.